CHAPTER 4 - WATER BALANCE IN AREAS OF SNOW ACCUMULATION

4-01. INTRODUCTION

4-01.01 General. - In areas of snow accumulation, water stored in the snowpack provides a time delay between precipitation and runoff. Hence, by proper evaluation of the variables affecting runoff, volumetric forecasts of snowmelt runoff may be made several months in advance of the runoff event. Runoff forecasting methods have developed largely on the basis of statistical correlations of runoff with either prior precipitation amounts or snowpack water equivalent. Additional indexes of soil moisture, winter temperature, and winter runoff are also sometimes used. (A discussion of volume-of-runoff forecasting procedures is contained in chapter 11.) In most cases, the period of record for the necessary basic data is too short to ensure an adequate statistical sample of the range in variation and probable long-term mean values. Consequently, there may be serious bias in forecasting procedures derived by statistical methods from data unrepresentative of long-term normals. Such a bias results in unrealistic weighting of the independent variables in the forecasting equation. A knowledge of the water balance for a given area is essential in order to select appropriate forecasting parameters and interpret their reliability. Moreover, the water-balance technique may be used as a forecasting procedure by quantitatively balancing runoff against precipitation, change in snowpack water equivalent, and losses. This procedure is particularly useful for areas where hydrometeorological records are short. In addition, knowledge of the water balance is necessary for rate-of-flow forecasting and for design flood computations.

4-01.02 The principal deterrent to computing water balances for project basins has been the lack of adequate basic information. Scanty areal sampling and unrepresentative measurements of precipitation and accumulated snow, as well as inadequate information on losses, have made water-balance determinations very indefinite. In contrast, the snow laboratories provide relatively dense areal sampling of the prime hydrometeorologic variables. These areas are also relatively free from hydrologic uncertainties involved in computation of losses. One of the primary considerations in the implementation of the snow laboratory program was to provide adequate data for obtaining hydrologic balances in areas of snow accumulation. Such balances for each of the snow laboratories, for its period of complete record (4 to 5 years each), are given in this chapter. This chapter also discusses the individual components of the water balance and the methods of derivation of monthly values for each laboratory.

4-01.03 Even with the dense sampling and better-than-average quality of data for the laboratory areas, errors in measurement of the larger and more important components of the water balance (e.g., precipitation and runoff) may obscure the effect of lesser components. Since the computed areal mean value of any given component cannot be considered to be precisely the "true" value, these computed values must be adjusted to provide the most reasonable balance, using the best information available.

4-01.04 Definition of components. - Because of the varied interpretations of the components of the water balance, each will be specifically defined as used in this report. Observed runoff is defined as the gaged volume of water passing a gaging station on a river or stream. Generated runoff is observed runoff corrected by recession analysis for transitory storage in the soil, ground, and stream channels. Total basin precipitation, in the form of either rain or snow, is defined as that hypothetical precipitation which falls above tree-crown level. Net precipitation is that portion of the total precipitation which reaches the ground or snow surface, after partial interception by the forest canopy (see next paragraph). Since the amount intercepted by vegetation is affected by the form of precipitation, the total precipitation is divided into total snowfall and total rainfall. Interception amounts are computed separately, and values of net snowfall and net rainfall are derived. The water equivalent of the snowpack is exactly what the term implies: the volume of water stored in the snowpack (both solid and liquid forms). Melt is defined as the net decrease in water equivalent of the snowpack after allowance is made for increases as a result of precipitation; thus it excludes water which re-freezes or is retained as liquid water within the snowpack.

4-01.05 Loss is defined as that part of total precipitation which is permanently lost to runoff by evaporation, sublimation, transpiration, and retention as stored soil moisture. (Soil-moisture storage differs from ground or channel storage in that water stored as soil moisture can be removed from the soil only by evaporation and transpiration, whereas water in ground or channel storage is temporarily stored while in transit and will ultimately appear as runoff.) Soil moisture is further subdivided into available and unavailable soil moisture; the latter is not subject to transpiration or evaporation under normal field conditions. In addition to the losses occurring after precipitation has reached the ground or snow surface, other losses result from the interception of precipitation by the forest cover. Interception loss consists of the evaporation and sublimation of intercepted precipitation from vegetation surfaces. It therefore represents the net difference in precipitation within a large forest opening and that received beneath the tree crowns. Since precipitation measurements are generally made in the open, areal interpretation of precipitation or snowpack measurements must consider forest effects. To facilitate computations, losses are grouped in three categories: (1) evapotranspiration losses, which include evaporation from the ground or snow surface, transpiration from all types of vegetation, and sublimation of water from snow surfaces to the atmosphere; (2) soil-moisture changes, which are computed on the basis of assumed fixed capacities of the soil to act as a storage reservoir which must be filled before runoff occurs and upon which evapotranspiration may draw; (3) interception loss, which is measured as the net difference between precipitation in the open and precipitation reaching the ground beneath the forest canopy. Delay to flow through ground and channel storage is accounted for by recession analysis. Deep percolation of water in underground channels, not measured as runoff at the gaging stations, is assumed to be negligible for snow laboratories.

4-01.06 Water-balance equations. - The water year selected for the purpose of this report ends on 31 August, at which time the snowpack water equivalent is zero or negligible for the laboratory areas.* Using the terms as defined in the preceding paragraphs, the water balance for a complete water year may be expressed as follows:

$$Q_{gen} = P - L \tag{4-1a}$$

where Q_{gen} is the generated runoff, P is the basin precipitation and L is the loss, all expressed in inches over the basin. For basins in which there is some carryover of snowpack water

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* This special period was selected to better define the annual precipitation regime for the snow laboratories since, hydrologically, this is the most quiescent time of the year. The conventional October-through-September water year would serve almost as well, however.

equivalent from year to year, or at the beginning or end of any other shorter period during the snow season, the above equation must be expanded to include a snowpack water equivalent term. A water-balance equation for any such area or for any portion of the year would thus be expressed as follows:

$$Q_{gen} = P - (W_2 - W_1) - L \qquad (4\text{-}2)$$

where $\underline{W_1}$ and $\underline{W_2}$ are the initial and final water equivalents respectively, expressed in inches. Losses, $\underline{L}$, may be subdivided as follows:

$$L = L_i + Q_{sm} + L_{et} \qquad (4\text{-}3)$$

where $\underline{L_i}$ is the interception loss, $\underline{Q_{sm}}$ is the change in available soil moisture, and $\underline{L_{et}}$ is the loss by evapotranspiration. The basin precipitation term may be subdivided into the following equations:

$$P = P_r + P_s = P_n + L_i = P_{rn} + L_{ri} + P_{sn} + L_{si} \qquad (4\text{-}4)$$

where the subscripts $\underline{r}$, $\underline{s}$, $\underline{n}$, and $\underline{i}$ refer to rain, snow, net, and interception, respectively. The interception loss may be excluded from total loss provided that total precipitation is replaced by net precipitation. Equation (4-2) may then be written as follows:

$$Q_{gen} = P_{rn} + P_{sn} - (W_2 - W_1) - Q_{sm} - L_{et} \qquad (4\text{-}5)$$

Since $\underline{P_{sn}} - \underline{(W_2-W_1)}$ represents melt, the water-balance equation may be written in the following form:

$$Q_{gen} = P_{rn} + M - Q_{sm} - L_{et} \qquad (4\text{-}6)$$

where $\underline{M}$ is the snowmelt in inches over the basin. Ground and channel storage is not explicitly included as a separate term in the above water-balance equation, since generated rather than observed runoff is used. In water-balance equations where observed runoff is involved, the equation would be changed by the following relationship:

$$Q_{gen} = Q + Q_g \qquad (4\text{-}7)$$

where $\underline{Q}$ is observed runoff and $\underline{Q_g}$ is change in ground and channel storage. Substituting in equation (4-1a) and transposing terms,

$$Q = P - L - Q_g \tag{4-8}$$

4-01.07 Water-balance components. - The preceding paragraphs have listed and defined the components to be considered and given the equalities to be used in water balances for areas of snow accumulation. The following sections deal with the technical aspects of the measurement of each of the variables, the reliability of measurements, and the general methods used in determining basin amounts from the point measurements. The elements so discussed are precipitation, interception loss, snowpack water equivalent, evapotranspiration, soil moisture, and ground-water storage and runoff.

4-02. PRECIPITATION

4-02.01 General. - The precipitation term in the water balance presents problems not ordinarily encountered in its use in the various index methods of determining basin precipitation amounts. The value of index methods in the determination of precipitation is dependent more upon their reliability as indexes than upon their representation of the true quantity. That is, an index of precipitation is acceptable if its deviation from the true value is a constant proportion of the true value. On the other hand, precipitation values used in a water balance must represent true quantities. A logical advantage claimed for precipitation indexes is that less rigorous treatment is required than if the true quantity were being determined. However, the claim of such an advantage can be justified only if it can be shown that the index is proportional to the actual amount. Unless the true value is known, however, the reliability of an index can be judged only by the results of its use. Runoff values obtained by statistical procedures often show considerable deviations from actual values; therefore, the reliability of the independent variables, including the precipitation index, remains questionable.

4-02.02 Problems of basin evaluation. - The discussion of precipitation in chapter 3 described the two basic problems involved in obtaining estimates of basin precipitation: (1) to obtain accurate values of point precipitation at the gage sites and (2) to estimate basin precipitation from these corrected station values. Deficiencies in gage catch are due primarily to turbulence and snow capping. Corrections for turbulence may be made on the basis of wind measurements, while loss of gage catch due to snow capping may be corrected by double-mass-curve analysis. Derivation of basin precipitation for mountainous regions from corrected station values is usually accomplished by the isopercentual method using a normal annual isohyetal map.

4-02.03 In the absence of a direct means of obtaining the annual basin precipitation, indirect methods may be employed. Using the basic water-balance equation,

$$P = Q_{gen} + L \qquad (4\text{-}1b)$$

the annual basin precipitation may be computed from observed runoff and estimated losses. Loss estimates may be made from accepted evapotranspiration equations (see section 4-05), and generated runoff is derived from direct observations of runoff by means of recession analysis (see section 4-07). In this method, there is no independent check on the water balance as a whole. On basins where precipitation is large in comparison with losses, the method provides reasonably accurate values of annual precipitation from which normal relationships between basin and station precipitation may be computed.

4-02.04 For basins where the quality of precipitation data is poor because of gage-catch deficiencies or poor areal coverage, true mean basin precipitation may be estimated by the use of adjustment factors derived from other components in the water balance. For example, during periods when precipitation is known to be in the form of snow over the entire basin and there is little or no loss or snowmelt, the basin precipitation must be the same as the change in the basin snowpack water equivalent. Computations based on monthly values will show seasonal variations in adjustment factors for converting observed station values to basin means. This would represent a refinement of the use of a constant adjustment based on normal annual values. The use of average relationships to relate station to basin precipitation, based on the average seasonal or annual point-to-basin relationship, was discussed in section 3-06. The use of average relationships is valid in areas where precipitation distribution is normally characterized by a relatively fixed pattern, as in the mountains of western United States.

4-02.05 Form of precipitation. - In the computation of monthly water balances, the basin precipitation must be separated into rainfall and snowfall. In areas where precipitation is predominantly in the form of snow throughout the winter period (as in the case of UCSL), this is a relatively minor problem. Occurrences of both rain and snow during the same month are usually confined to the fall and spring months. In the case of WBSL, however, the separation between rain and snow is a critical problem throughout the year except during summer, when precipitation amounts are so small that they are inconsequential in the annual total. Paragraph 3-02.03 described relationships which can be used in separating the rain and snow forms of precipitation.

4-03. INTERCEPTION LOSS

4-03.01 General. - The snow courses and precipitation gages, by which the hydrologist aims to sample basin precipitation, are usually located in the open and thus give no measure of the precipitation reaching the ground in that area of the basin which has a vegetation canopy. Part of the precipitation reaching tree-top level is intercepted by the vegetation surfaces and returned to the atmosphere by evaporation. This part of the basin precipitation is called interception loss, and refers to a permanent loss of precipitation to runoff; thus, it does not include temporarily stored water or snow which later reaches the ground by falling from overloaded branches or by melting and dripping. Although studies of interception loss usually concern trees or shrubs, even low-growing herbaceous plants may reduce the quantity of precipitation reaching the ground surface.

4-03.02 Interception loss warrants careful consideration because of its importance in the water balance. For example: for a moderately dense coniferous forest in an area with annual precipitation of thirty to fifty inches, conservative values of snowfall interception loss usually range between 15 and 30 percent of the total winter precipitation. Loss rates for summer rain usually range between 20 to 40 percent of the summer precipitation. Closer study of amounts of interception loss for different areas shows a wide range due to variation in the type and density of vegetation cover and in the type, magnitude, intensity, and frequency of storms. Consequently, interception-loss percentages should be chosen from studies done for areas of similar vegetation and climate. In water balances prepared without considering interception loss, the resulting error has been obscured by the use of unrepresentatively low precipitation values uncorrected for gage deficiencies. There is, however, increasing recognition of the advantages to be derived from using corrected precipitation data (i.e., data not affected by gage-catch deficiencies and unrepresentative gage or snow-course location). Consequently, there is an increased need to take account of interception loss in the water balance. Since studies of interception loss are now an integral part of present-day research in watershed management, it may be expected that more information on interception loss will be available in the near future.

4-03.03 Interception terminology. - In terms of the effect of vegetation on precipitation reaching the ground surface, total precipitation is divided into throughfall (precipitation which reaches the ground either by falling through spaces between branches and leaves or by dripping from vegetation surfaces), stemflow (precipitation which reaches the ground by flowing down stems after having been temporarily intercepted), and interception loss (intercepted precipitation which is returned to the atmosphere and does not reach the ground). Temporarily intercepted precipitation is referred to as drip and is part of the throughfall.

4-03.04 Interception loss is commonly expressed in terms of the percentage of loss to total precipitation for the selected time unit. The use of loss-percentage for relating measured interception loss to precipitation for a given season or water year has the disadvantage of being limited in application. A loss-percentage can be accurately applied only to the particular study area or to areas with closely similar climatic regimes as well as similar vegetation cover, as was previously mentioned. Presently available studies show an approximately linear relationship between loss and storm size for storms above a minimum size of about one-half inch. Below this minimum storm size, interception loss increases proportionally as storm size decreases. Virtually all of a light shower falling on vegetation is intercepted and evaporated. More useful forms of expressing findings on interception loss, such as the average amount of loss for storms in selected size ranges, are subsequently given herein.

