HYDROLOGIC MODELING OF SMALL WATERSHEDS

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PRECIPITATION

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INTRODUCTION

The input to most hydrologic models is precipitation, and rain and snow are the forms of precipitation of primary interest in the hydrologic modeling of small watersheds. Reasons for precipitation modeling include estimating annual and seasonal water yields, engineering design based on predicting flood peaks, erosion, sedimentation and chemical transport, and estimating crop yields from dry and irrigated croplands, and from range and pasture lands.

The proposed use of a hydrologic model dictates the needed detail and complexity of precipitation input. Economic considerations usually determine whether the desired sampling detail is actually achieved. For example, data from a single standard raingage may be sufficient to determine average annual or seasonal rainfall on a small watershed. A single recording raingage may provide enough information to predict average annual erosion and surface water yield. A network of recording gages is needed to describe the variation of precipitation in time and space. Data from a network of recording gages may be needed to estimate flood peaks, erosion, and sedimentation from individual events, or spatial variability of runoff production. Other hydrologic measurements, like temperature, humidity, solar radiation, evapotranspiration, and antecedent soil moisture, may be needed as well as precipitation for accurate water balance calculations or accurate crop yield estimates.

In this chapter, we will describe rainfall and snowfall models and rainfall and snowmelt as input to more complex hydrologic models. We tried to identify some hydrologic models that are widely used as well as some models or modeling efforts, that are less widely used but seem to show potential for future development. We made no effort to describe all hydrologic models or models which include precipitation.

RAINFALL

Significant Features

Rainfall is extremely variable, both in time and space. The extreme variability in mean annual precipitation and seasonal patterns of precipitation are illustrated in Fig. 3.1 by typical seasonal distribution graphs for

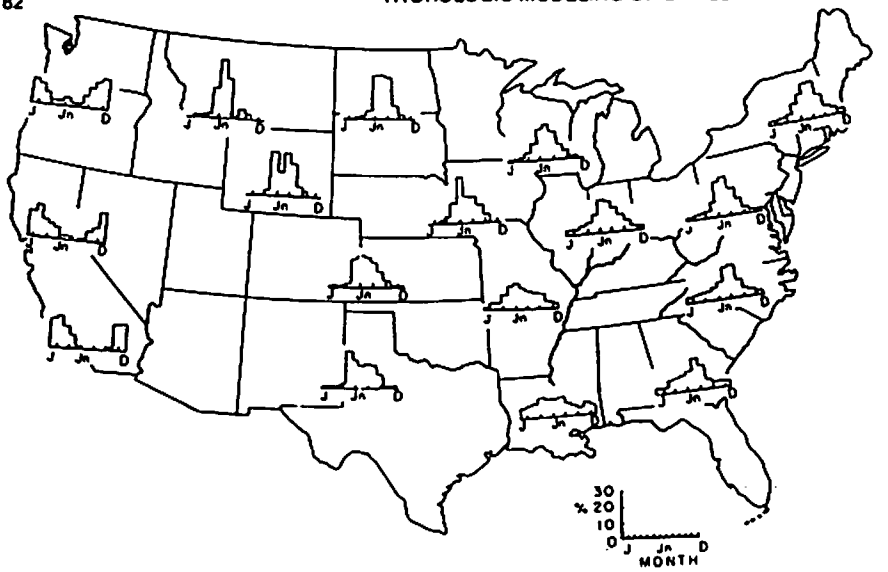


FIG. 3.1 Typical monthly distribution of precipitation in various climatic regions (after Linsley, et al., 1949).

selected stations in the United States. The spatial variability of precipitation amounts for a given event is illustrated in Fig. 3.2. Obviously, a model which describes rainfall, even for the most general low intensity rain, would be extremely complex. Therefore, the description of rainfall must be simplified to be useful in modeling. The nature of the rainfall, the required sophistication of the output, and the available resources will determine the amount of simplification.

There are three major directions in hydrologic analysis of rainfall: (a) determining the optimum sampling in time and space to answer specific questions, (b) determining the accuracy of precipitation estimates based on existing sampling systems, and (c) simulating precipitation patterns in varying degrees of complexity based on existing sampling systems for input to hydrologic models for gaged and ungaged watersheds. Input to

FIG. 3.2 Spatial variability of precipitation amounts for a given event (after Osborn, et al., 1974).

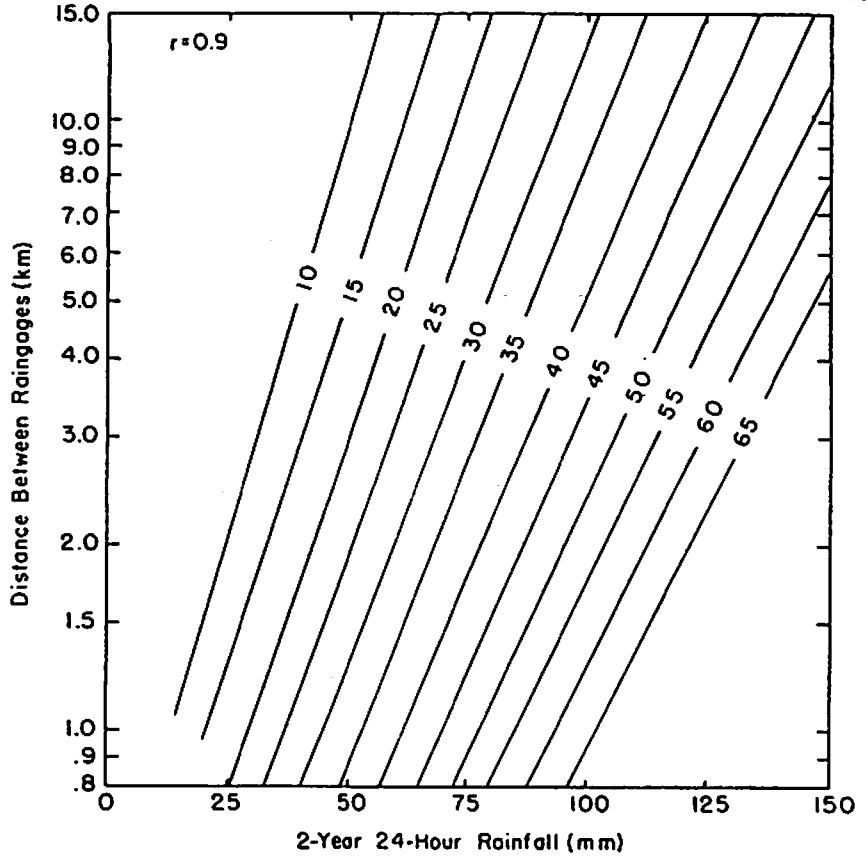


FIG. 3.3 Diagram for estimating the distance between gages as a function of the 2-yr 24-h and 2-yr 1-h rainfall (after Hershfield, 1965).

hydrologic models can be based on existing sampling systems (when available) or on simulated input. Most rainfall models are simulation models that have been developed based on data from existing sampling systems.

Optimum Sampling

In general, most analyses to determine optimum sampling systems to answer specific hydrologic questions are based on depth-area-duration relationships of rainfall. In the past, much anguish could have been avoided by thoroughly analyzing the sampling requirements for a planned hydrologic project, and by objectively choosing between available funds and sampling needs early in the project life. Many researchers have investigated network design: a sampling of these efforts follow.

Hershfield (1965) analyzed the spacing of 15 raingage networks in different climatic regions in the United States. He selected 15 major storms for each of the 15 watershed networks. Gage densities varied from 3 per 1 km^2 to 1 per 16 km^2. From these data, Hershfield developed a relationship (based on an arbitrary, but reasonable standard, $r = 0.9$) as a function of the 2-yr, 24-h and 2-yr, 1-h rainfall to aid in establishing gage density (Fig. 3.3). The relationship indicated that gages should be more closely spaced, as short duration rainfall intensities increase. Although the relationship was developed from a limited amount of data, it can be used as a first approximation of raingage spacing.

At times, the desired accuracy of rainfall measurement for a specified project or study may be unreasonable. For example, Osborn et al. (1972) analyzed records from a relatively dense raingage network in southeastern Arizona to determine the required spacing to accurately estimate the spatial variability of maximum 15-min storm rainfall (which is highly correlated to peak runoff from small watersheds). Using an arbitrary correlation standard ($r = 0.9$) between gages, the required gage spacing was 300 m. This spacing would have required 1400 raingages on the 150 km^2 watershed, which would have been completely unmanageable. In this case and others, there must be a compromise between desired and actual sampling. Usually, the compromise results in sparser sampling for a longer period of time.

Eagleson (1967) used harmonic analyses and the concepts of distributed linear systems to study the sensitivity of runoff peak discharge to the characteristic spatial variability of convective and cyclonic rainfall. He determined theoretical general relations for optimum rainfall network density for flood-forecasting purposes. He found that including watershed characteristics in the overall network design reduced the number of necessary gages, and that in simpler cases, for example when mean annual rainfall was required for a 3240 km^2 watershed, only two gages were needed.

Hendrick and Comer (1970) found statistical correlations among gages on a northern Vermont watershed based on distance and azimuth between gages, rainfall amount, and season. They found no correlation with elevation within a range of 400 m. They developed a correlation field function with which to determine raingage density and configuration for similar watersheds and climatic conditions.

Stol (1972) investigated correlations between rainfall gages in the Netherlands. He used negative exponential distributions utilizing both linear and quadratic distances between gages. Although the records from gages were often highly correlated, he found that extrapolating from one gage to another did not result in a correlation of 1.

Most efforts have been based primarily on large basins or regions, and are unnecessarily complex for small watershed design. However, observations on space and time correlation between number of gages and network watershed geometry is valid for all but the smallest watersheds in most regions, and probably valid for any but the smallest watersheds in regions where thunderstorm rainfall produces most of the runoff.

Effect of Rainfall Variability on Streamflow Simulation

Nash (1958) stated that the relationship between rainfall and runoff can be considered in three parts: (a) the relationship between volume of storm rainfall and the resulting volume of storm runoff, (b) the more complex manner in which the distribution of the rainfall in time affects the resulting runoff, and (c) the relationship between frequency of all rainfall occurrences and the occurrences of runoff-producing rainfall. The effects of spatial variability in rainfall on runoff might be considered as a fourth and most difficult relationship to define.

In discussing computer models, Linsley (1967) stated that with adequate amounts of the proper kinds of hydrologic data, streamflow hydrographs can be reproduced which are as accurate as the input supplied. For small watersheds, the input with the greatest variability is rainfall. Therefore, the accuracy of streamflow simulation depends primarily on how well this variability can be defined in a specific case.

Dawdy and Bergmann (1969) used data from a 15 km² watershed with three recording raingages to study the effects of data errors on simulations of flood hydrograhs and peaks. Their model required input of daily rainfall and evaporation (to evaluate antecedent conditions) as well as storm rainfall and a "R" factor for estimating rainfall excess. They found that the combined effects of differences in the time distribution of rainfall at different points as well as the spatial distribution over the watershed limited the accuracy of simulation.

