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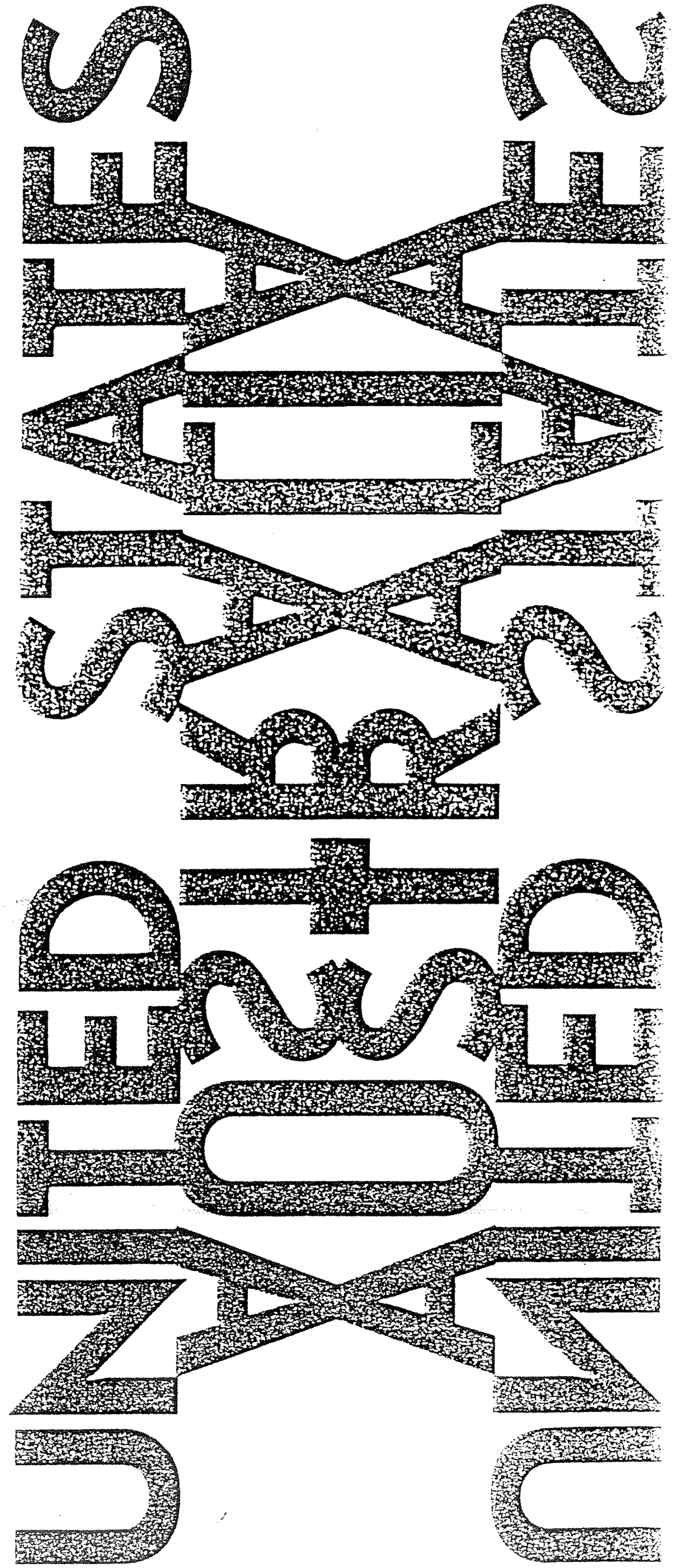

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Precipitation on Intermountain Rangeland in the Western United States

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Precipitation in its many forms is crucial to the water supply which maintains life of the Earth. As a result, scientists have devoted much attention over the years to collecting information about precipitation. The continents of the Earth receives about 72 cm/yr of precipitation (Sellers, 1965) and the United States receives 76 cm/yr, very near the continental average. Unfortunately, most of the rangeland areas of the United States receive considerably less and have unique problems of seasonal and orographic distributions which limit forage production.

In this paper, we did not attempt the impossible task of providing an exhaustive catalog of all recent progress in precipitation analysis. Rather, we selected those sectors about which we, the writers, were most knowledgable with the danger that this sample is doubtlessly biased. We believe that the sample is sufficiently large to reflect the current state-of-the-art, and further, point out some of the difficulties in analyzing precipitation with the tremendous temporal and spatial variability encountered in the basin and range topography of the Western United States.

Often, precipitation total, as measured with a recording gage, is not the parameter which is informative to the range user; rather, it is the amount of infiltration which is potentially available for plant growth. However, such parameters are not widely available. Most of the precipitation analysis and models developed by hydrologists are intended as input to hydrologic models, i.e., models intended to generate streamflow. Many of these findings can, however, be used for range management.

Because the physical processes involved in precipitation generation are not completely known and are exceedingly complex, the analyst often resorts to using statistical tools. Franz (1971) said that ". . .rainfall characteristics are often very difficult to mimic with statistical tools currently available. Empirical adjustments and a proliferation of parameters must often be used to obtain an acceptable level of performance. Considerable judgment and trial-and-error testing will be required for some time to come in the development and in the application of these models."

The characteristics of precipitation in the rangeland areas of the United States are varied and depend on the atmospheric moisture source, season of the year, and elevation, among other things (Figure 1). Thus, much winter moisture in the U.S. western rangeland areas (generally west of the 100th meridian) results from Pacific Ocean moisture carried into the area by prevailing westerly winds (Battan, 1974). These winds may result in orographic precipitation patterns with more rainfall and greater snowfall at high elevations. Figure 1 illustrates the marked change in seasonal pattern (reflecting the different moisture sources from north (Spokane) to south (Tucson)).

At other times of the year (especially in the Southwestern U.S.), the atmospheric moisture results from a slow air movement from the Bermuda high pressure area toward a thermal low pressure area, often called the "Las Vegas thermal low" by meteorologists. Thunderstorms are prevalent during such atmospheric conditions and produce the summer peak like that shown for Tucson on the distribution graph (Figure 1). Thunderstorms come in many sizes, shapes, and structures and fall into two broad categories: local (air-mass) and organized (frontal) thunderstorms. Local storms are fairly isolated storms with a short lifetime, high intensity rain, and limited areal distribution. One or a group of organized thunderstorms implies a longer lifetime than the local storm. Organized storms form in lines or bands of thunderstorms, sometimes called "squall lines." They often initiate along, or ahead of, a cold front and nearly parallel to it.

Such differences in precipitation types and moisture sources require different analyses. Thus, the subsequent discussions are divided between Southwest thunderstorms and non-thunderstorms (interior northwest precipitation) with a review for each.

THUNDERSTORM PRECIPITATION

The local type (air-mass thunderstorms often lead disappointed ranchers and farmers to state that it has

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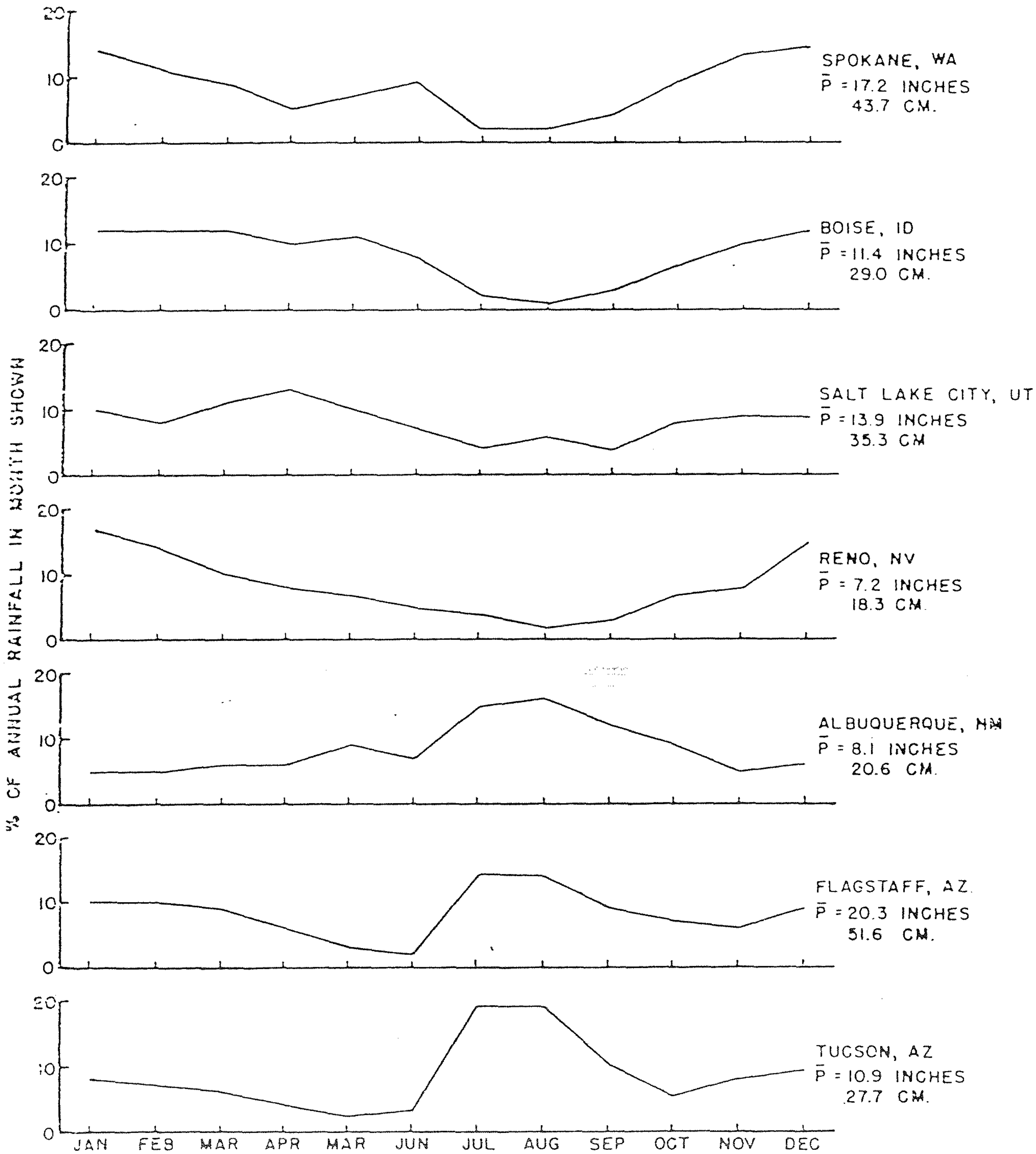

Figure 1. Distribution of annual precipitation at select stations in intermountain areas of the Western U.S.

rained everywhere but on their ranch. Exclamations like "we were completely surrounded by storms but somehow they all veered around us and we didn't get a drop all afternoon" are quite common. McDonald (1959) described this thunderstorm illusion with a graphical model similar to the following discussion. Fifty thunderstorms were randomly located in a 25,900 km^2 (10,000 mi^2) area (161 x 161 km or 100 x 100 mi) using a table of random numbers. In level country, true air-mass thunderstorms are apparently randomly located. Because McDonald was concerned simply with the impression left in the observer's mind, it was not necessary to specify whether the storms or each observer's observation occurred simultaneously. Rather, the storms may have occurred at randomly distributed times over some duration, like an afternoon. Osborn, Lane, and Kagan (1971) observed that on the Walnut Gulch Experimental Watershed in southeastern Arizona, the air-mass thunderstorm beginning times are normally distributed with mean and standard deviations of 1700 and 3.5 hours, respectively.