4-03.05 In theory, interception loss is considered to be the difference between precipitation reaching tree-top level and that reaching the ground surface; in practice, however, interception loss is usually the measured difference between precipitation catch in an open area and catch beneath the vegetation canopy. These two differences are not necessarily synonymous. Differences in deposition are added to the difference in catch due to interception loss. A major factor in deposition difference is the difference in wind velocities in the open and in the forest. Environmental differences in gage-catch deficiency may aggravate the deposition difference and thus further obscure the true interception loss. (See chapter 3 for other factors in the micro-environment of the precipitation gage which influence precipitation-gage catch.) In measuring interception loss as the difference between gage or snowboard catch in the open and catch under canopy, the resulting values may be too high if stemflow is not also measured. Stemflow amounts to only about one to three percent of precipitation in coniferous trees, but may amount to more than 30 percent for deciduous trees or shrubs. (Values of stemflow are given in the table in par. 4-03.11.) If values of snowpack water equivalent measured at snow courses located in the open and in the forest are used to determine interception loss, there is less initial difference in catch due to environmental differences than there is if precipitation gages are used. However, here the differences in the environment result in differences in melt. Because of the many differences between the environment in the open and under canopy, many of the measured interception data are actually measures of what is more appropriately called "catch difference" than interception loss.

4-03.06 Measuring interception. - The basic instrumentation for measuring interception loss or "catch difference" consists of precipitation gages or snowboards installed beneath the canopy with similar control gages or snowboards located in an adjoining clearing. Where winter rain is negligible, snowpack measurements for sampling points under canopy and in the open may provide all or part of the data. The large possible variations in catch beneath even a single tree crown (due to random concentration of drip or complete shelter by overhanging canopy) necessitates use of many gages to assure representative data. More complete experiments may include: devices to measure stemflow (such as a water-tight collar around the tree trunk with connection to covered collection can); devices to measure throughfall (such as an impervious surface installed on ground beneath vegetation canopy with covered collection can for resulting runoff); and precipitation gages installed at tree-top level.29/ Correlative data for analyzing interception loss should include measurements of canopy density as well as a description of the type of vegetation. Other useful data include: water-surface evaporation, humidity, wind velocity, soil-moisture, runoff, and temperatures of the air, soil, and vegetation surfaces. Comparative studies may use vegetation as a variable by making observations on both untreated plots and treated plots altered by clearing out the forest understory, keeping the ground surface bare around the trees, or cutting trees according to selected densities.

4-03.07 Interception storage. - The major determinants of interception loss are the interception storage capacity of the vegetation and the evaporation opportunity. Interception storage capacity is the maximum quantity of water (or snow) which can be stored on the leaves and branches of a specific type and density of vegetation. It is usually expressed in the same units as precipitation, that is, inches depth of water or water equivalent over the area. Determinations of interception storage are made by analysis of precipitation and interception-loss data. Amounts thus determined for coniferous forests range from 0.01-0.12 inch for rainfall and 0.01 to 0.34 inch (water equivalent) for snowfall. (Interception storage of snowfall is not necessarily three times greater than rainfall—0.34 vs. 0.12—since these maximum values are for different areas.) Few studies give data on interception storage capacity for both snow and rain for the same area. In a study on Sierra Nevada ponderosa pine, 29/ interception storage capacity was determined as 0.09 inch for snow and 0.12 inch for rain. These data are not a conclusive measure of the comparative storage of rain and snow, because the data for snow are for storms designated as snowstorms if only 50 percent of the total precipitation was snow. Interception storage capacity appears to be primarily a function of canopy density. Other important determinants are: branching type (whether branches are essentially horizontal, as in many coniferous trees, or slanting); foliage type (shape and plane of leaves, and whether foliage is evergreen or deciduous); and vegetative type (tree, shrub, or herb) and height.

4-03.08 Canopy density. - Canopy density is probably the single most important parameter in the determination of interception loss, when considering interception loss in the same climatic region. In this report, canopy density refers to the percentage of the forested area which is covered by a horizontal projection of the vegetation canopy. It does not refer to all the area beneath the periphery of a tree canopy unless all the area is sheltered. Until recently, estimating canopy density has been tedious and comparatively subjective. Recently, however, several instruments have been developed which make possible an objective numerical measure of canopy density.28/ 14/ 18/ These canopy-density meters consist basically of a convex silvered glass surface on which a grid may be placed or cast. The instrument is located at the sampling point and levelled, and a reading is made of the number of points on the grid which are shaded by the canopy. Readings may be satisfactorily duplicated by different observers. A meter of this type, called a "ceptometer," has been used in detailed measurements of forest influences on snow accumulation and melt reported on by Ingebo 14/ (see plate 4-11, fig. 6). A present drawback to the use of a canopy-density meter, as well as other methods of estimating canopy density, is the lack of standardization on the size of the solid angle to be included in the measurement.

4-03.09 Basin canopy cover. - Estimates of basin mean canopy cover may be made as follows. First, the proportion of forested area to total basin area is computed, using aerial photographs or large-scale vegetation maps such as the forest-type maps of the U. S. Forest Service. From point measurements or estimates of forest canopy density, a mean canopy density for the forested part of the basin is determined. Basin canopy cover is the product of the percentage of forested area and the canopy density within the forested area (see table 2-1 for examples).

4-03.10 Evaporation opportunity. - Evaporation opportunity, which determines how much of the intercepted precipitation can be evaporated, varies greatly between regions and at different seasons in the same region. Furthermore, evaporation of intercepted precipitation differs in several respects from evaporation from more continuous water or snow surfaces. In the first place, the evaporation of intercepted precipitation can take place only as long as there is intercepted precipitation remaining on the vegetation. Secondly, the environment of water or snow stored on vegetation surfaces differs from that of water or snow resting on the ground. For example, vegetation surfaces have lower albedoes and warm more rapidly than water or snow surfaces. In addition, the intercepted precipitation is freely exposed to air circulation. Estimates of the evaporation of snow intercepted by conifers range from less than 5 to more than 20 times as high as snowpack evaporation (cf. Kittridge17/).

4-03.11 Interception related to storm-type. - Studies of interception loss for single storms show that the total amount of interception loss is closely related to the frequency of occurrence of precipitation-free intervals during the storm periods. For a storm with continuous precipitation, interception loss is limited to little more than the amount of precipitation stored on the vegetation at the end of the storm. During a storm including precipitation-free intervals, the interception loss may be several times as large as the interception storage capacity. This is graphically illustrated by studies by Rowe and Hendrix29/

and Hamilton and Rowe13/, among others. Summarized below are the results of studies which show the disposition of precipitation in areas having forests and shrub growth:

Experimental Area Vegetation	Bass Lake,* Sierra, Nevada Ponderosa pine	San Dimas, ** San Gabriel Mts. Chaparral Evergreen shrubs		North Fork,** Sierra Nevada Chaparral Deciduous shrubs
Interc. storage capacity (inches)	0.12	0.08		0.03 (winter & fall only)
Length of storm (hours)	70	73	27	23
Number of pcpt.-free intervals	5	12	1	1
Length of pcpt.-free intervals (hours)	28	36	4	1½
Total pcpt. (inches)	3.19 (100%)	3.35 (100%)	2.61 (100%)	3.14 (100%)
Throughfall (inches)	2.76 (86%)	2.69 (89%)	2.20 (84%)	1.87 (60%)
Stemflow (inches)	0.12 (4%)	0.28 (9%)	0.25 (10%)	1.17 (37%)
Interception loss (inches)	0.32 (10%)	0.38 (11%)	0.16 (6%)	0.10 (3%)

* Rowe and Hendrix29/
** Hamilton and Rowe13/

4-03.12 Interception-loss analyses. - There are two basic methods used in determining snow interception, both of which involve differences between measurements made under the forest canopy and at an open site. They are as follows: (1) measuring increments of snowfall (new snow) for individual storms; (2) measuring snowpack (accumulated) water equivalent. The snowfall data may be readily analyzed in terms of volume of snowfall for individual storms, an important variable in interception loss determination. The snowpack data used in interception analysis are usually for the maximum annual snowpack water equivalent. Interception-loss amounts computed by these two methods are not comparable unless it can be assumed that no melt has occurred since the beginning of snow accumulation, or that melt rates during the entire accumulation period are identical in the open and under canopy. In the analysis of interception loss for a series of individual storms, the results are often expressed in terms of the linear regression equation:

$$L_i = \frac{bP}{100} + a \qquad (4\text{-}9)$$

where L_i is the interception loss in inches per storm, P is the precipitation in inches for the storm, a is the interception storage capacity of the vegetation cover, and b is the percentage loss of precipitation during the storm. The effect of canopy cover is inherent in the data; hence, the constants should not be used for areas with canopy cover which is much different from that of the study area. Simple linear equations giving the relationship between canopy density and snowpack accumulation are also used to express interception loss. In this case, the dependent variable gives the snowpack water equivalent under canopy in percent of the snowpack water equivalent in the open, and the independent variable is the canopy density in percent of complete cover. The effect of storm size is inherent in such relationships; hence, the constants should not be used for areas which are climatically different from the study area.

4-03.13 Snowfall and rainfall interception measurements. - A summary is given in table 4-1 of snowfall or rainfall interception loss for different storm amounts and canopy densities. This table summarizes an unpublished study by Munns 25/ of forest influences in the San Bernardino mountains of southern California. These data are for a stand of Jeffrey pine at an elevation of 6000 feet; the density of the stand as a whole is 0.8. Total amounts of precipitation are given in the table for each of the storm-intensity classes to indicate the size of the sample for each class. Interception-loss percentages are expressed by the formula

$$L_i = \frac{P_{open} - P_{canopy}}{P_{open}} \qquad (4\text{-}10)$$

where L_i is interception-loss percentage, P_{open} is precipitation catch in the open, and P_{canopy} is the precipitation catch under canopy. Reference is also made to work of, Johnson 15/ and Wilm and Neiderhof 38/ for evaluation of interception by storm sizes.

4-03.14 Snowpack interception-loss measurements. - A graphical summary of selected data on interception loss as measured by snowpack data is shown in figure 5 of plate 4-11. Interception loss here refers to the difference between snowpack water equivalent under canopy and that in the open, expressed as a percentage of the water equivalent in the open. Most data are for maximum seasonal values of snowpack water equivalent. These percentage-losses are plotted against canopy data. Part of the scatter in the plotted points is due to the difference in the methods used by the various authors in measuring and expressing canopy density. Qualitative expressions of canopy density are shown as a line extending over the probable range of the qualitative term.

4-03.15 The most conclusive information yet available on the influence of canopy cover on snowpack accumulation is from data collected in the upper Columbia River basin by Ingebo for hundreds of snow sampling points intentionally located to sample various conditions of forest cover. A unique feature of the study is that canopy cover data were obtained for each sampling point by the ceptometer, an instrument which gives a numerical measure rather than an estimate of the cover directly above the point (see par. 4-03.08). Additional analyses were made of these data by the Snow Investigations.* Preliminary results of the correlation between canopy density and water equivalent are given below. (X is the canopy density in percent of complete cover and Y is the snowpack water equivalent for the various canopy densities, expressed in percent of the snowpack water equivalent in the open.) Graphical plots of the relationships are shown on figure 6, plate 4-11.

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* Basic data and preliminary analyses made available to Snow Investigations Unit through courtesy of the Missoula Research Center, Intermountain Forest and Range Experiment Station.

Year and no. of sample points	Regression equation	S_{yx}	D	r
1951 only (383)				
By individual points	$Y = 95.5 - 0.387\ X$	19.1	0.257	0.507**
By means of 10% canopy-density classes	$Y = 96.8 - 0.401\ X$	2.6	0.996	0.998**
1949, 1950 (340)				
By individual points	$Y = 99.9 - 0.359\ X$	19.1	0.237	0.487**
By means of 10% canopy-density classes	$Y = 99.9 - 0.366\ X$	5.9	0.805	0.897**

** = significant at the 99% level (highly significant)

4-04. SNOWPACK WATER EQUIVALENT

4-04.01 General. - Quantitative values for basin snowpack water equivalent must be used in a water balance. Indexes of basin snowpack water equivalent cannot be used for the same reasons that indexes of basin precipitation have no place in a water balance (par. 4-02.02). The geographical variation in snowpack accumulation over a given area has the same general pattern as the areal variation in total precipitation, since deposition of snowfall and of rainfall are similarly affected by the terrain of the area. In addition, the distribution of the snowpack is affected by factors which have no effect upon basin precipitation. These factors are the difference in forms of precipitation and the variation in melt rates. During the accumulation period, only the deposition effects upon distribution are appreciable, and usually an elevation parameter will adequately express differences in snowpack water equivalent. Data from snow courses which adequately sample a drainage basin with respect to elevation may then be related to basin amounts, providing no consistent bias results from other terrain factors. During the melt period, on the other hand, areal variation in melt rates tends to make snow course data unrepresentative of the basin water equivalent.

4-04.02 Snow chart. - The difficulty of evaluating basin water equivalent favors the use of an index as a means of evaluating the accumulation of snow. However, the validity of the water-equivalent index is questionable, not only for the reasons mentioned in connection with precipitation indexes (par. 4-02.02), but also for the additional reasons in the preceding paragraph. Accordingly, the index methods are generally inadequate to derive a measure of water equivalent which can be checked against independently derived values for the other terms in the water balance. Variation of the snowpack with elevation is a primary consideration when evaluating the snowpack water equivalent during the accumulation season (see chapter 3). The variation due to other terrain factors (e.g., orientation, slope and exposure) is usually less important and tends to be relatively constant from year to year. The snow chart, therefore, which has elevation as one of its ordinates, is an effective means of integrating snow-course measurements into basin mean snowpack water equivalent (see par. 3-08.04). For basins having relatively few snow courses, the difficulty of determining the water equivalent of the various elevation zones reduces the reliability of the results obtained by use of the snow chart. Also, its use is generally confined to areas within which there is a relatively consistent pattern of climatic conditions from year to year.

4-04.03 When using the snow chart to determine the mean snowpack water equivalent of a basin area, the volume represented by a line of best fit with respect to the plotted points can be considered to be a fixed percentage of the true value. The percentage correction factor may be derived from the water balance as a whole. This is most readily done by analyzing periods when precipitation is entirely in the form of snow and when snowmelt and losses are negligible, since basin precipitation can then be compared directly with water-equivalent change. If net precipitation values are used, as distinguished from total precipitation, the correction factor implicitly includes the effect of interception loss on the snow accumulation. (Snow courses are generally located in the open.) Also, since snow courses are usually situated in areas where local terrain favors above-average snow accumulations, values for the correction factor are generally less than unity, ranging from 0.75 to 0.90.

