FACTORS AFFECTING SNOW EVAPORATION IN THE WHITE MOUNTAINS OF ARIZONA

by

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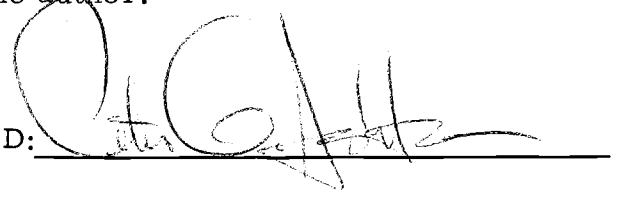
SIGNED: 

APPROVAL BY THESIS DIRECTOR

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ABSTRACT

Statistical analytical techniques were used to describe the relationships of quantifiable factors to snow evaporation and to compare evaporation rates among three selected forest sites.

The statistical analysis was divided into two phases. In the first phase a stepwise multiple regression and a correlation analytical technique was used to determine the effect of selected independent variables on snow evaporation. In the second phase a one way analysis of variance and the Duncan's Multiple Range Test were used to compare evaporation rates on three forest densities.

It was concluded that wind-vapor pressure deficit interaction, solar radiation, vapor pressure, and wind were the factors most closely correlated with snow evaporation. An equation was formulated using the above terms. With the use of U. S. Weather Bureau instruments and records, snow evaporation can be predicted within the area of study.

Snow evaporation rates were found to be higher on the cleared site than on the thinned and unthinned sites. Evaporation rates decrease with an increase in basal area.

INTRODUCTION

In arid regions water is a limiting factor in almost every sphere of human activity. Much interest is centered on obtaining more water for a variety of human uses and purposes. Snow accumulations in high elevated forested regions account for a large water yield potential. Any water loss reduces the amount of water for irrigation, power, industry, recreation and domestic use.

Vast ponderosa pine timber forests in Northern Arizona yield water to streams and ground water only when the forest consumes less water than they receive from snow and rainfall. Therefore, that portion of the annual precipitation that passes through the forest and becomes available for streamflow and ground water depends to a great extent on crown cover, age of stand, silvicultural practices, and other factors. The White Mountains of Arizona have large stagnated secondary growth stands of ponderosa pine sometimes known as "dog-hair thickets." Much of the precipitation is wasted by evaporation in crown cover interception and evapotranspiration by the trees and other vegetation.

The magnitude of water yield availability by snow melting is often determined by silvicultural practices. Regulation of thinning

and reproductive methods to atmospheric conditions can increase snow pack accumulations and reduce snow evaporation.

The purpose of this study was to provide a quantitative index of the relation between snow evaporation with air movement, vapor pressure deficits, incoming short wave radiation and temperature differences above a snow surface. It was also the intent of the study to compare the snow evaporation rates on natural, thinned and clearcut stands of stagnated ponderosa pine.

The study was conducted on plots in representative stand conditions at 6, 900 feet in Northern Arizona during the winter of 1965-66.

LITERATURE REVIEW

The main purpose of this review of literature is to organize and present relevant material regarding the study subject. For purposes of classification and organization, this review has been divided into two major sections: Previous Studies and Analytical Techniques.

Previous Studies

The energy balance of the natural snow cover has been the subject of considerable study for the past two decades and has generated too large a literature to review in a short paper. Much of the work has concerned the wet or thawing snow cover which is of immediate interest to the practical hydrologist.

Sublimation is the process of converting a solid into a vapor and condensing the vapor to a solid again. Substances exist in three different physical states that are in equilibrium. Thus, water has an equilibrium between vapor, liquid, and solid. The general characteristics of these three equilibra are illustrated on Figure 1, which temperature is located along the horizontal axis and pressure along the vertical axis. The curve, OA, is a plot of experimentally measured vapor pressures of water over a range of temperatures. The curve

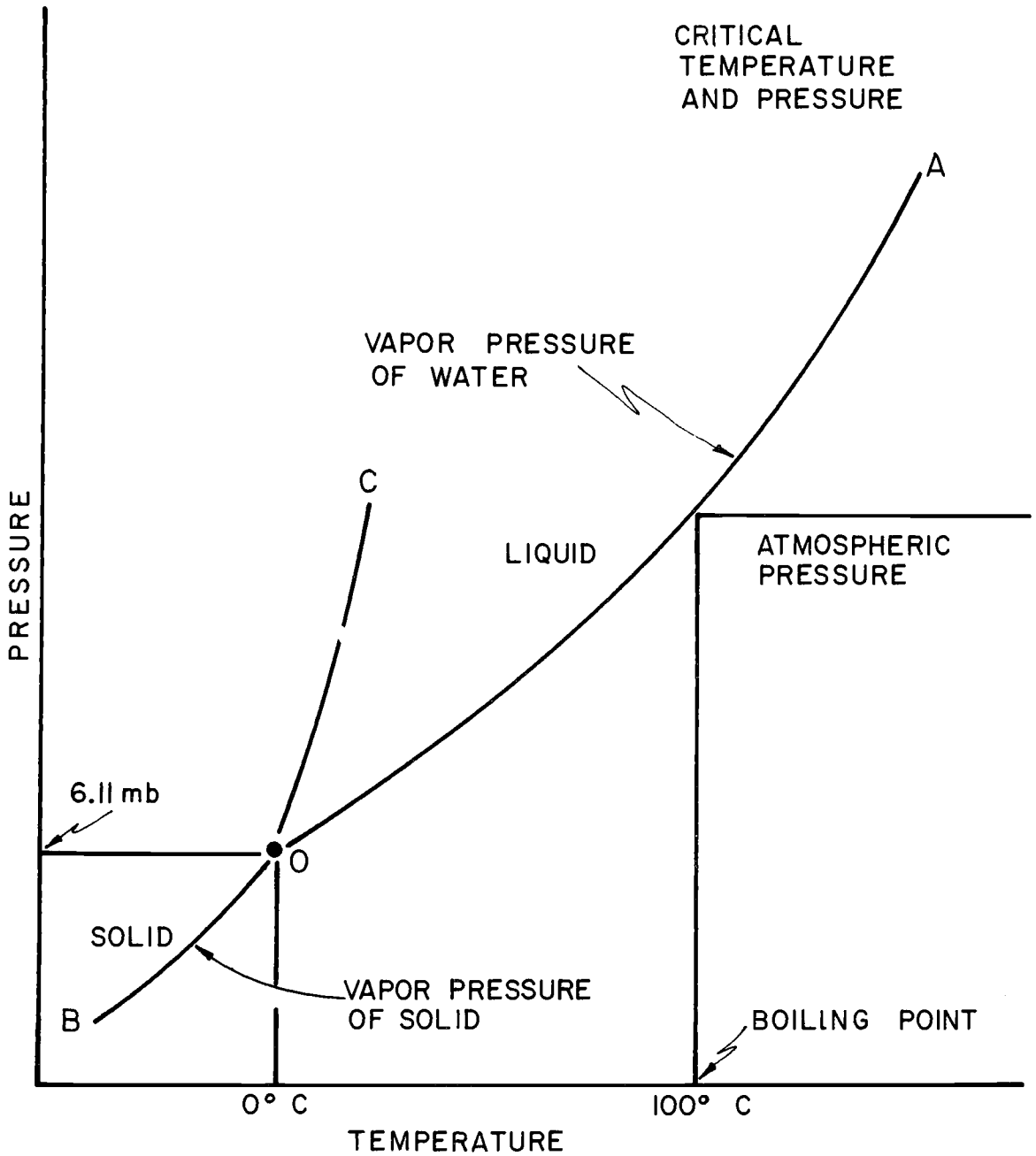

FIGURE 1.--ILLUSTRATION OF THE EQUILIBRIA OF THE THREE STATES OF MATTER FROM MACK(1958).

ends at point A which is the critical temperature that water can no longer exist in a liquid state. The curve, OB, is the vapor-pressure curve of the solid, called the sublimation curve. At point O, where the two curves intersect, is the freezing point of the liquid under its own vapor pressure which the vapor pressures of the solid and liquid are equal. At this intersection the solid, liquid, and vapor can coexist only at this temperature and pressure. If an outside pressure is applied to this existence at the intersection, and is greater than the pressure, the vapor will condensate leaving the liquid and solid state. If the applied pressure is further increased, the result is that the freezing point or melting point is raised. This effect is indicated by the line, OC, which represents the equilibrium between melting solid and freezing liquid. If the outside pressure is lowered below the vapor state pressure, the equilibrium will either shift from a solid to a liquid and evaporate or shift from a solid to a vapor and sublimate. If the temperature is raised by outside heat into the system, the equilibrium will shift in such a way as to absorb the heat and the result will be an evaporation from a liquid and an increase in the vapor state. Sublimation results from a decrease of outside vapor pressure while the solid is maintained at a constant temperature.

Light (1941) used Sverdrup's (1936) equation for heat and water flux to develop an equation for predicting theoretical rates of

melting or evaporation.

Sverdrup concluded that evaporation from snow will take place only when the vapor pressure of air is less than vapor pressure of snow over which air is moving.

Diamond (1953) computed, using Sverdrup's equation, the snowmelt and snow evaporation potentialities for a series of assumed conditions: air temperatures, relative humidity, vapor pressure, wind velocity, and snow surface temperature. His computations show that significant amounts of evaporation from snow occur at low temperatures, not far above freezing when the vapor pressure gradient is from snow to the air. At higher temperatures and relative humidities, snowmelt tends to exceed by far the loss by evaporation.

Atmospheric conditions, contributing to high evaporation rates, tend to produce turbulent transfer of considerable heat to the snow cover. When such quantities of heat are being received by the snow, some evaporation and melting occurs. Since the quantity of heat required to raise the temperature of snow is extremely low, the low thermal conductivity of snow causes the snow surface to retain most of the heat the snowpack receives from the atmosphere. This approximate quantity of heat transfer to snow by the turbulent process, and the amount of evaporation or sublimation from the snow surfaces may be computed from Sverdrup's formula:

(1) $$0 = \frac{Cp\ pK_o^2\ Ua(t-ta)}{\ln \frac{a}{zo}\quad \ln \frac{b}{zo}}$$

(2) $$F = \frac{0.623\ p\ Ko^2\ Ua(e-e_s)}{P\ \ln \frac{a}{zo}\quad \ln \frac{b}{zo}}$$

0 = Heat transfer in cal cm^{-2} sec^{-1}.

F = Evaporation in grams cm^{-2} sec^{-1} ($cm\ sec^{-1}$).

Cp = specific heat of air at constant pressure, 0.24 cal $gram^{-1}$ $C^{o}.^{-1}$

p = density of air, 1.276×10^{-3} at 1000 mb and 0° C.

Ua = wind velocity at height a, $cm\ sec^{-1}$.

t = air temperature at height b, C^{o}.

t_a = snow surface temperature, C^{o}.

s = vapor pressure of snow at t_s, mb.