On the storm map (Figure 2), five observers were randomly located with a table of random numbers with the constraint that each observer be at least 40.25 km (25 mi) from any boundary point of the square (each observer was presumably able to detect storms 40.25 km from his location), so the observer's circle of shower detection lies within the model area.

Inspection of Figure 2 shows that each observer except #1, will observe storms fairly well distributed around the horizon. To illustrate this, the storm positions were measured with a protractor and plotted in the lower portion of Figure 2. The number of observed showers ranged from 4 (observer #1) to 12 (observer #2) (Figure 2). In addition to seeing fewer storms, observer #1 was surrounded by two quadrants without any storms, with observer #3 having one quadrant (SW) without any storms. McDonald showed with probability theory that only 8 percent of all observer-quadrants will be storm-free in the model.

The expected value for the number of storms detected per circle is 9.82, whereas the mean for the five observers was 9.4. However, only 5.9 percent of the entire model area received rain, i.e., the probability was only 0.059 that rain will actually fall upon any observer.

Hershfield (1962) showed that the coefficient of variation (standard deviation divided by the mean) of annual precipitation is generally larger in the western U.S., as illustrated in Figure 3. The paucity of gages in this mountainous area may tend to smooth the actual variability somewhat but it does illustrate the role of thunderstorms in contributing to such variability. A series of monthly, water-year, or growing-season rainfalls would be most useful to range scientists planning range utilization programs. Hershfield stated, "the ideal procedure then would be to use a theoretical distribution. . .to construct a series of maps, with isohyetal patterns at least as complex as those of the mean map, for several probability levels. However, this would be a rather expensive project, which no one had considered important enough to undertake."

This later objective was partially fulfilled in a Regional Research Report prepared by members of the staffs at each land-grant institution in the west (Gifford, Ashcroft, and Magnuson, 1967). They obtained the probability of various precipitation amounts for weekly periods in various time intervals of a year at select stations in the Western U.S. Rather than constructing iso-probability lines, they estimated precipitation for each station and distance for which the data can be reliably extrapolated. They used a smoothing technique with a weighted 3-week moving average in which double weight was given to the week under consideration. This technique eliminated some of the random variations in the probabilities that result from using a short 30-year record for each station. This same Regional Research Committee also produced a report (Heerman, Finkner, Hiler, 1971) giving the probabilities of sequences of wet and dry days which may be helpful to range scientists.

THUNDERSTORM PRECIPITATION MODELING

Thunderstorm precipitation can best be described by realizing that three elements are necessary: describing the distribution of (1) rainfall events, (2) depths, and (3) the areal distribution pattern. Because of the complex physical laws involved in rainfall processes, hydrologists generally use a probabilistic description of a local variable to predict the statistical properties of future rainfall. Small rainfall amounts are important to most rangeland plant species and, therefore, many precipitation models designed to predict large basin runoff present an incomplete picture of the precipitation distribution (e.g., Lane and Osborn, 1972, Duckstein et al., 1972).

Figure 4 illustrates the spatial variability of a thunderstorm on the Walnut Gulch Experimental Watershed in Southeastern Arizona, and shows the annual precipitation totals at certain gages, which include numerous thunderstorms and some winter frontal storms. Each small circle represents a recording rain gage on the 150 km^2 (58 mi^2) area and its immediate vicinity with each gage approximately 1.4 km (0.9 mi) apart.

The storm of July 16, 1967, had 6.83 cm (2.69 in) at the center in 69 min, with 2.82 cm (1.11 in) occurring in 20 min. The maximum intensity of 18.3 cm/hr (7.2 in/hr) for several minutes is typical of these thunderstorms. Variability like this leads to highly variable annual totals in relatively shorth distances with annual maximum and minimum often only a few miles apart and the maximum often twice the minimum. In 1967, the maximum was only 60 percent more than the minimum. No definitive pattern

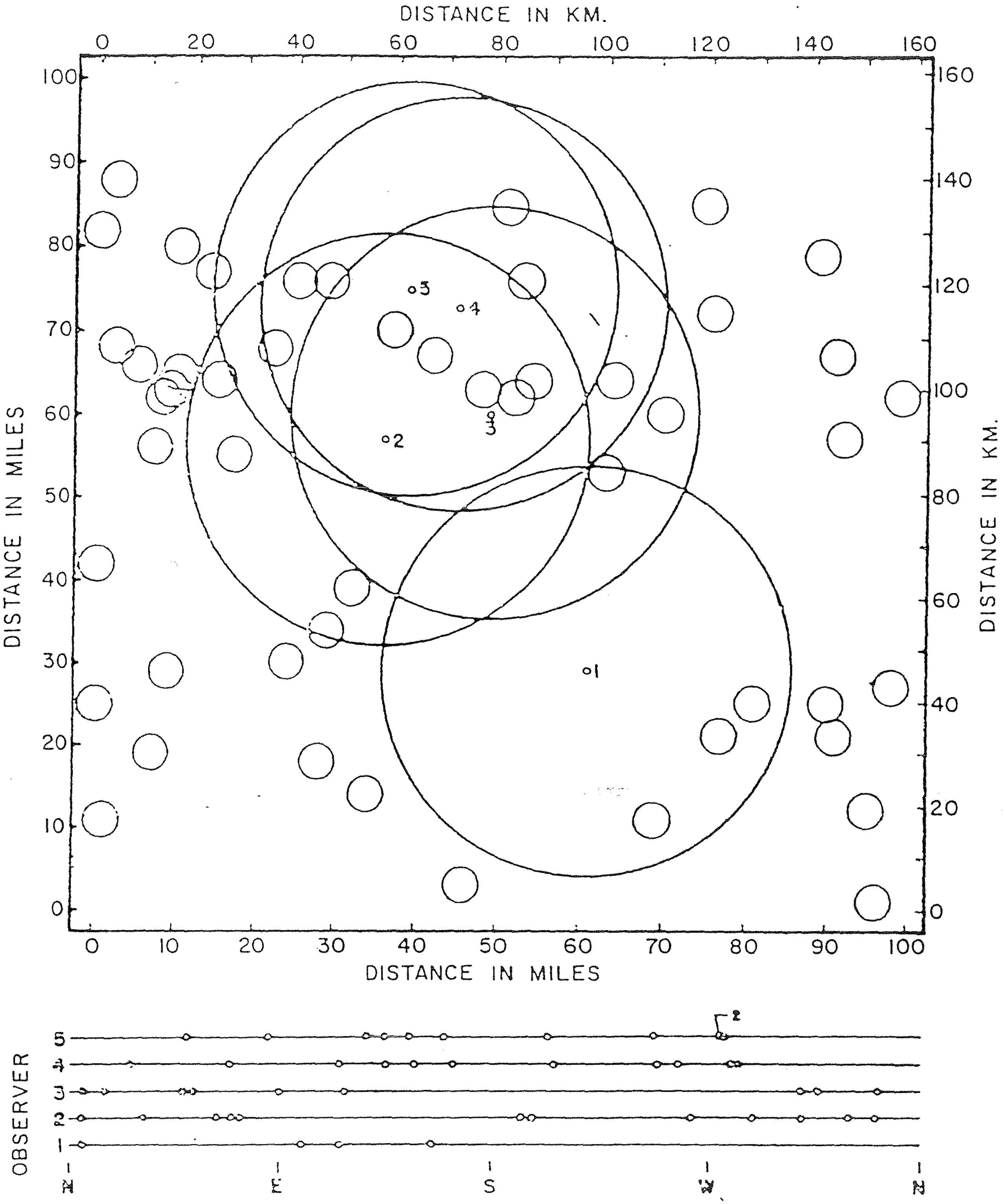

Figure 2. Hypothetical storm map showing the location of 50 thunderstorms with five observers randomly located. Each observer circle was 25 miles radius while each storm had a 1-mile radius. The lower portion shows the location of the storms in relation to each observer. (After McDonald, 1959).

for the occurrences has been discernible statistically on Walnut Gulch.

Distribution of Rainfall Events

Weiss (1964) and others found sequences of wet and dry periods are best described by a Markov chain probability model. Smith and Schreiber (1973) like Hershfield (1970) showed that a Markov chain adequately described air-mass thunderstorm rainfall in addition to the frontal-type situations. They also showed that in addition to describing the beginning of the summer "rainy" season in southeastern Arizona, the Markov model also gave a good fit to the cumulative distribution of wet days per season for the Tombstone rain gage (Figure 5), with 73 yr of data. Figure 5 also shows that the Markov model better fitted the historical data than did the independent (Bernoulli) daily model. The Bernoulli model also consistently overpredicted the probability of the starting day of the "monsoon" season. The Markov model with segmented nonhomogeneity was obtained by partitioning the wet and dry probabilities during the season which improved the fit of the historical data as compared with using the average wet and dry probability throughout the season.