4-05. EVAPOTRANSPIRATION

4-05.01 General. - Part of the water which enters the soil is removed and returned to the atmosphere by evapotranspiration. This loss occurs not only while water is being supplied to the soil, but also as long as stored soil moisture is available (see section 4-06). In addition, evaporation may take place from the snow surface itself as well as from water surfaces and water intercepted by vegetation (condensation may also occur, it being considered negative evaporation). Evapotranspiration, like interception, represents a permanent loss to runoff. Average annual evapotranspiration losses for humid mid-latitude regions range between 15 and 30 inches, with smaller amounts for arid or alpine areas and larger amounts for areas with long growing seasons and an ample water supply during the growing season.

4-05.02 Knowledge of the amount of evapotranspiration loss is important to the water balance of an area in several ways. First, as one of the components of the water balance, it provides a partial check on the other components. This is particularly useful in the evaluation of precipitation. Since net precipitation is the sum of runoff and evapotranspiration loss, an estimated value of net precipitation can be determined in this manner for comparison with the computed value of net precipitation. Such a check is especially useful for a basin with heavy precipitation or with a significant part of the precipitation falling as snow. Here, gage-catch deficiencies or errors in determining basin precipitation from point measurements may go unsuspected if there is no such check. The value of the check results from the relative magnitude of evapotranspiration and precipitation. Because evapotranspiration is usually one of the smaller components in the water balance for areas of significant snowmelt runoff, the errors in computing evapotranspiration are relatively small in comparison with the errors in computing precipitation, the largest item in the water balance. A second way in which a knowledge of the amount of evapotranspiration is important to the water balance is in computing the soil-moisture deficit, as discussed in paragraphs 4-05.11 and 4-06.19.

4-05.03 Evapotranspiration terminology. - Definitions of evapotranspiration differ as to which of the component parts of total evaporation are included. In this report, evapotranspiration is considered to include transpiration by plants, evaporation from soil particles, and evaporation from the snow surface. The other components of total evaporation, not included as evapotranspiration in this report, are interception loss and evaporation from lakes or other water bodies. Transpiration and soil evaporation are included in one term since most experimental data combine them because of the difficulty of measuring transpiration separately from soil evaporation.

4-05.04 Potential vs. actual evapotranspiration. - Potential evapotranspiration is the amount of water which would be lost by transpiration and evaporation if sufficient water were available in the soil at all times to meet the demand. Potential loss is determined by the energy supply, without reference to the water supply. Actual evapotranspiration refers to the actual loss resulting from the combined effects of the demand and the available water supply. On an annual basis, actual loss is almost invariably less than potential loss, since even in areas with high annual precipitation, the summer water supply (precipitation plus stored soil moisture) is usually not large enough to meet the demand throughout the entire summer. In general, there is much less areal variation in heat supply than there is in water supply, particularly in the mountain watersheds where, due to orographic effects, the areal distribution of precipitation is characterized by large variations. As a result, areas with large ranges in normal annual precipitation usually have much smaller ranges in actual evapotranspiration loss. This is especially true of areas where much of the precipitation falls in winter, the time when the potential evapotranspiration loss is at a minimum.

4-05.05 Transpiration. - Transpiration refers to the loss of water in vapor form from living plants. This loss is not to be confused with the evaporation of water from the outer surfaces of the plant (which is termed interception loss); transpiration loss occurs from within the leaves of the plant. Most of the transpiration loss occurs through stomata (very small openings in the lower surfaces of leaves). Water-vapor loss ordinarily occurs only during the daylight hours while the stomata are open. The vapor-pressure gradient is almost always directed outward from the leaves, resulting in loss of water molecules from the leaf. (Because the leaf temperature is usually warmer than the surrounding air during the day, its saturated vapor pressure is greater than that of the air, even for air with 100 percent relative humidity.) Because of the arrangement of cells within the leaf, the internal surface of the leaf is many times larger than the external surface.10/ The diffusion of water vapor through stomata can take place at a high rate. In general, transpiration is a very efficient means of water loss. Botanists have recorded annual transpiration losses of more than 100 inches of water.16/

4-05.06 Soil evaporation. - Unlike transpiration, soil evaporation is limited by the difficulty of moving the water stored in the soil up to the evaporating surface. In transpiration, water is withdrawn by the roots and transported inside the plant up to the evaporating surface in the leaf. In soil evaporation, water must be transported up through the soil to the evaporating surface. Since the permeability of the soil decreases sharply as the water content of the soil decreases, even though a steep vapor-pressure gradient may exist at the soil surface, soil evaporation may be restricted because capillary rise of water in the soil is slow. As a result of this retarding effect of permeability upon capillary rise, evaporation becomes decreasingly effective with increasing distance of the water from the soil surface. Consequently, plant roots usually remove stored soil moisture to a considerably greater depth than soil evaporation alone (see par. 4-06.08).

4-05.07 Evapotranspiration formulas. - Since it is not practicable to install and service the instrumentation necessary to measure evapotranspiration directly in all areas where such data are needed, it must be estimated by means of an appropriate formula. Many formulas have been developed to express the relation between observed evapotranspiration data and the concurrent hydrometeorological conditions. A formula used to compute evapotranspiration amounts in the water balance should meet the following requirements: good agreement with measured quantities; applicability to climate and vegetation of basin area; basic data ordinarily available as to variety and detail; basic time period of one month or less; and, if possible, quick computation. A formula which meets each of these requirements at least moderately well is that of Thornthwaite.33/34/ Whereas other formulas have been shown to reproduce measured loss more accurately for specific sites, these formulas require more data than is ordinarily available. Such formulas include those of Penman26/ (data required: duration of bright sunshine, air temperature, air humidity and wind speed) and Halstead12/ (data required: maximum and minimum air temperature). A promising method of computing loss for large regions, using radiosonde data, is based on the net increase in water-vapor content of the air in passing over a given region. 2/ 3/ This mass transfer method appears practicable only for large regions; it has been applied satisfactorily to regions as small as the Ohio River basin. 5/

4-05.08 Thornthwaite's evapotranspiration method. - From an analysis of the use of water by many kinds of vegetation, Thornthwaite concluded that climate was the principal determinant of evapotranspiration loss and that the type of vegetation and the character of the soil made relatively little difference. Limiting himself to a consideration of the climatic elements for which data are generally available, he found that the potential loss in any area could be evaluated satisfactorily by an empirical formula using only air temperature as a variable. In addition, an adjustment must be made for length of day or number of hours of possible sunshine (which vary with latitude and season). Thornthwaite's method also includes a monthly bookkeeping method by which monthly actual evapotranspiration is obtained by balancing potential loss against supply (precipitation and available soil moisture). The required basic data are latitude of station, monthly mean air temperature, and monthly precipitation. In addition, information on the average storage capacity for available soil moisture within reach of plant roots is required. Thornthwaite suggests that an average value of four inches of water may be used in default of specific local information.

4-05.09 In Thornthwaite's evapotranspiration method, the effect of latitude and season are standardized to a standard month of 30 days with 12 hours of possible sunshine each for convenience of computation. Using the formulas presented in the following paragraph, values of "unadjusted" potential evapotranspiration are computed on the basis of the standardized month. These values are then adjusted for the number of hours of possible sunshine for the given latitude and month.

4-05.10 Thornthwaite's specific formula for computing potential evapotranspiration postulates that evaporation and transpiration vary with temperature as expressed in the general formula*

$$e = ct^{a} \tag{4-11}$$

where $\underline{e}$ is the monthly potential evapotranspiration in cm, $\underline{t}$ is the monthly mean air temperature in °C, and $\underline{c}$ and $\underline{a}$ are coefficients which relate evapotranspiration to monthly mean air temperature. The coefficients $\underline{c}$ and $\underline{a}$ are both functions of an annual heat index, $\underline{I}$, which is the summation of monthly indexes $\underline{i}$ for the twelve months of the year. The monthly heat indexes are computed by the formula

$$i = (t/5)^{1.514} \tag{4-12}$$

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* Thornthwaite's usage of symbols is followed here; the reader is cautioned that the symbol $\underline{e}$ has been used elsewhere in this report to denote vapor pressure.

where $\underline{i}$ is the monthly heat index (dimensionless), and $\underline{t}$ is the monthly mean air temperature in $^{\circ}C$. For the range of $\underline{I}$ from 0 to 160, the exponent $\underline{a}$ ranges from 0 to 4.25; $\underline{c}$ varies inversely with $\underline{I}$. From the above relations, and incorporating results from his earlier work on temperature-evaporation relations, Thornthwaite derived the following specific formula for potential evapotranspiration:

$$e = 1.6 \ (10t/I)^a \qquad (4\text{-}13)$$

The author supplies tables and a graph which make it easy to compute potential evapotranspiration, as given by the above equation, and also to determine from this, the "adjusted potential evapotranspiration." (See Research Note 20.)

4-05.11 To compute so-called actual evapotranspiration, values of adjusted potential evapotranspiration are used in the month-to-month bookkeeping method presented by Thornthwaite. Demand (adjusted potential evapotranspiration) is balanced against supply (precipitation and available soil moisture). An average value of four inches (10 cm) of water is given for the storage capacity for available soil moisture within reach of most plant roots. As long as demand is met by precipitation or precipitation-plus-soil-moisture, actual evapotranspiration is equal to potential evapotranspiration. When precipitation and the water stored in the soil are insufficient to meet demand, actual evapotranspiration is less than potential. When the water stored in the soil has been depleted, it must be recharged (at a rate not exceeding the infiltration-capacity of the soil).

4-05.12 The quality of actual-evapotranspiration values computed on a monthly basis is affected by the within-month variation in temperature or precipitation. This variation is particularly important in the case of precipitation. Even though the total precipitation recorded for the month may be more than adequate to meet the total demand, it may not be available for use throughout the month. For example, most of the monthly total may fall in the last few days of the month, thus satisfying demand for only these days; or it may fall in a few high-intensity storms with little infiltration. The quality of monthly actual evapotranspiration values may be considerably improved by computing the loss on a daily basis. The probable improvement is most significant for months when stored soil moisture at the beginning of the month is not adequate to fill the demand during the month. An inspection of daily precipitation records for such months will usually show a few periods when supply and demand do not overlap. Computation on a daily basis is advisable for such periods.

4-05.13 Several minor modifications in the Thornthwaite method have been suggested for application to snow-covered areas. During the snowmelt period, values computed for potential evapotranspiration may be somewhat low. This is because the air-temperature data on which the formula was based were generally measured over snow-free surfaces; the air temperature in a snow-covered basin is usually measured above a snow surface. Reference is made to Research Note 17 for a discussion of the relation between temperature data measured above a ground surface and above a snowpack surface. In brief, while there is an energy-absorbing snowpack on the ground, air temperature does not represent the incoming energy supply as it does during other times. The potential evapotranspiration loss for snow-covered areas is thus probably greater than that computed by Thornthwaite's method since, for the same insolation, the measured air temperature is not as high as it would be over snow-free ground. Another minor modification which should be made when computing actual evapotranspiration for snow-covered areas is to include snowmelt as a supplementary source of water supply. Thus, precipitation and snowmelt should both be used to satisfy the potential demand before drawing on stored soil moisture. Equal consideration should be given to both the amount of snowmelt during the month (setting the limits to the quantity which could be used) and the mean area of snow cover (limiting the area where the supply is available).

4-05.14 In computing evapotranspiration loss by Thornthwaite's method, certain periods are critical, namely, times when appreciable demand (warm air temperatures) and water supply (precipitation and available soil moisture) overlap. (Hence soil moisture storage capacity is also critical in the determination of the available moisture supply). Soil moisture storage capacity is particularly important for areas with marked summer drought. On the other hand, quality of winter precipitation data is not significant in computing actual evapotranspiration for most areas where snowmelt is important, because winter precipitation usually exceeds the low potential evapotranspiration amounts of winter. The Thornthwaite formula presupposes measurable water use to begin at monthly mean air temperatures above 32°F, so no precipitation data are required for months with mean air temperatures at or below 32°F.

4-05.15 Thornthwaite has pointed out that when the albedo of the vegetation surface is higher or lower than average, the potential loss rate will be lower or higher, respectively. 34/ Also, the potential loss rate applies only to closely-growing vegetation. The effects of decreased density and increased exposure cannot be assessed in general terms. In view of the empirical basis of his formula, Thornthwaite advises for any area where it is used that the formula be tested against reliable measured evapotranspiration data. Such comparisons have shown the formula to be satisfactory in many areas.22/

4-05.16 The results of the use of the Thornthwaite formula with data from snow laboratories are not conclusive. Comparing computed evapotranspiration with the net precipitation minus runoff, results were good for UCSL. Since at WBSL the value for precipitation was not independent of loss, agreement between computed and residual values was not significant. At CSSL, while computed potential loss agreed well with the difference between net precipitation and runoff, actual loss (computed using the assumed soil-moisture storage capacity) was considerably lower than this residual. This lack of agreement may be due to incomplete measurement of outflow, as a result of unmeasured deep-percolation (see par. 4-10.08). Good results were obtained using the Thornthwaite method to compute evapotranspiration for the 438-square mile basin of the North Santiam River above Detroit (see Research Note 22).

4-05.17 Thermodynamics of transpiration at WBSL. - An independent evaluation of transpiration in the heavily forested WBSL during active snowmelt was made on the basis of energy-balance computations for periods of local climate (when advection of energy by the airmasses was known to be negligible). For this case, the only external source of heat energy is solar radiation. The measured quantity of insolation may be balanced against the energy used for snowmelt, transpiration, and loss through the atmosphere by longwave radiation. (The energy required for photosynthesis is negligible, having been estimated to be less than 3 percent of the energy absorbed by the tree crowns.)24/ Because the amount of snowmelt is a measured quantity and the amount of longwave loss can be estimated from theoretical considerations, the energy required for transpiration may be treated as the residual in the basinwide energy balance. Transfer of heat by convection from the needle surfaces to the adjacent air need not be considered. Considering the area as a whole, it represents merely an intermediate process in the transfer of heat to the snow surface. When dealing with a snow-covered area, the energy balance as outlined above does not involve the negligible changes in the storage of heat in the ground. The energy-balance computations for the WBSL were made for a five-day clear-weather period in May of 1949. The residual energy, expressed in terms of transpired water, represents the potential transpiration rate for that time of year. Details of the study are presented in Supplement to Research Note 19 and summarized below.

