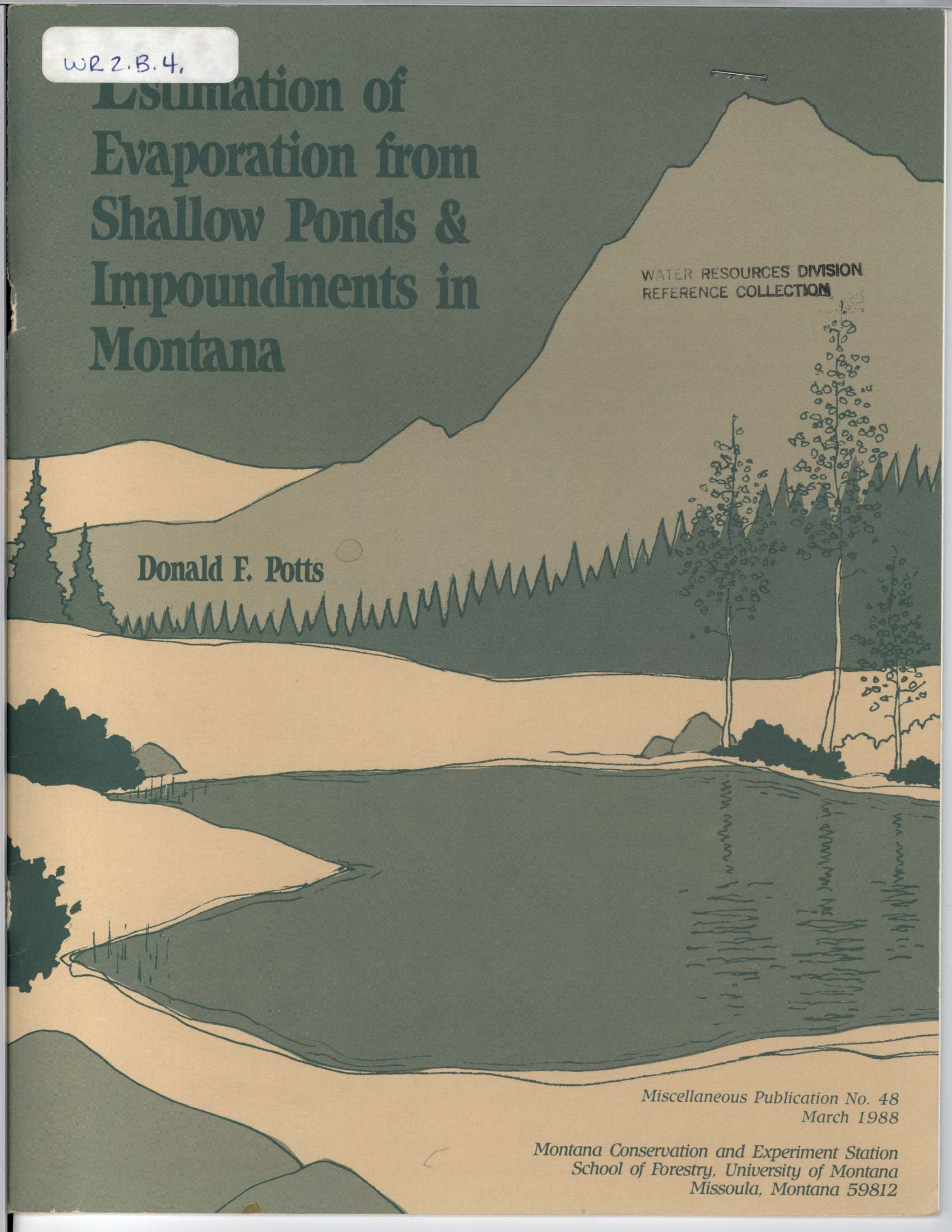
Estimation of Evaporation from Shallow Ponds & Impoundments in Montana

Donald F. Potts



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March 1988

Montana Conservation and Experiment Station
School of Forestry, University of Montana
Missoula, Montana 59812

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Donald F. Potts is an associate professor of watershed management at the University of Montana School of Forestry. He teaches courses in forest hydrology, watershed management, snow hydrology and forest meteorology. He earned a BA in chemistry at the State University of New York (Buffalo) and a PhD in forest meteorology at the New York State University College of Environmental Sciences and Forestry.

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Introduction

The past 50 years have witnessed a tremendous increase in the number of artificial lakes for storage of irrigation water, municipal supply, stock watering, power generation, industrial cooling, waste treatment and recreation in Montana. Evaporation from both artificial and natural water surfaces represents a continuing loss from the state's water supply. Engineers, planners, economists and others who require reliable short- and long-term estimates of water supply should be aware of the net loss of water from these open water surfaces.

Evaporation is a process in which water changes state from liquid to vapor and is then transferred into the atmosphere. The process occurs when some of the liquid water molecules obtain the energy to overcome surface tension and break away into free air. The greatest sources of energy available for this process are usually sunshine and warm air. Therefore, evaporation is strongly related to such things as latitude, season, time of day and cloudiness.

When water molecules leave a liquid surface, they produce a vapor pressure in the air over the surface. Water temperature is a measure of the energy of the water molecules. The higher the temperature, the greater the rate at which water molecules will escape. Consequently, vapor pressure at the surface is directly related to the temperature of the surface. Some molecules of water vapor already in the air will be going the other direction however, and condense on the liquid surface. The net gain or loss of water at the surface depends on the differences in vapor pressure between the atmosphere and the surface. The rate of evaporation or condensation is directly proportional to the magnitude of those vapor pressure differences. In semi-arid Montana, evaporation is much more common than condensation, and the simple term "semi-arid" suggests that evaporation usually proceeds at a fairly rapid rate.

If the air above a water surface is still, and if evaporation from the surface continues long enough, the air above the surface becomes saturated. There are no longer vapor pressure differences, and evaporation stops. If evaporation is to continue, the surface air layer must be constantly removed and replaced with unsaturated air. Over lakes and ponds, wind is usually responsible for this removal and replacement of air. The faster the wind is blowing, the faster evaporation will take place.

Detailed study of evaporation requires a fairly complete understanding of meteorology and physics. Similarly, methods to estimate or measure evaporation accurately usually require detailed measurements and complex computations. Fortunately, the need in many disciplines for knowledge about evaporation has encouraged researchers to develop simpler approaches and methods.