Distribution of Rainfall Depths

Rainfall depths for periods of 1 day and longer are generally widely available in publications, like the climatological reports of the National Weather Service. Accordingly, many hydrologists and meteorologists have investigated expressions for the maximum precipitation depths and have developed simulation models to predict the depth as input to runoff models. Table 1 shows examples of such approaches. Some of the differences in the distributions found to describe the maximum precipitation depths (Table 1) are undoubtedly associated with differences in the precipitation source involved (i.e., frontal versus air-mass thunderstorms) and with differences in precipitation type (i.e., snow versus rain).

Mixed distributions have been suggested by Hawkins (1971), Singh (1968 and 1971), and others to describe hydrologic processes which are notoriously non-normal. The problem has an old and honorable history dating back to the eminent statistician Pearson (1894) but because of the laborious calculations necessary for accurate solution, it has received only limited application.

That hydrologic variables are not normally distributed should not be surprising. Hawkins (1971) quoted Reich (1969) on this problem:

> Nature has no back room boy dictating that flood series (or precipitation depths) should follow a particular law. . . Rather let us visualize. . .mathematical functions for what they are—merely a continuation of man's efforts at curve fitting.

In an extension of their 1973 work on thunderstorm occurrence, Smith and Schreiber (1974) showed that the seasonal rainfall depth for three gages located in southeastern Arizona was describable with a compound or mixed exponential distribution of the form shown in Figure 6. All of the curves shown (Figure 6) may be approximated by two segments divided by some point of inflection, X_c, or a mixed exponential of the form

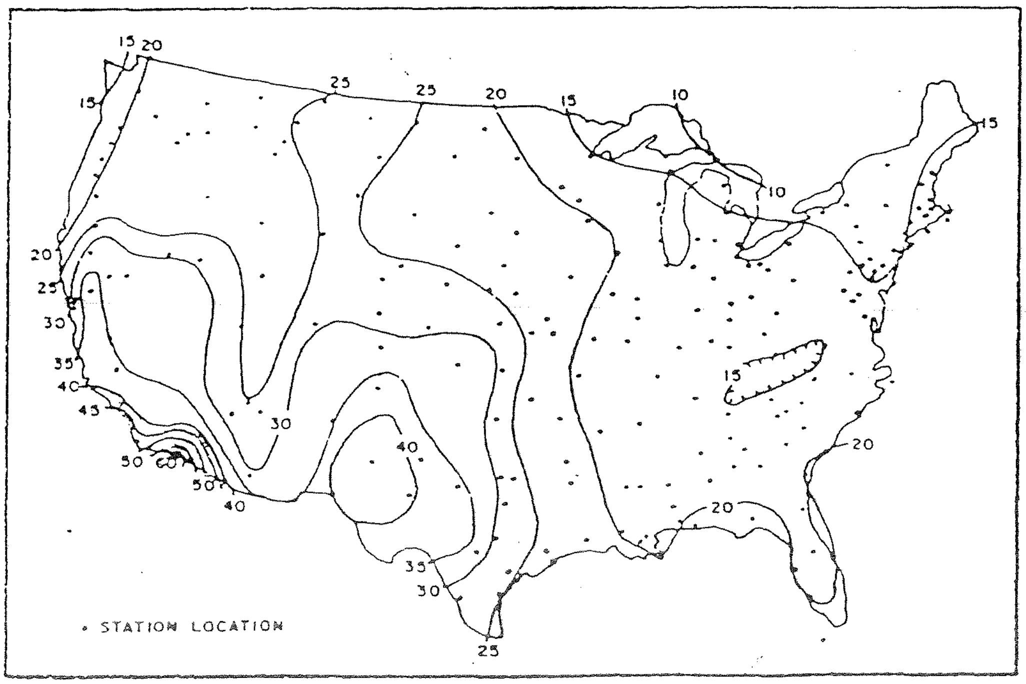

Figure 3. Coefficient of variations of annual precipitation in percent. (Hershfield, 1962).

$$P(X \geq x) = 1 - ae^{-\lambda_1 x} - (1\text{-}a)\, e^{-\lambda_2 x} \qquad \ldots\ldots(1)$$

(Woolhiser, 1975). The very skewed shape of the density function increases the uncertainty of the sample probabilities as depth of rainfall increased.

Smith and Schreiber (1974) stated that this model, ". . .could be used in practical watershed management for this hydrologic region to estimate the probability of extreme drought or wet seasonal rainfall, which is relatively indeterminate from short records." Further work is presently underway at the Southwest Watershed Research Center to extend this model beyond the area of its limited initial testing.

Hydrologic variables (especially precipitation) are almost always the result of countless causes or factors, like the phenomena causing thunderstorms vs. phenomena causing general low-intensity, frontal rain and snow. Such variables, therefore, can be measured as sampling either the combined effects of several phenomena in a single sample or sampling different phenomena separately over time in a combined sample like with precipitation.

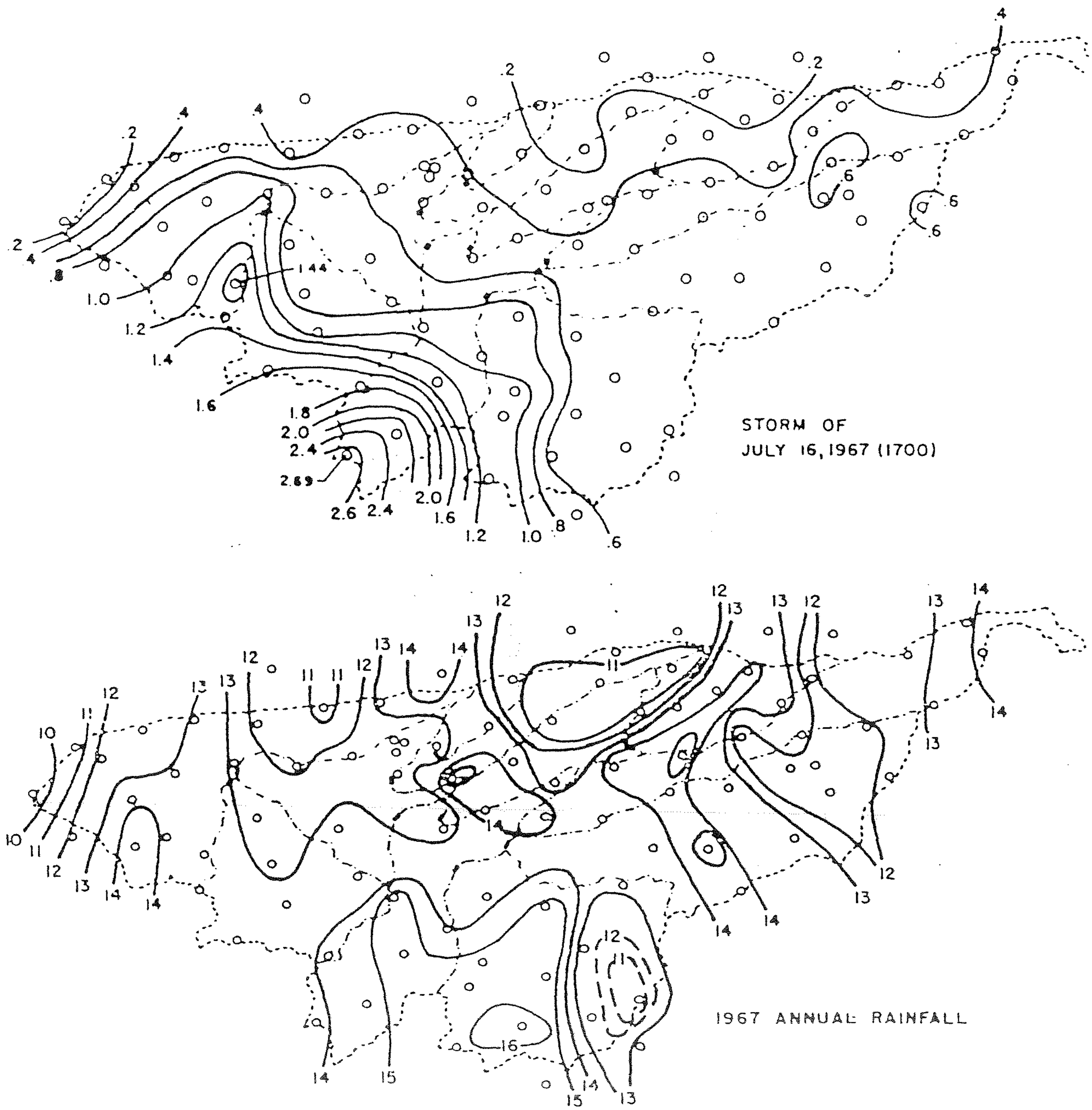

Figure 4. Isohyetal maps of the July 16, 1967, storm, and the annual precipitation (in) for 1967.