Ko = Von Karman's constant, 0.42.

Zo = roughness parameter, 0.25 cm for snow, from Sverdrup (1936).

P = surface pressure, assumed 1000 mb.

a = height of temperature and humidity element, assumed 2 m to approximate height of instruments in shelter.

b = height of aneomometer, assumed 6 m.

The evaporation of F gram cm^{-2} sec^{-1} of snow requires a quantity of heat equal to 680 F cal cm^{-2} sec^{-1} to satisfy the heat of vaporization requirements. If there is insufficient heat to satisfy the heat of vaporization requirements, the snow surface temperature will be lowered with a lowering of vapor pressure of the snow. This decreases the vapor pressure difference between snow and air and reduces the amount of heat required to establish a vapor pressure equilibrium.

Diamond explains heat that is transferred to a snow surface from air above 0^{o} C has the following processes:

1. A very small amount of heat will be used to raise the temperature of the snow surface to 0^{o} C if it is below that temperature.

2. A portion of the heat will be used in the evaporation of snow, which melts prior to evaporation. Ice at 0^{o} C must pass through the liquid state before evaporation takes place. In the process the snow may appear to pass directly from the solid state to the vapor state without the appearance of a visible liquid state. The snow crystals melt slightly on the surface and acquire a microscopic liquid film from which evaporation takes place. Whether the ice actually melts or only develops a liquid surface film prior to evaporation, the heat required in the transfer of a given quantity of snow to the vapor state is the same.

3. The remainder of the heat is used to melt snow, which may be either stored in the liquid state or refrozen at lower depths in the pack or discharged as runoff. This process is melting.

If heat is being transferred from air to snow when the temperatures of the former is below 0^{o} C, the heat is used in the following processes:

1. A portion of the heat is used to raise the temperature of the snow to the air temperature.

2. The remainder of the heat is used in the evaporation of snow.

If insufficient heat is transferred from air at temperatures below 0^{o} C to snow to satisfy the heat of vaporization requirements, the snow surface temperature will remain lower than the air temperature.

The amount of heat which a snow surface may receive by turbulent transfer from air of different temperatures as computed by equation 1 is presented in Table 1. The heat received by a snow surface from solar radiation during the spring months varies from 100-200 cal cm^{-2} day^{-1}, which is much less than the amount of heat transferred from air to snow under conditions assumed here (Miller 1955). It is assumed that the air having the temperatures listed in Table 1 has a relative humidity of 20 percent to simulate dry conditions. The amount of

snow which can be evaporated into air of 20 percent relative humidity at different temperatures is computed from equation 2 and presented in Table 2. The decrease of evaporation with rising temperatures shown in Table 2 is in agreement with the observations by Kaitera (1939), which showed a decrease in the amount of evaporation from snow at higher air temperatures.

Evaporation from a snow surface occurs only when the vapor pressure gradient is from snow to air. Since the saturation vapor pressure of air increases with increasing temperature, warm air has a lower relative humidity than cold air in order to have the same vapor pressure as the cold air. The vapor pressure over snow at 0^{o} C is 6.11 millibars. Air must have a lower vapor pressure than 6.11 millibars for evaporation to occur. The maximum relative humidity which air can have at different temperatures to have a vapor pressure less than 6.11 millibars is listed in Table 3.

The amount of snow evaporated is proportional to the vapor pressure gradient. Evaporation from snow to both cold and warm air is equal when the vapor pressure of each air is the same, and this occurs when the relative humidity of warm air is less than that of cold air. Table 4 lists the humidities required for different air temperatures to have the same amount of evaporation.

As the air temperature rises, more heat is transferred to the snow; but if the relative humidity remains the same, the evaporation potential of air decreases and more heat becomes available for melting. For the evaporation potential of air to remain the same with increasing air temperatures, the relative humidity must become extremely low. The decrease of evaporation potential with increasing air temperatures for a relative humidity of 20 percent and snow temperature of 0^{o} C is illustrated in Figure 2.

Miller (1955) states that when the mean air temperature is less than the snow surface temperature there is no heat transfer from air to snow, and the only major heat source for evaporation and melting is solar radiation. The amount of heat transferred to the snow surface from below is very small because of the low thermal conductivity of both snow and soil. During the spring months, the snow cover has a net radiation balance of about 100-200 cal cm^{-2} day^{-1}. While the snow surface does not retain all this heat, some passes to layers below the surface. This amount of heat will cause the evaporation of only 0.2 to 0.4 grams of snow. Therefore, appreciable evaporation will occur when there is turbulent transfer of the heat from air to snow.

The solar radiation intercepted by the earth will either be absorbed and used in any number of energy driving processes or returned

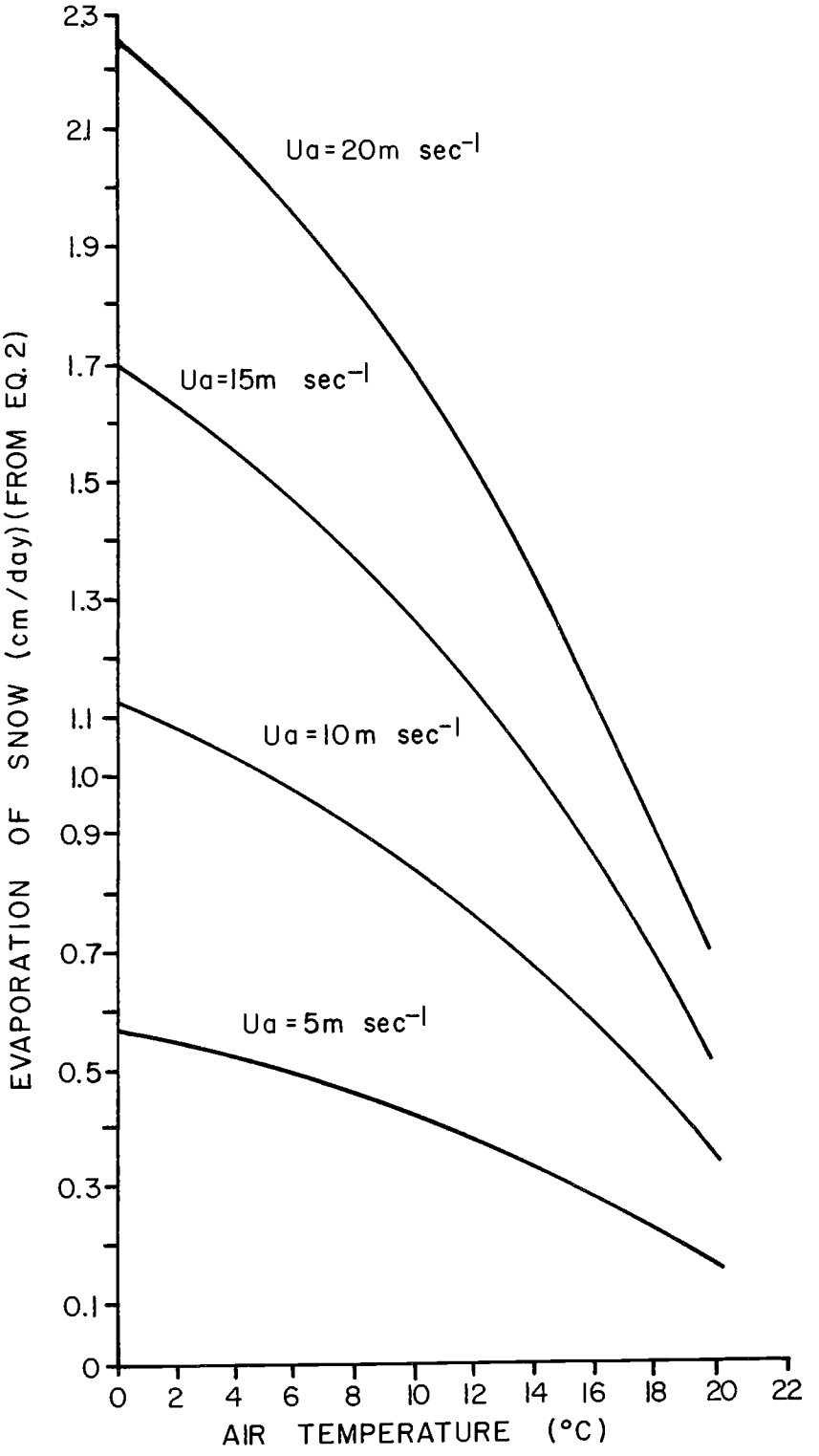

FIGURE 2.--DECREASE OF EVAPORATION POTENTIAL WITH INCREASING SNOW-AIR TEMPERATURE DIFFERENCE AT RELATIVE HUMIDITY 20% FROM DIAMOND (1953).

to space by scattering or reflection. Almost 99 percent of the sun's radiation is contained in the wavelength range 0.15 to 4.0 μ (short wave radiation). Of this, 9 percent is in the ultraviolet, 45 percent is in the visible, and 46 percent is in the invisible. Incoming solar radiation is accounted for by the following: cloud absorption 2 percent, atmosphere absorption 16 percent, cloud reflection 24 percent, atmospheric reflection 7 percent, backscatter by surface 4 percent, direct absorption by surface 22 percent, and surface absorption by diffuse sky and cloud radiation 24 percent. Almost half of the solar radiation intercepted by the earth (47 percent) eventually reaches the earth (Sellers 1965).

Geiger (1957) states that all of the solar radiation incident on the earth's surface is not absorbed. A certain amount is reflected, depending upon color and composition of the surface. The reflected radiation is lost to space.

Fresh snow has a high albedo (percentage of the incident short wave radiation reflected). The short wave albedo, over snow free surfaces, is a function of the solar altitude with the maximum occurring near sunrise and sunset. At midday the solar radiation penetrates into the snow pack. At low sun angles, during morning and evening, the penetration is shallow and reflection occurs. Cloud covers

reduce this proportionately thereby increasing the diffused radiation. There is very little dependence of the ultraviolet albedo on the solar altitude, because most of this radiation comes from the sky as scattered solar radiation, rather than from the sun.

Research by the Corps of Engineers (1956) indicated that since all the thermal energy involved in melting and evaporating a snowpack has its ultimate source in the solar radiation reaching the earth, the other processes serve merely as intermediate means of heat transfer. Yet, it is the measurement of some of the products of solar energy, such as air temperature and vapor pressure, that are most commonly used as indexes of snowmelt and snow evaporation. The earth's atmosphere is warmed but slightly by solar radiation passing through it. Over snow fields this absorption is increased since a portion of the reflected beam is also absorbed. The degree of heating of the air by solar radiation is relatively small. Of greater consequence is the heating of air which results from air passing over lands and objects which are heated by solar radiation. These give their heat to the air by the processes of convection. Similarly, water surfaces and ground surfaces containing water may be heated by solar radiation and by the transfer of heat from the air by the process of convection to give off energy in the process of evaporation. Therefore, water vapor is

added to the air and it may condense upon the snow surface or regulate snow evaporation by vapor pressure gradient.