Fogel (1969) reported on the effects of storm rainfall variability on runoff from small watersheds in the Southwest. He pointed out that runoff is a complicated process at best, and becomes much more complicated when the input is high intensity, short duration thunderstorm rainfall of limited areal extent. Again, although thunderstorm rains are more significant on small semiarid rangeland watersheds, they also produce significant runoff in more humid regions. Fogel pointed out that most current methods for estimating runoff volumes require knowledge of only total rainfall depth, which can lead to significant errors in estimating runoff.

Obsorn and Lane (1969) studied the relative sensitivity of rainfall variables and watershed characteristics on runoff from intense, short duration thunderstorm rainfall. They found that for four very small watersheds (less than 5 ha) runoff volume was most strongly correlated to total rainfall, that peak runoff rate was best correlated to maximum 15-min rainfall, that flow duration was best correlated to watershed length, and that lag time was best correlated to watershed area. Watershed characteristics did not add significantly to estimates of peaks or volume of runoff. In other words, for the data analyzed, the variability in rainfall dominated the relationships and indicated the difficulty in identifying significant variables other than rainfall in modeling runoff from small rangeland watersheds.

Wei and Larson (1971) presented a comprehensive analysis of the effects of areal and time distribution of rainfall on runoff hydrographs from small watersheds in southern Minnesota. They worked with a 2-phase model (Fig. 3.4) with precipitation as the input to the land phase portion, phase 1, of the model. Direct precipitation input to the channel, phase 2, can be considered insignificant for small watersheds, and only phase 1 is considered here. Five different triangular-shaped patterns of excess rainfall were selected to study the effects of time distribution, while three different rainfall patterns (concentrated on the upper, middle, and lower zones, respectively) were used to study the effects of areal distribution on runoff hydrographs. The results of the study were described in detail, and generally indicated significant difference in peak discharges for varying time and areal distributions of rainfall. For most small watersheds and design practices, this level of sophistication probably is not necessary, but in cases where relatively small differences in estimates of peak discharge can have economic impact, the work by Wei and Larson would be worth studying in detail.

The time and space distribution of heavy storm rainfall in Illinois were investigated by Huff (1967, 1968). His investigation was based on a network of 49 recording raingages on 1000 km² with subareas of 130, 260, and 520 km². His criteria for "heavy" storm rainfall was a mean depth of at least 12 mm, and a point value equal to or greater than that of a 2-yr frequency.

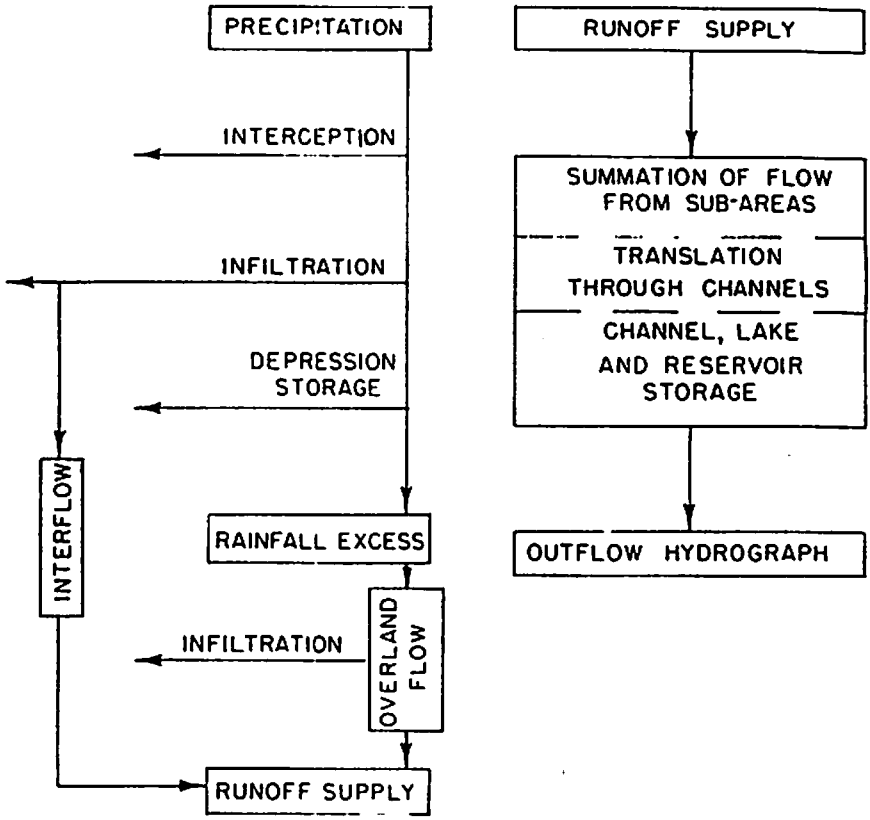


FIG. 3.4 The surface runoff process in two phases (after Wei and Larson, 1971).

For time distribution, he found that the relations could be represented by relating the percent of storm rainfall to storm duration and grouping the data according to the section in which rainfall was heaviest. These results were applicable to hydrologic modeling in the Midwest. For spatial distribution, he tested eight different statistical distributions for best fit. The most sensitive variables were: area, mean rainfall, and storm duration. The results were presented as probability distributions for different sets of conditions with respect to area, storm duration, and rainfall amount. Both investigations would be useful for hydrologic modeling for summer rainfall in Midwestern watersheds.

Effect of Elevation

Generally, elevation differences have little influence on rainfall occurrence and magnitude on small watersheds, or at least possible rainfall differences are difficult to identify on small watersheds with elevation differences of less than 500 m (Chang, 1977). For most hydrologic modeling of small watersheds, possible elevation effects on rainfall can be ignored. For mountainous terrain, both the number of events and amount of rainfall per event generally increase with elevation. Duckstein et al. (1973) used an event-based stochastic rainfall model and empirical data to investigate elevation effects on summer rainfall in the Santa Catalina Mountains in southern Arizona. Their results indicated an increase in the number of storms

and the amount of rainfall per storm with deviation and apply to many Western mountain ranges where summer thunderstorm rainfall produces significant runoff. Therefore, for long, narrow mountain watersheds with elevation differences of about 500 m or more, elevation may be considered as a significant variable affecting rainfall.

Sensitivity Analysis

In developing models to solve hydrologic problems, many variables may influence the model output significantly. Generally, however, a model is far more sensitive to a few variables (or possibly just one) than to others. For example, within narrow ranges, differences in total storm rainfall may have considerably less effect on runoff estimates than differences in the maximum amount of rainfall for a shorter duration within the storm, or changes in watershed characteristics, like vegetation, may have much less influence on runoff than rainfall intensity.

Sensitivity analysis should be a part of every effort in hydrologic modeling of small watersheds. For precipitation, sensitivity analysis requires varying selected parameters individually through an expected range of values and then comparing the range of output values from each input variable. The relative sensitivity of parameters is important in all phases of modeling — formation, calibration, and verification. Several persons have addressed themselves to this problem. McCuen (1973) used a variety of simplified models to demonstrate the importance of sensitivity analysis in all phases of modeling. Osborn et al. (1972) showed the relative sensitivity of rainfall parameters in runoff prediction. These efforts were only two of the many examples illustrating the often overlooked value of sensitivity analysis as an aid in hydrologic modeling.

Frequency Analysis

The primary objectives of frequency analysis are to determine the return periods of events of known magnitude, and then to estimate the magnitude of events for design return periods beyond the recorded history (Kite, 1975). The statistical tool used to extrapolate the known record is the probability distribution. An empirical probability distribution is often used with observed precipitation data to determine precipitation amouns for design return periods. The frequency of occurrence of rainfall amounts are determined by ranking the observed data, computing a plotting position, and plotting the rainfall amount and position on probability paper. Several plotting position formulas are available. Some have been described by Chow (1964) and Yevjevich (1972). The most practical and widely used plotting position formula is

$$P(y) = \frac{m}{N+1} \qquad [3.1]$$

where

y = the rainfall value
$P(y)$ = is the plotting position
m = the rank of the rainfall value, and
N = the number of observations.

The return period, or recurrence interval, is given by the inverse of P(y).

When a few data series must be analyzed various frequency papers can be examined and knowledge of probability and processes used to glean maximum design information (Reich, 1976, 1978). When data from many

stations are used to calculate the parameters for a preselected probability distribution, or to determine the best disribution (Alexander et al., 1969; Goodridge, 1976) for the particular data type, electronic computers are essential.

The selected distribution is then used to extrapolate to return periods greater than the record period. The predicted values are subject to considerable error, depending primarily on the length of record available (Bell, 1969).

There is no general agreement as to which distribution, or distributions, should be used for rainfall frequency analysis. The National Weather Service (NWS) used a modified version of the Gumbel extreme value distribution (W.M.O., 1974) regressed against topographic features surrounding all gages to develop a rainfall frequency atlas (NOAA Atlas 2) for the United States. It reduces some of the areal smoothing in Hershfield's (1961) TP40. The atlas is widely used to select point rainfall amounts for varying storm durations and return periods for small watershed runoff design. More recently, Frederick et al. (1977) produced maps for 5- through 60-min rains for the 37 eastern United States by mathematically fitting this distribution, which has the synonym Fisher-Tippett Type I. Examinations of single-state data include those by Goodridge (1976) and Reich et al. (1970) for California and Pennsylvania.

Bell (1969) pointed out that the design floods from small watersheds were generally the result of high intensity rains of short duration and limited areal extent. Rainfall frequencies are based on point records, and do not represent either the maximum rainfall that may occur on a watershed or the average watershed rainfall (Fogel and Duckstein, 1969). The NWS has published (also in NOAA Atlas 2) point-to-area relationships as a family of curves to be used to indicate the reduction in average areal rainfall for a given point value. These curves are generally valid for watersheds larger than 2 or 3 km^2, but should be used with caution on long, narrow watersheds or in regions where runoff is dominated by air-mass thunderstorms.

For regions dominated by air-mass thunderstorm rainfall (like southern Arizona), Osborn et al. (1979) published a family of curves indicating significantly larger reductions in average watershed rainfall than those published in the NOAA Atlas. Also, typical relationships between point and watershed maximum rainfall for given return periods were developed. For all but the smallest watersheds, the maximum rainfall that can occur about once in 100 years somewhere on the watershed, is significantly greater than that expected at any specific point within the watershed for the same return period.

HYDROLOGIC RAINFALL MODELS

Most rainfall models were not developed specifically for small watershed design. However, such models usually are applicable to small watersheds, and in fact, are often better suited to the smaller watersheds because of the difficulty of explaining rainfall variability in space and time on large watersheds. Most rainfall models are designed with runoff prediction in mind.

Recent modeling efforts have centered on autoregressive and Markov chain methods to describe persistence in time series of rainfall. Each model assumes one or more probability distributions to fit the stochastic distribution of observed rainfall. Both point occurrence and amount of rainfall have been simulated for varying time intervals. Early models were simple

and assumed constant correlation coefficients and homogeneous rainfall populations. Current efforts are directed towards segmented multipopulation models.