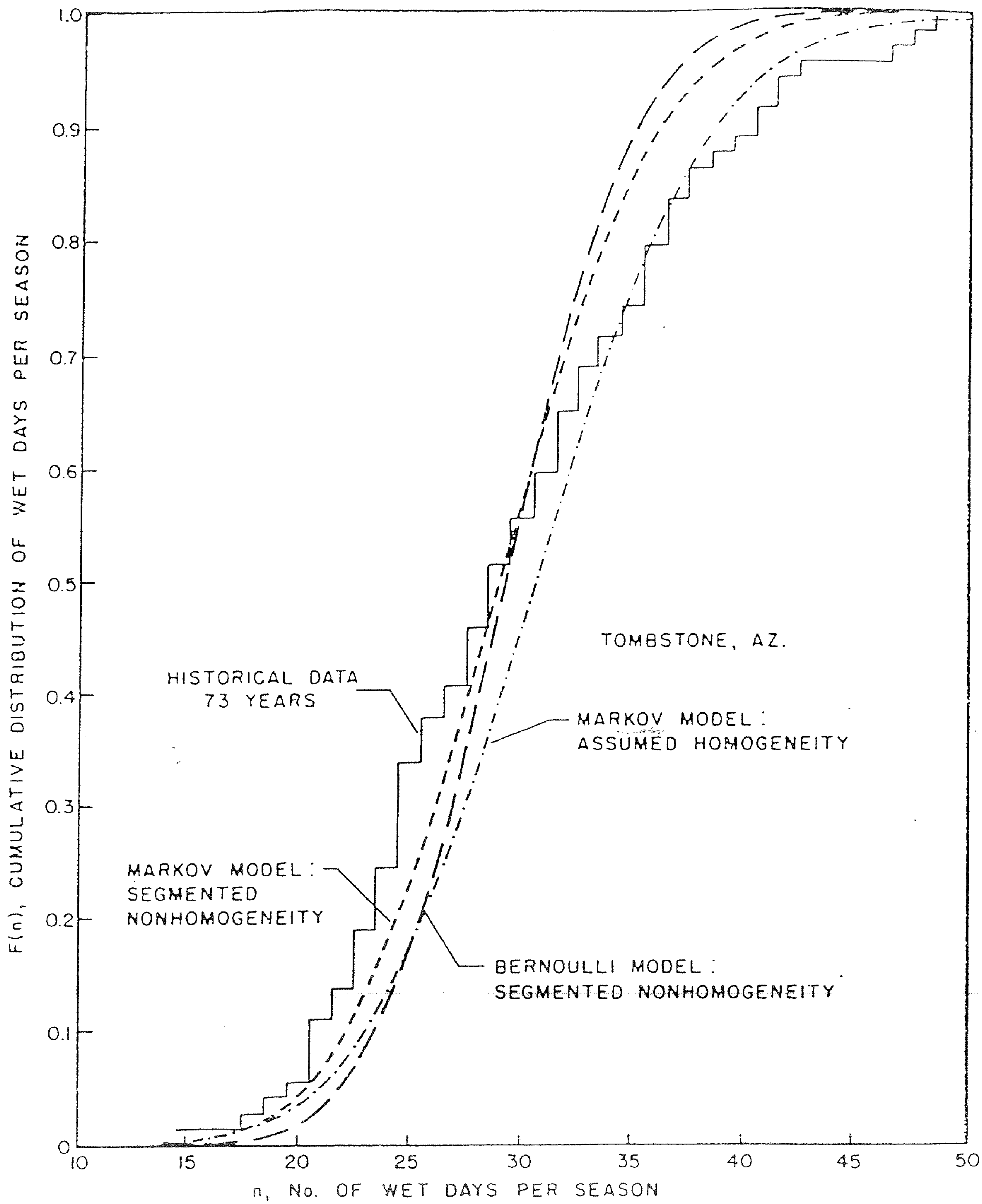

Figure 5. Predicted and observed cumulative distribution of the number of wet days per season at Tombstone, Arizona. (Smith and Schreiber, 1973.)

A distinction should be made between mixed distributions and mixed variables. In mixed distributions, the sample or measurement taken is from describable sources or populations, i.e., storm rainfall or annual flood peaks. For mixed variables, the measurement or sample is taken of components already in a combined state, e.g., streamflow measurement containing surface runoff and groundwater flow.

A two-component mixed distribution as used by Hawkins (1971) was defined as:

$$f(X) = a_1 f_1(X) + a_2 f_2(X) \quad \ldots\ldots\ldots\ldots\ldots (2)$$

where $a_1 + a_2 = 1$, $0 < (a_1, a_2) < 1$, and f_1 (X) and f_2 (X) are normal distribution functions described by μ_1, σ_1, μ_2, and σ_2, their means and standard deviations. a_1 and a_2 are the relative weights of each component distribution. The subpopulations are assumed to be normal and have no skewness, although the mixture may, indeed, be skewed.

The probability density function for the two-component mixed normal distribution is thus:

$$f(X) = (a_1/\sigma_1\sqrt{2\pi})\exp[-(X-\mu_1)^2/2\sigma_1^2] + (a_2/\sigma_2\sqrt{2\pi})\exp[-(X-\mu_2)^2/2\sigma_2^2] \quad \ldots (3)$$

Hawkins (1971) then demonstrated the utility of the mixed distribution model on the maximum 24-hr storm intensity for Farmington Warehouse, Utah (Figure 7). Although this example is not as dramatic as what might be obtained for some runoff stations, the bimodal peak on the density graph illustrates a distribution which is hard to fit with one function. The mixed exponential by Woolhiser (1975) mentioned earlier should also fit this data. Seemingly, further work along this line is warranted, especially since the wide availability of digital computers makes the computations easier.

Thunderstorm Patterns (Depth-Area Relation)

Court (1961) presented an early and noteworthy treatment of the description of a thunderstorm's areal pattern comparing the formulas set forth earlier. These efforts at depth-area relations were often peripheral to the main object of investigation. All the relationships reported by Court were empirical (usually obtained graphically) and related the average rainfall inside an isohyet to various powers of the area, or to its logarithm or exponential.

Court (1961) then proposed the bivariate Gaussian (or 'normal') distribution be used as a possible representation of the depth-area relation of storm rainfall. He compared

Table 1. Daily or shorter precipitation generation models.

Author	Date	Storm Depth Simulation
Brakensiek	1958	Log-probability and Gumbel extreme value distributions.
Fogel, Duckstein & Sanders	1971	Geometric probability for point depth with a Poisson distribution for the number of rainfall events per season.
Franz	1971	Multivariate normal distribution with persistence provision using a Markov sequence.
Grace & Eagleson	1966	Multi-stage model with storm length selected from a distribution which in turn was related to storm depth.
Khanal & Hamrick	1971	Markov Chain (first-order).
Kotz & Newmann	1963	Gamma distribution.
Nicks	1971	Maximum daily rainfall on a network was a skewed normal distribution with a Markov Chain for occurrence.
Osborn, Lane & Kagan	1971	Complex process using number of cells (Poisson distribution with a three cell minimum) with the center depth approximated by a negative exponential. Separation empirically determined from sample data. Point storm depths are the sum of the depth for all cells.
Pattison	1965	Sixth-order Markov Chain using hourly rainfall states (each state implies a depth range).
Sariahmed	1969	Weibull distribution for storm duration (depth a linear function of duration).
Skees & Shenton	1971	Censored Gamma distribution and numerous transformations to approximate a normal or Gamma shape.
Smith & Schreiber	1973	Mixed exponential distribution for storm depth with Markov Chain for occurrence.
Todorovic & Woolhiser	1971	Exponential for daily rainfall with the total amount of precipitation in an N-day period from a Markov Chain or a binomial distribution.
Wiser	1971	Mixed distribution generated from: 1) a storminess parameter (second-order autoregressive scheme) and 2) local process deterministic component (e.g., orographic effects) with a random selection that determines the simulated precipitation.

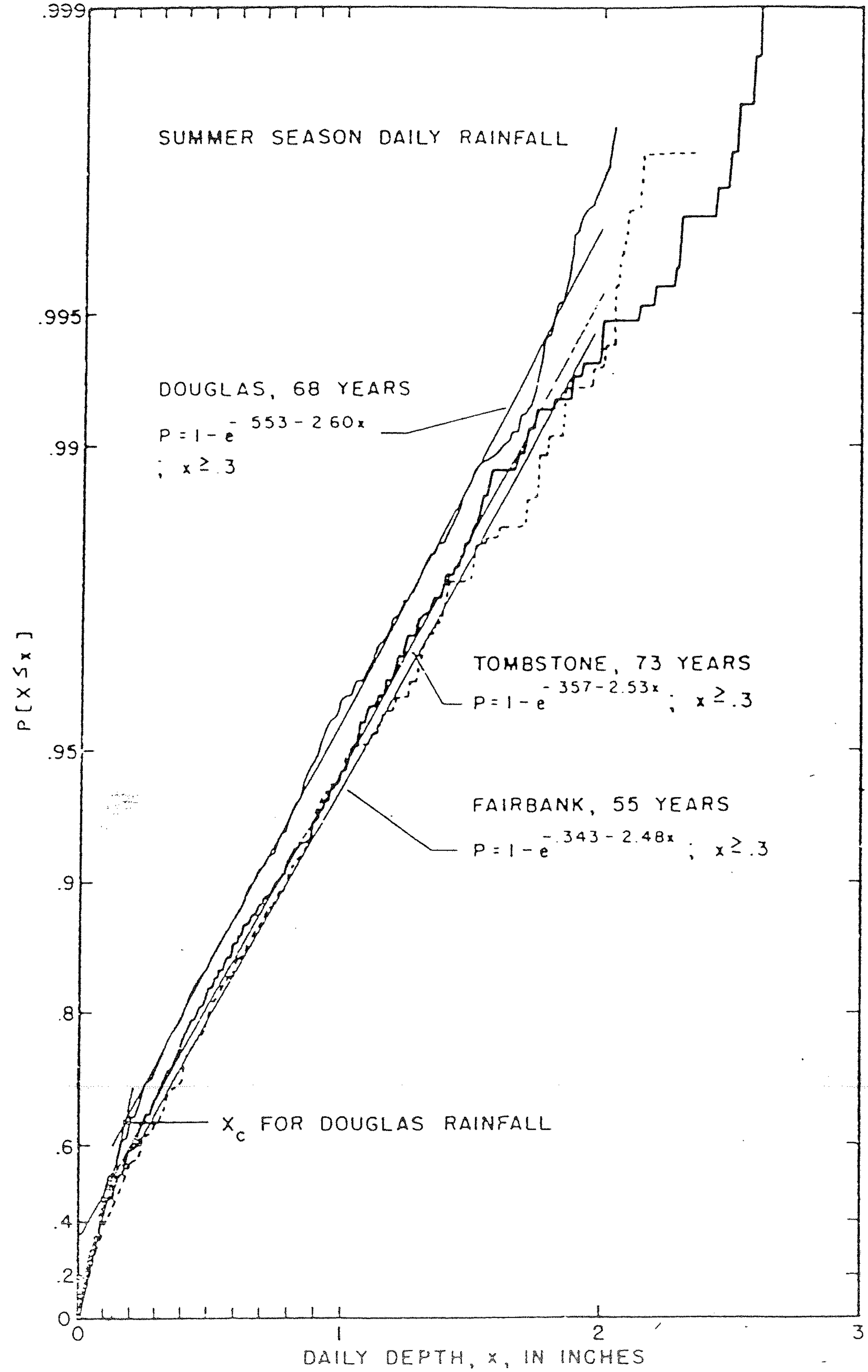

Figure 6. Distributions of summer seasonal daily rainfall depths for three sampling stations in southeastern Arizona. (Smith and Schreiber, 1974.) (in x 2.54 = cm)

his Gaussian (normal) model with that of other investigators as shown in Figure 8.