The rates of melting and evaporation of snow under atmospheric and vegetative influences are discussed.

Church (1911), in reporting on progress of the Mount Rose Observatory, Nevada, 1906-1912, discussed the importance of evaporation loss to the conservation of snow. He reported that timber screens, by checking the wind, would tend to reduce the evaporation of snow, which under the influence of wind movement of 33 miles per hour, and despite the fact that the snow was frozen, reached in a single night a total of 0.1 inch of moisture content.

Baker (1917) performed a series of experiments on evaporation from snow surfaces in Utah in the spring and winter of 1915-1916 over a period of about 180 days. The snow-water equivalent for the same locality for the winter of 1915-1916 was 21.91 inches, from which, according to Baker's measurements, $\pm$ 3 inches of evaporation occurred as measured by two different methods. Thus, Baker's measurements of evaporation indicated that about 14 percent of the total snowfall was evaporated into the air. It is noted that Baker made no measurements during periods of storms, high winds, or during the spring thaw period. The evaporation losses from the snow, as reported by Baker, were

based upon measurements during calm, clear days during the winter only.

Croft (1944) described a study of evaporation which was conducted for a 10 day record, 6 of which were complete, from snow under 3 site conditions. This included full insolation and free air movement, shade with free air movement, and full shade with no air movement on a drainage basin in the Wasatch Plateau in central Utah. Cones of snow 6 inches in diameter and 6 inches deep were used. Records of snow surface temperature, wet and dry bulb temperatures at 1/2 foot, 4 1/2 feet, and 12 feet above the snow surface and wind speeds at 1/2 foot and 4 1/2 feet above the snow surface were observed. Vapor pressures were calculated. Neither shade nor mean daily air temperatures were found to be related to evaporation, but air movement were found to be related to evaporation, but air movement and evaporation were very closely related. Differences in vapor pressure of snow and air were found to be very significant in relation to snow evaporation losses. Converting Croft's data to monthly losses, the results indicate that the average for his study would amount to about 1.2 inches per month. The loss for the highest days of evaporation would be the equivalent of about 1.6 inches per month. Where air movement was prevented, the equivalent loss would amount to about 0.7 inches per month.

Kirschmer and Rimkus (1946) constructed weighing lysimeters, with an exposed area of 6 1/4 square meters, for the direct measurement of snow disappearance by weight. The snowmelt was caught in receptacles in a pit under the lysimeter pan. They concluded that the gradual reduction of the snow cover during the winter is not due to evaporation but to continual melting of the bottom layers. Because even at air temperatures of 30 degrees below zero, the earth under the snow cover shows a temperature slightly above zero degrees C. They concluded also that the volume of mountain stream floods is not proportional to the total amount of winter snowfall. The conclusions of the two investigators have not been substantiated elsewhere, especially their conclusion that the melting of snow occurs from the bottom of the pack rather than from the top.

The melting of snow from the top has been proven by numerous investigators through the use of dyes, such as fuchsin, which remains a black powder in the presence of dry ice crystals but change to an intense purple solution in the presence of water. Its use permits the tracing of melt waters both vertically and horizontally through the snow packs and the technique has been used extensively to follow the progress of the ripening of the snow pack and the release of water from the melted snow.

Kittredge (1953) presents a very comprehensive analysis of evaporation in the ponderosa-sugar pine-fir zone of the Central Sierra Nevada in California. Kittredge is quoted as follows:

> The outstanding characteristic of the evaporation studies in the present work on the west slope of the Sierra Nevada is the small magnitude of the measured losses. The explanation is probably to be found in the geographic and physiographic location on the west facing slopes of the mountains which are exposed to the moisture bearing winds of the Pacific Ocean. Thus, the humidity of the atmosphere is high, and the vapor pressure difference between a snow surface and the air above tends to be low or negative; consequently, the evaporation is also low. The question has been raised as to whether it is possible to obtain a reliable measure of evaporation from snow by the use of any container in view of the fact that solar radiation penetrates 5 inches or more through the snow and thus tends to heat the walls of the container. This heat would be transmitted to the snow in the containers and would obviously accelerate melting.

West (1962) estimated the total loss of water due to evaporation on the Castle Creek Watershed, located in the Central Sierra close to the Donner Summit. Using data from a 3 year average seasonal evaporation loss, it was calculated that 428 acre feet of water, or 2.0 inches depth, would be lost to snow evaporation. The snow evaporation losses accounted for 2.7 percent of the total precipitation.

A special sequence of strip cutting called a wall-and-step cutting has been designated as an ideal forest for maximum snow accumulation and delayed melt (Anderson 1956). If maximum accumulation of the

snow pack is the management objective and strip cutting is silviculturally desirable and economic, cut the forest in strips oriented across slopes perpendicular to the direction of maximum solar radiation. Generally the strips would be east-west on north and south slopes, northeast-southwest on east slopes, and northwest-southeast on west slopes. Successive strip cutting (Figure 3) would proceed generally to the southward, that is toward the maximum solar radiation. Once through a cutting rotation, a wall-and-step forest would be established with the wall to the south providing shade and the steps to the north would give the least back radiation (Figure 4). The width of the cut strip would depend on the slope of the terrain, on the height of the trees expected at the next rotation, and on objectives. For maximum delayed snowmelt, strips one-half times the tree height on the steep south and west slopes, one to two times the tree height on level areas, and one to four times the tree height on steep north slopes.

When the wall-and-step forest (Figure 3) has been established, calculations indicate an additional 3 inches of snow water at the time of maximum accumulation as compared with the results from simple strip cutting (Anderson 1960).

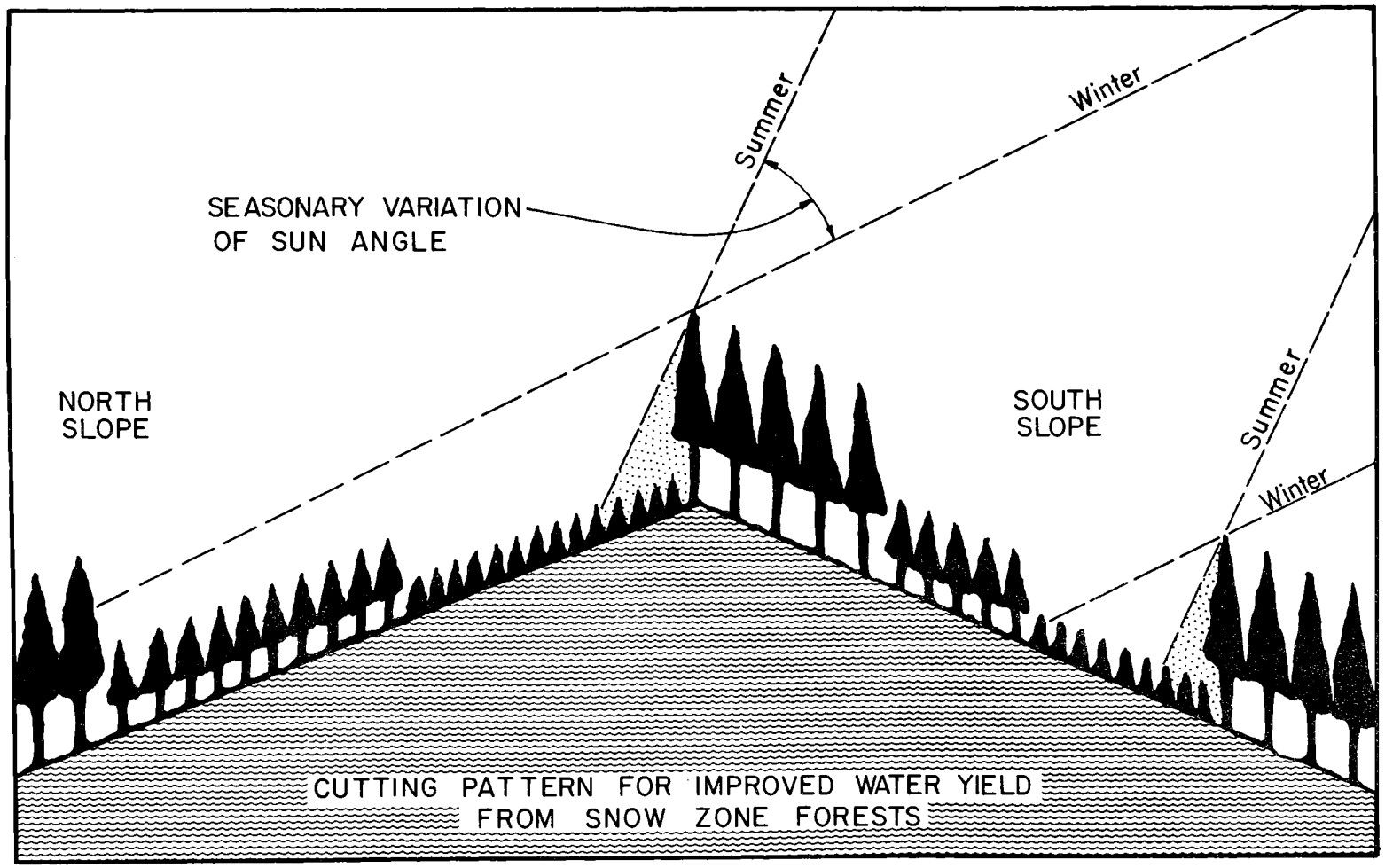

FIGURE 3.--WALL-AND-STEP FOREST FOR MAXIMUM SNOW ACCUMULATIONS AND DELAYING MELT FROM ANDERSON (1963).