4-05.18 The diagram shown on plate 4-12 illustrates schematically the daily mean balance of energy exchange for the snow-covered area of Mann Creek basin, WBSL, for the period 9-13 May 1949. The net allwave energy input of 490 lys per day was divided almost equally between transpiration and melt: 263 lys per day were used directly for transpiration; 227 lys per day were transferred to the snowpack by longwave radiation and convection, resulting in snowmelt. In addition, about 45 lys per day were transferred to the snowpack as a result of the condensation of water vapor transpired from the forest. This heat of condensation constituted a secondary heat supply for melting the snow. The net generated runoff for the period was 1.24 inches per day, representing an energy equivalent of 244 lys per day (heat of fusion of snow approximately equal to 198 cal per inch of resultant melt). The net transpiration loss to the atmosphere was 0.14 inches per day, representing an energy equivalent of 246 lys per day (heat of vaporization of water approximately equal to 1520 cal per inch of water evaporated). The gross transpiration rate, including water vapor condensed on the snow surface, was 0.17 inches per day—the maximum potential transpiration rate for the specified conditions. The potential evapotranspiration rate as computed by Thornthwaite's method for this condition is 0.165 inches per day. From the results of an energy-balance analysis therefore, the potential transpiration rate for this time of year as computed by Thornthwaite's method, appears to be reasonable.

4-06. SOIL MOISTURE

4-06.01 General. - The soil functions as a reservoir, storing water when available to be used during periods when potential evapotranspiration exceeds current supply. Under average conditions the depth of water stored as soil moisture available for use is about four inches.35/ In extreme cases, however, it may be less than one inch or more than 20 inches. Such a wide range in possible amounts makes accurate evaluation of the soil-moisture capacity of individual basins difficult. From the standpoint of computing basin soil-moisture storage capacity, data on soil-moisture storage and movement are inadequate, and empirical values of soil-moisture storage capacity which may be used in the actual evaluation of soil moisture are generally lacking. A brief review of soil and soil moisture is included here in order to assist in the interpretation of available information on soils.

4-06.02 Only a part of all the moisture in the soil is involved in the water balance: the stored soil moisture which can be removed by plant roots and natural evaporation. Since this available soil moisture is not measured directly, even in point measurements, it must be indirectly estimated. Methods by which this can be done are reviewed briefly in the following paragraphs. The terminology used is not that of the soil scientist, but that in most common use in hydrological studies. The discussion is a simplification of the complex interaction of forces controlling soil-moisture movement and content. Reference is made to the text, Applied Hydrology 19/, which contains a survey of the field of soil-moisture theories and a review of soil physics. Reference is made to other texts which discuss more specific aspects of soil moisture with respect to forest soils 21/ 16/ and with respect to the hydrologic cycle. 6/

4-06.03 Soil-moisture terminology. - The term soil is used here in its agricultural or soil-science usage: the surface layer of the earth, adapted by soil-forming processes to support plant life. Soil as thus defined is only the weathered top layer of the total mass of earth materials of concern in soil mechanics. This top layer is the zone from which stored water may be removed by transpiration and evaporation. Soil is made up of (1) a relatively inert "skeleton" of larger unweathered mineral particles, primarily sands and silts; (2) a physically and chemically active part consisting of tiny, plate-like clays, super-clays, and colloids, plus particles of humus; (3) water; (4) gases. In forest soils, the surface layer, consisting of partly decomposed vegetation (litter or duff), is usually at least several inches thick. The soil profile is the vertical section from the surface down to the unaltered parent material. A systematic vertical variation in texture and composition is typical for soils which have been subjected to seasonal variations in heat and water supply. Soils are commonly grouped into texture classes on the basis of the proportion of particles within specified size ranges (for example, sandy loam). The water storage capacity of soil is principally determined by its texture. This storage capacity is, however, affected by other factors such as the chemical activity of the soil particles, the shapes and arrangement of the particles, the proportion of admixed humus (decomposed vegetation), and the stoniness of the soil. Consequently, considerable variation is possible even in the storage capacity of soils of the same texture group.

4-06.04 The part of the soil moisture which is in permanent storage, and which cannot be removed from the soil by plant roots or evaporation under natural conditions, is the water content that exists at the permanent wilting point (commonly abbreviated as PWP). Although terminology varies, the terms wilting percentage and wilting coefficient may be assumed to refer to water content of the soil at which plants wilt beyond recovery. Although there is some variation among different species and for different stages of growth, the PWP is approximately the same for all plants in a given soil. Both plant roots and evaporation processes in the soil exert about the same maximum force to remove water films from soil particles; consequently, the PWP itself for a given soil is not affected by the presence or absence of plant cover.* The water content left in the soil at the PWP is appreciable. It ranges from less than one-half inch to more than two inches of water per foot depth of soil, increasing with increasing fineness in soil texture. The soil moisture in the soil at PWP is held tightly in the soil. In a laboratory, for example, in order to remove this remaining moisture, soil must be heated to a temperature above the boiling point of water for 24 hours. The PWP for a given soil is determined by growing plants under specified conditions. As a rough approximation, it is equal to about half the field capacity or moisture equivalent, discussed in the next paragraph. An approximate measure of PWP, used when laboratory data are available, is the water content when the tension in the soil sample is at 15 atmospheres. 7/21/

4-06.05 The field capacity (or field moisture capacity) is the upper limit to the amount of water which can be stored in the soil. It is the amount of water left in an initially saturated soil with unobstructed drainage after the downward movement of soil moisture has "materially decreased." Field capacity thus includes the soil moisture below the PWP as well as the available soil moisture. Field capacity is hypothetically equivalent to the capillary-moisture-holding capacity of the soil, or to the total amount of water which can be held against the force of gravity under natural conditions. Actually, gravity is only one of the directional forces acting on water in the soil. The total water content of the soil is the net result of all the directional forces or tensions affecting soil-moisture movement at a given time. Field capacity is an arbitrary measure which, like PWP, is widely used because it represents a useful quantity,

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* Since plant roots commonly penetrate deeper than does effective soil evaporation, the total quantity of soil moisture removed seasonally is usually considerably larger when a plant cover is present. See 4-06.08.

notwithstanding the fact that it is not a true equilibrium point on the curve of moisture depletion versus tension. For the purposes of the water balance, it may be assumed that, after field capacity has been reached, there is little additional downward movement of soil moisture to the ground-water table. The time required for soil to drain to field capacity is about one to five days or more, being shorter for sandy soils and longer for fine-textured soils. Two to three days is the period commonly accepted for medium-textured soils.21/ The amount of soil moisture at field capacity ranges from about one inch to over four inches of water per foot depth of soil, for very coarse and fine-textured soils, respectively. A laboratory measure of water content, which is approximately equal to field capacity, is moisture equivalent (M.E.). It is the water content remaining when an initially saturated soil sample is centrifuged under standardized conditions. For most fine-textured soils, M.E. is nearly the same as field capacity; for sandy soils, it is lower. An approximate measure of field capacity is the amount of water in a soil sample when the tension in the soil is at one-third atmosphere; for moisture equivalent, one-half atmosphere.

4-06.06 Values of soil moisture. - Empirical values of storage capacity for different soil texture classes are given below. (Storage capacity for available soil moisture is equal to the difference between field capacity and capacity at the permanent wilting point.) The values given provide only rough approximations for any given soil. In using these values to compute basin soil-moisture storage capacity, it should be kept in mind that in many soils the texture of the surface is often very different from that of underlying horizons. All values are in units of inches of water per foot depth of soil.

Texture Class	Field Capacity			Permanent Wilting Point			Available Soil Moisture		
	a/	b/	c/	a/	b/	c/	a/	b/	c/
Sand	1.2	-	-	0.3	-	-	0.9	-	0.5
Fine Sand	1.4	1.6	-	0.4	0.8	-	1.0	0.8	-
Sandy Loam	1.9	-	-	0.6	-	-	1.3	-	-
Fine Sandy Loam	2.6	-	-	0.8	-	-	1.6	-	1-1.5
Loam	3.2	-	-	1.2	-	-	2.0	-	-
Silt Loam	3.4	3.2	-	1.4	1.6	-	2.0	1.6	-
Light Clay Loam	3.6	-	-	1.6	-	-	2.0	-	-
Clay Loam	3.8	-	-	1.8	-	-	2.0	-	1.5-2
Heavy Clay Loam	3.9	-	-	2.1	-	-	1.8	-	-
Clay	3.9	4.4	-	2.5	2.3	-	1.4	2.1	-

a/ Mean values for texture classes from graph "Typical water-holding characteristics of different-textured soils," USDA Yearbook of Agriculture, 1955, p. 120.37/

b/ Computed from percentage weight of moisture data for field capacity or moisture equivalent, Lutz and Chandler, p. 294;21/ specific gravity of dry soils from ASCE Hydrology Handbook, p. 137.1/

c/ ASCE Hydrology Handbook, p. 134.1/

4-06.07 Soil moisture and the water balance. - The measure of soil moisture needed for the basin water balance is the usable storage capacity for available soil moisture. Available soil moisture is usually expressed in inches of water per foot depth of soil. To estimate the total storage capacity for available soil moisture, the amount per unit depth is multiplied by total depth of soil. The essential item for the water balance, however, is not the total depth of stored water but only the part which is within reach of plant roots and soil evaporation, as discussed in the next paragraph. A preliminary step is the estimation of total capacity. Since information on depth of soil is often given in qualitative terms, some commonly used quantitative equivalents are given below. Two groups of figures are given because the same qualitative terms are commonly used in spite of the fact that there is considerable difference in the mean maximum depth of the soil for different areas. In general, forest soils (that is, the soils in present forest areas) are usually not as deep as agricultural soils. The depths given below for forest soils are in use among foresters; 21/ those for agricultural soils are average values for agricultural areas in general. Local usage may differ considerably.

Descriptive terminology	Depth of soil in inches	
	Forest soils	Agricultural soils
Very shallow	<6	<12
Shallow	6 - 12	12 - 20
Moderately deep	12 - 24	20 - 36
Deep	24 - 48	36 - 60
Very deep	>48	>60

4-06.08 How much of the total storage capacity will be used is largely determined by the depth to which plant roots penetrate the basin soil. Evaporation from the soil is usually insignificant below the top foot or two, and rarely penetrates further than plant roots. Available information on root depths is incomplete and in part contradictory. In general, water-absorbing roots are concentrated in the top two to three feet of the soil. Pending better information, the following values indicate the range in average normal root depth for major vegetation types:

Vegetation type	Root depth (feet)
Coniferous trees	2 - 5
Deciduous trees and evergreen broad-leafed trees	3 - 6 or more
Evergreen shrubs (chaparral)	2 - 6 or more
Deciduous shrubs	2 - 6
Tall herbaceous vegetation (principally grasses)	2 - 5 or more
Low-growing herbaceous vegetation	1 - 2

Root depth may be restricted by impervious hardpans, high ground-water tables, or the shallow soils often characteristic of steep slopes in mountainous areas. Within the ranges suggested, plants in summer-dry areas commonly have deeper roots than plants in summer-rain areas. The range of average maximum depth for specific types of vegetation is, of course, larger than the above values.

4-06.09 The light, porous forest litter on most forest soils intercepts and retains for later evaporation some of the precipitation which reaches the forest floor. On the other hand, it retards soil evaporation. Although the moisture-holding capacity of the litter is high in terms of its weight, the amount of water involved is usually small. Under average conditions of depth and porosity, the water-holding capacity of the litter is only about 0.1 inch of water. An extremely deep accumulation of very porous litter can hold as much as two inches of water.16/ Litter is not considered to be a source of moisture supply to plant roots; where it is deep, it may be considered an additional component of interception loss.

4-06.10 Measuring soil moisture. - There are two basic methods of measuring the quantity of water in soil: (1) sampling methods which require soil samples for each laboratory analysis and (2) non-destructive methods which measure soil moisture at a given point by means of meters which remain in the soil. If a quantitative measure of the water content is needed, soil-moisture meters must be calibrated by laboratory analysis of the soil at the location and soil depth where the meter unit is to be installed. Meters may, however, serve as indexes of soil moisture without prior establishment of the relation between meter readings and the actual quantity of water in the soil.

4-06.11 Laboratory measurements. - The basic method of laboratory analysis of soil moisture is known as the gravimetric method. It determines the percentage weight of the water relative to the weight of the dried soil. Soil samples are weighed, oven-dried, and re-weighed. The soil-moisture percentage is the ratio of the decrease in weight to the weight of the dried soil. Because the water-holding capacity of soil is usually different for the various horizons or layers within the soil, it is necessary to take samples at intervals throughout the depth of the soil. Soil-moisture percentages may be assumed to be in percentage by weight unless specifically stated otherwise.

4-06.12 The percentage of soil moisture by volume (and consequently, the derived value for inches depth of water per unit depth of soil) is rarely measured directly. Instead, data for percentage soil moisture by weight are used to compute volumetric measures of soil moisture. Because of the considerable range in density of soil, the adjustment for weight per unit volume adds further sources of error. The difficulty in removing and moving samples without compressing the soil is the principal reason for this error. Since natural, uncultivated soils usually have a more open structure than agricultural soils of the same texture class (not having been subjected to compaction), a considerable error in true volume may result from compression of soil samples during removal and transportation to the laboratory.

4-06.13 <u>Field measurements</u>. - Many different types of instruments make measurements of soil-moisture variation without removing samples for each observation.[20/] [11/] Two types currently in common use are tensiometers and electrical resistance-type soil-moisture meters. Electrical meters in common use measure the range of available soil moisture. Tensiometers are not relevant here because they operate in the limited range of moisture between field capacity and saturation and do not include the range of available soil moisture. The operating principle of the electrical resistance-type meters makes use of the variation in the electrical resistance of a porous non-conductor with its moisture content. The greater the water content, the smaller the electrical resistance. The meters consist of two parts: (1) the sensing unit (or soil unit) which remains in place in the soil and consists of a porous non-conducting block with embedded electrodes and lead wires extending up through the soil surface; (2) the metering unit which measures the electrical resistance of the buried block.