To date, none of the many available models has had wide usage. Some models are scholarly efforts which may lead to practical designs in the future. Others are used by individuals or groups in one locality or region. All models have parameters whose values must be evaluated from rainfall data. A sample of the available models along with a brief history of their development follows.

Gringorten's (1966) observation, that persistence is often as important as variability in rainfall occurrence, has guided most of the efforts to model precipitation.

Rainfall for short time intervals, like a day or an hour, has been difficult to model because of the sequential persistence between rainfall amounts, and because the time series are dominated by zero values (intermittent process). The occurrence or nonoccurrence of rainfall for such short intervals have normally been described by Markov chains. With the Markov chain approach, the probabilities of transition from one state (wet or dry) to the other state are determined. A wet-dry sequence is then generated using the transition matrix. When a rainfall occurrence is determined, the precipitation amount is drawn from a probability distribution describing precipitation amounts given the occurrence of rainfall.

Gabriel and Neumann (1962) seemed to be the first to successfully describe the occurrence or nonoccurrence of daily rainfall with a Markov chain model. Additional evidence of the feasibility of using a Markov chain to describe the occurrence of sequences of wet or dry days was given by Caskey (1963), Weiss (1964), and Hopkins and Robillard (1964). Feyerman and Bark (1965) suggested that the matrix of Markov chain transition probabilities should be estimated to reflect seasonal variations.

Gringorten (1966) demonstrated that a simple Markov chain could be a useful device for making estimates of the frequencies of a large variety of weather events for durations ranging from several hours to several weeks. He used a normal or Gaussian distribution y (N|0, 1) where y has a mean of zero and variance 1.0 and cumulative probability P(y) so that

$$P(y) = \frac{1}{2\pi} \int_{-\infty}^{y} \exp\left(-\frac{t^2}{2}\right) dt \qquad [3.2]$$

He assumed that successive hourly values of y are generated by a stationary Markov process with constant correlation ρ between successive values. The i~th value of y becomes

$$y_o = N_o \qquad i = 0$$

$$y_i = \rho y_{i-1} + \sqrt{1 - \rho^2} \cdot N_i, \; i \geqslant 1 \qquad [3.3]$$

where N_i is the i~th normal number selected at random from the population. If $\rho = 0$, the m-hour minimum has a cumulative distribution F(y) so that

$$F(y) = 1 - [1 - P(y)]^m \qquad [3.4]$$

To solve practical problems, Gringorten found it necessary to simulate probability distributions by a Monte Carlo method. He presented eight examples which he felt illustrated the value of the method. Only one of the eight examples concerned precipitation, and that was an estimate of the frequency of heavy snowfall in January at Boston, MA. His examples do illustrate the wide variety of stochastic processes that can be modeled.

Later efforts are more or less similar to Gringorten's work. Several groups of investigators have developed models to simulate daily point rainfall occurrences, amounts, daily maximums within selected periods, and cumulative amounts of daily rainfall. Such models have wide applicability, but generally should be used only in the climatic regime in which they were developed.

Todorovic and Woolhiser (1974) described an application of stochastic processes for the description and analysis of daily precipitation. The total amount of precipitation during an n-day period was assumed as a discrete parameter stochastic process. The most general form of the distribution functions, mathematical expectation, and variance were determined. Special cases were considered for (a) the sequence of independent identically distributed random variables, (b) the sequence of independent random variables, and (c) the Markov chain. Presented were numerical examples based on rainfall records from Austin, TX, assuming daily rainfall amounts were exponentially distributed.

Todorovic and Woolhiser (1975) carried their work one step further by developing a stochastic model of n-day precipitation. General expressions were derived for the distribution functions of the total amount of precipitation, and the largest daily precipitation occurring in an n-day period. They compared two cases — one assuming persistence (Markov chain) and the other assuming no persistence (Bernoulli) — and found the persistence model gave a better fit to the data.

Hanson et al. (1974) coupled a stochastic model for daily rainfall with a threshold value for runoff production to develop a stochastic model for runoff volume. The ratio of daily runoff to daily rainfall, assuming the rainfall threshold to produce runoff is exceeded, was modeled using the beta distribution. This ratio was then multiplied by the daily rainfall to produce a sequence of synthetic runoff data. The synthetic runoff data were then used as input data in simulation studies to design stock ponds. This analysis provides an excellent example of the use of stochastic models in hydrologic design problems for agricultural structures.

Allen and Haan (1975) and Haan et al. (1976) pointed out that the design of many water resources projects requires knowledge of possible long-term rainfall patterns. To help in design of such projects, they developed a stochastic model based on a first-order Markov chain to simulate daily rainfall at a point. The model used historical data in Kentucky to estimate Markov transition probabilities. A separate matrix was estimated for each month of the year. The model is capable of simulating a daily rainfall record of any length, based on the estimated transition probabilities and frequency distributions of rainfall amounts. Although based on Kentucky rainfall records, the model probably is applicable within a larger region in the mideastern United States.

Raudkivi and Lawgrin (1972, 1974) developed a technique to simulate rainfall sequences based on 10-min time units. The serial correlation of the historical data was modeled by an autoregressive scheme, and the skewness described by the Pearson Type 3 distribution functions. The model was tested with rainfall data at Aukland, New Zealand, so the extent of applicability is uncertain. The method could be tested in other areas.

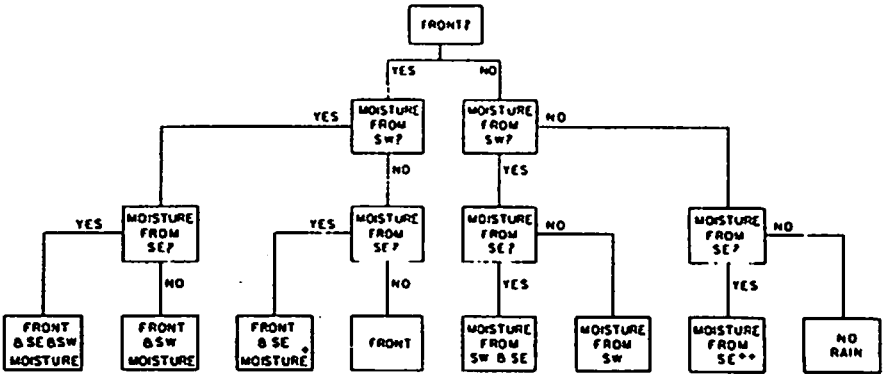
Smith and Schreiber (1974) investigated daily point rainfall records in southeastern Arizona to see how these daily records could be associated with short duration thunderstorm rainfalls. They found that a segmented Markov chain model gave a good fit to historical data from three independent point records. They also pointed out that yearly variations in the process require additional probabilistic description, indicated by annual variance in number of rain days and significant changes in autocorrelation properties, before the model could be used in simulation.

Chin (1977), in a very ambitious project, looked at daily rainfall occurrences from records of 25 yrs or longer at over 100 stations in the United States. He investigated the use of increasing orders to Markov chains to model daily rainfall occurrences. He found that the orders depended primarily on season and geographical locations, which in turn could be related to storm type. He concluded that the common practice of using the Markov chain order as the only model was unjustified without further testing. At the same time, however, he admitted that short records could mislead one into using a more complex model than justified. A specific example of a case in which a third-order Markov chain is applicable is given.

Osborn and Davis (1977) developed a three-parameter model to simulate rainfall occurrence in Arizona and New Mexico. The model was an effort,with simplifying assumptions, to follow what actually happens physically to produce rainfall in Arizona and New Mexico. The three parameters, latitude, longitude, and elevation can be determined easily for any point or small watershed. A flow diagram of the model (Fig. 3.5) follows through a logical sequence in determining if rainfall occurs. The model also allows for differences in storm types (frontal, convective, and frontal-convective) which may be important in areas where significant precipitation occurs from more than one storm type.

In Arizona and New Mexico, the principal sources of moisture for runoff-producing rainfall are from the Southeast — the Gulf of Mexico (SE), and the Southwest—the Gulf of California/Pacific Ocean (SW). Cold fronts also can trigger precipitation in the Southwest, and the combination of a cold front and warmer moist air from either the Southeast or the Southwest can result in exceptionally heavy rainfall. The model has eight outcomes based on the two sources of moisture and frontal activity. All probabilities for each of the three systems were determined independently (Fig. 3.5), and the "combination events were assumed to represent the less frequent, exceptional rainfalls." The model can be used to estimate rainfall occurrence for input to a more complex rainfall/runoff model or for a water balance model, as well as to indicate the variability in daily, seasonal, and annual water supply for agricultural users. The model is regional only; other types of models may work better in other regions.

Bras and Rodriguez-Iturbe (1976) pointed out that most rainfall models concentrate on storm exteriors or single fixed points. Very few models attempt to generate exterior and interior rainfall characteristics in space and time, and those that do have limiting assumptions of stationary behavior at all levels of storm activity. Unfortunately, simplification is essential in developing practical models of natural processes.



* AREAL RAINFALL GENERATED WITH ALAMOGORDO CREEK THUNDERSTORM RAINFALL MODEL.
** AREAL RAINFALL GENERATED WITH WALNUT GULCH AIR-MASS THUNDERSTORM RAINFALL MODEL

FIG. 3.5 Simplified schematic diagram of summer rainfall occurrence in Arizona and New Mexico (from Osborn and Davis, 1977).

Several investigators have developed models that include spatial distribution of rainfall. Bras and Rodriquez-Iturbe (1976) suggested a nonstationary multidimensional rainfall generator capable of simulating storm rainfall over an area, assuming the validity of Taylor's hypothesis (Taylor, 1937) of turbulence within the storm's interior. The method is fairly complex, but may be practical in the future if both watershed and storm characteristics can be better defined.

Areal representation of rainfall becomes more important in regions where convective storms of short duration and limited areal extent produce significant runoff-producing rainfall, and where it may be necessary to predict differences in runoff due to changes in watershed characteristics (like urbanization). Duckstein et al. (1972) introduced a stochastic model of runoff-producing rainfall for summer-type storms in the southwestern United states. They pointed out that modifications in runoff occur either naturally or through human influences, and that in either case, rainfall input must be properly modeled to determine actual changes in runoff with changes in watershed characteristics. They considered summer precipitation as an intermittent stochastic phenomenon, and obtained the probability distribution of areal rainfall by convoluting a Poisson distribution number of events with a geometric or negative binomial probability of rainfall amount. They then used their rainfall model in several rainfall-runoff relationships to illustrate the practical value of the method.