This report presents five methods for estimating evaporation from lakes and ponds; these techniques were chosen after a thorough review of the literature. The methods are particularly applicable to semi-arid environments in general and to Montana in particular. Each method is discussed in terms of its capabilities, limitations and data requirements. Procedures to obtain required data from records, measurements or extrapolation are also presented. Finally, the procedures are compared with a common input data set. The order in which the procedures are presented in this report is not a ranking of their capabilities.

Types of Approaches

It is very difficult to actually measure evaporation from a lake or a pond. An alternative approach, however, is to measure all the inflows, such as precipitation, and outflows, such as stream discharge, and calculate evaporation as a residual. This approach is called the **Mass-Balance** or **Water-Budget** method. The major problem is quantifying the subsurface inflows and outflows to and from the body of water. Additionally, most lakes have irregular shorelines, so measurements of depth change over time can provide only rough estimates of change in lake volume.

The **Evaporation Pan** was developed as a surrogate for measuring evaporation from lakes. The U.S. Weather Bureau Class A pan is made of unpainted, galvanized iron, supported on a wooden frame a few inches off the ground. It is four feet in diameter, ten inches deep and filled initially with eight inches of water. Unfortunately, the water can absorb considerable energy through the sides and bottom of the pan. The result is that evaporation measured from the pan is almost always higher than that from the open body of water adjacent to it. To account for this, **Pan Coefficients** were developed. Unfortunately, pan coefficients are location-specific and vary throughout the year. A further difficulty is that the relationship of pan evaporation to lake or pond evaporation also depends on the size and depth of the body of water. Most regions have average annual pan coefficients of between 0.70 and 0.75, but considerable errors are possible if the annual coefficient is used to estimate daily, weekly or monthly open-water evaporation.

The **Energy Budget** approach quantifies the energy flows to and from a body of water and determines the energy lost in evaporation as a residual. All modes of energy transfer - convection, conduction and radiation - must be accurately determined. The evaporation energy residual has a mass (and volume) equivalent because we know the value of the latent heat of vaporization: Approximately 580 calories are required to vaporize one cubic centimeter of water. The primary drawbacks of the energy budget approaches are equipment expense and the spatial and temporal variability of the energy fluxes.

The **Mass Transfer** method requires very accurate measurement of the vertical profiles of wind and humidity. More often, however, both the mass transfer and energy budget approaches are dealt with semi-empirically; that is, some of the more complex formulations have been parameterized, and simple coefficients replace those components that are either relatively invariable or, more often, the most difficult to determine. Typically, the more empirical a formulation is, the less reliable it becomes.

The Chosen Procedures:

The following five procedures are identified by number and named after the developer or primary investigator. These procedures are field-practical methods for estimating evaporation from lakes and ponds in Montana.

1. HARBECK (1962) - Calculation of daily evaporation (semi-empirical mass transfer)

Reference - Harbeck, G.E. 1962. A practical field technique for measuring evaporation using mass-transfer theory; U.S. Geological Survey Professional Paper 272-E. Washington, DC.

Tested applicability - Reservoirs ranging from less than .01 to greater than 100 sq. kilometers.

Accuracy claims - "acceptable accuracy" (Harbeck 1962) Up to 25% error possible if the mass-transfer coefficient, N, is estimated empirically as a function of reservoir area, or if net seepage, S, is extrapolated erroneously. But rapid and cheap estimates of the mass-transfer coefficient and seepage can be made using techniques first presented by Langbein et al (1951) and further elaborated upon by Harbeck (1962).

Data requirements

1. delta H - net change of water surface elevation adjusted for inflow and outflow and for precipitation. (cm)
2. vapor pressure of the water surface, e_s. (mb)
3. vapor pressure of unmodified air (ambient), e_a. (mb)
4. total daily wind run at 2 meter height (km/day)

Procedure

Nearly all mass-transfer equations for estimating evaporation have one thing in common: evaporation is considered proportional to the product of the windspeed, u, and the vapor pressure difference between a surface at saturation and unmodified air away from the surface. The simple semi-empirical equation is:

$$E = N\, u\, (e_s - e_a) \qquad (1a)$$

where E = evaporation in cm/day,
N = a constant known as the mass-transfer coefficient
u = total daily wind run 2 m above water (km/day)
e_s = vapor pressure at the water surface (mb)
e_a = vapor pressure of the surrounding air (mb)

Values for N may be estimated from the surface area of the impoundment according to the relation

$$N = 0.000169 \; A^{(-0.05)} \tag{1b}$$

where, N has units of (cm/day)/(km/d X mb),
A is the area of the impoundment (sq.km.)

Alternatively and preferably, N may be determined experimentally for a given impoundment. If there is no precipitation or other surface inflow to the impoundment, any change in surface elevation is due to a combination of evaporation and net groundwater seepage:

$$\text{DELTA } H = E + S \tag{1c}$$

where DELTA H = water level change (cm/day)
E = Evaporation (cm/day)
S = net groundwater seepage (cm/day)

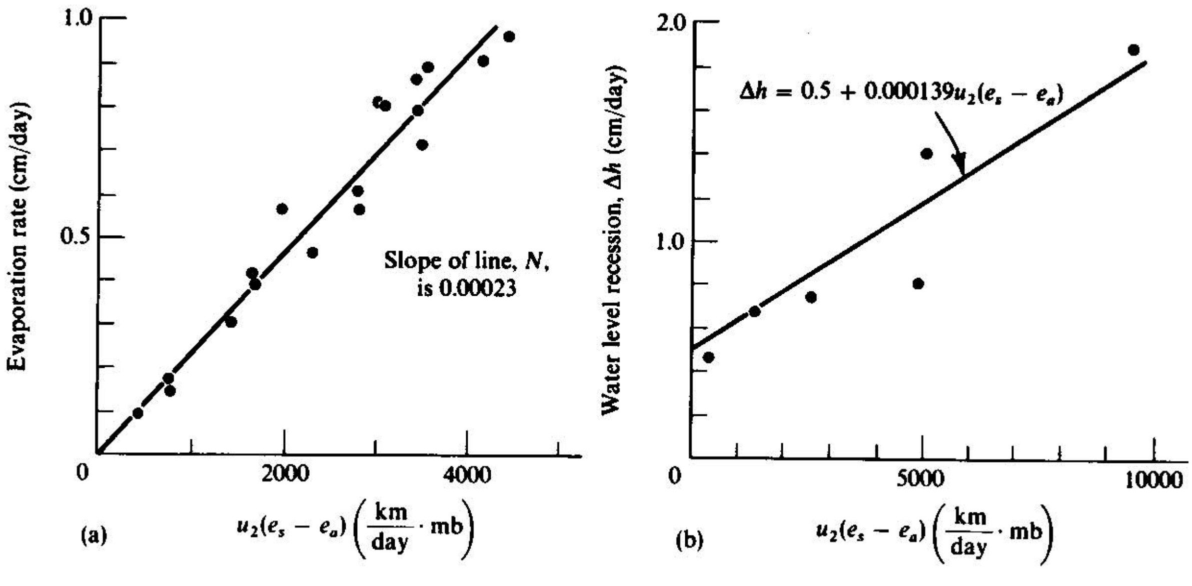
If S is already known, substitution into equation 1c provides a measured rate of evaporation from a measured water level change. When various measured-daily rates of evaporation are plotted against the corresponding products of daily wind run and vapor pressure deficit that produced them, the slope of the resulting line is the value of N, the mass-transfer coefficient (see figure 1a).