Interestingly, the models reflect the differing storm patterns studied from the limited extent air-mass storms described by Woolhiser and Schwalen (1959) in Arizona to the broad expanse storms of Huff and Stout (1952) in Illinois.

Recently, Smith (1974) investigated the areal properties of thunderstorms in Arizona and described the storm

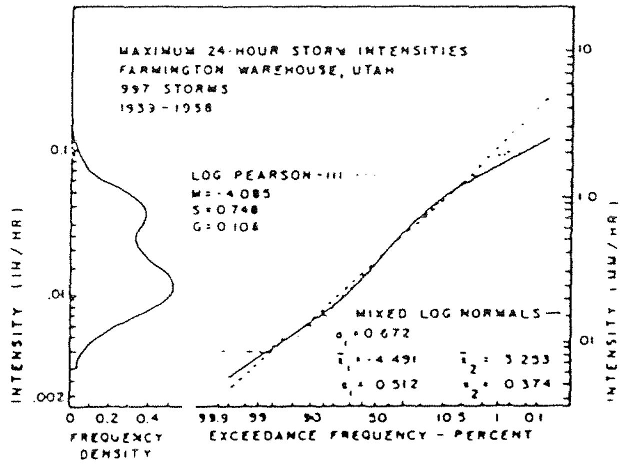

Figure 7. **Frequency curves for precipitation data, comparing mixed normal distributions to Pearson-III distributions. (Hawkins, 1971.) (in/hr x 2.54 = cm/hr.)**

pattern with a monotonic dimensionless depth-area relationship. He also expressed the depth-area relations for air-mass thunderstorms proposed by three other investigators in dimensionless form as shown in Figure 9. Assuming storms are occurring randomly, uniformly distributed in space, the rainfall population at any point may be considered to be composed of samples taken with equal likelihood from any point within the associated storm. With probability statistics he developed from this assumption, a general relation was presented between normalized storm isohyetal pattern, center depth probability, and point rainfall probability. This general relationship and the dimensionless depth-area relationships were used to obtain a record of the point rainfall depths which compared quite favorably with the historical record at the Tombstone rain gage (Figure 10).

INTERIOR NORTHWEST RANGELAND PRECIPITATION

Rangeland areas in the interior Northwest have maritime air from the Pacific Ocean as their moisture source for precipitation. The maritime influence is particularly noticeable during the winter, with greater average cloudiness, greater frequency of precipitation, and mean temperatures which are above those at the same latitudes east of the continental divide. The north-south mountains which dominate the Northwest result in a high percentage of the maritime moisture falling on western slopes, whereas the area east of these mountains receives only small amounts of precipitation. The sum result is an almost typical upland continental climate in summer, but

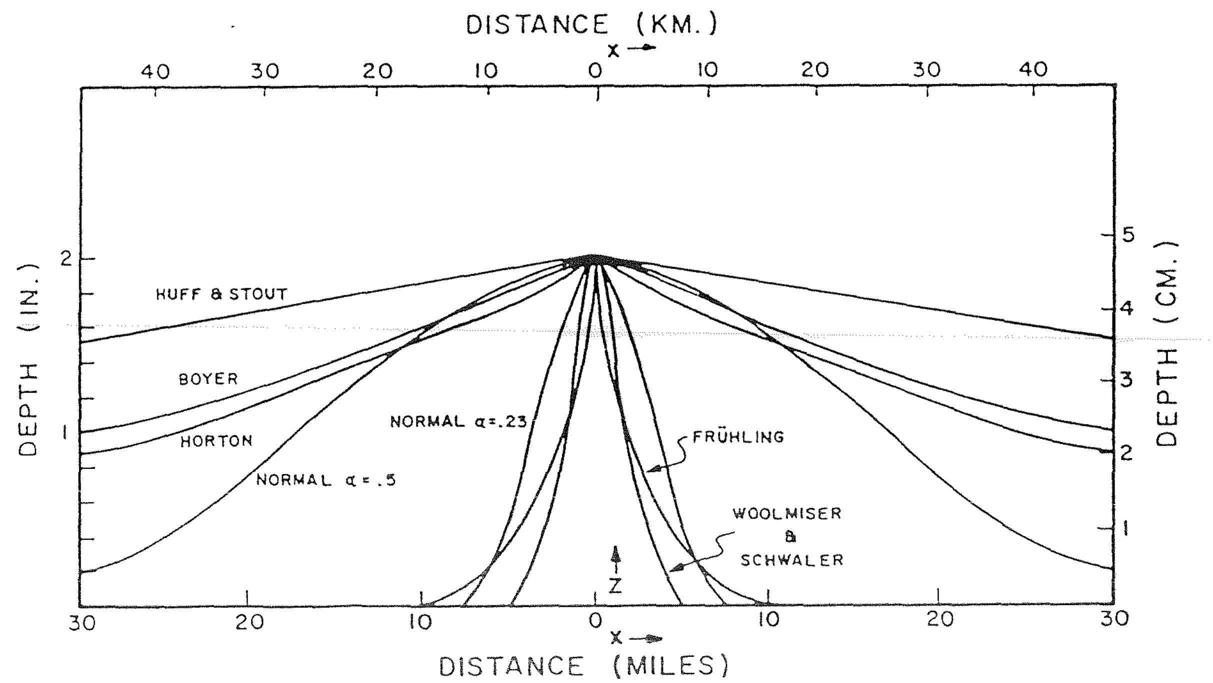

Figure 8. **Variation of isohyetal value z (in) with distance x (mi) from storm center, assuming central precipitation to be m = 2 in (5.08 cm), according to various formulas. (Court, 1961.)**

one tempered by periods of cloudy or stormy and mild weather nearly every winter. The normal precipitation in the Boise, Idaho, area shows a winter maximum and a very pronounced summer minimum (Figure 1). Within the interior Northwest region, elevation is the primary cause of great differences in precipitation over very short distances.

Annual precipitation on the Reynolds Creek Watershed, operated by the USDA, ARS, Northwest Watershed Research Center, varied from 254 mm (10 in) at the lower elevation of 1097 m (3600 ft) to near 1270 mm (50 in) at the highest elevation near 2134 m (7000 ft). Figure 11 shows the monthly distribution at three elevation sites in the watershed. Figure 12 shows the spatial distribution of annual precipitation over the watershed.

Even though winter precipitation predominates in the Reynolds Creek rangeland watershed, there are summer thunderstorms. However, they are very infrequent and since the previous discussion in this paper has covered them thoroughly, thunderstorm precipitation will not be discussed further.

Low Elevation Winter Precipitation

At the lower elevations, less than 1524 m (5000 ft), winter precipitation comes as both rain and snow. At these elevations, the most severe floods come between December and March (Johnson and McArthur, 1973). The usual antecedent conditions are persistent periods of extreme cold, which freeze the soil to considerable depth, and a shallow snow cover. Warm, moist, unstable air masses, accompanied by strong southwest winds, produce rain and cause rapid melting of the snow on frozen ground. The amount and intensity of the rainfall, the amount of snowmelt, and the imperviousness of the frozen soil combined to affect the flood severity.

Because the snow cover and soil frost depths at lower elevations change from day to day, devising a data

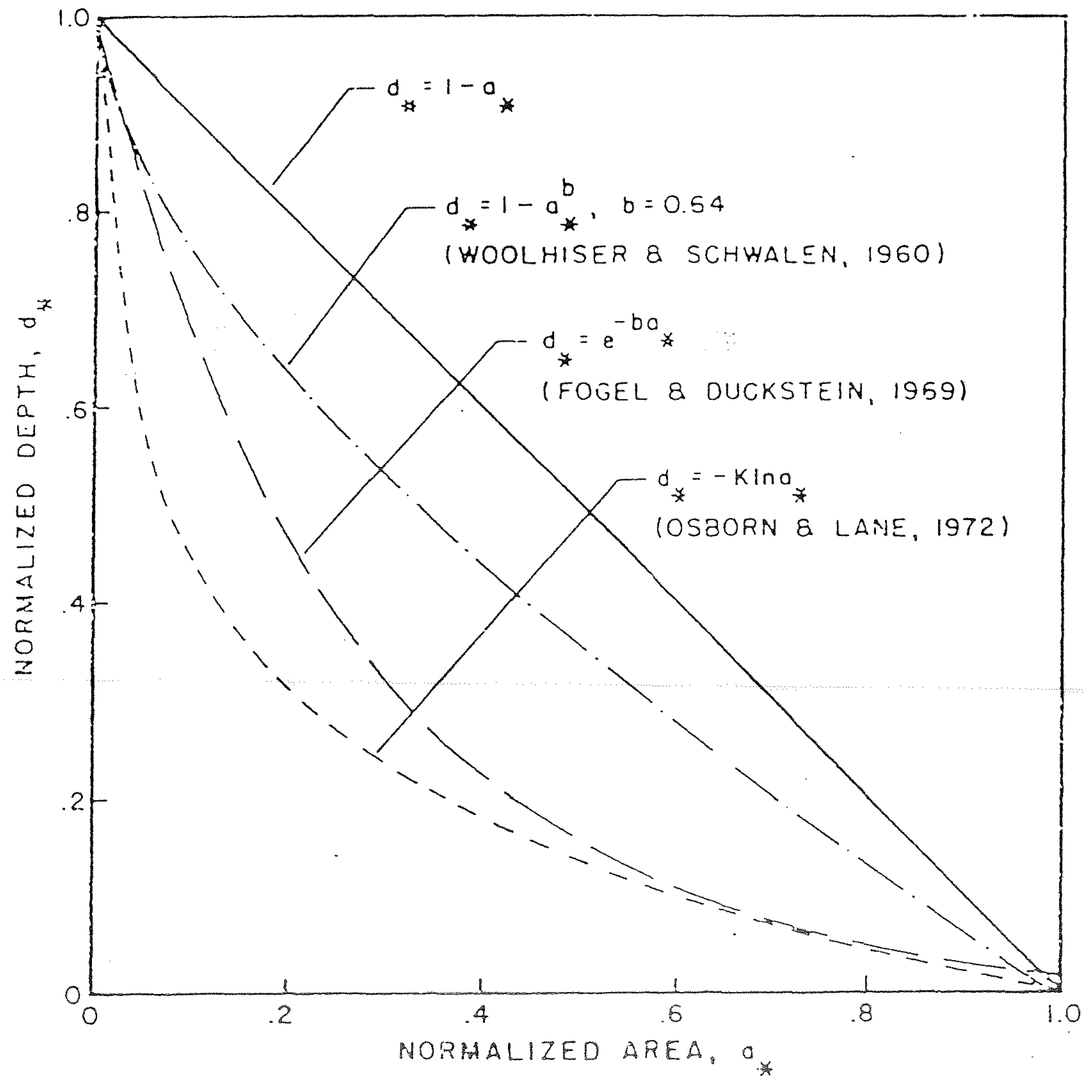

Figure 9. Normalized depth-area relations for air-mass thunderstorms proposed by three investigators. (Smith, 1974.)