In the first example, they successfully used a linear rain model of their own design (Fogel and Duckstein, 1970) to illustrate the accuracy of their rainfall model. In the second, and most interesting, case they looked at the Soil Conservation Service (SCS) formula (SCS, 1964):

$$Q' = \frac{(R-A)^2}{(R-A)+S} \qquad [3.5]$$

where

A = initial abstractions,
S = watershed factor,
R = rainfall, and
Q′ = runoff volume,

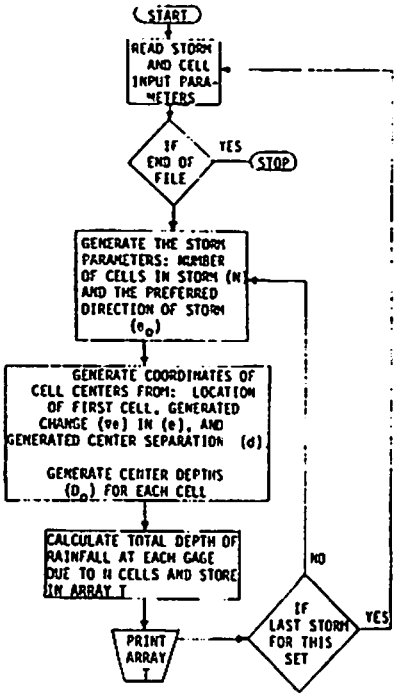


FIG. 3.6 Flow chart for simulation of individual air-mass thunderstorm rainfalls (from Osborn, Lane and Kagan, 1974).

and showed how their rainfall model could be combined successfully with the SCS model. In the final example, they used rainfall data from New Orleans to suggest that their model was more than regional.

Osborn et al. (1974) developed a simplified stochastic model (Fig. 3.6) based on airmass thunderstorm rainfall data from a dense network of recording raingages in southeastern Arizona. Probability distributions were used to model random variables — storm cell number, spatial distribution, and center depth. Storm rainfall could be simulated for any length of record. The principle purpose of this model was to predict peak discharge from rangeland watersheds. A Bernoulli random variable based on seasonal occurrences described the occurrence of the runoff-producing events. When more persistence was included in the model, the major events became too closely spaced — illustrating a modeling problem, i.e., it may be difficult to model all rainfall occurrences with equal accuracy for all uses. This model has been combined with the occurrence model (Osborn and Davis, 1977) to provide input to rainfall/runoff models to predict peak discharges from rangeland watersheds.

Smith and Schreiber (1974) proposed a probabilistic relationship among the point depth of rainfall, the local probability distribution of storm cell, maximum rainfall depth, and the dimensionless expression of storm depth-area pattern for air-mass thunderstorms in the Southwest. In a sample test, the expression was successfully used to reproduce point rainfall depth probability from storm maximum depth distribution and depth-area data from Tombstone, AZ.

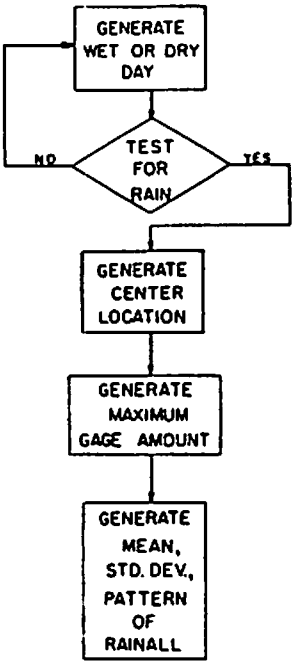


FIG. 3.7 Flow chart of the rainfall generating system (from A. D. Nicks, 1974).

Nicks (1974) used a four-stage stochastic generation technique to synthesize the daily rainfall for a 4000 km² area in central Oklahoma. Spatial patterns of rainfall input to a hydrologic model were constructed by stochastically generating: (a) the occurrence or nonoccurrence of rainfall on each day, (b) the location of the central or maximum rainfall amount within the area, (c) the maximum rainfall amount, and (d) the pattern of rainfall over the area corresponding to the central rainfall amount. Each phase in the generating sequence is shown in Fig. 3.7. Tests were presented for the representativeness and consistency of the generated data. The Markov chain model used for generating rainy day sequences and the method of generating mean rainfall were both satisfactory. The author is further refining this model.

Richardson (1977) used the multivariate normal distribution to model daily precipitation over an area (Fig. 3.8). Square roots of daily precipitation at a point were found to approximate a sample from a univariate normal distribution that had been truncated to zero. Zero daily precipitation amounts were considered negative amounts of unknown quantity. The method of moments could not be used to estimate the mean and standard deviation of the normal distribution because the data were truncated at zero. A method for estimating the mean and variance of truncated normal populations was used to estimate the parameters. The model was tested on a study region in Texas. Seasonal variation of the means and standard deviations were described with Fourier series. Fourier coefficients were related to position within the study region. Lag-one autocorrelation coefficients were found to be a function of interstation distance. The model is capable of being used to generate sequences of daily precipitation over a network of rain gages at any location in the study region given only the latitude and longitude of each station. The model can also be used for areas outside the study region if the model parameters are defined using historical data. Richardson (1978)

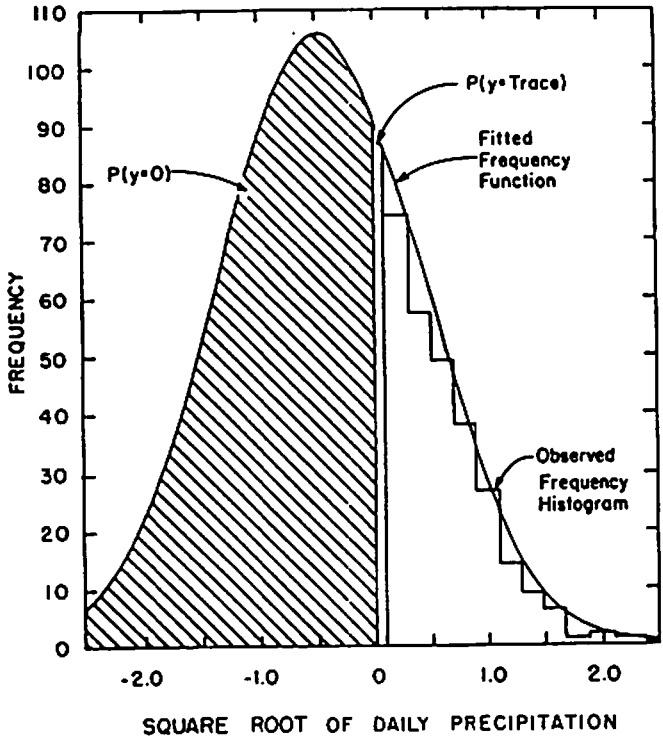


FIG. 3.8 The truncated normal distribution of daily precipitation with a square root transformation (from Richardson, 1977).

demonstrated the applicability of the model by generating precipitation sequences for an area in Texas and an area in Nebraska. The generated data contained many of the statistical characteristics observed in the historical data that are important from hydrologic design considerations.

Finally, Corotis (1976) presented a "universal" stochastic mesoscale model (STORM) based on regeneration of convective rainfall-producing cells, to be used to predict thunderstorm activity, generate rainfall for watershed models, like the Stanford watershed models (Crawford and Linsley, 1962; Ross, 1970), or contribute to a more complete theoretical understanding of the thunderstorm process. The user of the STORM program must define the basin or watershed area, and stipulate whether the storm is to be an air-mass or band-type thunderstorm. The beginning time and number of cells are simulated from a Poisson distribution. The program has been written to allow the user to specify any of several decay formulas for the depth-area relationship of the cells. The model was developed for hourly increments. A flow diagram of the STORM program is given in Figs. 3.9 and 3.10, and a complete listing of the Fortran deck, along with a description of the input, may be obtained from the writer. The user may wish to work with the author to shorten the rainfall interval to more accurately define air mass thunderstorm rainfall.

None of the proposed areal rainfall models should be considered for practical use without thorough study. The models fall roughly into two groups — those based on specific data from specific locations, and those that are primarily theoretical. Both types are relatively complex and are not easily verified. The value of the data-based models may be limited to the regions in which they were developed, and the theoretical models may be most valuable in the future as research tools.

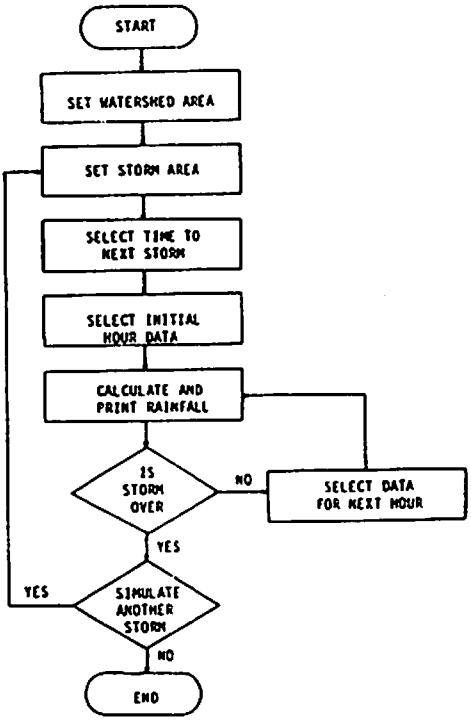


FIG. 3.9 Overall flow diagram for STORM program (from Corotis, 1976).

RAINFALL AS PART OF HYDROLOGIC MODELS

Rainfall data are required as input for most hydrologic or erosion models. The model output largely determines the form of the rainfall data input. For example, annual water or sediment yield models generally require estimates of annual or seasonal rainfall amounts. Models used to predict runoff peaks require rainfall inputs of relatively short duration (minutes or hours). Storage models require time dependent rainfall (or snowmelt) input. Models that predict erosion, sediment transport, and chemical transport on a storm basis usually require rainfall input on a storm basis.

Amounts of rainfall for given durations are usually based on point rainfall records, and some estimation procedure is used to compute average rainfall over a watershed area. On the other hand, models used to predict or estimate events of specified frequencies do not require input of actual rainfall data. In those cases, rainfall is usually based on regional estimates for the desired frequency (e.g. the 100-yr storm) and the output is that amount which will occur, on the average, once during the interval. Finally, some models may include simulation of time dependent rainfall amounts as input to deterministic models. These models predict a hydrologic series resulting from the simulated rainfall. The output of such models is stochastic.

Many hydrologic and erosion models have been developed that use rainfall data as input. Several of the most widely used models will be described to illustrate the type of rainfall data that are needed for various models.

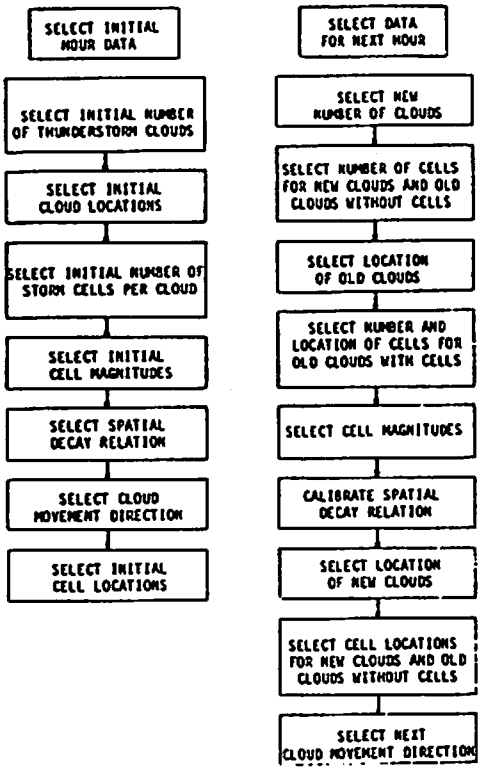


FIG. 3.10 Flow diagram for (a) Initial hour of STORM program; (b) Hourly computation of STORM program (from Corotis, 1976).