If S is not known, the plot of DELTA H against $u\ (e_s - e_a)$ yields a regression line with a slope that is the value of N and a Y-axis intercept that is the seepage rate (see figure 1b).

Daily evaporation rates can then be calculated via equation 1a and used in a daily mass balance for the impoundment or summed over weekly, biweekly or monthly time intervals.

FIGURES 1 a and b. (from Dunne and Leopold, 1978)

a) Plot of measured or calculated evaporation rate against $u\ (e_s - e_a)$.
b) Plot of measured change in water level against $u\ (e_s - e_a)$ for a pond with seepage losses.



2. LAMOREUX/KOHLER (1962) - Calculation of daily evaporation (semi-empirical / energy budget)

References - Lamoreux, W.W. 1962. Modern evaporation formulae adapted to computer use. Monthly Weather Review 90: 26-28.

Kohler, M.A., Nordenson, T.J., and W.E. Fox. 1955. Evaporation from pans and lakes. U.S. Weather Bureau Research Paper 38. Washington, DC. 21pp.

Tested applicability - One of the standard procedures. Often used as the standard against which comparisons are made.

Accuracy claims - "The results indicate that the (predictive) relation is universally applicable" (Kohler et al. 1955)

Data requirements -

1. mean daily dew point temperature (degrees F)
2. total wind run (miles/day)
3. mean daily air temperature (degrees F)
4. total daily global radiation (langleys/day)

Procedure

Kohler et al. (1955) produced a graphical method to compute lake evaporation from meteorological factors, based on the Penman formula:

$$Elake = 0.70 (Qn (S/S+g) + Ea (g/S+g)) \qquad (2a)$$

where

Elake is the evaporation rate,
0.70 is the standard pan-to-lake coefficient,
Qn is net radiation exchange,
Ea is pan evaporation,
S is the slope of the saturation vapor pressure curve at mean daily air temperature,
and, g is the psychrometric constant.

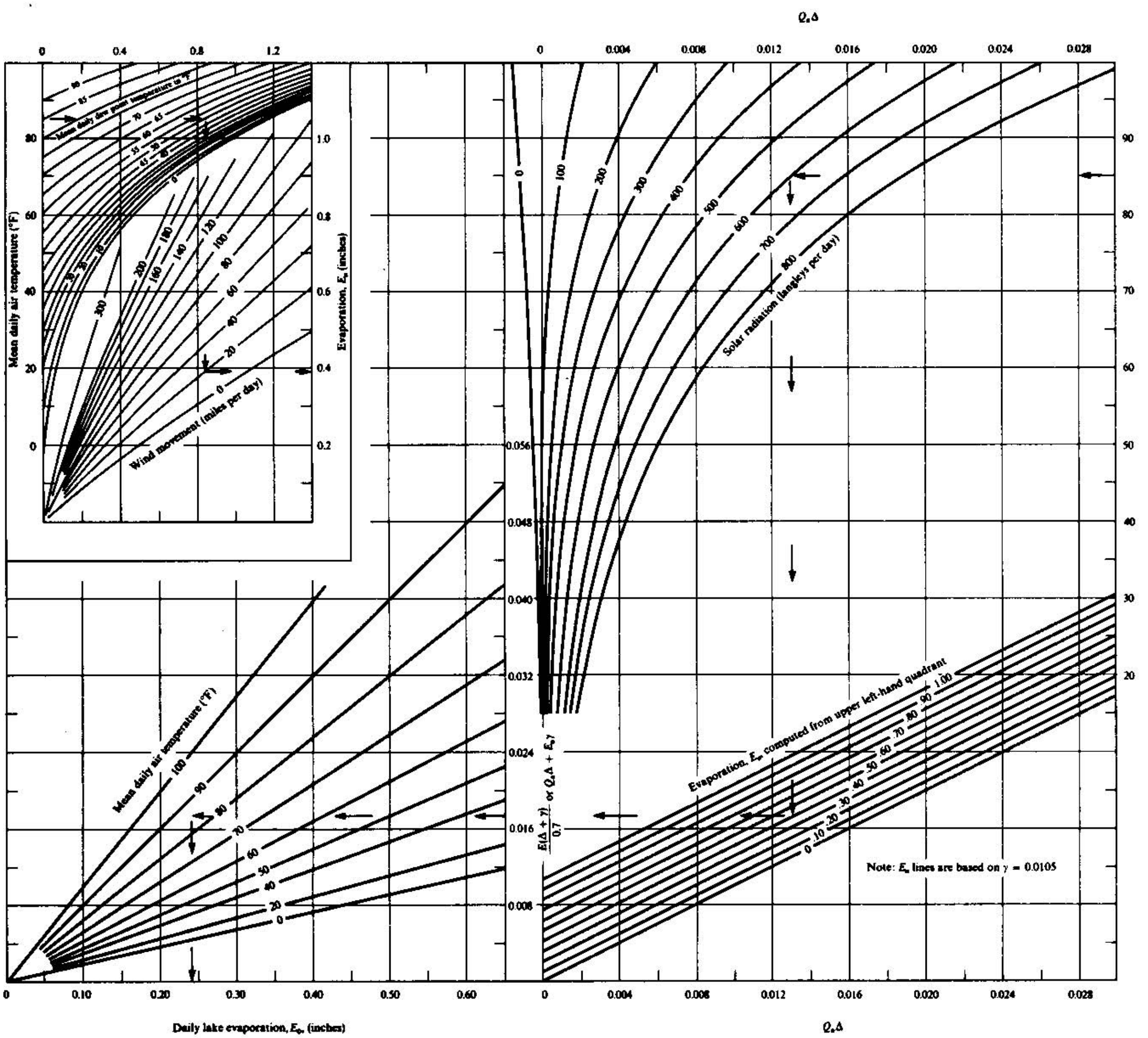
Note that this version of the Penman equation has a slightly different form from the one to be discussed in the Penman-Linacre method, but nevertheless produces the same results.