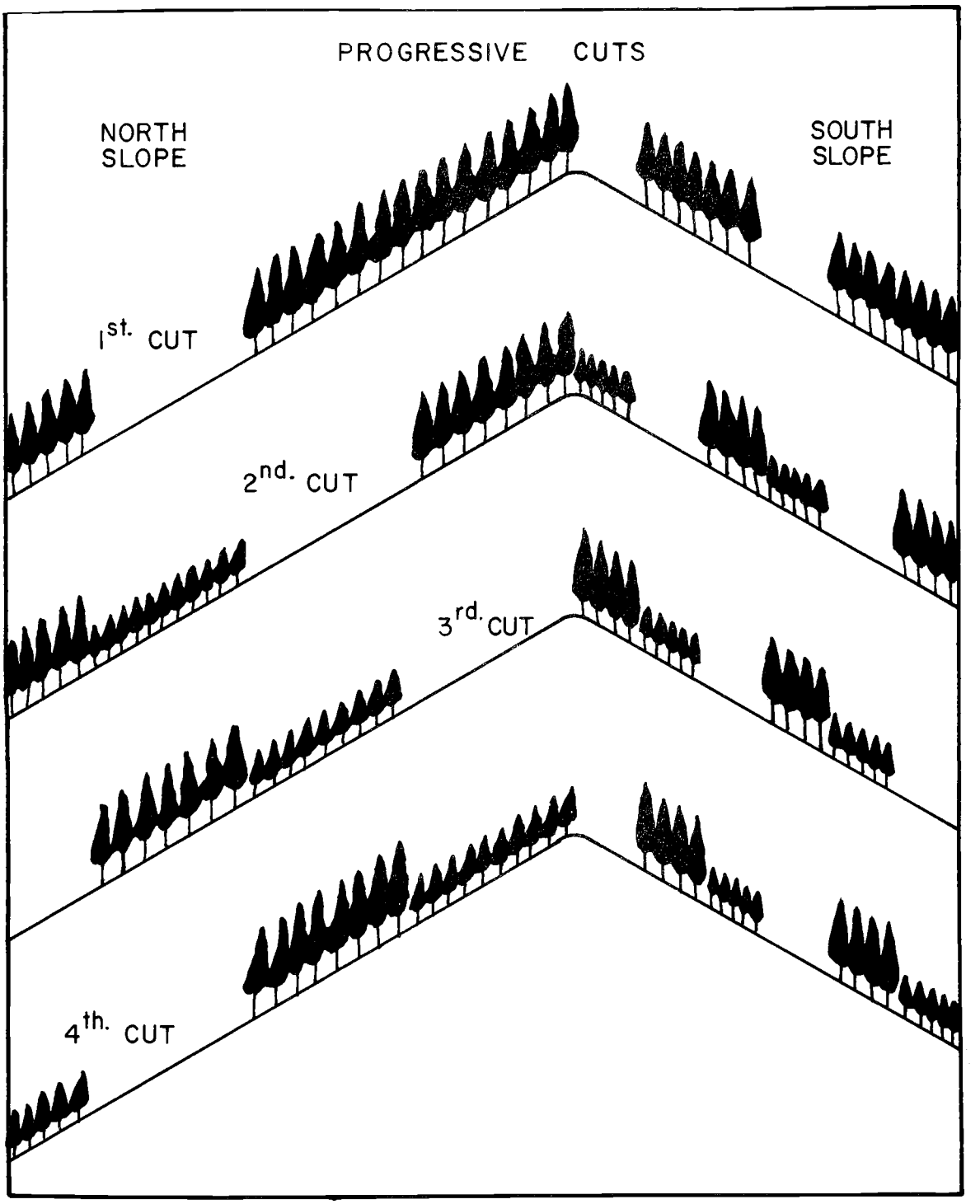

FIGURE 4.--PROGRESSIVE CUTS ESTABLISH WALL-AND-STEP FOREST IN AREA ROTATION FROM ANDERSON(1963).

Analytical Techniques

Many different types of analytical techniques are applicable to a study of this kind. Much literature has been published regarding the various types of analytical techniques which might be used to reduce the available data accumulated. This portion of the review of literature will present only relevant material directly concerned with the specific types of analyses used in the study.

One of the first to find the solution of the estimated correlation coefficient and determination between independent variables was Student (Fisher 1950). The usefulness of the correlation coefficient is made apparent in the early stages of inquiry when exploration is being conducted as to associations and relations between factors thought independent yet appearing related.

To perform multiple regression and correlation analyses, a series of simultaneous normal equations must be solved. The Doolittle technique is one method of solving normal equations and obtaining the various constants necessary in regression and correlation analyses. This method of solving equations, developed by M. H. Doolittle, involves slight modifications of the least squares technique of solving equations discovered by Gauss and revised by Fisher (Ezekiel 1959).

The Doolittle method of equation solution is intricate and involves many computations. Therefore, other possible methods for normal equation solutions have been offered by various statisticians and mathematicians. The offerings of methods for solutions which are usually only shortcuts on the Doolittle method, have been by Dwyer, Horst, and Waugh. The square-root method for solving simultaneous equations was developed by Dwyer. The "Abbreviated Doolittle" technique for solving simultaneous equations has been credited to Horst and Waugh (Fisher 1950).

Regression and correlation analyses may be applied to many types of hydrologic problems involving different variables limited through interacting relationships. The relation of snow evaporation to wind velocity, vapor pressure deficits, short wave radiation and temperature differences profiles is one area of the hydrologic cycle in which regression and correlation analyses play an important role. Regression and correlation analyses will help to determine the best significant combination of variables that will produce an equation describing snow evaporation.

DESCRIPTION OF STUDY AREA

The study was conducted on the Lakeside Ranger District, Sitgreaves National Forest, Lakeside, Arizona. The general terrain of the Sitgreaves National Forest slopes downward from the White Mountains towards the southwest into the westward flowing Salt River.

Highest elevations near Lakeside are reached in the White Mountains, which rise to 10, 115 feet on Greenspeak and 11, 590 feet on Baldy Peak. To the west-northwest of the town lies the broad forested Mogollon Plateau which stretches northwestward for more than 100 miles, rising to an average elevation of more than 7, 500 feet along its southern rim. The saddle connecting this plateau with the White Mountains dips to a minimum elevation of about 6, 500 feet ten miles northwest of Lakeside. Further to the north of the town the terrain descends gradually into the semi-arid valley of the Little Colorado River. Lakeside is located in one of the most extensive ponderosa pine forests in the world. These forests are interspread with open mountain meadows. With decreasing elevation to the south and north of Lakeside, the prevailing vegetation of ponderosa pine (Pinus ponderosa) changes first to juniper (Juniperus osteosperma) and Pinyon (Pinus edulis) and then to annual and perennial grass along the Little Colorado

River and to chaparral (Ceanothus olegantus) and oak woodland in the Salt and Black River Canyons.

Lakeside and vicinity is one of the few areas in Arizona that receive enough precipitation during a normal year to support a luxuriant growth of vegetation. The average annual precipitation is 24.70 inches of which 26 percent occurs as snow fall and amounts to 94.30 inches. Drought conditions usually exist only in May and June. About 80 percent of the annual precipitation occurs in three summer months and four winter months. Most of the warm season precipitation falls from thunderstorms that form over the tall peaks of the White Mountains, practically every afternoon, from early July until early September. Lakeside has a cool summer climate.

Winter precipitation, December through March, generally lasts longer and is much more widespread than that of summer. It is associated with middle latitude storms that move eastward from the Pacific Ocean. These storms are capable of producing unusually long periods of heavy precipitation, particularly when they move further south than normal or when they form or intensify off the coast of California. Winters at Lakeside are characterized by cool days and cold nights. Afternoon temperatures normally lie in the middle forties, however during warmer periods they may rise into the sixties, while during cold

spells they may not get even into the thirties. Early morning temperatures are practically always below freezing in mid-winter, generally lying in the high teens. Readings below zero degrees have occurred in the five coldest months and are commonly recorded on at least one or two days each winter. About 80 percent of the cool season precipitation at the town falls as snow. During the colder and wetter winters, snow may accumulate to a depth of several feet, particularly in depressions and on the northern slopes. Snow melting, during the winter precipitation, is the principle source of streamflow.

Vegetation consists mainly of ponderosa pine (Pinus ponderosa) with an understory of annual and perennial grasses.

Study Sites

The study sites were three, one-quarter acre, level areas located in ponderosa pine stands at an elevation of 6,890 feet.

Site 1 was a clearing about 500 feet in diameter located at the ranger station (Figure 5). The site received direct sunlight from about 0800 to 1600 hours. It was then shaded by a shadow cast by trees to the west. Timber surrounding the clearing was primarily stagnated growth ponderosa pine which graded into mature trees. In the stagnated growth timber, average height was 6.09 meters. In the mature stand heights ranged from 23-30 meters.

Figure 5. Cleared site, Site 1

The two remaining sites were located one half mile north from the ranger station. Site 2 was situated in a stagnated stand of ponderosa pine that had been thinned to a basal area of 65 (Figure 6). Average height of the stand was 6.09 meters. The forest floor of the site received sunlight from about 0900 to 1600 hours. Site 3 was an unthinned stagnated ponderosa pine growth stand with a basal area of 190 which permitted partial sunlight on the forest floor from about 1100 to 1400 hours (Figure 7).

Figure 6. Thinned site, basal area 65, Site 2

Figure 7. Natural site, basal area 190, Site 3

METHODS AND PROCEDURES

Snow evaporation is dependent upon meteorlogical factors as they may be modified by vegetation. Four independent meteorological variables were chosen for correlation with evaporation primarily on the basis of accessible standard U. S. Weather Bureau recordings and instruments. These were: (1) solar radiation, (2) temperature gradients, (3) wind velocity, and (4) vapor pressure deficit. These variables were measured only on the cleared forest site.

The time intervals between observation periods varied from 30 minutes to 9 1/2 hours. Snow evaporation was measured simultaneously with the independent variables on the cleared site and twice daily on the thinned and unthinned sites.

Analysis of the correlation data for this study was concentrated on the period of 30 January 1965 through 4 February 1966 (Table 5). During the entire snowfall period, November 1965 through March 1966, evaporation and condensation measurements were obtained twice daily and on each forest site by the Lakeside District Ranger during each snow ablation period.

Snow Evaporation Measurements

Evaporation from snow between observation periods was measured on the three sites by weighing snow in white polyethylene food containers (Figure 8). The containers were 10.9 by 10.9 by 9.5 cm with almost vertical sides. Three samples were taken on each site for every evaporation interval.

Undisturbed snow samples were obtained by forcing a container laterally into the snow pack. The containers were then pushed upward until the tops were level with the original snow surface.

The containers with samples were withdrawn, dried by a towel, and weighed on a triple beam balance to the nearest 0.05 grams. The procedure was repeated for each evaporation interval.

Within each site, three impressions were made by pressing an empty container into the snow pack to the depth where the containers were level with the snow surface. The containers with samples were placed into the impressions. The purpose was to simulate an unbroken snow surface.

The net gain or loss in snow weight was recorded in grams. Samples were changed and reweighed when the level of snow in the containers dropped below 1.3 cm. Prior to precipitation, the containers were weighed and the procedure was started after the precipitation stopped.

Figure 8. Snow evaporation measuring procedure

Snow evaporation was recorded in grams per unit of time and converted to centimeters per second. Three samples on each site were averaged for every evaporation time interval as follows:

$$(3)\quad \frac{\text{difference in average weight of containers between snow weighings (grams)}}{\text{cross section of container } (cm^2) \times \text{density of water } (gm/cm^3)} \times \frac{1}{\text{time lapse between intervals (sec)}}$$

$$= \text{evaporation or condensation (cm/sec)}$$

Meteorological Measurements

Solar Radiation

Short wave radiation incident on a horizontal surface was measured continuously during the study with a standard U. S. Weather Bureau Epply pyrheliometer mounted about 121 cm above the snow surface (Figure 5). Output from the pyrheliometer was reduced with a voltage divider and recorded with one point of a 12 point millivolt recorder (Figure 9). Reduction in output was necessary in order to record temperature profiles simultaneously on the 11 remaining points available on the recorder.