Stanford Watershed Model

The Stanford Watershed model is a digital computer program used to synthesize continuous streamflow hydrographs (Crawford and Linsley, 1962; Ross, 1970). The model has many applications and has been used to produce continuous hydrographs, evaluate runoff coefficients, evaluate the effects of urbanization on flood peaks and volumes, and estimate infrequent flood peaks on natural watersheds. The program is relatively simple to operate and can give useful practical output.

Input to the model includes hourly rainfall amounts at selected points over the watershed. Rainfall is assumed evenly distributed in space and time. Other model parameters are varied to produce a best fit to known runoff data, and then the model may be used on similar ungaged watersheds. Some parameter fitting may actually offset rainfall variability, in which case simulation for an ungaged watershed will have added uncertainty. Obviously, the greater the temporal and spatial variability of rainfall, the greater the uncertainty. Users of the model should be aware of the probable errors in simulation because of the input errors, and should allow for these in their analyses.

USDA Hydrograph Laboratory (USDAHL) Watershed Model

The mathematical model of watershed hydrology developed at the USDA Hydrograph Laboratory, Beltsville, MD, is a collection of subroutines designed to estimate runoff from precipitation data on small watersheds (Holtan et al., 1975). Variables include precipitation, land use, infiltra-

tion, evapotranspiration, and routing coefficients. Input precipitation to the model consists of a continuous record subdivided between obvious break points of rainfall or snowfall weighted over the watershed. Unknown variations in areal distribution must be accepted as error, although these errors may be reduced by dividing larger watersheds into small areas and independently applying the model to rainfall measurements on each small area. Usually, however, precipitation is estimated from gages outside the watershed, so input is still uncertain even on very small watersheds. The model can be run for any period of time.

Universal Soil Loss Equation

The Universal Soil Loss Equation (USLE) (Wischmeier and Smith, 1965) is widely used for estimating average annual erosion from field-size watersheds (Stewart et al., 1975). The equation parameters account for soil, vegetation, climate, and cultural practices. The USLE is given as

$$A = RKLSCP \qquad [3.6]$$

where

A = estimated soil loss, and
R = rainfall factor.

The other parameters describe the soil, vegetation, topography, and cultural practices.

A good estimate of the R factor is essential. Wischmeier and Smith (1965) presented an iso-erodent map of the eastern United States. Ateshian (1974) developed a method for estimating R for the entire United States based on two distributions and 2-yr, 6-h point rainfall amounts. Average annual values of the rainfall-erosivity factor, R, for the entire United States were developed by Wischmeier in 1975 (Stewart et al., 1975). The USLE, based on R values from iso-erodent maps, is acceptably accurate in many cases for estimating average annual erosion rates for long periods of time. However, for shorter periods (like 1 or 2 yr or for individual events) in the eastern United States, and for any length of record in many areas of the western United States, estimates of erosion based on R values from regional or national maps may be very inaccurate (Wischmeier, 1976; Renard and Simanton, 1975).

The inaccuracies are due primarily to the short duration and extreme spatial and temporal variability of intense runoff-producing thunderstorm rainfall. To compensate for the time variability in rainfall intensity, Wischmeier and Smith (1958) suggested using rainfall energy (E) times rainfall intensity (I) for more accurate estimates of R in the USLE when recording raingage records were available. More recent work (Renard and Simanton, 1975) indicated that, particularly for intense widely scattered thunderstorm rainfall when recording raingage records are available, the EI index gives a much better estimate of R than available maps of average R values.

Wischmeier and Smith (1958) provided a table of kinetic energies for given rainfall intensities (Table 3.1), and explained their method in detail. Rainfall intensities (I) are determined for the shortest reasonable durations and energy values are determined from the table and accumulated for each storm. The total accumulated energy (t-m/ha/cm) is multiplied by the maximum 30-min intensity (cm/h) and the product divided by 100 to give the R factor for the storm.

TABLE 3.1 KINETIC ENERGY OF NATURAL RAINFALL (METRIC TON-METERS PER HECTARE PER CM) (MODIFIED FROM WISCHMEIER AND SMITH, 1958)

Intensity cm/h	0.0	0.1	0.2	0.3	0.4	0.5	0.6	0.7	0.8	0.9
0	0	121	148	163	175	184	191	197	202	206
1.	210	214	217	220	223	226	228	231	233	235
2.	237	239	241	242	244	246	247	249	250	251
3.	253	254	255	256	258	259	260	261	262	263
4.	264	265	266	267	268	268	269	270	271	272
5.	273	273	274	275	275	276	277	278	278	279
6.	280	280	281	281	282	283	283	284	284	285
7.	286	286	287	287	288	288	289	289	290	290
8.	291	291	292	292	293	293	294			

Unfortunately, this method provides a good estimate of R only if a recording raingage is located on or very near the watershed. The accuracy of erosion estimates for short periods of record, based on single-gage precipitation records, decreases rapidly with distances between gages and watershed. Serious errors may occur if gages and watershed are separated by as much as 2 km. For watersheds without recording raingages, average annual R values can be used (Fig. 3.11), but erosion estimates must be for relatively long periods (Stewart et al., 1975).

Rational Method

There are many empirical rainfall-runoff models of similar form that require input of rainfall estimates for storms of given frequencies. Possibly the best known is the simple and aptly named Rational formula (Linsley et al., 1949). This model is the most widely used empirical equation for predicting peak discharge from a small watershed. The equation is:

$$q = CiA_i \qquad [3.7]$$

Where

q = peak discharge in ft^3/s
i = rainfall intensity in in./h for the given frequency
A_i = the area in acres, and
C = a runoff coefficient.

(Because of the unique 1 to 1 relationship between i and A_i, the rational equation is not neatly converted to metric, although it can be by adding a constant.) The equation is rational and useful as long as the rainfall intensity is for a duration equal to the time of concentration and the area is small enough to ensure relatively homogeneous rainfall and watershed characteristics. Analyses suggested that, although the method can result in large errors in any given case, it may, on the average, give reasonable design results (Shaake et al., 1967).

SCS Methods

SCS has developed two methods for estimating volume and rate of runoff from agricultural watersheds in the United States. The older method (NEH-4), which is described in the National Engineering Handbook, Section 4, Hydrology, is now used generally for watersheds larger than 800 ha (SCS, 1972). The more recent method (TR-20), which is described in SCS-TP-149, is now used for establishing conservation practices for drainage on individual farms and ranches up to 800 ha (Kent, 1973). The two methods are closely allied.

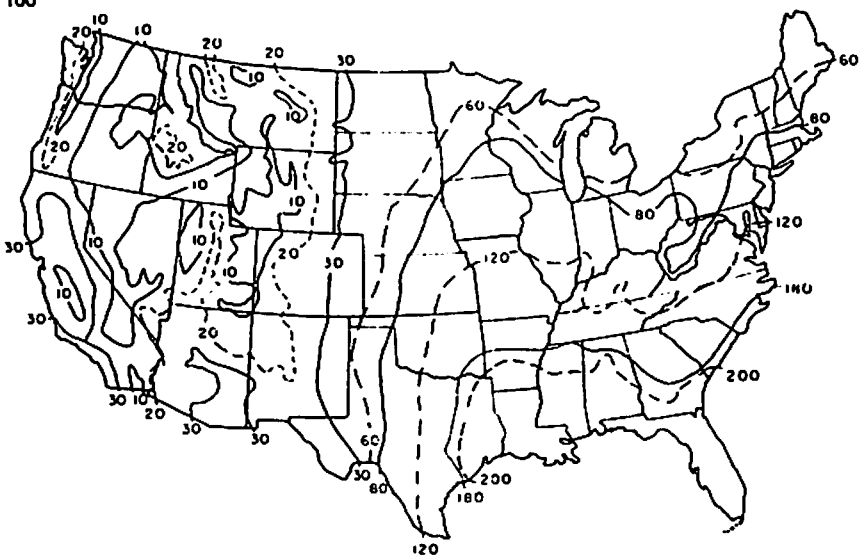


FIG. 3.11 Average annual values of the rainfall-erosivity factor, R (from Stewart et al., 1975).

The NEH-4 method is based on 24-h rainfall, since almost all rainfall data available at the time the method was developed were from standard nonrecording raingages. Both methods still use 24-h rainfall as the base for calculating watershed input. Both methods are fairly complex, involving estimates of watershed characteristics including slope, area, soil type, and cover. The NEH-4 method provides for development of a complete hydrograph, whereas TR-20 has been simplified with computer programming to develop a series of graphs which can be used directly for most small watershed designs. The methods leading to the development of the graphs are described in detail in SCS-TP-149 (1972). The graphs can be used for drainage up to 800 ha, peak discharges of 0.14 to 56 m^3/s, 24-h rainfall depths from 25 to 300 mm, curve numbers (based on watershed characteristics) from 60 to 90, and three general slope classifications, flat (1 percent), moderate (4 percent), and steep (16 percent). One can interpolate between the three given slopes.

There are three detailed examples in TP-149 of a more complex method to aid in determining peak discharges in cases that do not fit into the general categories. In most cases, however, the differences one obtains in peak discharge by following through the more complex and tedious method are not large enough to warrant the extra effort. Use of this method in a few specific cases will soon give the planner confidence in using the more simplified method.

It was recognized that 24-h rainfall is not evenly distributed in time and space, and that some adjustments were necessary to simulate reasonable peaks and volumes of runoff. The "adjustments" in 24-h rainfall are of primary interest here.

The adjustments were confined to the time distribution of rainfall. It was believed that areal variability on small watersheds was small enough to be ignored. More recent research suggests that in many cases, areal variability will have a greater affect on peak discharge than the time distribution of rainfall. However, much more information is available on time distribution at a point than on the variability of rainfall, both in time and space over an area.