Figure 2 shows the various graphs used in the Kohler method to compute lake evaporation. Lamoreux (1962) simply fit mathematical functions to the various graphical relations and combined them linearly into an equation which is readily adaptable for performing repetitive calculations on a computer:

$$\text{Elake} = [\exp((T - 212)(0.1024 - 0.01066 \ln \text{GLOBAL})) - 0.0001 + 0.0105\,(e_s - e_a)\,0.88\,(0.37 + 0.0041\,\text{Up})] \times [0.04686\,(0.0041\,T + 0.676)^7 + -0.01497]^{-1} \qquad (2b)$$

where T is average daily temperature (degrees F),
GLOBAL is global radiation (langleys/day),
Up is total wind run (miles/day), and
$(e_s - e_a)$ is the average vapor pressure deficit (in.HG)

FIGURE 2. Computation of lake evaporation from meteorological data. To use diagram: (1) enter upper left diagram with mean daily air temperature; (2) at mean daily dew point temperature, read down to wind measurement; (3) read horizontally to right scale of Ea; (4) enter upper right diagram with mean daily air temperature, move left to value of solar radiation; (5) move downward to previously computed value of Ea in lower diagram; (6) thence downward to read answer, daily lake evaporation. From Kohler et al. 1955 , in Dunne and Leopold 1978.



3. STEWART AND ROUSE (1976) - Calculation of daily evaporation (semi-empirical / energy budget)

References - Stewart, R.B. and W.R. Rouse. 1976. A simple method for determining the evaporation from shallow lakes and ponds. Water Resour. Res. 12(4):623-628.

Priestley, C.H.B. and R.J. Taylor. 1972. On the assessment of surface heat flux and evaporation using large-scale parameters. Mon. Weather Rev. 100:81-92.

Tested applicability - Lakes and ponds with depths between 0.5 and 2 meters.

Accuracy claims - The Priestley and Taylor model can be used to estimate daily evaporation from shallow lakes within 5%. The model, as modified by Stewart and Rouse, estimated shallow lake evaporation within 10% for periods of two weeks to a month.

Data requirements

MINIMALLY
1. daily total global radiation
2. average daily air temperature at screen height

OPTIMALLY
1. daily total net radiation

Procedure

Priestley and Taylor (1972) showed, and numerous researchers have verified in recent years, that potential evaporation from a water surface can be estimated accurately on a daily basis from the expression:

$$LE = 1.26 \ (S / S + g) \ (Q_n - G) \qquad \text{(MJ/m2-day)} \qquad (3a)$$

where S is the slope of the saturation vapor pressure-temperature curve at the mean daily air temperature,
g is the psychrometric constant,
Q_n is the daily total net radiation flux density, and
G is subsurface heat flow.

The ratio, (S / S + g), is determined at mean daily temperature. The value is usually found in "look up" tables in standard meteorology texts.

Simplifying assumptions

It has often been observed that over the course of a 24-hour day, conduction or subsurface heat flow has a net value of approximately 0. In other words, for shallow ponds, water and the underlying surface gain and lose energy (therefore, temperature) very efficiently. Energy gained during daylight hours is lost at night, and the result is that mean daily water temperature doesn't change much from day to day. Small increments do add up, however, so mean daily water temperature in July is much warmer than in March, for example. Stewart and Rouse (1976) contend that little error in evaporation is produced by assuming that $G = 0$ for time intervals from one to two weeks.

Net radiation, Q_n, is usually measured with an expensive radiometer positioned one to two meters over the water surface. Obviously, net radiation data will not be routinely available for field application. Alternatively, Robinson, et al. (1972) derived an accurate empirical expression for net allwave radiation, Q_n, over mid-latitude water surfaces:

$$Q_n = 0.368 + 0.823(R_{net}) \quad \text{(MJ/m2-day)} \qquad \text{(3b)}$$

where R_{net} is net shortwave (solar) radiation.

Assuming an <u>average</u> water surface albedo of 0.20, then:

$$R_{net} = 0.80 \text{ (GLOBAL)} \quad \text{(MJ/m2-day)} \qquad \text{(3c)}$$

where GLOBAL is daily total incoming beam and diffuse shortwave radiation. Global radiation may be measured directly and summed over the daylight period, or may be estimated by the Bristow and Campbell (1984) procedure, which is discussed elsewhere in this report.

Combining equations 3b and 3c,

$$Q_n = 0.368 + 0.658 \text{ (GLOBAL)} \quad \text{(MJ/m2-day)} \qquad \text{(3d)}$$

When equations 3a and 3d are combined, the total energy utilized in daily evaporation is determined thus:

$$LE = (S / S + g) (0.463 + 0.829 \text{ (GLOBAL)}) \quad \text{(MJ/m2-day)} \qquad \text{(3e)}$$

Total depth of water evaporated daily (mm/day) is determined by dividing LE by L, the latent heat of vaporization, and assuming the density of water to be 1 gr/cm.

4. PENMAN/LINACRE (1977) - Calculation of evaporation over any averaging period using a simplified PENMAN approach (semi-empirical energy budget)

References - Linacre, E.T. 1977. A simple formula for estimating evaporation rates in various climates using temperature data alone. Agric. Meteorol. 18:409-424.

Rosenberg, N.J., Blad, B.L., and Verma, S.B. 1983. Microclimate - The Biological Environment, 2d Ed. John Wiley and Sons, New York. 495 pp.