Calibration of the reduced pyrheliometer output involved the attachment of a separate potentiometer to read the actual output directly. The relationship between the actual output and the recorder

Figure 9. Epply pyrheliometer

readings were determined by linear regression analysis and plotted (Figure 10). The regression coefficient was -6.49 with the intercept not varying significantly from 0. The correlation coefficient was -0.999.

Due to the large number of recorder outputs of the pyrheliometer, a systematic selection of every third point was used during each evaporation interval. These points were averaged over their corresponding evaporation intervals and were converted to millivolts by the use of a graph (Figure 10).

The Epply pyrheliometer has been standardized with reference to the International Pyrheliometric Scale (1956) and has been found to develop an emf of 7.00 millivolts/cal cm^{-2} min^{-1}. The recorded millivolts of solar radiation were converted to langleys per minute as follows:

$$(4)\ \frac{7.00 \text{ millivolts/cal } cm^{-2}m^{-1}}{\text{chart readings}} \times 1 \text{ cal } cm^{-2}m^{-1} = \text{langleys per minute}$$

Temperature Gradient

Local air temperatures were measured at 0.0, 7.62, 15.24, 30.48, 60.96, 91.44, 121.92, 152.40, 213.36, and 274.32 cm above the snow surface by a movable copper constantan thermalcouple mast (Figure 11). Outputs from the thermalcouples were recorded on the

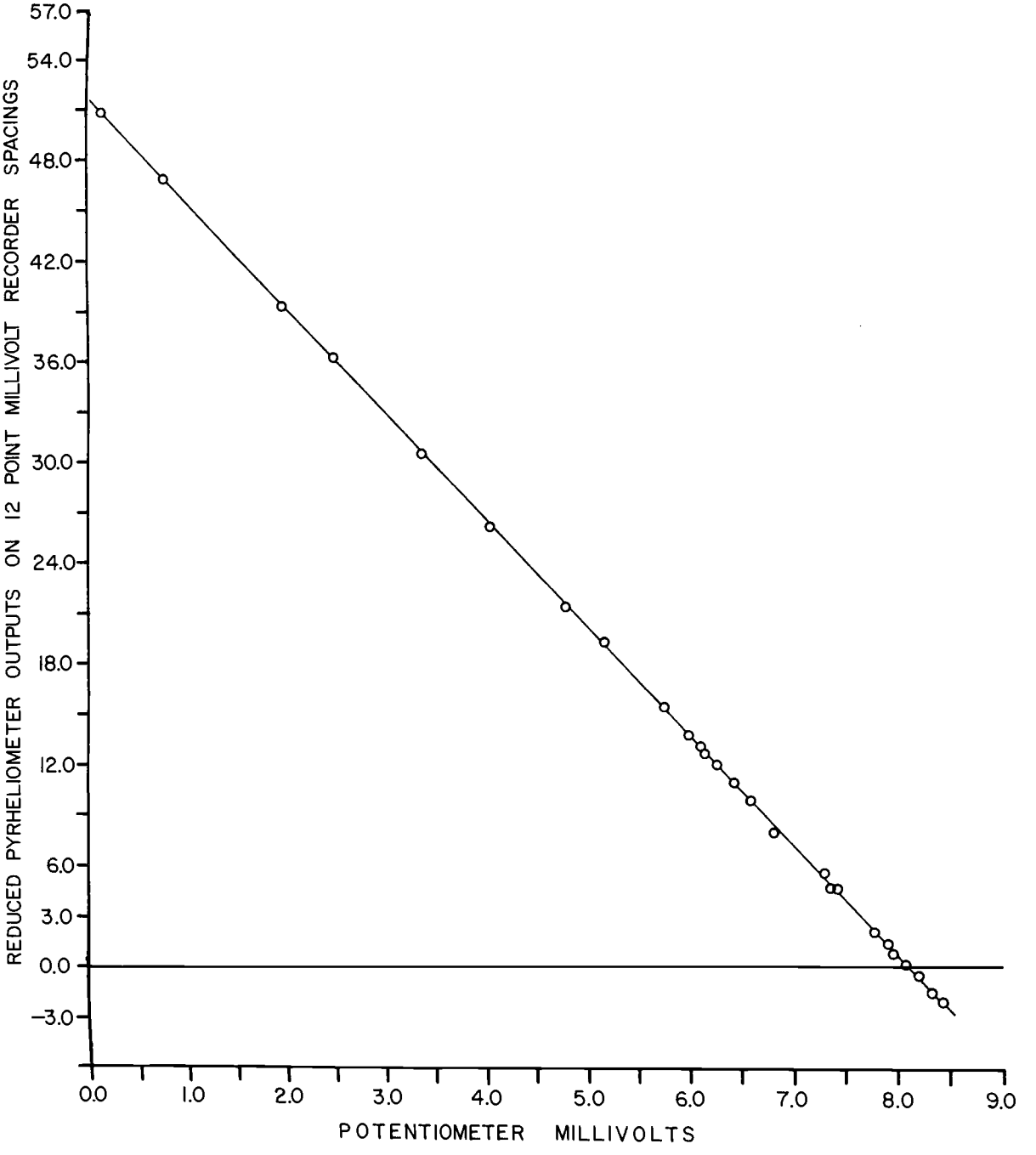

FIGURE 10.--CALIBRATION OF THE REDUCED EPPM PYRHELIOMETER TO A SEPARATE POTENTIOMETER

Figure 11. Movable copper constantan thermal-couple mast

12 point millivolt recorder at five second intervals per thermalcouple.

A cold junction was not used, but temperature differences between the thermalcouples could be readily obtained. Nine temperature differences between the snow surface and each thermalcouple reading were tabulated from the recorder chart with a plexiglas grid duplicating one recorder chart division. The differences were recorded either minus or positive between spacings from the snow surface. Minus signified warmer temperatures than the snow surface and positive were colder temperatures. Because of the large number of points made by the recorder for each temperature difference level, a systematic selection of every fourth point was used during each evaporation interval. These points were averaged for their corresponding evaporation interval.

The 12 point millivolt recorder (Figure 12) was calibrated with a selected chart span by applying a known emf to the recorder and noting the subsequent chart readings. The difference between two known emf inputs was 0.2 millivolts. The respective differences on the recorder chart was 10 subdivisions of the one large division (e.g., 50.0 - 51.0).

Figure 12. Variable potentiometer and 12 point millivolt recorder

Using copper constantan millivolt equivalent temperature tables, one degree fahrenheit temperature difference between the range of -10 to 70°F gave an average millivolt difference of 0.021. The spacings were converted to degrees centigrade as follows:

(5) $\frac{\text{chart spacing difference (0.02 MV)}}{\text{copper constant difference (0.021 MV)}} \times 1^{\circ}F \times 5/9 = {}^{\circ}C$ difference between thermal-couples

Wind Velocity

Wind velocity was measured with a standard U. S. Weather Bureau totalizing anemometer mounted approximately 91 cm above the snow surface (Figure 13). Wind velocity was recorded as miles of wind in events per unit of time by an Angus event recorder.

For each evaporation interval, the wind speed as miles per hour, was averaged and converted to meters per second as follows:

(6) $\frac{\text{total wind past measuring point (miles)}}{\text{time lapse between measurements (Hours)}} \times \frac{\text{feet}}{\text{miles}} \times \frac{\text{hours}}{\text{seconds}} \times \frac{\text{meter}}{\text{feet}}$

= wind velocity meters/second

Vapor Pressure Deficit

Vapor pressure deficit was measured as the difference in pressure at the snow surface and the air pressure as a constant height

Figure 13. Totalizing anemometer

above the snow surface. During each evaporation interval, wet and dry bulb temperatures were obtained every 60 minutes with a hand aspirator psychrometer at 91 cm above the snow surface. Hygrothermographs were used to check and follow the trend between the psychrometer readings (Figure 14).

Vapor pressure deficit V was determined by subtracting the saturated vapor pressure E_s over ice in millibars from the vapor pressure E at 91 cm above the snow surface, e.g., $V = E_s - E$. The vapor pressure at 91 cm above the snow surface was computed by multiplying the average relative humidity RH for each evaporation interval to the saturated vapor pressure at its occurring temperature in degrees centigrade, e.g., $E_{mb} = (RH)(E_s)$.

At temperatures above 0^o centigrade, the saturation vapor pressure is essentially a constant 6.107 millibars. For temperatures below 0^o centigrade, the saturation vapor pressure decreases with a decrease in temperature and was obtained from Smithsonian Meteorological Tables (List 1958). Units of millibars were converted to millimeters (mm) of Hg by multiplying millibars by 0.7500616.

Figure 14. Instrument shelter with hygrothermograph

ANALYSIS

The analysis was divided into two phases. The first phase consisted of determining the relationships of independent meteorological variables to snow evaporation and to develop a prediction equation that that could use records commonly available from the U. S. Weather Bureau. In the second phase a comparison was made of snow evaporation rates among the three selected sites.

Phase I

A stepwise multiple regression and correlation analyses were used to accomplish the objectives of Phase I. All computations were made on the 7330 IBM computer located at the Numerical Analysis Laboratory, University of Arizona. The results of the computer analyses were checked by comparison with prior calculations on a desk calculator.

The stepwise procedure is helpful for screening a large number of independent variables in order to select those that are likely to give a good fit to the sample data. It should be noted, however, that the procedure is strictly exploratory. The probabilities associated with tests of hypotheses that are selected by examination of the data are not what they seem to be. Significance tests made in this way do not have

the same meaning that they have when applied to a single pre-selected hypothesis. The stepwise procedure will frequently lead to the linear combination of the independent variables that will result in the smallest residual mean square, but it does not always do so. This can only be done by fitting all possible combinations and then comparing their residuals. Here again, tests of significance may be informative, but the exact probabilities are unknown.

The stepwise regression analysis ranks each independent variable according to its highest correlation with the dependent variable. As the variables are introduced, the multiple coefficient of determination R is increased while the degrees of freedom are decreased and the significance of the overall analysis as indicated by the ratio is decreased (Table 6).

The multiple coefficient of determination is obtained by dividing the sum of squares of deviations due to regression in the analysis of variance of the multiple regression problem by the total sums of squares, or

$$(7) \quad R^2 = \frac{(\Sigma x y)^2}{\Sigma x^2 \Sigma y^2}$$

with the $H_o : B_1 = B_2 \ . \ . \ . \ B_k = 0$.

Each variable is held constant and compared to an introduced variable and then the new variable is kept constant and compared to the

first variable with each introduced variable. These variables are referred to as partial regression coefficients. Each partial regression coefficient provides an estimate of the rate of change in the dependent variable associated with a one unit change in the independent variable. T-values are derived for each variable coefficient by dividing the coefficient by the standard error of the coefficient with n-2 degrees of freedom. A 2-tail T test is used to test the hypothesis about partial regression coefficients; that is, $H_o : B = 0$.