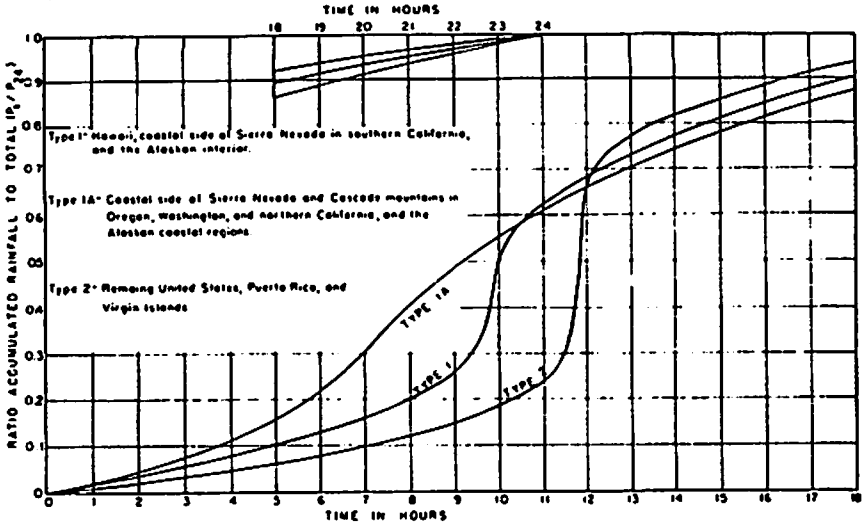


FIG. 3.12 Twenty-four-hour rainfall distributions (after SCS TP-149, 1973).

Two major climatic regions were identified in TP-149. Runoff from small watersheds in the first region was assumed most strongly influenced by maritime climate and rainfall was classed as Type I (Fig. 3.12). Runoff from small watersheds in the second region was assumed most strongly correlated to thunderstorm rainfall which was classed as Type II (Fig. 3.12). In the first case (Type I), the maximum 30-min rainfall was concentrated at about 10 h within a 24-h period of rainfall, while the second (Type II) was concentrated at about 12 h, and was more intense. The selection of the period of maximum intensity for both distributions was based on design considerations, rather than on meteorological factors (Kent, 1973).

In practice, the location of the intense rainfall within the 24-h period makes little difference, since peak discharge for most small drainages is highly correlated to the maximum 30-min rainfall. Using the Type II distribution will give somewhat larger peaks, and would represent a more conservative design. Since thunderstorms occur almost everywhere, the user must decide whether or not they occur often enough in a particular area to use the Type II distribution rather than the Type I.

The sparse runoff data available in the western United States when the original SCS method was developed limited the verification of curve numbers for conditions in arid and semiarid regions. For example, analysis of rainfall-runoff data in the Southwest indicates that the high intensity and the limited areal extent of thunderstorm rainfall, which are not included in the SCS method, are dominant factors in predicting runoff production from semiarid rangelands (Simanton et al., 1973).

Others

The synthesis of a flood hydrograph is often based on characteristics of a design storm. Time distributions of 1,623 flood producing rains of 30 min to 48 h were studied (Kerr et al., 1970) at 46 sites across Pennsylvania. After examining many combinations, the storms were classified into eight dimensionless patterns based upon storm amount, duration, and season. Geographic differences could not be verified. The resultant design procedure was incorporated into Lee's et al. (1974) double triangle unit hydrograph manual.

Analysis of 69 large flood hydrographs from 17 varied watersheds of 10 to 500 km^2 suggested that a 1-h single triangular hydrograph can be developed from the area and percentage of wooded area. Reich and Wolf (1973) suggested tentative design hydrographs may thus be estimated in hilly terrain, like that in the Appalachians with long summer humid continental or east coast continental climates, and mean annual precipitations between 850 through 1200 mm. Some generalization for estimating 1-h maximum rainfalls, which would be necessary input to this approximate flood-runoff model were generalized for remote parts of the world (Reich, 1963), and have been improved meanwhile through data analysis.

SNOW IN SMALL WATERSHED MODELING

Snow and snowfall play a significant part in the hydrologic regime of agricultural areas in many parts of the world.

In this section, we will discuss those aspects of snow hydrology that are important in modeling accumulation and melt on small watersheds — primarily shallow packs in non-mountainous areas, since both deep and shallow packs in mountainous areas are already well covered in the literature (Corps of Engineers, 1956; Leaf and Brink, 1973; Anderson, 1973).

Snow has received attention as a water resource, primarily in the northern parts of North America, Europe, and Asia. All of Canada receives snow in hydrologically significant amounts (Gray, 1968; McKay and Thompson, 1968), while some portions of the United States receive very little snow, particularly in the Southeast.

Potter (1965) defined snow cover as 25 mm or more of snow on the ground without regard to the water content of the snow. For hydrologic purposes, the water content is more important than depth, unless one is interested in the insulating properties of the snow as in soil freezing studies. Thus, maps, like those by Kuzmin (1963) and McKay and Thompson (1968) which show snow water equivalent and snow density, are more useful for hydrologic modeling. Frost on plants is also a significant factor in some regions where snowfall may not be present in large amounts. If enough days of frost formation occur, much of the water budget for an area may be unaccounted for unless frost formation and subsequent melt and sublimation are modeled (Makkink and van Heemst, 1975).

Snow Properties

Small watershed snow packs characteristically will be shallow (depth < 1 m), have relatively uniform density, and exhibit some degree of redistribution of snow during and after snowfall. The more important snow properties and characteristics, which are used in simulation, are snow water equivalent, density or specific gravity, depth, optical properties, and areal extent of the snow cover. The water equivalent of snowpack, W, is the depth of water contained in the ice and liquid water present in a snowpack. The density of the snowpack, P, is then defined as the mass of water per unit volume of snow (Martinec, 1976), but is conventionally expressed as a specific gravity and is measured simply by weighing a known volume of snow. In conventional snow surveys, both the depth, D, and water equivalent, W, are determined (Soil Conservation Service, 1972). Then

$$P = W/D \qquad [3.8]$$

Typically, the density increases with time as the pack settles (Ffolliott and Thorud, 1969; Gray, 1968; Garstka, 1944). Density values usually range between 0.15 to 0.45 (150 to 450 mg/m³), with the lower values early in the accumulation season or immediately after snowfall, while higher values prevail after a period of partial melting or blowing snow.

The density can be simulated by estimating the density of snow at the time it falls, and then by a bookkeeping technique for accounting for the water equivalent and depth of snow on the ground. Garstka (1964) gave a range of 0.01 to 0.15 units for newly fallen snow, while the average density for the United States is 0.10 units. In eastern Colorado, a 19-yr average was 0.119 units (Greb, 1975). In its most simple form, the density of newly fallen snow can be assumed to be 0.10 units, or an equation for specific gravity, like the following may be used to calculate newly fallen snow density (Hydrocomp, 1969):

$$P = P_o + (T/100)^2 \quad [3.9]$$

P = specific gravity when $T = 0°C$
P_o = specific gravity at -18 °C
T = current air temperature (-18 °C)

The remaining problem then is to determine if the precipitation is rain or snow. This can be done in several ways. The most simple is to assume that all precipitation is either rain or snow based on a temperature of 0 °C or some similar value. The SCS Model TR-20 (1965) for river basin planning uses 1.7 °C as the dividing line between rain and snow. It is reasonable to assume an air temperature greater than 0 °C at the ground level since the snow is formed at considerable heights in the atmosphere.

Another method used by Shih et al. (1972) apportions a percentage of the day's precipitation to rain, and the remainder to snow when the model time period is 1 day and the mean daily temperatures are used. Hydrocomp (1969) uses a combination of dry bulb and dew point temperatures to determine if rain or snow is falling.

Anderson (1976) used the basic equations governing mass and energy transfer to derive equations that express the increase in density of snow on the ground due to the increase in weight of the snow with new snowfall and crystal metamorphism. These equations, while more exact, require extensive calibration, and would apply to only one point. Thus, they are not really practical, but deserve further study as more sophisticated models are developed and better results needed.

It is generally accepted that the areal density of snow in shallow packs does not vary as much as the depth. Ffolliott and Thorud (1969) measured 195 points within a 170-ha watershed with an average water equivalent of 84 mm. The 5 percent confidence interval was 9 mm, whereas the density was 0.24 ± 0.01 units. The figures for the peak melt season was 0.36 ± 0.07 units. McKay and Thompson (1968) present a map of Manitoba and eastern Saskatchewan showing an average density of 0.16 to 0.22 units. McKay (1968) says that areas that have uniform topography and vegetation could be measured using a large number of snow stakes and a few density measurements. This would indicate that for much agricultural land the density could be more accurately modeled than the water equivalent. Hegedus and Szesztaz (1969) found that for Hungary, the coefficient of variation (C_V) of the density was less than 0.08 for the entire winter, while for depth, C_V was 0.8 to 0.20 units. Adams (1976) also observed that over a 206-ha area the depth was much more variable (100-200 mm) than the density (0.27 - 0.30 unit).

The areal extent of the snow cover during the melt period is needed to calculate the effective contributing area of melt. The most common methods for doing this are to assume the area not covered with snow is a function of time since the last general snowfall, or as a function of a percentage of seasonal runoff.

The fraction area not covered with snow is given by Martinec (1960) as:

$$A = 1/(1 + e^{-bt}) \quad [3.10]$$

where

A = fraction of total area snow free
t = time since arbitrary origin
b = calibration coefficient.

An estimate of the total melt time is necessary. For small watersheds, this will normally be quite short, depending on the water equivalent, area, topography, and cover of the watershed. This may be only a few days after a snowfall event (Garstka, 1944; Druffel, 1973). If the snow covered area is related to the percent of seasonal runoff, the only initial estimate to be made is the total runoff expected (Corps of Engineers, 1972; Kim et al., 1974). Since the generated runoff could also be related to the water equivalent in the basin, the area could also be related to the water equivalent remaining on the watershed (Anderson, 1973). In either case, the shape of the curves relating area to either runoff or water equivalent must be determined for each basin, but each will often have an exponential shape.

Another snowpack property of prime importance to modelers is water holding capacity. Any water above this capacity will drain from the pack. The amount of water in the snowpack is thermal quality of the pack, B, and is defined as the ratio of heat necessary to produce a given volume of water from a snowpack to that required to produce the same volume of water from ice. This ratio also is the same as the fraction of the snowpack that is ice.

$$B = 1 - f_p \quad [3.11]$$

where

f_p = liquid water content of the snowpack as a fraction of W.

Values of B ranged from 0.78 to slightly over 1 in a 1.4 m deep pack in California (Gerdel, 1945). He also found that a 0.76 m pack probably could not hold more than 10 percent of the water equivalent as free water (f_p = 0.10). Anderson (1976) used values of f_{pmin} = 0.03 and f_{pmax} = 0.1, and interpolated between these values as a function of density. Kovzel (1969) found that the water holding capacity of snow was related to the snow density as (0.11/P) - 0.11 for a range of density of 0.13 to 0.45. Dunne et al. (1976) used a value of f_p = 0.08 to calculate water movement through ice and snow. Colbeck (1974) found that for a ripe pack (P - 0.56) f_{pmin} was not 0.07 times the pore volume. In many cases, shallow packs will not be able to hold much water, since even for a 1 m deep pack with f_p = 0.03 and P = 0.50, this would only be 15 mm of water, not a very large amount as compared with the 500 mm stored in the ice.

MODELS OF SNOWMELT

Most models of snowmelt use variations of the energy balance method pioneered by Wilson (1941) in which he outlined the sources of energy that cause snowmelt. In this section, the use of the energy balance method and its simplifications are first outlined, and secondly, the application of various techniques of snowmelt calculations as incorporated into currently used models are described.