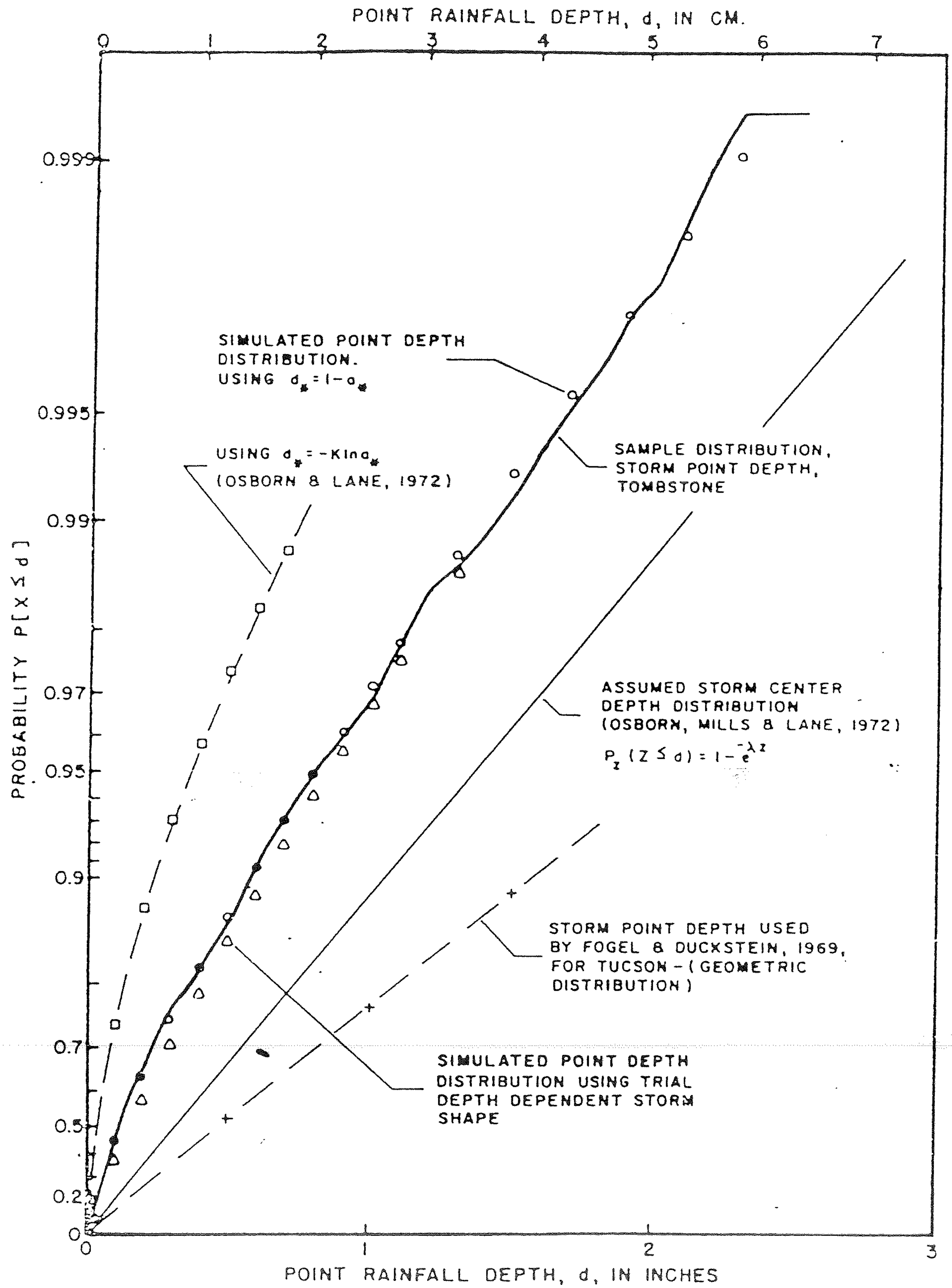

Figure 10. Storm center-depth distribution and several point depth distributions, including measured data for Tombstone, Arizona, and simulated distributions. (Smith, 1974.) (in x 2.54 = cm.)

collection network to provide a historical record and research data of antecedent conditions before a snowmelt-runoff event has been necessary. Such a data collection network was initiated on the Reynolds Creek Watershed during the 1971 to 1972 winter season. Data collected includes general snowline elevation, snow depths, snow-water equivalent, percent of snow cover, soil frost depth, and existing weather conditions. This data is being used to study snowmelt-runoff relationships at the lower elevations. Much of this water runs off rather than infiltrates the soil. Nonetheless, winter precipitation does recharge shallow water supplies that are utilized for stockwater tanks. In those areas of a deep soil profile, furrowing may be used to retain some surface runoff for soil infiltration. Such additional water harvested and infiltrated may prove crucial in these areas for establishing a more productive range. These areas are now used only in early spring and the principle forage is cheatgrass (*Bromus tectorum*). However, research is needed to evaluate this complex relation between winter stored soil water, negligible summer rainfall, and grass varieties that are adapted to the soil and water conditions.

Mid and High Elevation Winter Precipitation

At mid-elevation, 1500 to 2100 m (5000 to 7000 ft), precipitation comes as snow and is stored on the watershed until spring melt. Approximately 75 percent of the annual water yield from Reynolds Creek is from snowmelt. Discontinuous snow storage as massive drifts is prevalent on the watershed, which is typical of many millions of hectares of medium elevation sagebrush rangeland in the northwest. Research on the Reynolds Creek Watershed has quantified the magnitude of this water resource and has identified possible ways to manage this resource for

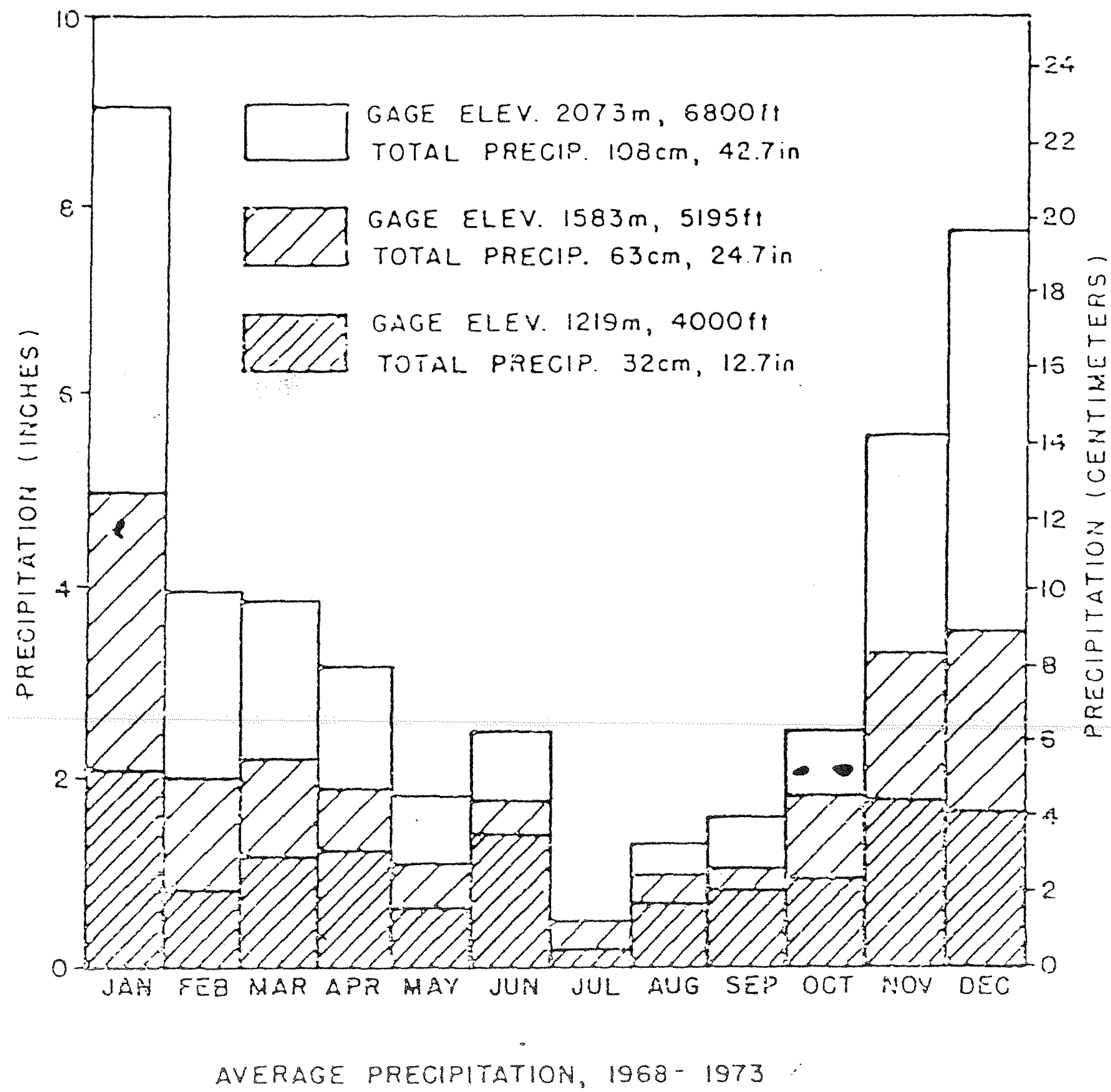

Figure 11. Reynolds Creek Watershed, Idaho. Average precipitation for 1968 to 1973.