Tested applicability - Tested against measured pan evaporation rates in a wide range of climates and geographic locations.

Accuracy claims - Average errors are less for longer periods of averaging, which is often the case when errors are random rather than systematic: 0.3 mm/day for annual means, 0.5 mm/day for monthly means, 0.9 mm/day for weekly means, 1.7 mm/day for a day.

Data requirements

MINIMALLY
1. elevation (meters)
2. latitude (degrees)
3. daily maximum and minimum temperatures (C)

OPTIMALLY
4. average net radiation or global radiation (cal/cm^2 - s)
5. ambient vapor pressure and saturation vapor pressure (mb) at daily mean temperature

Procedure

The Penman formula for the rate of evaporation from an extensive and uniform wet surface can be written:

$$LE = (Q_n + \rho c(e_s-e_a)/S\ r_a\ ) \ /\ (1 + g/S) \quad (cal/cm^2\text{-}s) \quad (4a)$$

where L = the latent heat of evaporation of water (580 cal/gm)

E = mass vapor flux (gm/cm^2-s)

Q_n = the net radiation flux density (cal/cm^2-s)

ρ = the density of air (about .0013 gm/cm^3)

c = the specific heat of air (about 0.24 cal/gm-C^o)

(e_s-e_a) = the average saturation-deficit of the air (mb)

r_a is diffusion resistance between water and air (s/cm)

S is the slope of the saturation vapor pressure curve (mb/C^o)

g is the psychrometric constant (.66 mb/C^o)

Note: Upon close inspection, this equation is a rearranged version of the formulation used by Kohler et al. (1955).

Initial simplifying assumptions

The term r_a in equation 4a depends on the wind, varying approximately according to the inverse square root. Thus, a range of wind speeds from 1 to 9 m/s alters r_a only between 1.8 and 0.6 s/cm and an intermediate value of 1.2 s/cm is widely representative.

The expressions $(e_s-e_a)/S$ and $(1+g/S)$ may be replaced by equivalent expressions involving temperature values alone:

$$(1 + g/S) = 2\ (1 - 0.0125\ T) \qquad (4b)$$

and

$$(e_s-e_a)/S = (T - T_d) \qquad (4c)$$

where T is the mean temperature for the period of interest and T_d is the dewpoint temperature.

If it is impossible to obtain the net radiation term, Q_n, by direct measurement, it may be adequately estimated by the empirical method discussed by Robinson et al. (1972) or by the relationship with global radiation, Q_s, (which assumes an average daily albedo for water of 0.20):

$$Q_n = (0.55)\ Q_s \quad (\text{cal/cm}^2\text{-s}) \qquad (4d)$$

If it is impossible to measure Q_s directly, it may be adequately estimated from daily maximum and minimum temperatures by the procedures outlined by Bristow and Campbell (1984) or alternatively by the method presented by Linacre (1969):

$$Q_s = T_m / 60\ (100 - A) \quad (\text{cal/cm}^2\text{-s}) \qquad (4e)$$

where T_m is the sea-level equivalent of the measured mean temperature (C) and A is the latitude (degrees) and

$$T_m = T + 0.006\ h \qquad (4f)$$

where h is the site elevation above sea level (meters).

Combining all of the equations into a single expression yields the following for the rate of evaporation from a lake:

$$E_o = ((550\ T_m / (100-A)) + 15\ (T - T_d)) / (80 - T) \quad (\text{mm/day}) \qquad (4g)$$

5. DALINSKY/KOHLER (1971) - Estimation of average monthly evaporation from a map of average annual evaporation depths.

References - Dalinsky, J.S. 1971. The sinusoidal function of regional monthly average relative pan evaporation. Water Resour. Res. 7(3): 677-687.

Kohler, M.A., T.J. Nordenson, and D.R. Baker. 1959. Evaporation maps for the United States. Weather Bureau Technical Paper No. 37. Washington, DC. 17 pp.

-or- U.S. Department of Agriculture, Soil Conservation Service. 1974. Evaporation pond design for agricultural wastewater disposal. Montana Technical Note: Environment No. 7. Bozeman, MT 9 pp.

Buffo, J. and Fritschen, L.J. (n.d.) Direct solar radiation on various slopes for 45 and 47.5 degrees north latitude. Internal Report 51, Coniferous Forest Biome, University of Washington, Seattle 30 p.

-or- U.S. Department of Commerce, Weather Bureau. 1950. Mean monthly and annual evaporation from free water surface for the United States, Alaska, Hawaii, and West Indies. Technical Paper No. 13. Wash., DC. 11 p.

Tested applicability - The Kohler map is a standard reference and is often used as the basis for comparison with other evaporation estimating procedures. The SCS Montana evaporation map was "developed from evaporation data at individual weather stations." Dalinsky actually tested his theory in Israel, but it has sound technical logic and should work well elsewhere.

Accuracy claims - none made

Data requirements

1. Average annual evaporation map for the state
2. Average relative monthly evaporation expressed as a sinusoidal function with an annual wavelength

-- alternatively--

3. The annual sinusoidal function of daily total global radiation at 47.5 degrees north latitude

Procedure

Dalinsky (1971) observed that in a relatively large climatically and physiographically heterogeneous region, the average relative evaporation in each month (expressed as a percentage of the annual total) was equal at all stations and could be expressed as a sinusoidal function. It is therefore possible to estimate the average evaporation rates for every location in the region by

using one parameter (the amplitude of the sine function) and a map of average annual evaporation depths. Dalinsky also suggested that the sinusoidal function of relative global radiation provides an adequate surrogate for the relative evaporation sinusoid when data are not available. As it turns out, this is the situation in Montana.
The sinusoidal function has the form:

% annual evaporation in any month = 8.33 % + A sin B

where,

8.33% is the average relative monthly percentage (100% / 12 mo.)
A is the amplitude which = (maximum % of total - minimum %)/2

and

B is the phase angle which increases by 30 degrees for each month (360 degrees / 12 months = 30 degrees/month)

Similarly, the sinusoidal function of relative potential global radiation has the form:

% annual Rad. on any day = (100/365 sin B

where A is amplitude and B has a value of about 1 degree/day into the period.