Energy Balance Techniques

The use of the energy balance technique results in a model which may be very close to being correct, but which may be unwieldly to use, except in very specialized, highly instrumented situations (Anderson, 1976; Corps of Engineers, 1972; Gray and O'Neill, 1974; Obled and Rosse, 1977). This section begins with an overview of the energy budget, and then goes on to deal with usable simplifications and problems encountered in modeling shallow packs on small watersheds.

Energy Budget

The energy budget for a snowpack is commonly given as (Gray and O'Neill, 1974; Corps of Engineers, 1956; Anderson, 1968; Kuzmin, 1973):

$$H = H_c + H_e + H_g + H_p + H_{rl} + H_{rs} \quad \text{[3.12]}$$

H = net heat transfer to snowpack from its environment
H_c = convective or sensible heat transfer from the air
H_e = latent heat transfer from condensation - evaporation - sublimation
H_g = conduction of heat across the soil-snow interface
H_p = heat transfer due to heat content of rain drops
H_{rl} = net longwave radiation exchange between the snowpack and its surroundings
H_{rs} = net shortwave radiation exchange between the snowpack and its surroundings.

For shallow snowpacks, Gray and O'Neill (1974) add a term, H_{gs}, the heat transferred to the soil surface by solar radiation penetrating the snowpack. If H is the total net change in energy, the melt, M, is calculated as H/L_f where L_f is the latent heat of fusion of ice. If liquid water is present in the snowpack ($B < 1$), then the calculated melt must be divided by B to determine true melt.

The sensible and latent heat transfer are often treated together because they have some elements in common. One of the more common methods for calculating evaporation from a snow surface is the mass-transfer method (Corps of Engineers, 1956 and 1960).

$$L_e = f(v)\,(e_s - e_a) \quad \text{[3.13]}$$

L_e = loss of mass from or to the snow by evaporation or condensation
v = wind speed
e_s = snow surface vapor pressure
e_a = air vapor pressure.

Other methods used for evaporation from water surfaces are also used for evaporation from snow. These can be characterized by the profile methods. An example is (Munn, 1966):

$$H_e = L_v k_e \rho_a \, \partial q/\partial z \quad \ldots \quad [3.14]$$

L_v = latent heat of vaporization
k_e = diffusivity coefficient
ρ_a = density of air
q = specific humidity
z = height above surface.

For a shallow pack site at Ottawa, Ontario, Gold and Williams (1961) found 25 percent of a pack, containing 134 mm of water, was lost to evaporation over a 2-wk period. Lemmela (1973) found 2.0 to 9.8 percent of total water equivalent was lost to evaporation. Norum et al. (1976), as well as other researchers, contend that sublimation is low for most snowpacks, and so can be ignored.

Since sensible heat transfer is similar to evaporation, the following equation from Anderson (1976) is:

$$H_c = -\rho_a C_p k_c \partial T_a/\partial_a \quad \ldots \quad [3.15]$$

C_p = specific heat of air at constant pressure
k_c = eddy transfer coefficient
T_a = air temperature.

Conduction of heat from the ground is many times ignored as the values are expected to be small. This may not always be the case, since Obled and Rosse (1977) reported 0.8 mm/day melt into a lysimeter from ground heat. The Corps of Engineers (1960) used 0.5 mm/day over an entire basin. Investigating a shallow prairie snow pack, Norum et al. (1973) felt that the net transfer from soil to snow may not be enough to cause appreciable melting. Most modelers assume a constant value for this heat movement both because of the difficulty of determining good values of the heat conduction coefficient of wet or frozen soil and the smallness of the heat value as compared with other heat fluxes.

The conduction of heat within a snowpack has an effect on the heat storage within a pack. During actual melt, the snowpack is isothermal, and there is no change in heat content of the snowpack due to temperature changes. However, in many cases during the accumulation season and for some packs during the melt season, there is a temperature gradient in that snowpack — particularly for shallow snowpacks and for the top 30-50 mm of deeper packs, where the liquid water will refreeze each evening (Obled and Rosse, 1977).

Allen (1976) divided the snow temperature calculations into three classes based on water equivalent. In all cases, the snow surface temperature is assumed less than or equal to the air temperature, but never greater than 0 °C. If the water equivalent is less than 50 mm, the mean snowpack temperature is equal to the mean daily air temperature, but never more than 0 °C. If $50 < W < 114$ mm, then the snowpack temperature is 1.5 times the snow surface temperature plus the previous day's ground surface temperature. For $W > 114$ mm, the diffusion equation is used, and the snowpack is not assumed to be isothermal. He found the thermal diffusivity, ∞, equal to $0.025/(2.75 \cdot \rho_s)$ cm^3/s/k.

The purpose of calculating snowpack temperature is to calculate the cold content of the snowpack. This is the amount of energy it would take to bring the temperature to 0 °C, and is calculated as (Eagleson, 1970):

$$H_{cc} = \int T_s \rho_s c_p \, dz \quad [3.16]$$

H_{cc} = heat deficit (cold content) of snowpack.

For a shallow pack, it should be possible to assume some average temperature. The amount of cold content is not great, since an average value for nighttime radiational cooling is about 2 mm of water (Eagleson, 1970). The depth of water corresponding to H_{cc} is:

$$H_{cc} = L_f \rho_w d_{cc} \quad [3.17]$$

d_{cc} = depth equivalent of water at 0 °C
ρ_w = density of water.

For a pack with a depth of 320 mm, with conditions of a pack at an average temperature of −16.4 °C, and density of 0.173 gm/cm³, H_{cc} = −45 cal/cm² or 5.7 mm of water at 0 °C would have to be added to the pack to bring its temperature up to 0 °C (Kuzmin, 1972). This calculation is always necessary to obtain a complete heat balance of a snowpack since this energy must be supplied before melting can begin. This is particularly important with energy and advection to the snowpack by rain.

Net radiation is the sum of all the radiation fluxes, and is the most important component in the energy balance of a snowpack during accumulation and periods of continuous snowcover (Allen, 1976; Gray and O'Neill, 1974; Male and Gray, 1975). Net radiation is:

$$H_r = H_{rs} - H_{rl} \quad [3.18]$$

For research watersheds with continuous snowcover and little relief, it is possible to measure H_r, since both short and longwave radiation tend to be fairly uniform over large areas. Usually, however, L_i, shortwave incoming radiation, or R_d, longwave incoming radiation, or both, are measured, so several approximations must be made — H_{rs} can be calculated as $H_{rs} = L_i(1 - a)$ where a = albedo of the snow. The albedo is usually calculated as a decay function since last snowfall. Only the surface "active layer" (about top 100 mm) has an effect on the albedo and absorption of solar radiation (Anderson, 1976; O'Neill and Gray, 1973a; Obled and Rosse, 1977). For deeper snowpacks, the Corps of Engineers (1956 and 1960) recommends that the albedo be decreased from 0.80 to 0.40 after about 20 days. Eggleston et al. (1971) developed a decay function for albedo as a function of time, whereas Anderson (1976) uses an equation based on snow surface density. Bergen (1975) also suggested that for surface grain sizes over 1.5 mm, the albedo is a function of density rather than grain size. O'Neill and Gray (1973a) found only a small variation (0.65 to 0.85) in albedo during the non-melt season, but a much sharper decrease in albedo during the melt season than that indicated by the Corps of Engineers (1956). This was perhaps due to a much shallower snowpack. O'Neill and Gray (1973b) showed little effect of snow depth on albedo for depths greater than 30 or 40 mm, which is the depth 5 percent of the incident radiation could penetrate. Gerdel (1948) found that 5 percent of the radiation penetrated to a depth of 180 mm in the Central Sierras. Usually, a is decreased to 0.4 after a rainfall.

Kuzmin (1973), Allen (1976), and Eagleson (1970), as well as many others, extensively discussed solar radiation calculation on inclined surfaces, like hills and gullies. Without measured values, these calculated values may be used, but can have large errors associated with them, especially for areas with some dust or cloud cover. Allen (1976) does not allow the shortwave radiation on any slope to decrease below 5 percent of the measured value.

Longwave radiation can be calculated from the Stephen-Boltzmann Law. Since this radiation is radiated uniformly from the sky, the effect of topography is not felt as in shortwave radiation. For clear sky conditions, the emissivity is commonly taken as a constant, whereas the moisture in the air is ignored (Bengtsson, 1976; Corps of Engineers, 1956). For areas with crop or forest cover, the computations are carried out separately for the covered and uncovered area, with the canopy temperature usually taken as equal to the air temperature.

The melt due to rain falling on a snow pack is not very large, but is very important because of its roles of increasing the snow temperature to 0 °C and filling the available water holding capacity of the snowpack. For shallow packs, this capacity can be very small. The actual heat content of rain is (Anderson, 1976):

$$H_p = \rho_w C_v (T_w - T_s) P \quad [3.19]$$

where

C_v = specific heat of rain water
T_w = wet bulb of air (often assumed = T_a)
T_s = snow temperature
ρ_w = density of rain water
P = depth of rainfall.

If the precipitation is actually snow at a temperature below 0 °C, the temperature profile of the snowpack will be rearranged, and its cold content increased.

Simplifications

Several investigators have studied the relative importance of the various energy balance components. This greatly aids in simplifying the computations when the situation justifies it or more detailed data are not available.

Zuzel and Cox (1975) measured daily values of wind, air temperature, vapor pressure, net radiation, and melt at a point. They found that for an area with continuous snow cover, vapor pressure, net radiation, and wind run explained 78 percent of the variations in melt, whereas air temperature and net radiation explained 60 percent. Temperature had a coefficient of determination of 0.51, and net radiation was 0.40.

Raffelson (1974) investigated the energy balance of isolated snowdrifts in Wyoming during melt. He found the sensible and latent heat components were about the same size, and both substantially larger than the radiation component. O'Neill (1972) and Gray and O'Neill (1974) found that net radiation was the predominant energy source for snowmelt for the Canadian Prairies when the snow cover was continuous, supplying 93 percent of the melt energy. For non-continuous cover, advection of heat from bare ground to isolated drifts caused 44 percent of the melt energy to be supplied by sensible heat transfer and 56 percent by net radiation. For an isolated drift,

Cox and Zuzel (1976) found that 69 percent of the energy available for melt and evaporation came from sensible heat input. The Corps of Engineers (1960) assigned a constant value to shortwave radiation during rain periods. King and Molnau (1976) noted that temperature index methods seem to work well for calculating snowmelt during overcast periods, indicating that radiation was relatively unimportant during those periods. Kuzmin (1973) explored five different simplifications of the basic energy budget method. He found that the use of temperature was possible for plains, but only if the mean daily temperature was greater than 2 °C.

Emphasis has been put on determining snowmelt by use of air temperature or a temperature index because of the ease of obtaining air temperatures and because temperature is the most easily extrapolated of meteorological variables. Usually, an air temperature equation has the form of

$$M = k(T_a - T_b) \quad [3.20]$$

where

k = degree day coefficient millimeter/day/°C
T_b = a base temperature, °C
M = melt, millimeter/day.