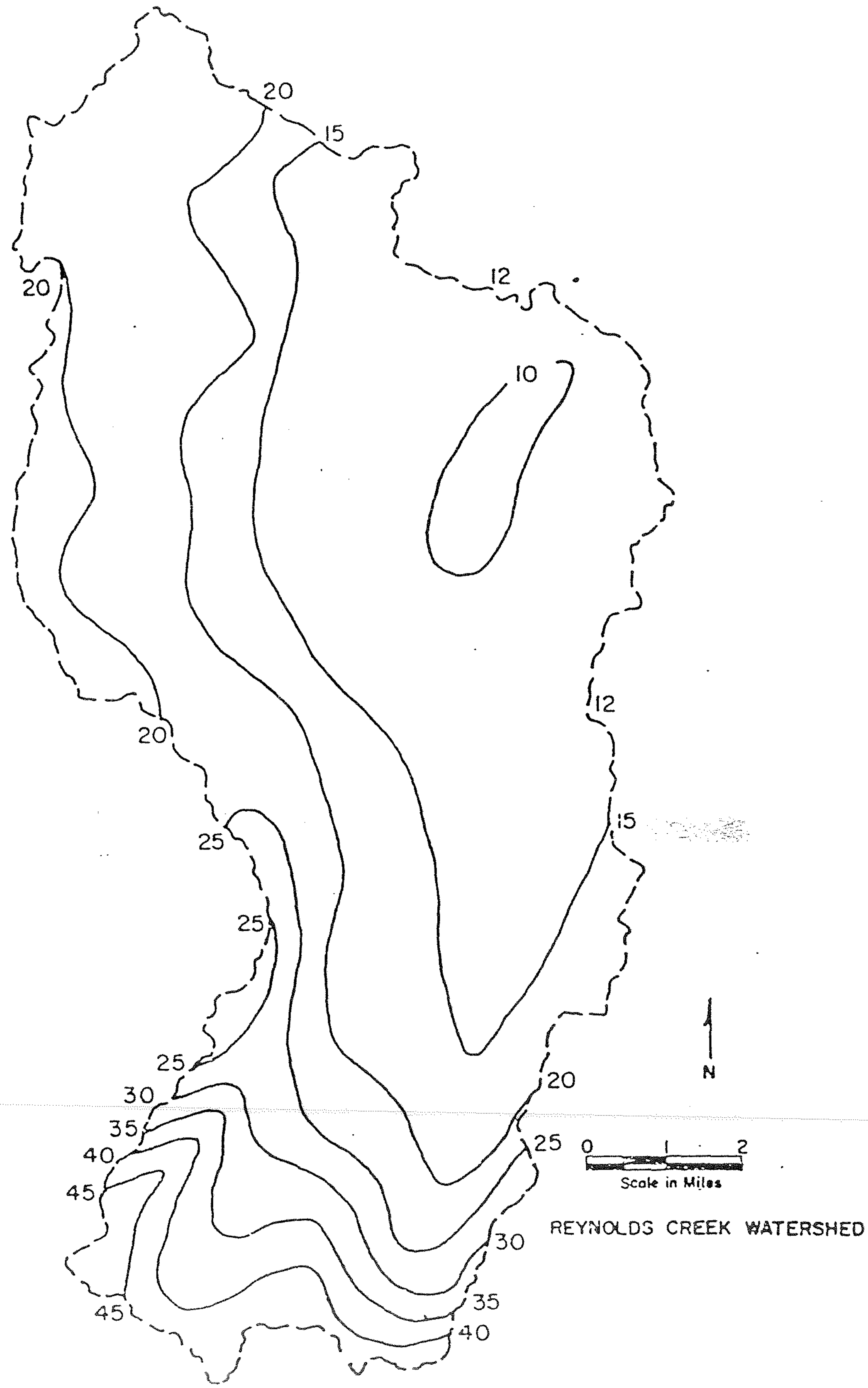

Figure 12. Annual rainfall [inches] at Reynolds Creek Watershed, Idaho.

more efficient use. The following section will discuss these natural systems, their measurements, and potential management.

Physical System

Snow, as it falls and where it initially settles, is moved and redeposited by an interaction of wind with topography, vegetation, and elevation. By the start of the melt season, the yearly areal distribution of snow is roughly the same regardless of the total snow volume that fell. This snow cover is present as large, isolated snowdrifts, located on north- and east-facing slopes which persist late into the melt season, with some remnants remaining even into late July and early August (Figures 13 and 14).

Many of these snow accumulation slopes are covered with a deep soil, and thus, in combination with the soil

Figure 13. Snowdrift sites, 6/10/74.

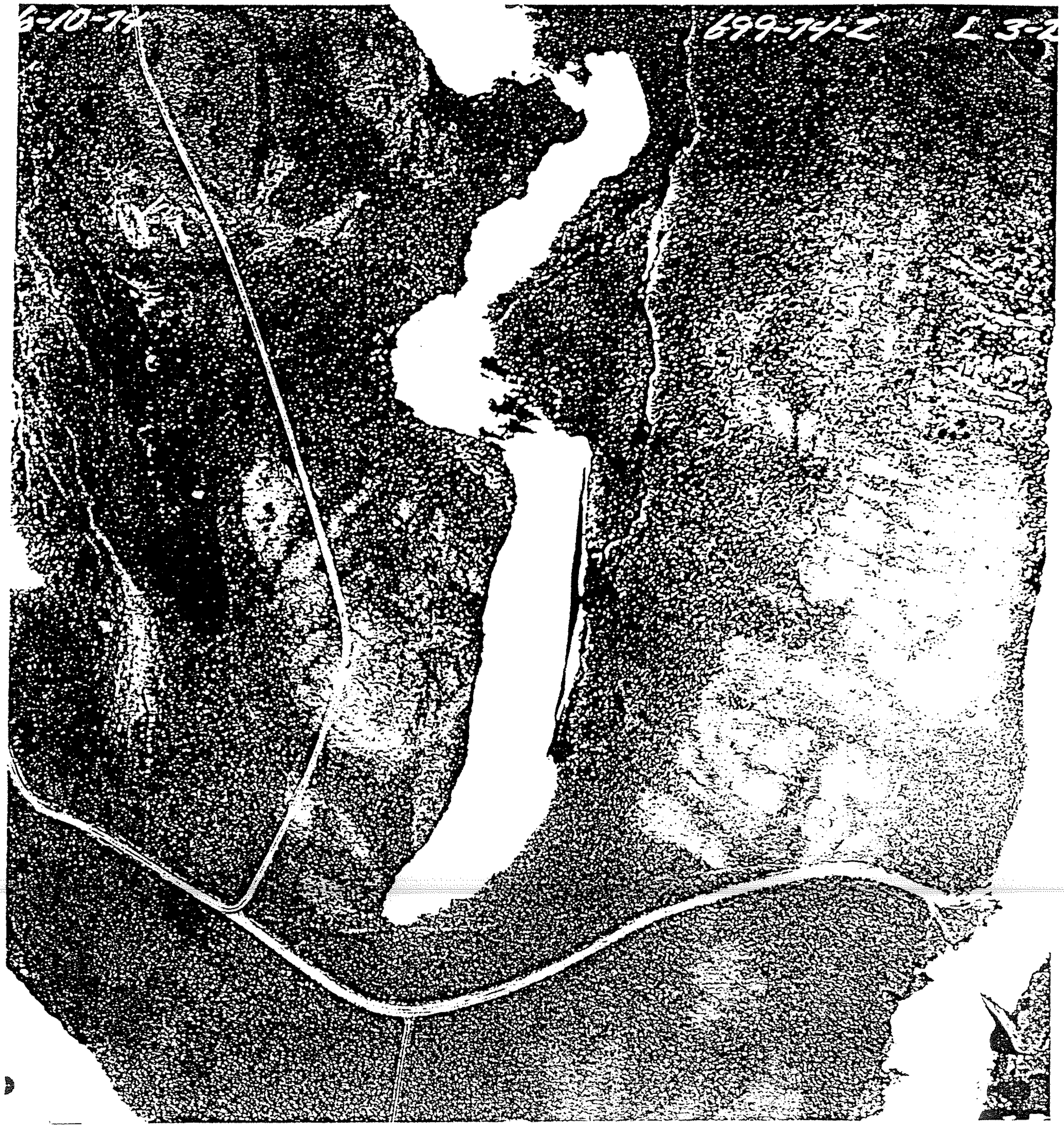

Figure 14. Segment of Figure 13 on 6/10/74. Lower fringes up to trees contains 28.400 m^3 (28 ac-ft) of water.

water supply, could be managed as high forage production sites. Approximately 30 percent (54 km^2 (21 mi^2)) of the mid-elevation part of the watershed are comprised of north- and east-facing slopes which presently are mostly aspen or sage brush covered.

Measurement by a Dual Gage System

Much has been written about errors in the catch of recording rain gages. Court (1960) stated that the largest source of error connected with raingage readings lies in the assumption that they represent the actual precipitation at the site. Several investigators have found that rain gages normally exposed with level orifices and placed approximately 1 m (3 ft) above the ground surface caught from 3 to 10 percent less rain than did a gage with the orifice at the ground. Individual storm or daily amounts could be as great as 50 percent (in windy situations) in error. Neff (1975) made similar comparisons for several locations in the Western U.S. and his results are summarized in Table 2.

The Reynolds Creek precipitation network consists of 46 dual-gage sites, (1 site/5.18 km^2 (1 site/2 mi^2)). Each is instrumented with two recording raingages, mounted on posts, with the collectors at 3.04 m (10 ft). One of the gages is shielded by a modified Alter shield, with the baffels constrained at 30 degrees from vertical to maintain a constant airflow across the collector. The second gage is unshielded.

The purpose of the dual-gage network is to develop a procedure for calculating "true" precipitation, especially in areas of snowfall, since the combination of wind and snow is the major source of error in gage catches. The method used at Reynolds Creek for computing "true" precipitation was developed by Hamon (1972), and requires the unshielded gage catch, shielded gage catch, and an empirically determined calibration coefficient as input parameters.