4-06.14 The electrical soil-moisture meters in most common use at present are those developed by Bouyoucos and Mick 4/ and by Colman and Hendrix 6/. In the Bouyoucos meter, electrical resistance is observed by the null balance of a modified wheatstone bridge. The porous non-conductor in the basic Bouyoucos block is plaster of paris (gypsum); later modifications of the basic unit use nylon and plaster of paris. The Colman meter measures resistance by means of a battery-operated alternating-current ohmmeter; dial readings are in resistance. The Colman meter also includes a resistance-type thermometer in the soil unit. Using a separate circuit, readings of resistance for the thermistor element are made by means of the same metering unit. The temperature data are used to convert measured resistance in the porous block to a common temperature base. The porous material used in the Colman meter soil unit is fiberglass; it is sandwiched between monel-metal screen electrodes.

4-06.15 The sensing elements of the electrical soil-moisture meters appears to be the source of most of the disadvantages of the instruments themselves as far as hydrologic application is concerned. The size of the element must not be so large that there is an excessive time lag between the moisture changes in the surrounding soil and in the porous material between the electrodes. On the other hand, it must not be too small to maintain satisfactory contact with the soil throughout the random slight movements due to shrinkage from drying, etc., which occur seasonally. The porous material itself must be durable and should have an adequate range in resistance (or other electrical property) for the range in moisture content encountered in the soil. Although the material must not be so chemically active that it disintegrates in the soil, it must have sufficient "buffer effect" to insure that an accumulation of dissolved salts from percolating soil water will not significantly affect the electrical resistance between the electrodes.

4-06.16 A dozen different soil units were tested under field conditions at CSSL. Descriptions of the soil units used and results of the testing are presented in "A review of soil moisture measuring methods and apparatus" by Gerdel in Miscellaneous Report 2. The testing program was undertaken after it was found that some of the plaster-of-paris Bouyoucos blocks in use at CSSL had disintegrated after less than one year in the soil. Resistance readings from the Colman meter (1947 model) were occasionally erratic, when compared with both the Bouyoucos-block readings and the hydrological conditions. These errors were attributed to the lack of buffer effect in the chemically inert materials of the Colman units and to unsatisfactory contact between the soil and porous material of the block when soil shrinkage occurred. The Bouyoucos-block units gave readings which appeared more satisfactory; however, some of the blocks disintegrated in less than six months. The other experimental units, none of which are commercially available, had various disqualifying disadvantages. Subsequently alterations and improvements have been made in both the Bouyoucos blocks and Colman meters. In 1949, Colman meters were installed at nine stations at UCSL by the U. S. Forest Service as part of a forest-effects research program. On the basis of preliminary analysis, the data from these meters appeared to agree with concurrent hydrological conditions. The less satisfactory performance of the Colman meters at CSSL may be due to the more alkaline soils and to the use of the earlier model of the meter. In summary, the experience at the snow laboratories with electrical soil-moisture meters indicated that none of the meters used could be recommended for field use with reservations.

4-06.17 Calibration of soil-moisture meters. - Electrical soil-moisture meters must be field calibrated to relate meter readings to concurrent soil-moisture content for each soil-sampling site and each level of measurement at each site. Calibration consists of removing separate soil samples from the area adjacent to the meter and making laboratory determinations of the percentage moisture content by weight (gravimetric analysis). Samples of soil and concurrent resistance readings are taken throughout the cycle of saturation and drying. To define the relation between the electrical resistance of the porous block and the soil-moisture content of the soil, soil scientists recommend taking at least 20 sets of measurements.[20] The field moisture cycle cannot be satisfactorily duplicated by saturating and drying a soil sample in the laboratory. The magnitude of errors resulting from improper calibration procedures is discussed by Remson and Fox.[27]

4-06.18 Point-to-point variation in soil moisture makes it difficult to determine true basin soil moisture from actual measurements. Though point measurements may be accurately made, too few samples or a poor sampling plan may result in biased results. The significance of this source of error is proportional to the importance of soil moisture in the water balance as a whole. The areal variation referred to is for the same soil type, at the same depth in the profile. Even larger variation is possible between different soil types in the same area. In general, areal variation in moisture content is greater in uncultivated soils, particularly in mountainous areas, than in cultivated soils. Recent studies of areal-sampling methods and results are summarized in "Soil moisture measurements."[20] The topography and environment of soil-moisture measurement sites, as those of snow courses and ground-water wells, have a strong influence on the measured soil-moisture amounts. To mention a single determinant, the effect of slope on soil moisture is discussed by van't Woudt.[36]

4-06.19 Computation of soil moisture. - For areas where measured values are not available, the amount of available soil moisture at a given time may be computed from hydrometeorologic data. In a method presented by Mather, computations are made on a daily basis using the procedures from Thornthwaite's method of computing monthly amounts of "actual" evapotranspiration by the water-balance bookkeeping method discussed in 4-05.11.22/ The data required are daily precipitation and air temperature and the estimated quantity of usable storage capacity for available soil moisture. Beginning at a time when the amount of stored water in the soil is known--conveniently, either a maximum (field capacity) or a minimum (wilting point) value is used--, the quantity of soil moisture in the soil is computed by maintaining a daily budget of all additions (precipitation and snowmelt) and withdrawals (evapotranspiration). Published data show good agreement between computed and measured quantities of soil moisture.22/35/ In a method presented by Snyder,31/ the changes in storage of available soil moisture are computed as part of a procedure for computing daily runoff and its component surface and ground-water flow. Procedures are presented for analyzing streamflow and precipitation data in order to establish empirically the relationship between the ground-water component of flow and the initial loss to runoff attributable to soil-moisture deficiency. Together with an empirical formula for estimating evapotranspiration, this relationship is used in a method for computing daily runoff. Good agreement was shown between computed and observed amounts of runoff.

4-07. GROUND-WATER STORAGE AND RUNOFF

4-07.01 General. - From the hydrologic viewpoint, runoff may be considered the last phase of the hydrologic cycle and the end product of all that precedes it. Similarly, it is considered the dependent variable in mathematical expressions of the water balance. Runoff measurements are usually regarded as the most accurate of any variable in the water balance. This is because the measurement of runoff, unlike measurements of other variables which sample only points within an area, effectively integrates the entire area from which the measured flow originates. Even so, the measurement of runoff entails uncertainties in the water balance. These result from errors in the measurement of the runoff itself particularly during periods of ice effects, and from corrections for recession flow.

4-07.02 Deep percolation. - Knowledge of inflow into a basin or outflow from a basin through underground channels is, of course, vital to the determination of a water balance. The possibility of ground-water loss or gain should be given early consideration. Although there is no presently available means of directly evaluating deep percolation, water-balance studies for some areas indicate that a considerable quantity of water may pass from the basin through underground channels and emerge at some distance from the basin. On the other hand, in many mountainous regions, the basins appear to be relatively impervious, which is the case at UCSL and WBSL. There is some question as to ground-water outflow at CSSL. In general, it is believed that loss by deep percolation is small for most areas in the mountains of western United States. However, in some areas exceptions occur which completely invalidate water-balance computations for these areas.

4-07.03 Streamflow measurement. - Factors affecting streamflow measurements are well known and reference is made to books on hydrography for description of techniques used in determining streamflow in open channels. Specific reference is made to publications of the U. S. Geological Survey 9/ for details pertinent to establishment and operation of stream gages, and for compilation of basic streamflow data. In areas of snow accumulation, the quality of the discharge record may be adversely affected by the effects of ice in the channel and the gage installation. Special precautions must be taken in order to insure record of acceptable accuracy for these areas. All regularly established gaging stations operated by the Geological Survey are rated as to probable accuracy of measurement, for periods of both high and low flows.

4-07.04 Storage effect on streamflow. - Delay to runoff due to ground and channel storage is a basic hydrologic phenomenon. For the purpose of this report, ground-water storage is defined as the temporary storage of water in the ground, consisting of both the water under hydrostatic pressure and the water in transit through the soil under natural drainage. Direct evaluation of ground-water storage through the use of well records is impractical in mountainous areas because of the wide variability of conditions on a drainage basin. Streamflow-recession analysis provides an indirect means of evaluating both channel and ground-water storage. As previously mentioned, generated runoff is computed by adding the change in ground and channel storage to the observed runoff. Assuming that all inflow to a basin is suddenly stopped, all outflow subsequently passing the gaging station would result from depletion of ground and channel storage. A measure of this recession flow is therefore a measure of ground and channel storage. Thus, for any given period, the generated runoff may be obtained by adding to observed runoff the terminal recession flow volume and subtracting the antecedent recession flow volume.

4-07.05 Recession analysis. - Each of the several components of ground-water and channel storage has a recession flow that is essentially a decay-type curve (that is, a curve that recedes in a manner such that the incremental change in rate of flow is directly proportional to the rate of flow). Such curves may be defined by an equation of the form,

$$q = q_o \, C_r^{\,t} \tag{4-14}$$

where q is the flow at time t after the initial flow q_o, and C_r is the recession constant (ratio of the flow on any day to the previous day's flow). The recession constant must be evaluated in the same time units used for t. Decay curves for the recession flow components can also be expressed by an equation of the form.

$$q = q_o e^{-t/t_s} \tag{4-15}$$

where e is the base of Naperian logarithms, and t_s is the recession constant known as the "time of storage." In equation 4-15, when t equals t_s, $q/q_1 = 1/e$ (=0.368); hence t_s may be defined as the time required for the flow component to recede to 0.368 of its initial value. Moreover, the slope of the recession curve, dq/dt, at time zero equals $-q_o/t_s$; hence t_s may also be defined as the time required for a tangent to the decay curve at any point to reach zero flow. The relationship between t_s and C_r (of equation 4-14) is given by the equation,

$$t_s = \frac{-1}{\log_e C_r} \tag{4-16}$$

The time-of-storage concept is a very useful one in several aspects of hydrology, notably storage routing. It is further considered in a discussion by Snyder.[32] Integration of equations 4-14 and 4-15 gives,

$$S = -q_o/(\log_e C_r) \tag{4-17}$$

and,

$$S = q_o \, t_s \tag{4-18}$$

where S is the volume of the recession flow component.

4-07.06 The total recession flow for any drainage area can be represented as the sum of two or more decay-type curves as given above. Nominally, one such flow component may represent ground-water discharge, another interflow, and a third, surface or channel runoff. These components can be derived by plotting, on semi-logarithmic paper, the observed recession flow of a basin. Since the equations of the recession components plot as straight lines on semi-logarithmic paper, a tangent to the tail of the recession curve can be drawn, extending back under the observed curve. This tangent represents the ground-water (or most sluggish) flow component. Differences between the observed curve and the ground-water curve can then be read off and plotted on the same sheet of paper and another tangent drawn to the tail of this recession. This process is continued until the residual may be fitted by a single straight line. Usually two or three such lines are sufficient to define the observed recession curve. An analysis of this kind is especially useful in the analytical determination of the volume of water discharged by the recession flow. The volumes of the individual recession components can be determined using equations 4-17 and 4-18 and the total recession volume as the combined sum.

4-07.07 An alternative method of describing a recession curve is by using variable recession constants in equations 4-14 and 4-15. The derivation of the values may be done empirically for each stream during periods of no inflow. Recession curves and curves showing the variation of $t_{\underline{s}}$ with discharge for each laboratory basin are shown in chapter 2. These curves were determined by plotting flows on semi-logarithmic graph paper, during times of no inflow, for all available ranges of flows. From these plottings, a single recession curve was derived, utilizing near-maximum slopes for each range in flow. Variations due to unusual storm conditions were ignored. The recession curves so derived represent average conditions over the basin. While some seasonal differences occur, these differences are small.

4-07.08 The use of a single recession curve for all conditions of flow appears to be adequate for the laboratory areas. Since it combines the components of surface flow, interflow, and ground-water flow, it is basically assumed that each component contributes its proportional part to a given flow. When attaching recession curves to hydrographs, care must be exercised to assure that the point of attachment represents the true streamflow recession. For large areas, it may be necessary to separate ground flow recession from surface flow, depending upon the character and relative magnitude of ground-water flow.

4-07.09 The following equation was used go obtain generated monthly runoff (Q_{gen}) from observed monthly runoff volumes:

$$Q_{gen} = Q + Q_{rt} - Q_{ri} \tag{4-19}$$

where $\underline{Q}$ is the observed monthly runoff, $\underline{Q_{rt}}$ is the terminal recession volume, and $\underline{Q_{ri}}$ is the initial recession volume, all expressed in inches over the basin area. In order to facilitate these computations, volume-vs-flow curves were derived for relating the remaining runoff volume beneath the recession curve to the flow at the beginning time. This was done by incrementally summing the areas beneath the empirically-derived recession curves to the lowest value of the recession encountered in the analysis. Thus, the curves do not represent the total volume to zero flow.

4-08. WATER BALANCES FOR SNOW LABORATORIES

4-08.01 The three snow laboratories were situated in areas which represent three different climatic types found in the mountainous areas of the western United States. (These climates are described in detail in chapter 2.) In this chapter, water balances are derived for the years of record of each of the three laboratories to further illustrate the climatic differences between areas and to give firm examples of the different amounts and disposition of the precipitation that occurs in each area. Those differences are summarized in the following table which gives the annual values of the water-balance components for each of the laboratories for its period of operation, in inches depth over the drainage basin.

Laboratory (water year)	Total precipitation			Net precipitation			Loss	Runoff
	Rain	Snow	Total	Rain	Snow	Total		
UCSL								
1946-47	21.1	39.0	60.1	17.8	30.8	48.6	13.6	35.0
1947-48	22.2	35.2	57.4	19.1	28.1	47.2	13.7	33.5
1948-49	10.8	32.6	43.4	7.9	25.5	33.4	11.7	21.7
1949-50	22.2	45.4	67.6	18.9	36.2	55.1	14.0	41.1
Mean	19.1	38.1	57.2	15.9	30.2	46.1	13.2	32.9
CSSL								
1946-47	12.9	41.8	54.7	12.3	37.1	49.4	17.2	32.2
1947-48	14.9	57.4	72.3	14.1	50.1	64.2	18.9	45.3
1948-49	10.2	47.5	57.7	9.1	42.5	51.6	18.1	33.5
1949-50	9.4	68.5	77.9	9.0	61.2	70.2	15.0	55.2
1950-51	37.2	55.4	92.6	35.2	48.9	84.1	14.8	69.3
Mean	16.9	54.1	71.0	15.9	47.9	63.8	16.8	47.0
WBSL								
1947-48	69.5	58.2	127.7	58.7	51.0	109.7	18.1*	92.6
1948-49	43.4	72.2	115.6	34.6	63.4	98.0	15.7	82.3
1949-50	56.4	77.5	133.9	46.7	68.2	114.9	17.7	97.2
1950-51	66.4	68.4	134.8	56.9	60.1	117.0	14.3	102.7
Mean	58.9	69.1	128.0	49.2	60.7	109.9	16.4	93.7

* Includes one inch from soil-moisture storage carried over from previous year.