Results of the analysis also include a constant term B_o and is found by the equation:

$$(8) \quad B_o = \overline{Y} - (B_1\overline{X}_1 + B_2\overline{X}_2 \ . \ . \ . B_n\overline{X}_n)$$

where n = number of independent variables.

The nine temperature difference levels above the snow surface were plotted against snow evaporation. The temperature difference between the snow surface and 15.24 cm revealed the best correlation to snow evaporation and was chosen for the temperature difference independent variable.

Preliminary plotting of each independent variable against the dependent variable indicated that the relationship could be either linear or curvilinear. It was also apparent that the variables were interacted.

A model for predicting snow evaporation was assumed which included a linear and quadradic expression for each variable and all two way interactions between variables.

The multiple regression equation for determining the effects of the selected variables on snow evaporation Y was then written:

$$(9)\quad Y = B_o + B_1x_1 + B_2x_1^2 + B_3x_2 + B_4x_2^2 + B_5x_3 + B_6x_3^2 + B_7x_4 + B_8x_4^2 + B_9x_1x_2 + B_{10}x_1x_3 + B_{11}x_1x_4 + B_{12}x_2x_3 + B_{13}x_2x_4 + B_{14}x_3x_4$$

where:

Y = snow evaporation $cm/sec^{10^{-7}}$

Term 1 = x_1 = temperature difference in ^oC between snow.

Term 2 = x_1^2 = (temperature difference in ^oC between snow surface and 15.24 cm.)^2 .

Term 3 = x_2 = wind speed in meters per second.

Term 4 = x_2^2 = (wind speed in meters per second)^2 .

Term 5 = x_3 = radiation in langleys per minute.

Term 6 = x_3^2 = (radiation in langleys per minute)^2 .

Term 7 = x_4 = vapor pressure deficit in mm Hg .

Term 8 = x_4^2 = (vapor pressure deficit in mm Hg)^2 .

Term 9 = x_1x_2 = temperature difference x wind speed.

Term 10 = x_1x_3 = temperature difference x radiation.

Term 11 = x_1x_4 = temperature difference x vapor pressure.

Term 12 = x_2x_3 = wind speed x radiation.

Term 13 = x_2x_4 = wind speed x vapor pressure.

and

Term 14 = x_3x_4 = radiation x vapor pressure.

Phase II

A one way analysis of variance and the Duncan multiple range test were used to accomplish the objectives of Phase II. The analysis of variance at 5 and 1 percent significance levels were first used to test the $H_o : \mu 1 = \mu 2 = \mu 3 = 0$. After a significance was determined, each site was then compared with all sites for determination of significance.

RESULTS

The meteorological and evaporation data used in Phase I of the analysis are presented in Tables 6 and 7. The Phase II data presented in Table 8 and Figure 15 are the snow evaporation and condensation rates among the three forest sites during December 1965 through March 1966.

Phase I

Relationships between meteorological factors and snow evaporation as determined from the regression are presented in Tables 6 and 7.

An analysis of variance was performed to determine the significance of the overall reduction in sums of squares due to regression of snow evaporation Y on each entered independent variable. Analysis of variance showed all entering variables significant at 1 percent (Table 3).

The asterisks in Table 6 after each F-ratio indicate the levels of statistical significance of each value. One asterisk succeeding a F-ratio indicates a significance at the 0.05 level, two asterisks indicate significance at 0.01 level. The chances of making a Type I error (rejecting the H_o that there is no significant correlation between two variables when in fact there is not) are 5 in 100.

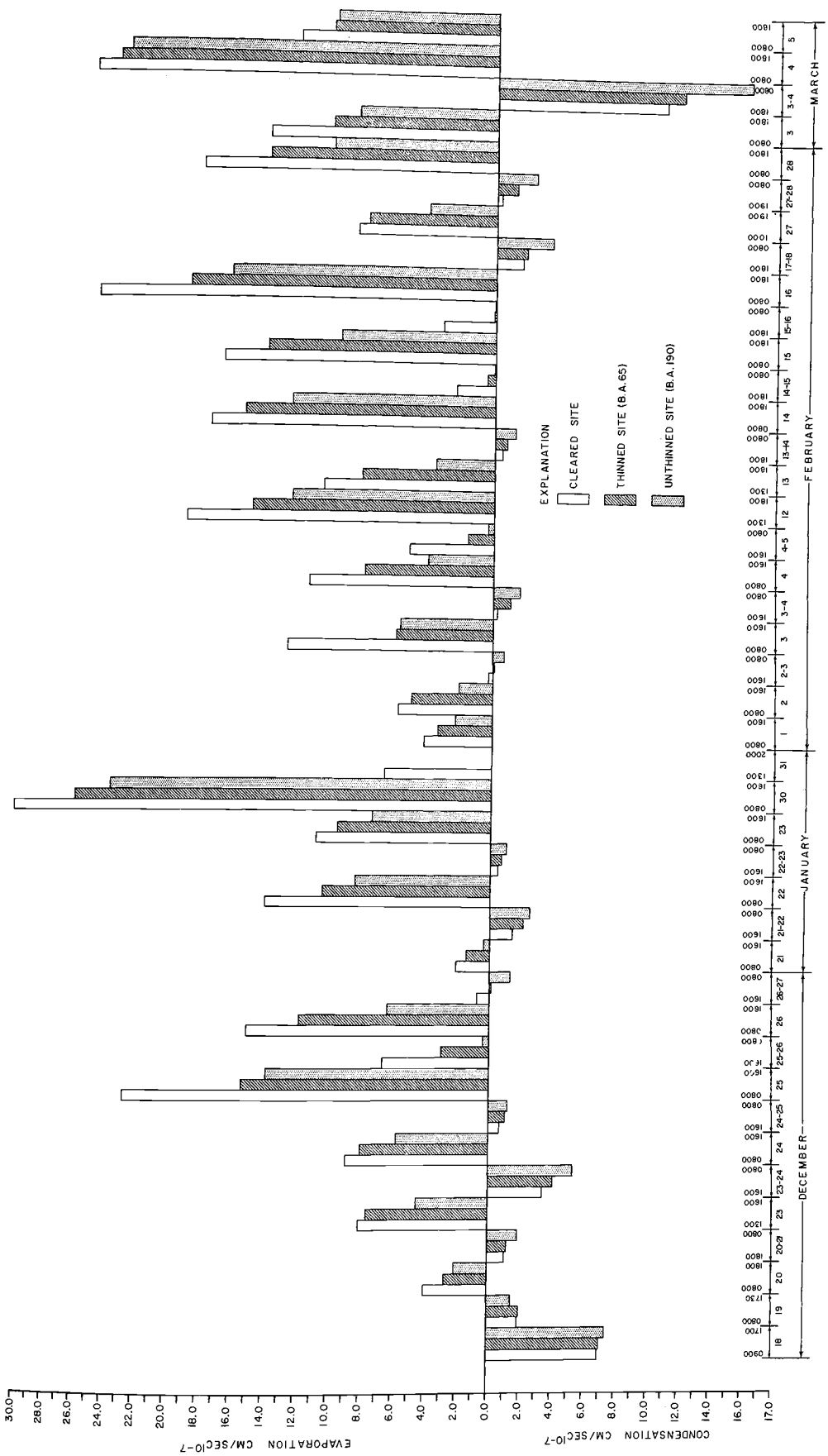

FIGURE 15.--SNOW EVAPORATION AND CONDENSATION RATES STUDIED ON THREE FOREST SITES, DECEMBER 1965 THROUGH MARCH 1966, LAKESIDE, ARIZONA.

A Type II error is made when it is stated that no correlation exists between two variables when in fact it actually does. Since the sample for this study is relatively small (n = 29), it is more important to avoid making a Type I error (stating that one variable is correlated with another variable when in fact it is not). The probability level for making a Type I error has been set at 0.05 (5 chance in 100).

The partial regression coefficients, T-values and the level of significance for each entering independent variable with its subsequent value changed for each new step in the analysis is presented in Table 7.

The wind-vapor pressure deficit interaction term showed the highest correlation with evaporation. It made a highly significant contribution to the regression in all steps of the analysis. Solar radiation showed the next highest correlation and made a highly significant contribution to the regression through Step 8. Vapor pressure deficit showed the next highest correlation and made a significant contribution to the regression through Step 5. Wind velocity showed a non-significant correlation in Step 4 but did not reduce the significance of all the previous entered variables. Vapor pressure deficit squared, Step 5, showed no correlation and made variables 3 and 7 non-significant. Temperature difference squared was entered as Step 6. It showed no significant correlation and made variables 3, 7 and 8 non-significant (Table 7).

Step 4 was chosen for the prediction equation because it represented three of the four linear independent variables and one interaction. It also has the largest number of significant variables.

The multiple regression equation showing the effects of the selected independent variables on snow evaporation Y is then:

$$(10)\quad Y = 3.926 - 3.742W + 11.398R - 3.091V + 5.743WV.$$

The coefficient of multiple determination R^2 of 0.842 shows that about 84 percent of the variability in snow evaporation was associated with or attributable to variation in the selected independent variables.

The partial regression coefficients B provide an estimate of the rate of change of the dependent variable Y associated with one-unit change in each independent variable as listed in Table 7.

The partial regression coefficients cannot be ranked according to their values because they are not in standard units of measure. However, the computed T-values of the partical regression coefficients may be ranked because they represent the value of the partial regression coefficient in terms of its standard error. The ranking of the partial regression coefficients is based on their respected T-values from Table 7.

The coefficients ranked in the order of their contribution to the regression are $B_{13} > B_5 > B_7 > B_3$.

The constant term B_o for each multiple regression equation in which snow evaporation was the dependent variable, is listed in Table 6.

The standard deviations of the partial regression coefficients were found by the following formula for the selected equation:

$$(11)\quad sbi = \sqrt{(Cij)\,(s^2y\text{-}1,2,3\ldots14)}$$

$$sb_3 = 2.494$$

$$sb_5 = 3.167$$

$$sb_7 = 1.561$$

$$sb_{13} = 5.525$$

Table 9 lists the snow evaporation, vapor pressure, short wave radiation, wind velocity means and the range of values encountered for each variable. The standard error of estimate ($sy\, x_1 \ldots x_n$) was computed as 3.14603 (Table 6). The standard error of the estimate is in the same units of measure as the original data. Therefore, approximately 95 percent of the snow evaporation observations made can be expected to vary from 9.8835 ± 3.14603 cm/sec$^{10^{-7}}$.

Phase II

The second phase of the analysis dealt with the determination of the effects of forest stand densities on snow evaporation.

An analysis of variance, one way classification, for snow evaporation on cleared, thinned and unthinned forest sites is prescribed in Table 10. The Duncan's Multiple Range Test was conducted on the mean evaporation rates on each forest site and is presented in Table 11. Snow evaporation on the cleared site was significantly greater than on the unthinned site. The evaporation on thinned site did not significantly exceed that on the unthinned site.