Measurement by Photogrammetry

Measuring snow depths and the areal distribution of snow by aerial photogrammetry on the 40 hectares (16.2 ac) subwatershed in the Reynolds Creek Experimental Watershed was shown as a practical method for obtaining detailed information on the distribution of snow in areas of complex relief (Cooper, 1965).

Briefly, the photogrammetric method for measuring snow consists of initially making a topographic map of the area and establishing horizontal and vertical controls at specified points. Then, at the desired intervals during the snow season, the control stations are remarked, the snow depth measured at each of the control stations, and the area rephotographed aerially to determine snow elevations where volume is computed. Although the snow depth and volume over an area may be accurately determined by this method, considerable error could be introduced when calculating the total water content of the snow cover because of the variation in snow density over the study basin.

Continued study of photogrammetric snow measurements on the Reynolds Mountain Study Basin have indicated that grid spacing could be increased from 7.6 m to 30 m or 100 m (25 to 100 or 325 ft) with only a 2.5- and 10-percent loss of accuracy in total volume of snow, respectively. The above change in grid spacing enabled the number of points processed to be decreased by 94 and 99 percent, respectively. Continued evaluation of snow density on the watershed has indicated that density varies according to aspect and drift locations. Some drift locations may have snow density 10 percent greater than some other sites.

Snowmelt Studies

The rate at which a snowpack will melt is dependent on the amount of heat it receives from three sources: 1) radiant heat from the sun, 2) latent heat of vaporization from the condensation of water vapor on the snow surface, and 3) heat by conduction from the ground, rainfall, or air in contact with the snow. The snowmelt process can be very complex because heat may be added to the snowpack by any or all of these sources, or the snowpack may be simultaneously gaining heat by one process and losing it by another. Actual snowpack melt is produced by a combination of all heat sources. Under different meteorological conditions, different heat sources will predominate in producing melt.

An early May 1972 study, when most of the ground surface was covered with snow, showed that net radiation supplied about 82 percent of the melt energy, with the

Table 2. Precipitation catch in ground level and 3 ft high orifice rain gages (from Neff, 1975).

Location	Type of Gage	Average Error	Range in Error
Reynolds Creek, ID	Belfort Recording	7%	0-50%
Pullman, WA	U.S. Weather Bureau Standard nonrecording	10%	0-50%
Ekalaka, MT	Fischer & Porter Recording	18%	0-75%
Sidney, MT	Fischer & Porter Recording	4%	0-50%

remaining 18 percent lost by evaporation. However, during the latter part of June, when only isolated snowdrifts were present, net radiation supplied 60 percent of the melt energy, sensible heat exchange accounted for 32 percent, and about 8 percent vapor exchange was recorded.

Exceptionally high snow ablation rates were measured on an isolated, late-lying snowdrift during May and June, 1974. Unseasonably high, average daily air temperatures, 16° C (61° F), were recorded at the 2072 m (6798 ft) elevation during late May and June and, thus, contributed to this high ablation rate. For one 9-day period during the latter part of June, 7.1 cm (2.8 in) of water were melted and 3.3 cm (1.3 in) were evaporated for a total loss of 10.4 cm/day (4.1 in/day). Energy exchange measurements showed that 54 percent of the energy available for ablation came from sensible heat transfer and 46 percent from radiant heat. This is contrasted with continuous snow cover conditions where practically all of the energy comes from radiant heat transfer. Drift profile studies showed that the top of the drift surface ablated 23 percent faster than the more abrupt face.

Snowmelt Forecasting

For a forecast model to be analogous to a simplified water balance equation, each coefficient must have hydrologic significance. To achieve this, each beta coefficient should represent a percentage of the total drainage area. The terms $\beta_i X_i \ldots \beta_n X_n$ in the forecast equation are analogous to a weighted average scheme and can then be expressed as

$$Y = a + \sum_{i=1}^{n} (\beta_i)(X_i) \qquad \ldots (4)$$

The hydrologic significance of β_i can be maintained by the additional requirements that $\beta_i \geqslant 0$ and $\sum_{i=1}^{n} \beta \leqslant 1$, provided that runoff and snow-water contents are expressed in the same units, (e.g. cm or in). The basis for this reasoning is that the snow-water content at a snow course is considered as a discrete sample of snow water content associated with an area within a drainage basin, and not simply as an index of runoff.

A general optimization program developed by TVA (Green, 1970) was used to generate beta coefficients and alpha values of Equation (4) for various forecast periods for three drainage basins located in southwest Idaho: Tollgate Drainage of Reynolds Creek Watershed, the Middle Fork of the Boise River, and the entire Boise River above Boise, Idaho (Zuzel, Robertson, and Rawls, 1975).

To verify the usefulness and accuracy of the forecast procedure, an optimized March-July forecast euqation was developed for the Tollgate Drainage of the Reynolds Creek Watershed, using 7 years of record from seven snow courses (1966 to 1972). The forecast equation was first solved for the fitting coefficient, alpha, using Thiessen weights as the beta coefficients. The solution resulted in a correlation coefficient (r) of 0.979 and a standard error of 1.93 cm (0.76 in) or 9 percent. The equation was then solved to optimize coefficients by using the Thiessen weights as initial estimates of the beta coefficients. The initial estimate of alpha was obtained from the Thiessen solution. Optimization resulted in a correlation coefficient (r) of 0.993 and a standard error of estimate of 1.12 cm (0.44 in) or 5 percent. Table 3 compares the coefficients and results of the Thiessen and optimized forecast equations. The optimization process eliminates snow courses that do not contribute to the forecast accuracy by assigning a weight of zero to them.

Additional progress is being made to improve forecast equations for short time periods of the snowmelt process; a better understanding of the relative importance of various meteorological parameters is needed. Factor analysis and regression analysis were used to determine the effectiveness of wind, air temperature, vapor pressure, and net radiation in predicting snowmelt (Zuzel and Cox, 1975).

Table 3. Tollgate drainage 54 km^2 (21 mi^2). March-July Forecast, developed from 1966-1972 data.

Snow Course	Thiessen Weighted	Optimum Weighted
144062	0.1881	0.2207
155054	0.1876	0.1936
163020	0.0424	0.1425
163098	0.0581	0.0
167007	0.1471	0.0
174026	0.0352	0.0861
176007	0.0786	0.0
Correlation Coeff.	0.978	0.993
Fitting Coeff.	-0.07	-9.53
Error Range	2% - 16%	1% - 17%
Standard Error	1.93 cm (9%)	1.2 cm (5%)
Average Runoff	20.47 cm	20.47 cm

Analyses of meteorological and snowmelt data collected at the Trinity Mountains in the Boise River basin in May 1973, showed that the standard error of daily snowmelt prediction could be decreased 13 percent by using vapor pressure, net radiation, and wind in predictive equations rather than just air temperature (Table 4).

Management of Rangeland Snow

Management of snowdrift shape by natural barriers or fences could, by reducing the flat top and increasing the slope, significantly reduce evaporation losses. Because air-borne particles have very high sublimation losses, barriers would reduce these losses by tying down the snow particles. An optimum shape of the drift would also prolong melt into the summer season, and extend soil water supplies and water yield to streams.

A trial planting of Monterey Knob Cone Pine trees was made at one drift site to test the possibility of using natural vegetation for snow management purposes. Two hundred trees were planted in two rows, .61 m (2 ft) apart at the toe of an existing snowdrift in the center of a mile-long drift area. Approximately 55 percent of the trees have survived to date.

Melt from the drifts not only provides most of the dependable water yields from watersheds, like Reynolds Creek, but also furnishes soil water to deep soil laid slopes. Nonproductive vegetation now uses this water. A real potential exists for increasing the productivity of these areas by modulating snowmelt.

Table 4. Correlation of meteorological factors with snowmelt, Trinity Mountain, Idaho.

Independent Variables	R	Standard Error (cm)[1]	Standard Error (%)
VP, NR, W	0.885	1.00	30
VP, NR, W, T	0.885	1.02	31
VP, NR	0.823	1.18	35
W, T, NR	0.827	1.20	36
T, VP, NR	0.824	1.21	36
W, T, VP	0.788	1.31	39
W, T	0.773	1.32	40
T, NR	0.773	1.32	40
T, VP	0.728	1.43	43
W, VP	0.720	1.45	43
W, NR	0.718	1.45	43
T	0.717	1.42	43
NR	0.631	1.58	47
VP	0.628	1.59	48
W	0.383	1.88	56

W = 24-hour wind run (km)
T = Average 24-hr air temperature (°C)
VP = Average 24-hr vapor pressure of air (mmbar)
NR = 24-hr net radiation (Ly)
R = Multiple Correlation Coefficient
NUMBER OF OBSERVATIONS for each analysis = 24

CONCLUDING COMMENTS

The range scientist is undoubtedly interested in the soil water available for forage production, which a hydrologist should be able to forecast from available precipitation data. Much of the historical work of hydrologists pertaining to precipitation was intended for estimating streamflow. The analytic tools developed in this work are available to range scientists and should facilitate estimating the water available for forage.

Precipitation variability is large in basin and range physiographic areas as well as in major mountain areas like those in most range areas of the Western U.S. Orographic precipitation patterns, redistribution of blowing snow and thunderstorms of limited areal extent, all contribute to the steep isohyetal gradients and limit the application of rain-gage data, except for the point in question. Unfortunately, additional gages are greatly needed at sites removed from the valley floors to facilitate quantifying the variability. The thunderstorm phenomenon which dominates the rainfall pattern in much of Arizona and New Mexico, also exists in other portions of the Western U.S. and produce the extremely large runoff events on small watersheds, which might be used for stock watering ponds. Additional rain gage network data are needed to supplement the information from the three large networks presently in operation in the Western U.S (Walnut Gulch in Arizona, Alamogordo Creek in New Mexico, and Reynolds Creek in Idaho). Also needed is a more concerted effort by hydrologists to translate hydrologic modeling outputs to soil water availability for forage production.

Future hydrologic research and application will probably lead to developing maps of the areal variation in values of transition probability, and values of α, λ in the daily depth distribution as discussed. These parameters can be described successully through the year with a Fourier series (only a few terms are needed), thus producing a space and time model of precipitation. Such information should be more comprehensively valuable than the currently used maps of various depths and probabilities for specified storm durations.

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