The detailed monthly data for the several years from which this summary was made are given in tables 4-2, 4-3, and 4-4 for UCSL, CSSL, and WBSL, respectively. Graphical presentation of these data is made on plates 4-3 and 4-4 for UCSL, 4-6 through 4-8 for CSSL, and 4-9 and 4-10 for WBSL. Mean monthly precipitation, snowpack, and runoff data for the period of laboratory record are summarized in plate 2-10, together with temperature and radiation data for the three laboratories.

4-08.02 The water balances are not all for the same period of record and for this reason they are not strictly comparable. There were also some differences in the methods employed in computing the balances for the three laboratories.

In general, each of the components of the water balance was computed separately. Adjustments to these computed values were then made, considering the water balance as a whole, to arrive at the adopted values of the components. Details of the methods employed are presented in the sections which follow.

4-09. WATER BALANCE FOR UCSL

4-09.01 General. - The records for Skyland Creek, UCSL, are used for deriving a water balance representing conditions in headwater areas in the upper Columbia River basin. Skyland Creek basin alone was used, rather than the combined Bear and Skyland Creek area, because of the generally better instrumentation and, consequently, the better definition of hydrologic variables in the Skyland Creek area. Monthly mean values of each component in the water balance were computed for the four water years 1946-47 through 1949-50. Since each component has inherent errors in measurement as well as errors resulting from computation of basin amounts from point measurements, adjustments must be made in the computed values in order to arrive at the most logical balance of all components, considering the water balance as a whole. Computations of individual water-balance components were performed insofar as practical by the procedures outlined in the previous sections of this chapter. There were, however, some problems peculiar to Skyland Creek basin which made some modifications necessary. The following paragraphs describe the methods and specific details of computation of each component of the water balance at Skyland Creek, UCSL. The results are given in table 4-2.

4-09.02 Basin precipitation. - Basin precipitation was computed by the isopercentual method described in chapter 3. Normally, all precipitation stations in and adjacent to a basin would be used for computing basin values. However, because wind records were used for making adjustments for gage-catch deficiency, only those stations having anemometers were used in the computation. Stations having wind records selected for use are 1-B, 10, 12, 18, 20, and 24; the locations of these stations are shown in figure 1 of plate 4-1.

4-09.03 Double-mass-curve analysis. - Double-mass curves of precipitation at station 1-B versus precipitation at each outlying station were plotted to check the reliability of the records at the outlying stations. Records at station 1-B are considered highly reliable because the station was regularly attended. On the other hand, outlying stations were attended at infrequent intervals and therefore their records were more subject to errors such as those resulting from gage malfunction or capping of the orifice. It was found that the month-to-month relation between station 1-B and the outlying stations was generally consistent and that there were no significant gage deficiencies which might be attributed to capping or gage malfunction.

4-09.04 <u>Gage-catch-deficiency corrections</u>. - The adjustments made for wind effect on gage catch were based largely on the degree of exposure of the precipitation gage to wind. Wind records at the outlying stations showed only the total miles of wind travel during the intervals between observations; consequently, it was necessary to use the daily records at headquarters to obtain mean monthly speeds at the outlying stations. From the studies on gage-catch deficiencies mentioned in section 3-05, a chart was prepared (fig. 4, pl. 4-2) showing turbulence correction factors for gage-catch deficiencies at various wind speeds and for various mean monthly temperatures at UCSL. Studies for UCSL have indicated that precipitation is almost entirely in the form of snow if the mean monthly temperature is 25°F or less, and that precipitation is largely in the form of rain if the mean monthly temperature is 40°F or greater. Therefore, the gage-catch deficiency for snowfall is indicated by the line for a mean monthly temperature of 25°F on figure 4. This relationship between windspeed and gage-catch deficiency for snow was established from observations reported on in Research Note 21. The line labeled 40°F represents deficiencies for precipitation in the form of rain; it was derived from a study by Wilson.40/ Lines representing gage-catch deficiencies for temperatures between 25°F and 40°F (that is, for various proportions of rain and snow) were drawn by linear interpolation.

4-09.05 <u>Precipitation distribution</u>. - An isohyetal map of mean annual precipitation for the four-year record was drawn for the basin, using the procedure described in chapter 3. The mean annual isohyetal pattern, together with station values, is shown in figure 4, plate 4-1. The isopercentual method was used to obtain annual basin precipitation for each of the years of study. The isopercentual maps which illustrate the year-to-year variations in precipitation pattern, are shown in figures 2, 3, 5, and 6 of plate 4-1. Having derived annual amounts of basin precipitation for each year of study, the monthly amounts of basin precipitation used in the water balance were computed by multiplying the 6-station average precipitation for each month by the ratio of the basin annual to the 6-station average annual amount. Basin snowfall and rainfall are also computed by the above relationship, using the monthly station amounts previously determined.

4-09.06 Snowpack water equivalent. - Basin snowpack water equivalent was computed by using the snow chart described in paragraph 3-08.04 and illustrated in figure 1 of plate 4-2. Where actual measurements were lacking, it was necessary to make estimates of end-of-month water equivalent. Daily snow stake readings and temperatures at station 1-B were used as aids in determining end-of-month values. A chart was prepared for each month, December through June, using stations 1-C, 10, 12, 18, and 20. These snow-course stations were selected on the basis of adequate records, general reliability of measurements, and location of adjacent precipitation gages. A preliminary line of best fit through the points was drawn on each chart. After careful study of the relationship of the individual points to the line, a fixed average relationship was established and the lines were redrawn accordingly. Since the snow courses are located in open areas, the average amount of snow on the courses exceeds the basin snowpack by the amount of the interception loss. In accordance with the snowfall interception loss of 20 percent established for this basin (see next paragraph), the preliminary basin snowpack water equivalent determined from the chart is multiplied by a factor of 0.80 to obtain corrected basin snowpack water equivalent. A sample computation of the basin snowpack water equivalent is shown in figure 1 plate 4-2.

4-09.07 Interception loss. - Approximately 90 percent of the Skyland Creek drainage area is forested; within this forested area, the canopy density is about 80 percent; the basin mean canopy cover is thus about 72 percent. For amounts of snow normally occurring in the Skyland Creek basin, accumulation of snow is approximately 30 percent less under the tree crowns than in the open (par. 4-03.15 and fig. 6, plate 4-11). The net snowfall interception loss over the basin is thus computed to be 21.6 percent (72 percent x 30 percent). A rounded value of 20 percent was adopted as the interception loss for snowfall. Interception of rainfall was determined on the basis of data quoted by Kittredge,16/ (from Munns' "Studies of Forest Influences in California") which are summarized in the following tabulation:

Rain per shower, inches	Percentage interception			
	At base of tree	Under heavy crown	Under light crown	Under edge of crown
0.01	100	100	100	81
0.06-0.10	94	84	68	48
0.11-0.30	74	48	27	5
0.51-1.00	53	33	16	4

Figures showing interception "under light crown" were used to compute interception losses. To facilitate computation, a number of months representing the complete range of monthly rainfall amounts were analyzed (by individual storms) to determine the expectable interception loss for given monthly rainfall totals. Results are shown in graphical form in figure 2, plate 4-2. Relationships of monthly rainfall and interception loss as shown by the graph were then used to obtain monthly amounts of interception loss of rainfall for the 4-year period of study.

4-09.08 Evapotranspiration. - Evapotranspiration losses were computed by the method developed by Thornthwaite (described in section 4-05). Although this method appeared to be the best of the various methods tested, one of the basic assumptions adopted in the method for computing actual evapotranspiration does not appear applicable to this area, namely the assumption of even distribution throughout the month for both precipitation and potential evapotranspiration demand. Accordingly, the computed values were modified to reduce the loss during summer months such that loss would not exceed the difference between available water and measured runoff.

4-09.09 Soil moisture. - As indicated in section 4-06, the quantitative evaluation of soil moisture is difficult. Observations of soil moisture under the snowpack were made at UCSL by use of Bouyoucos blocks and the Colman meter, both of which are electric resistance-type soil-moisture sensing devices. However, the data were not considered reliable enough to be used in the water balance. Therefore, indirect determinations of change in soil moisture were made on the basis of assumed capacity of the soil to hold moisture. A maximum value of four inches was adopted in accordance with that used by Thornthwaite, after checking its applicability to UCSL by computations based on data from other sources. Having established a maximum value for available water, the amount for any given month is calculated in the process of computing actual evapotranspiration losses by Thornthwaite's method.

4-09.10 Computed runoff. - From independent computations of snowpack water equivalent, precipitation, and losses, the water balance equation (eq. 4-2) was used to obtain computed values of generated runoff. Final values of computed generated runoff are entered in table 4-2.

4-09.11 Observed runoff. - Since stream-gage records are considered to be one of the most reliable quantitative measures in the water balance, they may be used as a check on the evaluations of the other components as integrated in the computed generated runoff. As previously pointed out, observed runoff measurements must be corrected for initial and terminal recessions for use in water balance. Figure 3 of plate 4-2 includes a curve for Skyland Creek basin relating the volume of recession flow to the observed discharge in cfs, based on the average recession curve for this area. Volumes are given for flows above an arbitrary base of 2.0 cfs. Generated runoff values computed from observed runoff values (i.e., observed runoff corrected for recession flow) are shown in table 4-2 for comparison with those computed by the water balance method. It will be noted that computed values are not entirely in agreement with observed generated runoff, particularly during the winter months of 1946-47 and spring months of 1950. A comparison of the ratios of monthly runoff values for Skyland Creek to those for other streams in the vicinity indicates that the observed Skyland Creek flow was too low during the winter months of 1946-47. Further substantiation of the low flow is obtained by comparing the 1946-47 flow with that of the following winter, when more runoff occurred even though meteorologic conditions were less conducive to high winter runoff. Similarly, a comparison of runoff from Skyland Creek with runoff from adjacent drainages for the spring months of 1950 shows a marked dissimilarity in runoff distribution, suggesting either an abnormal distribution of runoff for Skyland Creek in that period or the possibility of error in the observed runoff values.

4-09.12 Adopted values of water-balance components. - Because of the lack of agreement between observed and computed runoff values, other values were adopted where necessary, considering the water balance as a whole, to give more logical values of the various components. Changes were confined to the months when computed runoff failed to agree with observed runoff. In such cases a study was made of the hydrometeorological conditions during the month in question, in order to determine which components were incorrect. The computed values are based on procedures that will produce the best over-all results; these procedures will not necessarily give correct values for periods with unusual conditions. In a few instances a study of existing conditions failed to identify the incorrectly evaluated component, and in such cases the figures were arbitrarily changed to effect a proper balance. The outstanding examples of arbitrarily adopted values are those for the spring months of 1950. Although there were indications that the values for observed runoff were incorrect, an examination of the original hydrograph revealed no reason for changing the observed runoff values. Similarly, a recomputation of the snowpack water equivalent values during the melt season gave no indication that they were grossly in error. Accordingly, the adopted values of snowpack water equivalent and runoff were a compromise between the computed snowpack water equivalent and the observed runoff. Adopted values of all components of the water balance are shown beside the computed values in table 4-2, to permit comparison of the computed and adopted values. The adopted monthly water-balance components are shown in graphical form for each water year, 1946-47 through 1949-50, on plates 4-3 and 4-4.

4-10. WATER BALANCE FOR CSSL

4-10.01 General. - Although the methods used in computing the components of the water balance for CSSL are generally the same as those previously discussed for UCSL, the details of the methods differ somewhat due to the different nature of the area and of the data. Some of the components of the water balance were evaluated by methods suited to the hydrometeorological conditions occurring at the time rather than by the more general methods described in sections 4-01 through 4-07. The evaluations of the various components of the water balance are considered reliable, being based on an exhaustive study of the basic data and pertinent field notes. The water years 1945-46 through 1950-51 were selected for study, covering the entire period for which adequate data were available. The procedures used in evaluating the water-balance components are discussed in subsequent paragraphs under appropriate headings. Some of the columns corresponding to the columns in table 4-2 for UCSL and table 4-4 for WBSL are omitted in table 4-3 for CSSL.

4-10.02 Basin precipitation. - Basin precipitation was computed by the method used for UCSL, as described in section 4-02, with minor deviations resulting from differences in basic data. Turbulence correction factors, based on monthly mean temperature and wind speed, were applied from relationships shown in figure 9 of plate 4-5 and summarized here. For periods without melt or rainfall, the amount of precipitation at each station was compared with the increase in snowpack water equivalent at the adjacent snow course. If the two quantities were found to be in agreement, the precipitation data were considered correct and no further adjustment in station precipitation was made. Using the isopercentual method, as in the UCSL study, basin precipitation was computed for each month. The mean annual isohyetal pattern and isopercentual patterns for each year of the study are shown on plate 4-5, together with the turbulence correction-factor chart and the basin map.

4-10.03 Basin snowfall. - The separation of total precipitation into rain and snow was accomplished on a day-to-day basis, using snowboard and precipitation data for the headquarters station. In most cases the separation was clear. In marginal cases where observations of form of precipitation were lacking, the evaluation was made on the basis of air temperature, as previously discussed in section 3-02.

4-10.04 Snowpack water equivalent. - Water equivalent values used in the monthly water-balance computations were based on snow-survey data at 22 snow courses. The following data were used as aids in determining the end-of-month values of snowpack water equivalent at each course: precipitation, temperature, daily snow-stake readings at headquarters, and daily readings of the radioisotope snow gage. The snow chart was not used in computing the basin water equivalent for CSSL. Since about 80 percent of the basin area is within an elevation range of 800 feet, it is apparent that the effects of elevation upon the distribution of snow are minor and are overshadowed by the effects of other terrain features. Because the elevation range was too small to show an unequivocal increase in water equivalent with elevation, and because the areal density of snow courses was high, the basin snowpack water equivalent was based on the mean of the water equivalent depths at all the snow courses. This basic value was adjusted as follows. A comparison of snow-course sites with the basin's average topography and vegetation cover indicated that a 10-percent reduction in the value of mean snow-course water equivalent would approximate the basin snowpack water equivalent during the accumulation season (see following paragraph). During the depletion season, the rate of melt on the snow courses is about 10 percent greater than on the basin as a whole because of the predominance of southerly exposures and open sites at the snow courses. Adjustments to the end-of-month values during the depletion season consisted of determining the monthly ablation of water equivalent from the average of the snow courses, reducing the ablation by 10 percent and recomputing the end-of-month values on the basis of the adjusted ablation amounts.