DISCUSSION AND SUMMARY

Hypotheses

The hypotheses tested in this study were that relationships existed among factors affecting snow evaporation and that evaporation rates would differ between forest stand densities. Multiple regression and correlation techniques, one way analysis of variance, and the Duncan Multiple Range Test were used to test the hypotheses.

Procedure and Descriptive Analysis

Data from which this study was obtained as a secondary study on The Hydrology of Ponderosa Pine the White Mountains of Arizona (Croft 1965). The data was collected by the Lakeside District Ranger, U. S. Forest Service, and by the author.

Three major factors routinely measured by the U. S. Weather Bureau: wind velocity, solar radiation, and vapor pressure, were studied on a cleared site. Consistent and frequent measurements of these factors and evaporation were obtained during January 30 through February 4, 1966.

An equation was developed that enabled snow evaporation to be predicted with a high degree of accuracy from measurements of these factors.

The variable having the highest correlation in the equation was an interaction between wind and vapor pressure. Light (1941) explains the phenomenon by the theory of heat and water transport.

In a solid body, heat flows from regions of high temperature to regions of low temperature through the process of molecular conduction. The quantity of heat transported through a unit cross section is proportional to the product of the molecular heat conductivity of the material and the temperature gradient normal to the section. Similar to molecular diffusion, heat transported across a horizontal section of turbulent air results from vertical motion of eddies and is proportional to the product of the coefficient of turbulent exchange and the vertical temperature gradient. Likewise, since vertical movements of air particles are also responsible for diffusion of moisture from one layer to another, the rate of moisture transport is determined by the product of the vertical moisture gradient and the coefficient of turbulent exchange.

The coefficient of turbulent exchange or eddy conductivity depends on three factors: wind velocity, surface roughness, and the stability of the layer of air next to the snow. An increase of wind-velocity or roughness is accompanied by an increase in degree of turbulence.

Since vertical air motions in the atmosphere take place adiabatically, continued action of turbulence within a given air mass will eventually produce an adiabiabatic lapse-rate of temperature in that air mass. Conversely, any action which decreases the lapse-rate, such as downward heat-transport, stabilizes the air-mass and inhibits turbulence. Therefore, a cold surface tends to produce stability in a warm air mass flowing over it and thus acts to dampen turbulence. Rossley (1933) has shown that for an adiabatic atmosphere, or in other words, air which no stabilization influences are present, eddy conductivity varies directly with wind velocity and height, which produces a logarithmic distribution of wind-velocity with height. Vapor-pressure will also follow the logarithmic law assuming that the processes (momentum transfer and heat-and-moisture transfer) are the same as wind.

Vapor pressure exerts a force on a snowpack regulating the molecular movement of the water molecules on ice. The vapor pressure increases with a decrease in temperature and conversely decreases with an increase in temperature. The results show an increase in snow evaporation with a decrease in vapor pressure deficit and conversely a decrease in evaporation is influenced by an increase in vapor pressure deficit.

An increase movement of air detaches water vapor molecules from the snow crystals and suspends them into the atmosphere. Study results show an increase in snow evaporation with an increase in wind velocity.

The combination of increased wind velocity, disturbing a stable surface by eddy turbulent forces, and a decrease of vapor pressure allows the molecular movement of water vapor to escape the snow crystals and produce high evaporation.

Results of the study also show that increasing short wave radiation is associated with an increase in snow evaporation. Total radiation incident on the earth's surface is in the range of wave lengths between 0.15 to 4.0μ. Forty-seven percent of the original radiation is absorbed either as direct or reflected radiation. Radiation indirectly affects snowmelt and evaporation by the heating of air which results from air passing over lands and objects. These give their heat to the air by the process of convection and conduction. Moist soil and water surfaces may also be heated or cooled by water vapor condensation or evaporating at their surfaces. Snow evaporation may be decreased when the vapor pressure gradient is decreased by the addition of water vapor to the air or when water vapor condenses on the surface.

Light (1941) states that warm moist air flowing over a snow surface transfers heat to snow in two ways. First, there is the direct heat exchange due to the difference in temperature between the air and snow. Secondly, moisture is brought down to the snow surface and condensed, releasing latent heat condensation amounting to 600 calories per cc of water deposited. Since the heat of fusion of ice is 80 calories per cc, the moisture condensed on the snow surface melts 7.5 times its own weight of snow.

Three stand densities of 0, 65, and 190 basal area of ponderosa pine were selected for comparison of total daily evaporation and condensation rates during the entire winter. Evaporation measurements were established twice daily on each site.

Results showed that snow evaporation decreased with an increase in stand density. Evaporation on the cleared plot was significantly greater than that on the unthinned plot. Evaporation on the plot thinned to a basal area of 65 did not differ significantly from that on either the cleared or unthinned plots.

As the stand density increases, solar radiation is intercepted by the crown cover, prohibiting the absorption of heat for energy conversion and decreases the air temperature. As the air temperature decreases, the vapor pressure inversely increases, impairing

evaporation or creating condensation. Forest density increase restricts air movement, which reduces the eddy conductivity.

CONCLUSIONS

1. Snow evaporation can be predicted with a high degree of accuracy from standard U. S. Weather Bureau measurements of wind velocity, relative humidity and solar radiation. An empirical equation was developed for this purpose.
2. Snow evaporation is significantly affected by forest cover. Evaporation decreases with the extent of thinning on the order: cleared, thinned and unthinned.

APPENDIX

Table 1. Heat transfer from air of different temperatures to snow

Air Temperature ^{o}C	Snow Temperature ^{o}C	Wind Velocity m sec^{-1}	Heat Transfer Cal cm^{-2} day^{-1}
5	0	20	900
10	0	20	1790
15	0	20	2690
20	0	20	3590

Table 2. Evaporation from a snow surface at 0°C to air of different temperatures

Air Temperature °C	Relative Humidity %	Wind Velocity m sec^{-1}	Snow Evaporation gm cm^{-2} day^{-1}
5	20	20	2.02
10	20	20	1.69
15	20	20	1.25
20	20	20	.67

Table 3. The maximum relative humidity which air can have at different temperatures to have a vapor pressure less than 6.11 mb

Air Temperature °F	Relative Humidity %
0	99
5	70
10	49.7
15	35.7
20	26

Table 4. Conditions for equal evaporation from a snow surface at varying air temperatures

Air Temperatures ^{o}C	Saturated Vapor Pressure mb	Relative Humidity %	Vapor Pressure mb	Wind Velocity m sec^{-1}	Snow Evaporation gm $cm^{-2} day^{-1}$
5	8.719	20.0	1.74	20	2.02
10	12.277	14.2	1.74	20	2.02
15	17.054	10.2	1.74	20	2.02
20	23.391	7.4	1.74	20	2.02

Table 5. Snow evaporation rate, temperature difference, wind velocity, radiation and vapor pressure deficits studied on a cleared forest site, Lakeside, Arizona

Date	Time	Snow Evaporation cm/sec 10^{-7}	Temp. Difference $\pm$ °C 0-15.29 cm	Wind Velocity m/sec	Shortwave Radiation cal $cm^{-2}m^{-1}$	Vapor Pressure Deficit mm Hg
19 Dec 65	1020-1255	8.56	+2.330	0.601	0.836	1.779
19 Dec 65	1255-1445	6.44	+1.872	0.421	0.624	1.466
19 Dec 65	1445-1730	0.42	-0.213	0.060	0.091	1.329
30 Jan 66	1200-1635	34.54	-2.250	2.342	0.254	3.229
31 Jan 66	1300-1400	15.13	-0.622	1.341	0.753	1.175
31 Jan 66	1400-1510	15.96	-0.710	0.767	0.691	1.192
31 Jan 66	1510-1655	5.32	-0.729	1.022	0.411	1.227
31 Jan 66	1655-1950	1.97	+0.164	0.460	0.004	0.903
1 Feb 66	0810-1000	2.54	+0.294	0.488	0.171	0.243
1 Feb 66	1000-1100	6.98	-0.153	0.894	0.358	0.305
1 Feb 66	1320-1405	4.65	-1.179	1.192	0.448	1.133
1 Feb 66	1430-1500	13.97	-0.558	0.894	0.271	1.330
1 Feb 66	1500-1650	2.54	-1.029	0.488	0.190	1.250
2 Feb 66	0930-1135	5.31	-0.349	0.644	0.421	0.383
2 Feb 66	1135-1410	9.01	-1.280	1.049	0.571	1.030
2 Feb 66	1410-1525	9.31	-1.228	1.073	0.771	1.523
2 Feb 66	1525-1755	4.66	-0.166	0.894	0.116	1.599
3 Feb 66	0745-0935	5.08	-0.039	0.0	0.054	0.285
3 Feb 66	0935-1035	11.64	-1.551	0.0	0.674	0.487
3 Feb 66	1035-1220	14.76	-1.153	1.022	0.963	0.654
3 Feb 66	1225-1330	19.34	-3.100	0.826	0.933	2.703
3 Feb 66	1330-1710	10.16	-2.078	0.488	0.583	2.801
3 Feb 66	1710-2250	0.0	+0.937	0.158	0.003	1.333

Table 5--Continued

Date	Time	Snow Evaporation $cm/sec^{10^{-7}}$	Temp. Difference $\pm^{o}C$ 0-15.29 cm	Wind Velocity m/sec	Shortwave Radiation $cal\ cm^{-2}m^{-1}$	Vapor Pressure Deficit mm Hg
4 Feb 66	0745-1000	5.69	-0.032	0.0	0.196	0.115
4 Feb 66	1000-1145	10.64	-1.554	0.213	0.896	0.566
4 Feb 66	1145-1300	10.24	-3.257	0.715	0.891	2.795
4 Feb 66	1300-1415	19.18	-6.616	1.073	0.781	3.298
4 Feb 66	1415-1600	13.63	-1.446	0.766	0.541	3.237

Table 6. Stepwise multiple regression analysis of the relationships between the independent and dependent variables of snow evaporation

Step Number	Term Entering	Standard Error of Y	R^2	Constant	Partial Regression Coefficient	Standard Error of Coefficient	T-values	F-ratio
1	13	4.3755	0.656934	5.14015	3.4415	0.478630	7.19042	51.702**
2	5	3.4471	0.794899	1.05738	8.85933	2.11843	4.18203	50.383**
3	7	3.23206	0.826677	2.27595	10.2338	2.08717	4.90320	39.746**
4	3	3.14608	0.842349	3.92693	3.74187	2.42250	1.54463	32.059**
5	8	3.18506	0.845146	4.52148	-2.78017	2.87076	-0.968443	25.105**
6	2	3.20129	0.850365	4.58682	-8.79906	0.10047	-0.875991	20.837**
7	10	3.20265	0.857046	4.49814	-0.158735	0.123279	-1.28760	17.986**
8	9	3.12734	0.870181	4.76288	-1.81501	0.155752	-0.116532	16.758**
9	14	3.17078	0.873221	3.76513	3.36701	0.175586	0.191759	14.541**
10	12	3.18377	0.878908	4.47676	5.61746	0.177996	0.315595	13.065**
11	11	3.14028	0.888738	4.10762	-0.12848	0.231252	-0.555348	12.345**
12	1	3.22154	0.889793	3.59738	1.28490	3.28313	0.391363	10.765**
13	6	3.31657	0.890496	3.26068	1.58711	3.51753	0.451201	9.383**
14	4	3.42133	0.891238	3.27624	1.75061	3.66702	0.477392	8.194**