For most cases, T_b is assumed to be a constant. Granger and Male (1977) used $T_b = 0$ while observing that k increased during the melt season. They felt that this was due to the effect of radiation on air temperature during clear periods. Anderson (1973) used the degree-day method for clear weather melt periods and a simplified energy balance during rain periods when radiation factors can be more easily calculated. The melt factor is allowed to vary from a minimum on December 21 to a maximum on June 21, using a sine curve. For an Iowa watershed with no forest cover, the melt factor ranged from 7.3 to 3.6 mm/°C/day. Martinec (1960) developed a relationship between k and density for an open area in Bohemia.

McKay (1968) presents curves of degree-day factors for a shallow prairie snowpack. Gartska (1944) noted a strong correspondence between cumulative runoff and cumulative degree-hours above 0 °C. This relationship seemed consistent within a storm, but varied between storms. The SSARR model (Corps of Engineers, 1972) uses a table input of degree-day factor versus cumulative runoff. This did not work well for SSARR tests in the large Minnesota River basin (Kim et al., 1974). They used functions of the degree-day factor versus percentage of snowpack depleted as well as cumulative degree-days. Both methods seemed to work well for these large basins with shallow packs. King (1976) used the degree-day method on small watersheds in the Palouse Prairie. He used k as a function of cumulative degree-hours with good success. However, he found that different functions may be needed for basins with different aspects because of the rolling topography. Jolley (1973) also found that the summation of degree days gave good correlation to snowmelt runoff on a watershed near Ottawa, Ontario which was half grassland and half forested.

Bengtsson (1976) developed the idea of an equilibrium temperature to use in place of the base temperature. This is the temperature at which no net transfer of heat between the air and snow takes place. By equating the energy balance approach with the degree-day factor, he found that k could be determined as a function of wind speed for a forested watershed and a function of solar radiation for nonforested areas.

Following the degree-day method of simplified snowmelt calculation in popularity are the Corps of Engineer equations as outlined in their publications of 1956 and 1960. While developed using data from deep mountain snowpacks, these equations have been used in New Brunswick (Davar, 1970), Minnesota (Kim et al., 1974), as well as in areas of deeper snowpacks. Good results for these equations were reported in a comparison of several watershed models with snowmelt routines (NWS, 1972). These equations for open or partly forested areas include air temperature, a wind speed function, rainfall, and a constant radiational component for rainfall periods. For rain-free periods, forest cover, solar radiation, and computed long-wave radiation are added to the equations.

Full derivation and use of the equations can be found in various publications (Corps of Engineers, 1956, 1960). It is worth noting that the most complicated equation is one in which the land is open or partly forested, as in most agricultural lands.

In some operational and many research situations, it may be possible to measure melt directly, rather than rely on equations for the snowmelt input to mathematical models of small watersheds. These have also been used successfully in testing snowmelt equations. Lysimeters measure the meltwaters directly, and are reported by Thompson et al. (1975), Davar (1970), Haupt (1969), and Corps of Engineers (1955). Others have used a lysimeter-snow pillow to obtain both the change in snow water equivalent and water from the bottom of the pack, and used these measurements for actual forecasting (Cox and Zuzel, 1973; Cox, 1971; Molnau, 1971).

Vegetative and Topographic Influences

Vegetation and topography influence the accumulation and melt of snow. Attempts to manage the snow by use of windbreaks or snow fences have given new insight into the influence of natural features on melt of snowdrifts.

Saulmon (1973) found that water loss from induced drifts in eastern Montana averaged 50 percent, but total water yield was increased an average of 112 mm. McCool (1976) used snow fences to keep drifts from forming in high erosion hazard areas. If such drifts are formed on an area to be modeled, consideration must be given to models of drifts. The different cropping patterns or growth of windbreaks will also greatly influence snow catch and drifting (Greb, 1975; Tabler, 1975a; Tabler, 1975b; Willis et al., 1969; Frank and George, 1975). Since large drifts normally form in areas of rolling topography, the differential melting due to the varying influence of solar radiation on melt must be considered (Dunne and Black, 1971; Cox and Zuzel, 1976).

SNOWMELT IN HYDROLOGIC MODELS

Many hydrologic models include routines which will compute the amount of snowmelt by any one or combination of methods mentioned in previous sections. Very few models have been designed primarily as snowmelt models; normally, the snowmelt routine is added to the precipitation section where the water input to the main part of the hydrologic model is determined. Thus, the following sections describe snowmelt routines from several different types of hydrologic models which have been used in a wide variety of situations and may be applicable to small watersheds.

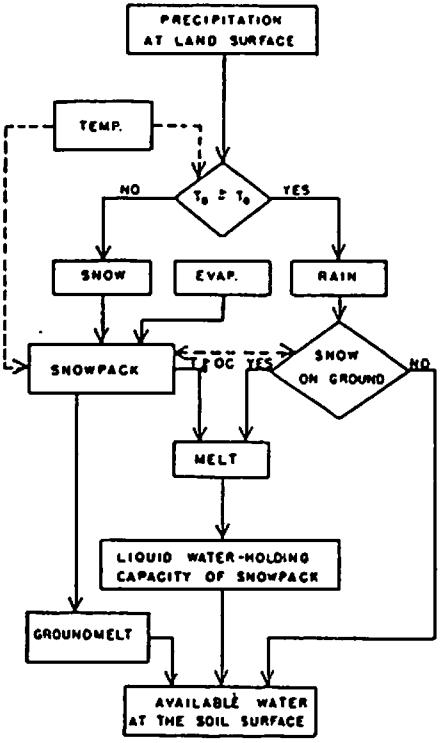


FIG. 3.13 Snow accumulation and ablation (from Riley et al., 1969).

In general, all comments in previous sections about areal variability of rainfall input to hydrologic models also apply to snowfall input. With few exceptions, some type of temperature function is used to determine the forms of precipitation. Snowfall is usually more uniform over an area than is rainfall because of the nature of the weather systems which cause snow to form. However, it is not the snowfall but the melt from the snow on the ground that is of interest. This snow may be very unevenly distributed, particularly in areas of rolling terrain or vegetation variation that can cause drifts to form. In these cases, the watershed may have to be divided up into areas corresponding to the major drift areas, rather than some other hydrologic unit.

US Department of Agriculture Hydrograph Laboratory (USDAHL)

The USDAHL model (Holtan et al., 1975) is an example of the use of a minimal amount of snow data. Precipitation input to the model must be tagged as either rain or snow, and the watershed is considered to be completely snow covered or not covered. Snow is melted by use of a single degree-day type equation. In an attempt to overcome the limitations of this equation, King and Molnau (1976) added a separate subroutine that would allow the program to use daily maximum and minimum temperature to determine if the precipitation was rain or snow, and treated each zone separately. The melt was based on a degree-day melt factor which varied with the accumulated degree days above 0 °C.

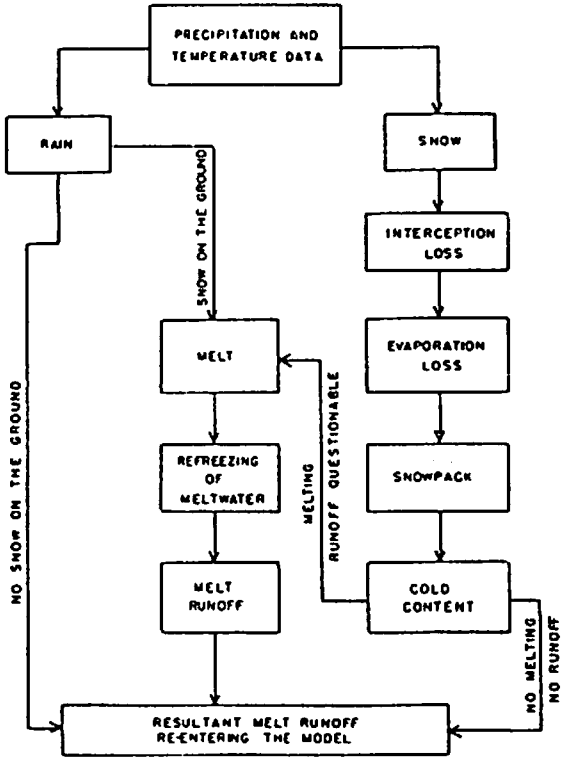


FIG. 3.14 Block diagram of snowmelt for the Ohio State University Model (from Ricca, 1972).

SSARR

The Streamflow Synthesis and Reservoir Regulation (SSARR) model, developed by the Corps of Engineers (1972) for use in forecasts in the Columbia River Basin, uses an extensive set of options for snowmelt. These include calculating melt on the entire basin, a split basin (snow on part only), or by elevation band. The methods of calculation can be by the generalized energy budget equations or by a degree-day equation. The degree-day factor may be entered in a table for variability throughout the year, or it may be a function of accumulated seasonal runoff. The snow covered area can be entered on a table of values or computed as a function of accumulated runoff. The University of Minnesota (Kim et al., 1974) modified SSARR to also include the option of varying the degree-day factor with the accumulated degree-day above the base temperature.

Utah Water Research Laboratory

The flow chart for this hybrid model (Eggleston et al., 1971; Riley et al., 1969) is shown in Fig. 3.13. This is a routine used in a hybrid computer model and illustrates some of the necessary steps in a mass budget of snow on the ground. This model has been used successfully in mountain snowpack situations, but there is nothing in its development which suggests it would not work on agricultural watersheds. It includes sections for dealing with steeply sloping areas where a radiation index is used to compute actual radiation on a slope.

National Weather Service

The National Weather Service model for forecasting streamflow has been tested on several types of watersheds, including those of primarily agricultural character (National Weather Service, 1972; Anderson, 1973; Kim et al., 1974). When the temperature is above freezing, a degree-day approach is used, but when more than 2.5 mm of rain falls in a 6-h period, an energy balance equation is used to compute melt. There is a negative melt factor included to lower the pack temperature when the air temperature is less than freezing.

Ohio State University Model

The Ohio State University Model (OSUM) (Fig. 3.14) is derived from the Kentucky Watershed Model (KWM) (Ricca, 1972; James, 1972). The OSUM includes a snowmelt routine developed specifically for agricultural watersheds and was tested on the Coshocton Watersheds. The model includes simplified versions of each of the energy balance terms and requires daily average dewpoint, wind run, solar radiation, and maximum and minimum temperature. A factor for increasing or decreasing the rate of cold content change, essentially a thermal conductivity index, is included, which is very important in shallow snowpacks.

Other Models

Based on the above models, there are other models which are used for various application. Some of these that may prove useful in small watershed studies are those of Allen (1976) and Obled and Rosse (1977). Development of a model for the prairies of Canada is underway at the University of Saskatchewan, which specifically addresses the problem of shallow snowpacks, frozen ground, and other problems associated with small agricultural watersheds (Male and Gray, 1975; Norum et al., 1976).

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