4-10.05 Interception loss. - Interception losses were computed separately for rainfall and snowfall. The method used for the UCSL water balance, described in paragraph 4-09.07, was also used for this laboratory. The forested area of CSSL covers 40 percent of the total basin area. Since the mean canopy density within the forested area is about 50 percent, the basin canopy cover is about 20 percent. Rainfall interception was computed in a manner similar to that for UCSL and is illustrated in figure 8 of plate 4-5. Snowfall interception was determined largely from snow-course data from CSSL. A comparison was made between water-equivalent data from snow-course sample points located under forest canopy and data from points located in the open. The results of the comparison indicated that the basin forest cover intercepts about 10 percent of the snowfall.

4-10.06 Evapotranspiration. - Computations of evapotranspiration loss for CSSL by Thornthwaite's method (section 4-05) resulted in values which were considerably smaller than the difference between net precipitation and runoff. For the five-year period as a whole, the mean annual evapotranspiration, computed as the difference between net basin precipitation and runoff, was approximately 17 inches. Computed by Thornthwaite's method, the mean annual potential evapotranspiration was 18 inches; but the mean "actual" evapotranspiration was only 10 inches. The climatological regime at CSSL probably results in actual losses greater than those computed by Thornthwaite's method because of the carryover of water stored in the snowpack to the spring and early summer when rainfall is less than the potential demand, and because of the opportunity for loss by evaporation from the snow surface in winter. Estimated monthly values of adopted evapotranspiration are shown in table 4-3 along with values of potential evapotranspiration according to Thornthwaite's method.

4-10.07 Soil moisture. - On the basis of the other components of the water balance, it is estimated that the storage capacity for available soil moisture amounts to about six inches over the CSSL basin. This value seems high by comparison to the four-inch value normally adopted for areas having deeper soil mantles. A part of the assigned value of six inches may be due to ground-water recharge. Castle Creek normally becomes dry early in the summer and ground-water levels continue to drop after than time. As a result, streamflow recession analysis does not properly account for the resulting ground-water deficit early in the fall. Inspection of ground-water and precipitation data shows that about two inches of rainfall are required to raise ground-water levels sufficiently to produce runoff.

Although one of the criteria in the selection of the laboratory area was the absence of losses by deep percolation, the possibility of such losses cannot be ignored.

4-10.08 Analyses of ground-water well data have shown that a considerable amount of water is depleted from the meadows of the laboratory area after the cessation of surface flow at the stream-gaging station below the meadows. Furthermore, as noted by field observations, a number of springs above the meadows furnish an additional supply of water, which, together with that contained in the water table, is lost by either evapotranspiration or deep percolation after the cessation of flow at the stream gage. Although the total volume of the supply is unknown, it may be greater than that which could potentially be lost by evapotranspiration, in which case the excess loss could be accounted for only by deep percolation. With presently available data, the losses of water on the CSSL area cannot be fully accounted for, and the losses attributed to evapotranspiration in the adopted water balance may be assumed to include possible loss by deep percolation. Under such circumstances the six-inch value assigned for available soil-moisture supply would be designated as available soil-moisture supply plus loss by deep percolation.

4-10.09 Observed runoff. - As for most basins, runoff from the CSSL is considered to be one of the most accurately measured components of the water balance. Except for the possibility of deep percolation, the only source of cumulative error in streamflow measurement on this area is leakage from the flume, a structure established to provide proper channel control. Periodic field checks of the structure indicated that about two percent of the total flow may have been unmeasured as a result of flume leakage. Generated runoff was computed from the observed runoff, using the method described in section 4-07, then multiplying the resulting values by a factor of 1.02 to compensate for the estimated flume leakage.

4-10.10 The water balance. - The water balance derived for CSSL comprises both computed and adopted values of the several components, as for UCSL. However, except for the values of precipitation, runoff, and snowpack water equivalent, the computed and adopted values are the same. The above cited components incorporated all the adjustments needed in order to arrive at a proper balance between all components for the basin as a whole. The interception losses determined from preliminary evaluations of rainfall and snowfall were accepted as the adopted amounts without further corrections, even though the proportions of rainfall and snowfall were revised slightly in making the final water-balance adjustments. The tabulated values of interception loss of snowfall thus vary somewhat from the 10 percent of total snowfall value previously established. The total annual loss by evapotranspiration was obtained by subtracting the computed generated runoff from net precipitation. The annual evapotranspiration amount is prorated by months, on the basis of monthly amounts of potential evapotranspiration, precipitation, and available soil-moisture. In addition, some evapotranspiration was assigned to the winter period, the amount depending upon meteorological conditions.

4-10.11 In order to complete the basin water balance, monthly values of net rainfall, net snowfall, melt, and change in water equivalent, must be determined. An over-all balance with the other components of the water balance must be made. Melt was computed by subtracting the net rainfall from the sum of the generated runoff, change in soil-moisture, and evapotranspiration loss. The melt may also be determined independently by algebraically subtracting the change in water equivalent from the net snowfall. The evaluation of melt, then, is dependent mainly upon the accuracy of separation of net precipitation into rainfall and snowfall. Final adjustments were made in the amounts of melt, net rainfall, and net snowfall, to achieve a balance between all the factors involved. The resulting adopted values are given in table 4-3. Adopted values of generated runoff are shown in column 16a of the same table. Graphical plots of adopted monthly values of each component of the water balance are shown on plates 4-6 through 4-8.

4-11. WATER BALANCE FOR WBSL

4-11.01 General. - The water balance for the WBSL was made for the basin as a whole; that is, for the entire drainage of the Blue River above station 14. It thus includes both the Mann and Wolf Creek drainages as well as the contributing area below the confluence of these creeks. Hydrometeorologically, this area differs in several important respects from UCSL and CSSL: a considerable part of the winter precipitation occurs as rain; appreciable snowmelt occurs in most winter months. As a result, heavy winter runoff is typical. The contrast between the WBSL climate and those of UCSL and CSSL is graphically shown on plate 2-10. The opportunity to sample occurrences of rain on snow was a principal reason for selecting WBSL as a snow laboratory. Compared to UCSL and CSSL, WBSL is situated at a relatively low elevation (between 1960 and 5364 feet msl). The frequent occurrence of rain during winter is a result of the relatively low elevation and the dominance of maritime airmasses, resulting from WBSL's nearness to the ocean and location on the windward side of the Cascade Range. In addition to the variation in form of precipitation during winter, there is also a marked variation in amount of precipitation within the basin. An orographic precipitation-distribution pattern characterizes WBSL, an area of extremely rugged terrain, heavy winter precipitation, and considerable elevation range. Precipitation stations at WBSL sample an elevation range of more than 3000 feet. An extremely dense coniferous forest covers WBSL except for a few small areas of meadow or bare rock.

4-11.02 Unfortunately, the same factors that differentiate WBSL from both CSSL and UCSL and made its data unique and desirable (rugged terrain, considerable elevation range, and dense forest), also made the data harder to collect and of poorer quality. Mixed rain and snow further complicate the making of precipitation (and other) measurements. Compared to UCSL and CSSL, the quality of the snowpack water-equivalent data was poor, due to errors in basic measurements ("short cores", especially) and less frequent snow surveys. Furthermore, it was difficult to evaluate the measurements in terms of end-of-month values, both because appreciable melt occurred during the accumulation season and because rain constituted an unknown part of the precipitation occurring between the end of the month and the last preceding snow survey. Hydrometeorological data for WBSL improved considerably in quality and coverage after the first two years of laboratory operation. Water balances were made for the water years 1947-48 through 1950-51. Procedures used in computing the various components are discussed in the following paragraphs. Reference is made to table 4-4 for monthly values of the various water-balance components for the above years.

4-11.03 Basin precipitation. - The isopercentual technique used in the other laboratories for computing basin precipitation was not used for WBSL. The more empirical methods used here were made necessary by the following aspects of WBSL precipitation data: first, precipitation-gage records for the first two years of the four-year period showed obvious irregularities such as evaporation or freezing of gage contents, capping of the orifice, and errors in servicing or records. Furthermorc, there were twice as many gages during the last two years of record; consequently, the use of the isopercentual method would not take advantage of the extra gages, since no four-year means would be available. (There were eight gages in the basin during 1947-48 and 1948-49, and 16 gages during the subsequent period). A second reason for not using the isopercentual technique was that most of the adjustments of precipitation records were made by comparison with the record of a single gage, the Friez at station 1B. As a result, the year-to-year variation in precipitation distribution within the basin would be partly masked, thus negating one of the principal purposes of the isopercentual method. A third reason for not using the isopercentual technique was that no adjustment could be made for gage-catch deficiency due to turbulence at individual gages, because of the lack of wind data.

4-11.04 The method used in computing basin precipitation for WBSL is as follows. First, a careful examination was made of the records of all stations with adequate records throughout all four years. (These stations were 1B, 2, 5, 6, 8, and 10.) Double-mass-curve adjustments were made for periods of missing or erratic records and monthly amounts were tabulated. The six stations were fairly well distributed throughout the basin, although there was a bias toward locations at above-average elevation. However, the slight bias toward higher elevation and consequently toward higher precipitation was probably more than compensated by gage-catch deficiencies. The mean of the six stations was used as an approximation of basin mean net precipitation, after a trial balance showed that for three of the four years, the six-station mean closely approximated the sum of basin runoff plus estimated evapotranspiration loss. For the 1950-51 water year, the six-station mean was adjusted (increased by less than five percent) to equal net basin precipitation computed as the sum of runoff plus estimated evapotranspiration loss. The basis for the adjustment was a comparison of annual totals of precipitation for all WBSL stations for their entire period of record. The comparison indicated that in 1950-51, the precipitation catch in the upper part of the basin, in terms of the basin as a whole, was relatively low as compared to the other years. Since the six-station mean was biased toward stations at higher elevations, the adjustment necessary for the 1950-51 water year was considered reasonable. Total basin precipitation for each of the four years was computed by working backward from the six-station mean, assumed to equal basin net precipitation. The difference in both gage-catch deficiencies and interception loss for snowfall and for rainfall made it necessary to evaluate total snowfall and rainfall separately in the evaluation of losses. Lacking specific data to determine gage-catch deficiencies due to wind at individual stations, an arbitrary average correction was used: a 10-percent increase in the observed quantity for snowfall and a 5-percent increase for rainfall. Monthly values of interception loss were computed, using different percentages of loss for snowfall, winter rainfall, and summer rainfall (discussed later in par. 4-11.08).

4-11.05 Form of precipitation. - The procedure used to estimate the proportions of rain and snow in the basin mean net precipitation was as follows. Using the 34° - 35° F surface air-temperature dividing line between snow and rain found for Donner Summit, California (see 3-02.03) and air-temperature data from station 1A (headquarters, WBSL), a curve was derived which relates base-station temperature to the proportion of basin snowfall in basin precipitation. A lapse rate of 3° F per 1,000 feet was assumed in deriving the curve. This curve includes the effect of both the proportion of the basin area within given elevation zones and the normal increase of precipitation with elevation. Thus it gives the proportion of snowfall in basin precipitation rather than the area over which snow is falling relative to the total basin area (see figure 2, plate 4-11).

4-11.06 Basin snowpack water equivalent. - As was done for UCSL, basin mean values of snowpack water equivalent were computed using the snow chart (described in chap. 3). The actual basin snowpack water equivalent was determined by multiplying the basin mean value from the snow chart (referred to hereinafter as the index value) by an adjustment factor representing the ratio of actual to index values. The actual value used in the adjustment factor was determined for WBSL by a preliminary water balance. Unlike UCSL, for WBSL it was not possible to make a direct comparison of basin snowpack accumulation and basin precipitation for periods of 100 percent basinwide snowfall, since few such periods occurred. Interpolation of snow-survey data to determine end-of-month values was made difficult because of the frequent occurrence of rain interspersed with snowfalls. This problem was especially acute because the changing elevation of the snowfall line (that is, the dividing line between rainfall and snowfall) usually fluctuated within the elevation range of the basin. In comparison with UCSL and CSSL, the basic snow-survey data were of generally poor quality; many "short cores" were noted during the first years of operation. Both the selection of stations used in plotting the snow chart and the weighting of the stations varied somewhat from year-to-year because of the greater number of snow courses in the later years. On the whole, the determination of snowpack water equivalent for WBSL was not rigorous; however, in spite of the above-cited weaknesses, it is probably considerably more accurate than most such basinwide snowpack water-equivalent determinations.

4-11.07 Basin snowpack values based on snow courses are measures of total-snowpack rather than net-snowpack values, since snow-survey courses are generally located in the open. However, the snowpack adjustment factor (which relates the water-equivalent index value to the actual basin snowpack water-equivalent value, as discussed in the preceding paragraph) may be derived so that allowance is made for interception loss. This was done for WBSL.

4-11.08 Interception loss. - Interception loss for WBSL was determined separately for snowfall, for winter rainfall, and for summer rainfall. Interception loss of winter rainfall and of snowfall at WBSL was computed as shown on figures 3 and 4 of plate 4-11. For summer rainfall, a constant percentage loss of 35 percent was used, modified from studies of summer-rainfall interception loss in Douglas fir in Washington.30/ Interception loss for WBSL is not as great as might be expected from a consideration of the denseness of the forest cover alone. The frequent occurrence of storms in this area reduces the evaporation opportunity and thereby the interception loss, despite the large interception-storage capacity that exists. Storm frequency is illustrated in the following table which shows the monthly average number of days with precipitation for the four-year period on which the WBSL water balance is based.

Month	Mean no. days with precipitation	Range of days with precipitation
Sept.	9	5-17
Oct.	19	13-25
Nov.	22	15-25
Dec.	24	23-25
Jan.	24	18-31
Feb.	24	22-26
Mar.	25	23-28
Apr.	16	4-28
May	16	10-24
Jun.	11	4-18
July	5	3- 7
Aug.	5	3-11

Interception losses did not enter into the water balance itself for this laboratory since net precipitation was calculated first, based on annual runoff and evapotranspiration. Total precipitation is given only to illustrate the approximate magnitude of the interception loss.

4-11.09 Evapotranspiration loss. - Evapotranspiration losses were computed by Thornthwaite's method 33/, as for UCSL. During the spring and summer months, the contribution of snowmelt as well as rainfall was considered in the month-to-month accounting of available water. A storage capacity for available soil moisture of 5 inches was considered more representative of this area than the 4 inches used at UCSL (and recommended by Thornthwaite for areas where no local information is available).