The potential global radiation sinusoid for 47.5 degrees north latitude (data from Buffo and Fritschen, n.d.) has a total estimated radiation of 178,802 cal/cm2-yr., for an average radiation of 490 cal/cm2/day or 0.00274 of the annual total. Maximum daily radiation is 834 cal/cm2-day (.00466 of total), minimum daily radiation is 136 cal/cm2-day (.00076 of total), and the amplitude of the relative radiation sinusoid is thus 0.00195 ((.00466 - .00076) divided by 2). Numerical solution and summing over 30 day (monthly) intervals provide estimates of the monthly percentage of annual lake evaporation, which appear in Table 1.

Recent estimates of long-term monthly means of potential evaporation for Montana are not available. Weather Bureau Technical Paper No.13 does, however, provide some summertime estimates for selected Class A stations prior to 1950. Data used for this analysis were from the Bozeman 6W Exp. Farm, which represents the only long-term record from the mountainous part of the state available at that time. Using the same procedure for estimation and accepting Kohler's estimate that 80% of annual evaporation takes place between May 1 and October 31 in most of Montana, monthly percentages of annual lake evaporation were estimated and also appear in Table 1.

Note in Table 1 that there is strong agreement between the estimates made by the two procedures. The Bozeman evaporation data provide slightly higher estimates of summertime loss and slightly lower estimates of wintertime loss than do the 47.5 degree north latitude potential radiation data. Note also, however, that the May - October evaporation estimated from the radiation data is only 71% of annual. This indicates that the summer percentages determined by the Bozeman data might be more accurate. The slightly lower wintertime loss estimates are also reasonable. Evaporation continues from an ice- and snow-covered pond, but additional latent heat is required to go from solid to vapor rather from just liquid to vapor (677 vs. 597 cal/gm).

TABLE 1. Percentage of annual lake evaporation per month determined from A.) relative potential radiation at 47.5 degrees north latitude and B.) relative pan evaporation at Bozeman, MT.

Month	A. % from rad. data	B. % from pan data
April	9	8
May	12	13
June	13	14
July	15	19
August	14	17
September	10	11
October	7	6
November	4	3
December	3	2
January	3	1
February	4	2
March	6	4

EXAMPLE: Both the Kohler and SCS maps of annual lake evaporation indicate that a location near Missoula experiences annual lake evaporation of 35 inches (890 mm). What is the expected (average) evaporation loss for the month of August?

By Method A. - 14% of 890 mm = 125 mm = 4.2 mm/day

By Method B. - 17% of 890 mm = 151 mm = 5.0 mm/day

OBTAINING DATA INPUTS FOR THE CHOSEN METHODS

Measurements

A. Physical - The only physical measurements (size and location) required by the various procedures are latitude, elevation and surface area. The first two may be adequately estimated from USGS 1:24000 maps, the latter by simple survey.

B. Meteorological - The various methods may require measurements or estimates of temperature, vapor pressure, wind, radiation or temperature dependent constants.

1. water temperature - this is needed primarily to estimate saturation vapor pressure at the pond surface (e_s). According to the literature, small shallow ponds usually mix well, resulting in little thermal stratification. The suggested procedure for measuring average daily pond temperature is to use a mercury-in-glass thermometer held a foot or two below the water surface as far as possible from the water's edge. Temperature should be measured in the morning and in the late afternoon and averaged to provide the estimate of mean daily water temperature.

2. air temperature (maximum, minimum, mean daily) - air temperatures may be measured with recording thermographs or with mercury-in-glass max-min thermometers housed in standard weather shelters, preferably situated on the prevailing down-wind side of the pond. A recording thermograph (or preferably a well-calibrated hygrothermograph) will allow determination of a weighted mean daily temperature and maintain a permanent record. The simple average of daily maximum and minimum temperatures will provide an estimate of mean daily temperature.

4. vapor pressure (relative humidity, saturation, ambient) Estimates of relative humidity at mean daily temperature may be made directly with a well-calibrated hygrothermograph. Ambient vapor pressure is an air mass property and therefore changes slowly over the course of a day unless there is an obvious frontal passage. Saturation vapor pressure is determined by temperature. A sling psychrometer used in the morning and afternoon may be used alternatively. In either case, familiarity with psychrometric tables or charts allows the estimate of any vapor pressure measure required by the various procedures.

5. dew point temperature - dew point is a conservative measure of atmospheric moisture content. It may be determined from measurement of ambient vapor pressure. Alternatively, daily minimum temperature, particularly at higher elevations in the summer months, provides a very good estimate of dew point.

6. wind - the mass-transfer method requires an estimate of total wind run during the day. This may be obtained with the use of a relatively inexpensive totalizing anemometer placed preferably at two meters over the pond, but more realistically

with the instrument shelter on the prevailing down-wind edge of the pond.

7. radiation - accurate radiation measurements require expensive radiometers and data recorders or data-logging systems. Net radiation over a pond may be measured with a net allwave pyrradiometer suspended at 1 or 2 meters over the water surface. Global radiation (the total of beam and diffuse shortwave) may be measured with a less expensive pyranometer. Albedo, or shortwave radiation reflectivity, is a function of sun angle. When the sun is nearly overhead, a smooth water surface may absorb up to 95% of incident radiation. Wind-caused ripples or chop can increase the albedo. At very low sun angles, water may reflect up to 70% of shortwave. A reasonable estimate of average daily albedo is about 20%.

C. Other - Some of the parameters in the estimating procedures are constants, such as the psychrometric constant (g), or variables which primarily show temperature dependency, such as the latent heat of vaporization or the ratio, S /(S + g). These temperature-dependent variables have values that may be found in standard textbooks which deal all or in part with evaporation processes and principles (e.g. Campbell (1978) An Introduction to Environmental Biophysics. Springer-Verlag).

Empirical Derivation

Empirical derivation of input data is the process of estimating variables that are difficult or expensive to measure by using simpler, more easily obtained surrogates. In the discussion above, the assumption that dew point temperature is approximated nicely by nighttime minimum temperature at higher elevations in clear weather is a very simple empirical derivation. Those involving the estimation of global or net radiation are less simple and often less exact. As stated previously, an instrument system to acquire radiation data is expensive and delicate. It is extremely advantageous to find methods that can provide acceptable information.