** Significant at 1%

Table 7. Partial Regression coefficients, T-values and the level of significance for each entering independent variable with its subsequent value change for each new step in the analysis

Step Number	Entering Term	Partial Regression Coefficient	T-values	Level of Significance
1	(13)	3.44155	7.19042	**
2	(5)	8.85933	4.18203	**
	13	3.36674	8.91724	**
3	5	10.2338	4.90320	**
	(7)	-2.21186	2.14093	*
	13	4.36348	7.46107	**
4	(3)	-3.74187	-1.54463	NS
	5	11.3897	5.26070	**
	7	-3.09101	-2.67497	*
	13	5.74372	5.42116	**
5	3	-2.78017	-0.968443	NS
	5	11.1256	4.98936	**
	7	-4.58194	-1.76761	NS
	(8)	0.544364	0.644542	NS
	13	5.23196	3.92043	**
6	(2)	-8.79906	-0.875991	NS
	3	-2.89735	-1.00307	NS
	5	11.6504	5.02191	**
	7	-5.04314	-1.89731	NS
	8	0.798986	0.890469	NS
	13	5.23126	3.90803	**
7	2	-0.158735	-1.28760	NS
	3	-3.34714	-1.14426	NS
	5	10.9711	4.53338	**
	7	-4.32271	-1.56801	NS
	8	0.496917	0.524164	NS
	(10)	-0.802503	-0.990652	NS
	13	5.51584	4.01939	**
8	2	-1.81501	-0.116532	NS
	3	-2.54414	-0.873787	NS
	5	0.27974	3.50786	**
	7	-5.22858	-1.89011	NS

Table 7--Continued

Step Number	Entering Term	Partial Regression Coefficient	T-values	Level of Significance
	8	0.517933	0.559417	NS
	(9)	1.78434	1.42252	NS
	10	-1.96638	-1.72786	NS
	13	7.08191	4.08349	**
9	2	3.36701	0.191759	NS
	3	-2.11866	-0.701865	NS
	5	11.5518	2.68407	NS
	7	-4.61259	-1.56389	NS
	8	0.736112	0.741457	NS
	9	2.41860	1.55661	NS
	10	-2.29495	-1.83256	NS
	13	6.94884	3.92726	**
	(14)	-2.25407	-0.675036	NS
10	2	5.61746	0.315595	NS
	3	-4.98620	-1.14650	NS
	5	10.0685	2.18270	**
	7	-4.51653	-1.52413	NS
	8	0.910395	0.897187	NS
	9	2.74928	1.71716	NS
	10	-2.38257	-1.88935	NS
	(12)	5.77287	0.919406	NS
	13	7.48041	4.00377	**
	14	-3.84925	-1.01962	NS
11	2	-0.128481	-0.555348	NS
	3	-8.22608	-1.63255	NS
	5	13.3366	2.5289	*
	7	-2.25450	-0.652187	NS
	8	-0.247109	-0.179569	NS
	9	5.45521	2.00963	NS
	10	-0.036445	-0.015964	NS
	(11)	-1.98828	-1.22558	NS
	12	10.0174	1.41175	NS
	13	9.27069	3.94237	**
	14	-6.64604	-1.52179	NS
12	(1)	1.28490	0.391363	NS
	2	-0.235291	-0.650545	NS

Table 7--Continued

Step Number	Entering Term	Partial Regression Coefficient	T-values	Level of Significance
	3	-8.11659	-1.56789	NS
	5	15.8020	1.90298	NS
	7	-1.60383	-0.409487	NS
	8	-0.50587	-0.324509	NS
	9	5.52455	1.97984	NS
	10	-0.316604	-0.129282	NS
	11	-2.62821	-1.12647	NS
	12	9.20801	1.21680	NS
	13	9.32846	3.85965	**
	14	-7.49743	-1.50536	NS
6	1	1.58711	0.451201	NS
	2	-0.241742	-0.648339	NS
	3	-8.74065	-1.53442	NS
	5	19.0028	1.41833	NS
	(6)	-3.56359	-0.310267	NS
	7	-1.41682	-0.347515	NS
	8	-0.561896	-0.347921	NS
	9	5.31149	1.79827	NS
	10	-0.725738	-0.255074	NS
	11	-2.57331	-1.06844	NS
	12	-9.56317	1.21449	NS
	13	9.44378	3.75376	**
	14	-7.52373	-1.46716	NS
14	1	1.75061	0.477392	NS
	2	-0.256053	-0.660814	NS
	3	-8.54653	-1.44616	NS
	4	-1.69662	-0.308945	NS
	5	20.1445	1.40804	NS
	6	-4.70057	-0.378872	NS
	7	-1.70980	-0.396579	NS
	8	-0.629225	-0.374490	NS
	9	5.36349	1.75760	NS
	10	-1.01307	-0.329035	NS
	11	-2.55897	-1.02977	NS
	12	10.7795	1.19417	NS
	13	10.8390	2.08094	*
	(14)	-8.21011	1.43091	NS

Table 7--Continued

* Significant at 5%

** Significant at 1%

( ) Entered Variable

NS Not Significant

Table 8. Snow evaporation and condensation rates studied on three forest sites, December 1965 through March 1966, Lakeside, Arizona

Date	Time	Cleared, Site 1 cm/sec 10^{-7}	Thinned (B.A. 65) Site 2, cm/sec 10^{-7}	Unthinned (B.A. 190) Site 3, cm/sec 10^{-7}
18 Dec 65	0900-1700	+ 6.694	+ 7.044	+ 7.422
19 Dec 65	0800-1730	+ 1.924	+ 1.936	+ 1.495
20 Dec 65	0800-1800	- 3.958	- 2.608	- 2.188
20, 21" 65	1800-0800	+ 1.0811	+ 1.214	+ 1.846
23 Dec 65	1300-1600	- 8.150	- 7.607	- 4.501
23, 24" 65	1600-0800	+ 3.493	+ 4.158	+ 5.322
24 Dec 65	0800-1600	- 8.878	- 8.004	- 5.821
24, 25" 65	1600-0800	+ 0.728	+ 1.048	+ 1.281
25 Dec 65	0800-1600	-22.762	-15.427	-13.971
25, 26" 65	1600-0800	- 6.695	- 3.056	- 0.291
26 Dec 65	0800-1600	-15.136	-11.934	- 6.413
26, 27" 65	1600-0800	- 0.800	+ 0.146	+ 1.310
21 Jan 66	0800-1600	- 2.183	- 1.455	- 0.291
21, 22" 66	1600-0800	+ 1.426	+ 2.183	+ 2.474
22 Jan 66	0800-1600	-14.117	-10.479	- 8.441
22, 23" 66	1600-0800	+ 0.437	+ 0.655	+ 0.946
23 Jan 66	0800-1600	-10.915	- 9.605	- 7.422
30 Jan 66	0800-1600	-29.835	-25.906	-23.722
31 Jan 66	1300-2000	- 6.653	-	-
1 Feb 66	0800-1600	- 4.366	- 3.493	- 2.328
2 Feb 66	0800-1600	- 5.967	- 5.094	- 2.037
2, 3 " 66	1600-0800	- 0.364	+ 0.109	+ 0.728
3 Feb 66	0800-1600	-12.865	- 6.025	- 5.821
3, 4 " 66	1600-0800	+ 0.218	+ 1.092	+ 1.746
4 Feb 66	0800-1600	-11.497	- 8.150	- 4.075
4, 5 " 66	1600-0800	- 5.312	- 1.601	- 0.291
12 Feb 66	1300-1800	-19.094	-14.903	-12.740
13 Feb 66	1300-1800	-10.711	- 8.383	- 3.725
13, 14" 66	1800-0800	+ 0.509	+ 0.800	+ 1.150
14 Feb 66	0800-1800	-17.691	-15.602	-12.108
14, 15" 66	1800-0800	- 2.495	- 0.499	- 0.016
15 Feb 66	0800-1800	-16.999	-14.204	- 9.780
15, 16" 66	1800-0800	- 3.326	- 0.166	- 0.049
16 Feb 66	0800-1800	-24.683	-19.094	-16.533
17, 18" 66	1800-0800	+ 1.746	+ 1.996	+ 3.659

Table 8--Continued

Date	Time	Cleared, Site 1 cm/sec 10^{-7}	Thinned (B.A. 65) Site 2, cm/sec 10^{-7}	Unthinned (B.A. 190) Site 3, cm/sec 10^{-7}
27 Feb 66	1000-1900	- 8.797	- 9.021	- 4.269
27, 28" 66	1900-0800	+ 0.358	+ 1.254	+ 2.597
28 Feb 66	0800-1800	-18.396	-14.204	-10.106
3 Mar 66	0800-1800	-14.204	-10.246	-8.615
3, 4" 66	1800-0800	+10.811	+11.976	+16.270
4 Mar 66	0800-1800	-24.916	-23.519	-22.820
5 Mar 66	0800-1800	-12.342	-10.246	-10.012

Evaporation (-)

Condensation (+)

Table 9. Range and mean of the independent and dependent variables

Variable	Range	Mean
Snow evaporation cm/sec. 10^{-7}	0.0 - 34.54	9.883
Vapor pressure mm Hg	0.115 - 3.298	1.465
Wind Velocity m/sec	0.0 - 2.342	0.751
Solar radiation cal cm^{-2} m^{-1}	0.003 - 0.963	0.472

Table 10. Analysis of variance of snow evaporation influenced by cleared, thinned and unthinned forest sites in Lakeside, Arizona

Source of variation	Degrees of freedom	Sums of squares	Mean squares	F-values
Among	2	354.574	117.287	3.947**
Within	78	3503.319	44.914	
Total	80	3857.893		

** Significance at 1% level

Table 11. Duncan's Multiple Range Test for snow evaporation rates on cleared (A), thinned (B), and unthinned (C) forest sites in Lakeside, Arizona

Mean	Varieties	Differences	SSR*	SSR**	Conclusions	
(A)						
12.455	A-C	4.114	3.818	5.018	S	NS
(B)						
9.612	A-B	2.843	3.624	4.811	NS	NS
(C)						
7.341	B-C	2.271	3.624	4.811	NS	NS

* Significant at 5%
** Significant at 1%
Standard error of mean = 1.29

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