4-11.10 Computed and observed runoff. - Runoff values were computed from the foregoing water-balance components by means of equation 4-2. For comparison, the observed values of monthly runoff were corrected by means of recession curves to represent monthly generated runoff. These data are presented in columns 16 and 17 of table 4-4.

4-11.11 Adopted values of water-balance components. - The values of computed and observed runoff of table 4-4 differ by the errors in the water-balance computations. Since observed runoff is probably the most accurate of the water-balance components, it seems unlikely that much of the error results from this source. It appears more likely that the errors result from the other components which, unlike runoff, must be estimated from point measurements. Accordingly, these other values of the water balance were adjusted to make computed runoff agree with the observed values of generated runoff. The values so adjusted are designated the "adopted" values and are identified in table 4-4 by the letter a following the column number. As for the other laboratories, the adopted values were based on the most reasonable values of the various elements, considering the water balance and the water year as a whole. These adopted values of the water-balance components are shown graphically in plates 4-9 and 4-10, for each of the four years of record.

4-12. REFERENCES

1/ ASCE, Hydrology Handbook, Manuals of Engineering Practice, No. 28, New York, 1949, 184 p.

2/ BENTON, G. S., and Jack Dominitz, "The use of atmospheric data in the evaluation of evapotranspiration," The Johns Hopkins University, Department of Civil Engineering, Water Vapor Transport Project, Scientific Report No. 3, Baltimore, Maryland, September 1954, 39 p.

3/ BENTON, G. S., and Mariano A. Estoque, "Water-vapor transfer over the North American continent," Journal of Meteorology, Vol. 11, No. 6, December 1954, pp. 462-477.

4/ BOUYOUCOS, G. J., and A. H. Mick, "An electrical resistance method for the continuous measurement of soil moisture under field conditions," Michigan Agricultural Experiment Station Technical Bulletin 172, 1947.

5/ CARNAHAN, R. L., and G. S. Benton, "The water balance of the Ohio River basin for 1949," The Johns Hopkins University, Department of Civil Engineering, Water Vapor Transport Project, Technical Report No. 1, Baltimore, Maryland, March 15, 1951, pp. 9-41.

6/ COLMAN, E. A., and T. M. Hendrix, "The fiberglass electrical soil-moisture instrument," Soil Science, Vol. 67, 1949, pp. 425-438.

7/ COLMAN, E. A., Vegetation and Watershed Management, The Ronald Press, New York, 1953, 391 p.

8/ CONNAUGHTON, C. A., "The accumulation and rate of melt of snow as influenced by vegetation," Journal of Forestry, Vol. 33, pp. 564-569, 1935.

9/ CORBETT, D. M., and others, "Stream gaging procedure," U. S. Geological Survey Water-Supply Paper 888, 1943.

10/ CURTIS, O. T., and D. G. Clark, An Introduction to Plant Physiology, McGraw-Hill, New York, 1950, 752 p.

11/ HAISE, H. R., "How to measure the moisture in soil," in Water, The Yearbook of Agriculture, 1955, U. S. Department of Agriculture, U. S. Gov't. Printing Office, Washington, D. C., 1955, pp. 362-371.

12/ HALSTEAD, M. H., "Theoretical derivation of an equation for potential evaporation," The Johns Hopkins University, Laboratory of Climatology, Micrometeorology of the Surface Layer of the Atmosphere, Interim Report No. 16, 1951, pp. 10-12.

13/ HAMILTON, E. L., and P. B. Rowe, "Rainfall interception by chaparral in California," U. S. Forest Service, California Forest and Range Experiment Station, in cooperation with Division of Forestry, Sacramento, 1949, 43 p.

14/ INGEBO, P. A., "An instrument for measurement of density of plant cover over snow course points," Proceedings, Western Snow Conference, 23rd Annual Meeting, 1955, pp. 26-28.

15/ JOHNSON, W. M., "The interception of rain and snow by a forest of young ponderosa pine," Trans. Amer. Geophys. Union, 1942 Part II, pp. 566-569.

16/ KITTREDGE, Joseph, Forest Influences, McGraw-Hill, New York, 1948, 386 p.

17/ KITTREDGE, Joseph, "Influences of forests on snow in the ponderosa-sugarpine-fir zone of the Central Sierra Nevada," Hilgardia, Vol. 22, No. 1, March 1953, 96 p.

18/ LEMMON, P. E., "The spherical densiometer for estimating forest overstory density," manuscr., Soil Conservation Service, Portland, Oregon (submitted in 1956 for publication in Journal of Forestry).

19/ LINSLEY, R. K., Jr., Max A. Kohler, and Joseph L. H. Paulhus, Applied Hydrology, McGraw-Hill, New York, 1949.

20/ LULL, H. W., and K. G. Reinhart, "Soil-moisture measurement," Occasional Paper 140, Southern Forest Experiment Station, U. S. Forest Service, New Orleans, La., 1955, 56 p.

21/ LUTZ, H. J., and R. F. Chandler, Jr., Forest Soils, John Wiley and Sons, New York, 1946, 487 p.

22/ MATHER, J. R., "Summary and conclusions," in "The measurement of potential evapotranspiration," edited by J. R. Mather, The Johns Hopkins University, Laboratory of Climatology, Publications in Climatology, Vol. VII, No. 1, Seabrook, New Jersey, 1954. pp. 210-217.

23/ MATHER, J. R., "The determination of soil moisture from climatic data," Bull., Amer. Meteorological Soc., Vol. 35, No. 2, February 1954, pp. 63-68.

24/ MILLER, David H., "Snow cover and climate in the Sierra Nevada, California," Univ. of Calif. Publications in Geography, Vol. 11, Univ. of Calif. Press, Berkeley, 1955.

25/ MUNNS, E. N., "Studies in forest influences in California," 1921, 101 p. unpubl. manuscr. on file at California Forest and Range Experiment Station, U. S. Forest Service, Berkeley, California.

26/ PENMAN, H. L., "Estimating evaporation," Trans. Amer. Geophys. Union, Vol. 37, No. 1, February 1956, pp. 43-46. Discussion, pp. 46-50.

27/ REMSON, Irwin, and G. F. Fox, "The displacement of calibration curves for electrical soil-moisture units," Trans. Amer. Geophys. Union, Vol. 36, No. 5, Oct. 1955, pp. 821-826.

28/ ROBINSON, M. W., "An instrument to measure forest crown cover," Forestry Chronicle, Vol. 23, 1947, pp. 222-225.

29/ ROWE, P. B., and T. M. Hendrix, "Interception of rain and snow by second growth ponderosa pine," Trans. Amer. Geophys. Union, Vol. 32, No. 6, December 1951, pp. 903-908.

30/ SIMSON, A. G., "The interception of summer rains by forest cover," U. S. Forest Service, Pacific Northwest Forest Experiment Station, Forest Research Notes No. 5, January 1931, p. 5.

31/ SNYDER, F. F., "A conception of runoff phenomena," Trans. Amer. Geophys. Union, 1939, Part IV, pp. 725-736.

32/ SNYDER, F. F., "Discussion of 'Storage and the unit hydrograph' by C. O. Clark," Amer. Soc. Civil Eng., Vol. 110, 1945, pp. 1419-1488.

33/ THORNTHWAITE, C. W., "An approach toward a rational classification of climate," Geographical Review, Vol. 38, No. 1, 1948, pp. 55-94.

34/ THORNTHWAITE, C. W., "A re-examination of the concept of potential evapotranspiration," edited by J. R. Mather, The Johns Hopkins University, Laboratory of Climatology, Publications in Climatology, Vol. VII, No. 1, Seabrook, New Jersey, September 1954, pp. 200-209.

35/ THORNTHWAITE, C. W., and J. R. Mather, "The water budget and its use in irrigation," in Water, The Yearbook of Agriculture, 1955, U. S. Department of Agriculture, U. S. Gov't. Printing Office, Washington, D. C., 1955, pp. 346-358.

36/ VAN'T WOUDT, B. D., "On a hillside moisture gradient in volcanic ash soil, New Zealand," Trans. Amer. Geophys. Union, Vol. 36, No. 3, June 1955, pp. 419-424.

37/ WATER, The Yearbook of Agriculture, 1955, U. S. Department of Agriculture, U. S. Gov't. Printing Office, Washington, D. C., 1955.

38/ WILM. H. G., and C. H. Niederhof, "Interception of rainfall by mature lodgepole pine," Trans. Amer. Geophys. Union, Part III, pp. 660-666, 1941.

39/ WILM, H. G. and E. G. Dunford, "Effect of timber cutting on water available for stream flow from a lodgepole pine forest," U. S. Dep't. of Agric. Technical Bulletin No. 968, Washington, D. C., Nov. 1948, 43 p.

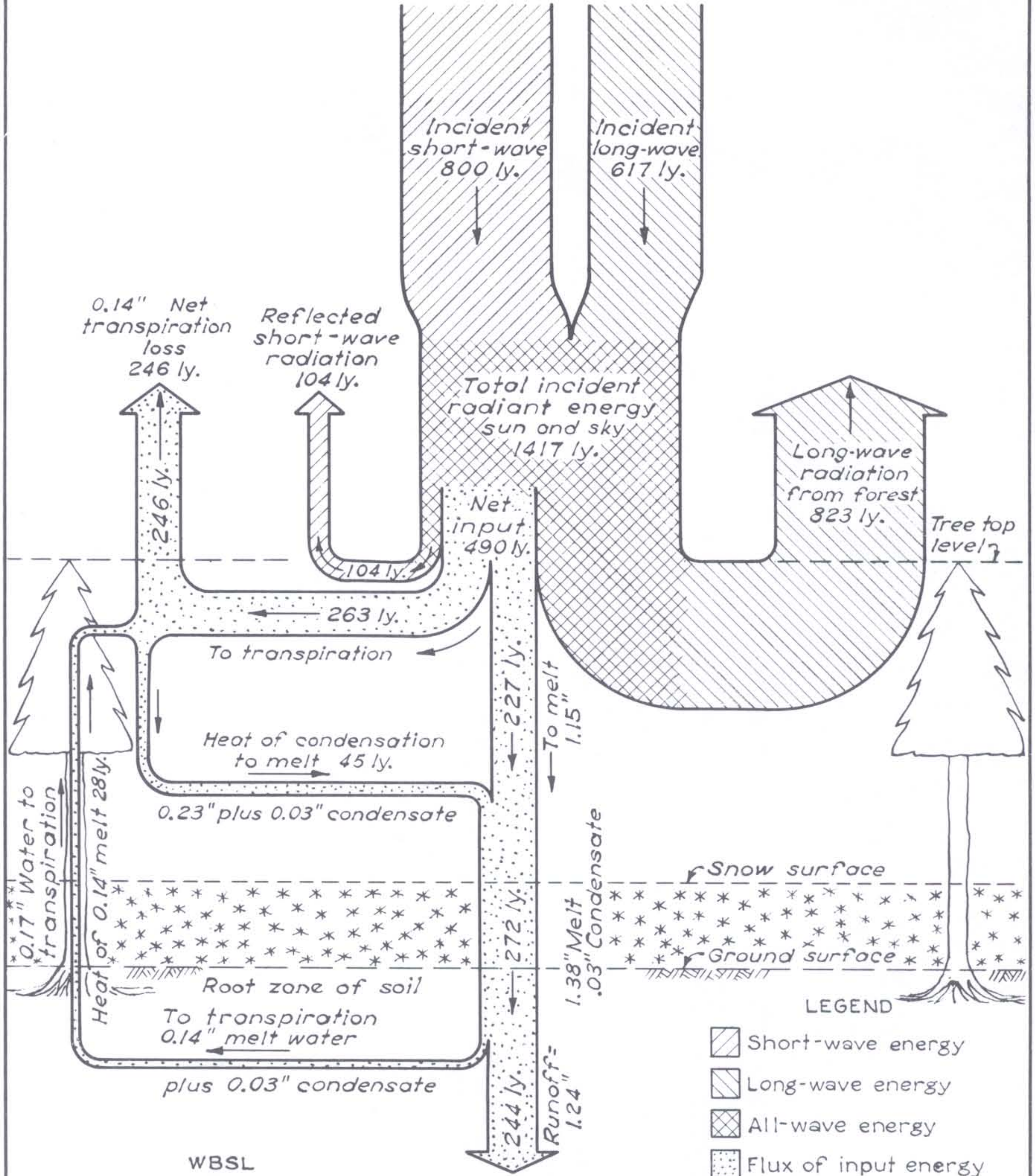
40/ WILSON, W. T., "Discussion of 'Precipitation at Barrow, Alaska, greater than recorded,' by R. F. Black," Trans. Amer. Geophys. Union, Vol. 35, No. 2, April 1954, pp. 206-207.

TABLE 4-1

PERCENTAGE INTERCEPTION LOSS BY STORM INTENSITY*

	Storm intensity classes in inches/day								
	.01	.02-.05	.06-.10	.11-.30	.31-.50	.51-1.00	1.01-2.00	2.01+	Total
	Percent of total precipitation								
Edge of crown	81	23	48	5	12	4	5	2	4.8
Light crown	100	67	68	27	32	16	14	14	18.6
Heavy crown	100	93	84	48	57	33	29	25	33.8
Base of tree	100	99	94	74	67	53	34	36	47.7
Mean	94	70	74	38	42	26	20	19	26.2
	Percent of rainfall only								
Mean	93	70	74	38	44	28	15	4	26.6
	Percent of mixed rainfall and snowfall								
Mean	100	73	76	39	42	26	22	24	25.7
Total precipitation in open, in inches	0.16	0.84	2.00	11.43	13.80	36.29	29.99	48.13	142.34

*Data are from Munns 25/



WBSL
9-13 MAY 1949
SUMMARY OF ENERGY — ly/day

		Down	Up	Net
Radiation	Short-wave	800	-104	+696
	Long-wave	617	-823	-206
	All-wave	1417	-927	+490
Transpiration		45†	-291	-246
Runoff				244

† Condensation

SNOW INVESTIGATIONS
SUMMARY REPORT
SNOW HYDROLOGY

DAILY ENERGY BALANCE IN HEAVY FOREST DURING ACTIVE SNOWMELT

OFFICE OF DIVISION ENGINEER, NORTH PACIFIC DIVISION
CORPS OF ENGINEERS U.S. ARMY

PREP. G.M.	SUBM. P.B.B.	TO ACCOMPANY REPORT DATED 30 JUNE 1956
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