1. Global radiation - Bristow and Campbell (1984) reported a procedure to estimate atmospheric transmittance of extraterrestrial radiation as determined by the daily maximum and minimum temperatures. Their empirical formula, developed from measured solar radiation data in Seattle, Pullman, and Great Falls, reportedly accounts for 70% to 90% of the variation in daily solar radiation. The model has the form:

$$T_t = A \left(1 - \exp(-B \, \text{delta}\overline{T}^C) \right)$$

where T_t is the daily total transmittance, delta$\overline{T}$ is the daily range of air temperature, and A,B,and C are empirically derived coefficients. The coefficients derived for Pullman were not vastly different from those in Helena, so it is reasonable to assume that the Helena coefficients are representative for both eastern and western Montana.

A is a coefficient that represents the maximum atmospheric transmission coefficient. It has a value of 0.77. The B coefficient has a value related to the monthly mean temperature range, delta$\overline{T}$, via the relation:

$$B = 0.036 \exp(-0.154 \text{ delta}\overline{T})$$

The authors found it adequate to hold C constant at 2.4. Therefore the final model for estimating transmisson of solar radiation is:

$$T_t = 0.77 \left(1 - \exp(-B \text{ delta } \overline{T}^{2.4})\right)$$

Values for total extraterrestrial radiation at any latitude on any day can be found in the literature (e.g. Buffo and Fritschen n.d. or Frank and Lee 1966). When multiplied by the value of Tt obtained from the daily range of temperatures, the estimate of total daily global radiation is obtained.

2. Net Radiation - The literature is filled with procedures that allow the estimation of net radiation to various surfaces. The problem is that each surface has different thermal and radiative properties, so there is no universal quantitative approach. Robinson et al. (1972) developed an empirical relationship between daily total net global radiation (incoming - reflected) and daily total net allwave radiation for water surfaces in mid-latitudes. The relationship has the form:

$$Q^* = 0.368 + 0.823 K^* \quad \text{(MJ/m2-day)}$$

where Q* is net allwave radiation and K* is net global radiation.

Global radiation may be measured as previously described or estimated by the Bristow and Campbell (1984) approach. An average daytime albedo of 0.20 is a reasonable estimate for a pond surface.

Extrapolation

Extrapolation of meteorological data cannot be done reliably over large vertical or horizontal distances in mountainous terrain. A possible exception to this is perhaps global radiation over a horizontal surface, particularly in the summer. At no time may wind velocity or run be extrapolated reliably.

Finklin (1983) states that the 24-hour average temperature smooths out local daytime and nighttime effects such that the overall lapse rate (cooling with increase in elevation) in the mountains is about 5.5 degrees C per 1000 m (3 degrees F per 1000 ft.). This is very close to the "climatic lapse rate" cited by Baker (1944) of 3.3 degrees F. per 1000 ft.. Thus, it is possible to estimate mean daily temperatures and, by extension, mean monthly temperatures in the mountains from nearby valley observations. Caution still needs to be observed, however, because mean temperature gradients are complicated by nighttime inversions and resulting "thermal belts." Very often, nighttime minimum temperatures average lower in canyon bottoms than on adjacent ridge tops and valley walls. Therefore, the straight use of a climatic lapse rate will result in an underestimation of average temperatures at higher elevations.

A Comparison of the Five Methods

A data set obtained during field research for a fire behavior modeling study was used to test and compare the various procedures. The data were obtained with standard meteorological instrumentation in a forest clearing 5 km northeast of Missoula during August 1984. The site is located at 5,360 feet and 47 degrees north latitude. These data appear in Table 2 with notes explaining any assumptions or derivation methods.

For additional reference, the Climatological Data Annual Summary, Montana, 1984 (U.S.Dept. of Commerce 1984), reported pan evaporation in August for Hungry Horse Dam at Coram and for the Western Montana Agricultural Research Station near Hamilton of 7.01 " (178mm) and 6.74 " (171 mm), respectively. Accepting the published pan coefficient of .72 for western Montana (Kohler et al. 1959), this corresponds to an estimated average lake evaporation of approximately 125 mm for August 1984 in western Montana.

Results of the Comparison

The evaporation estimates calculated from the climatological data set appear in Table 3.

The Harbeck method consistently provided the lowest daily estimates, and therefore the lowest total monthly estimated evaporation. Judging by any of the values of August evaporation found in the literature, these estimates are at least 50% lower than expected. According to the Climatological Data Summary, August 1984 was almost 3 degrees F. warmer than normal in nearby Missoula. The site (about 2000 feet higher than Missoula) had an average temperature for the month of 64.9 degrees F., while Missoula had an average temperature of 68.4 degrees F. If August does in fact produce 14 to 17% of annual lake evaporation, then normal annual lake evaporation estimated by the Harbeck method would be between 20 and 25 inches. Again, this is at least 50% below expectations.

The exact reason for the failure of the Harbeck estimate is unknown, but it is probably the fact that the pond size chosen for the example is out of the tested applicability range of the model. This in turn affected the estimated mass transfer coefficient, N. The mass transfer coefficient used in this test was also determined from Harbeck's mean relationship between pond area and evaporation in the arid southwest. Ficke (1972) and Meyboom (1967) found mass transfer coefficients nearly an order of magnitude higher than Harbeck's for very small impoundments in Indiana and western Canada, respectively. Therefore, this comparison does not discredit the use of the Harbeck procedure, but emphasizes the importance of developing empirical coefficients on a case-by-case basis as described in the procedure.

The Lamoreux/Kohler, Stewart and Rouse, and Dalinsky/Kohler methods all produce very similar estimates of August evaporation.

TABLE 2: WEATHER PARAMETERS FOR POND EVAPORATION ESTIMATION

(Data taken in Aug. 1984 at 5360 feet elevation, 47 degrees north latitude)

Day	Tmax		Tmin		Tavg		Wind		e	RH %	e	Q	K*	Q*
	F	C	F	C	F	C	mi	km	(mb)	@Tavg	(mb)	(W/m2)	(MJ/m2d)	
1	79	26.1	56	13.3	67.5	19.7	73	117	23.0	54	12.0	463.71	21.7	16.8
2	84	28.9	51	10.6	67.5	19.8	57	92	23.0	45	10.5	536.12	25.1	19.4
3	84	28.9	52	11.1	68.0	20.0	55	89	23.0	41	9.5	520.87	24.4	18.8
4	82	27.8	52	11.1	67.0	19.4	50	80	23.0	46	10.5	484.58	22.7	17.6
5	83	28.3	54	12.2	68.5	20.3	71	114	23.5	42	10.5	515.37	24.1	18.6
6	74	23.3	49	9.4	61.5	16.4	98	158	18.5	40	7.5	474.69	22.2	17.2
7	79	26.1	44	6.7	61.5	16.4	103	166	18.5	39	7.5	526.15	24.6	19.0
8	86	30.0	49	9.4	67.5	19.7	66	106	23.0	38	9.0	536.84	25.1	19.4
9	92	33.3	53	11.7	72.5	22.5	59	95	27.0	35	7.0	554.86	26.0	20.1
10	80	26.7	56	13.3	68.0	20.0	59	95	23.0	34	8.0	453.93	21.2	16.4
11	67	19.4	53	11.7	60.0	15.6	74	119	17.5	59	10.0	264.29	12.4	9.8
12	84	28.9	52	11.1	68.0	20.0	85	137	23.0	44	10.0	521.68	24.4	18.8
13	77	25.0	47	8.3	62.0	16.7	106	171	18.0	36	7.0	486.72	22.8	17.6
14	84	28.9	47	8.3	65.5	18.6	64	103	21.5	35	7.5	517.32	24.2	18.7
15	85	29.4	53	11.7	69.0	20.6	49	79	24.0	43	10.5	515.21	24.1	18.6
16	74	23.3	52	11.1	63.0	17.2	49	79	19.5	65	12.5	408.07	19.1	14.8
17	84	28.9	49	9.4	66.5	19.2	56	90	22.0	41	9.0	507.99	23.8	18.4
18	86	30.0	53	11.7	69.5	20.9	65	105	24.5	41	10.0	529.82	24.8	19.1
19	74	23.3	48	8.9	61.0	16.1	119	192	18.0	53	9.5	469.19	22.0	17.0
20	76	24.4	41	5.0	58.5	14.7	74	119	16.5	40	7.0	503.89	23.6	18.2
21	83	28.3	46	7.8	64.5	18.1	70	113	20.5	35	7.5	505.97	23.7	18.3
22	89	31.7	52	11.1	70.5	21.4	72	116	25.5	36	9.0	522.19	24.4	18.8
23	82	27.8	54	12.2	68.0	20.0	53	85	23.0	51	12.0	475.51	22.2	17.2
24	77	25.0	50	10.0	63.5	17.5	62	100	20.0	53	11.0	452.91	21.2	16.4
25	77	25.0	49	9.4	63.0	17.2	55	89	19.5	45	9.0	452.91	21.2	16.4
26	83	28.3	50	10.0	66.5	19.2	100	161	22.0	36	8.0	487.51	22.8	17.6
27	80	26.7	54	12.2	67.0	19.5	147	237	23.0	46	10.0	460.49	21.6	16.7
28	62	16.7	46	7.8	54.0	12.3	182	293	14.0	46	6.5	272.95	12.8	10.1
29	75	23.9	45	7.2	60.0	15.6	144	232	17.5	35	6.0	460.49	21.6	16.7
30	69	20.6	45	7.2	57.0	13.9	53	85	15.5	53	8.5	391.34	18.3	14.2

NOTE:
1. Wind is total wind run in 24 hours.
2. e is saturation vapor pressure at Tavg.
3. e is ambient vapor pressure at Tavg.
4. It is assumed that Tmin equals dewpoint temperature.
5. Q is the daytime average radiation flux density after Bristow and Campbell.
6. K* is based on Q for a 13 hour daylight period.
7. Q* is net radiation over water using Robinson's equation.

TABLE 3. Comparison of the various evaporation estimating methods using a field research data set. Daily evaporation estimates for August 1984 are presented in mm depth for each method.

ESTIMATING METHOD

DATE	1. HARB.	2. L/K	3. S/R	4. P/L	5. D/K
01	2.7	4.3	5.0	6.7	
02	2.5	5.1	5.8	7.4	S
03	2.6	5.0	5.7	7.4	E
04	2.1	4.5	5.2	4.6	V
05	3.2	5.1	5.6	7.3	E
06	3.7	4.6	4.8	5.9	N
07	3.9	5.1	5.3	6.6	T
08	3.2	5.3	5.8	7.6	E
09	4.0	6.1	6.3	8.6	E
10	3.0	4.5	4.9	6.8	N
11	1.9	2.2	2.7	5.0	
12	3.8	5.3	5.7	7.4	P
13	4.0	4.8	5.0	6.3	E
14	3.1	5.1	5.5	7.3	R
15	2.3	4.9	5.7	7.6	C
16	1.2	3.1	4.2	5.9	E
17	2.5	5.1	5.4	7.4	N
18	3.2	5.3	5.8	7.7	T
19	3.5	4.3	4.7	5.9	
20	2.4	4.4	4.9	6.1	O
21	3.1	4.9	5.3	7.2	F
22	4.1	5.5	5.7	8.2	
23	2.0	4.3	5.2	7.1	8
24	1.9	3.8	4.8	6.3	9
25	2.0	3.9	4.6	6.3	0
26	4.8	5.2	5.2	7.2	m
27	6.6	5.2	5.0	6.8	m
28	4.7	3.1	2.6	4.4	
29	5.7	5.0	4.6	6.1	
30	1.3	2.9	3.8	5.2	
TOTAL	95.0	137.5	150.8	200.4	151.1

1. HARB. is the Harbeck method
2. L/K is the Lamoreux/Kohler method
3. S/R is the Stewart and Rouse method
4. P/L is the Penman/Linacre method
5. D/K is the Dalinsky/Kohler method

The principal difference among the methods is, of course, that the first are capable of making real-time evaporation estimates on a daily basis, and the Dalinsky/Kohler method can provide only expected values of evaporation. Further, the estimates are only as good as the original annual evaporation maps. For the majority of the western part of Montana, detail in the evaporation maps is conspicuously absent.

The Penman/Linacre estimates are consistently the highest and would have been higher still if an average daytime water albedo of 0.20 had not been chosen. Albedo is a function of sun angle, however, and according to the values reported in the literature, 0.20 is a legitimate, realistic estimate of the average during the course of a day. Curiously, the calculated total evaporation of 200.4 mm for August is very close to the average pan evaporation for the region that month. Further investigation of the method reveals annual totals that are approximately 25% to 30% higher than those indicated for corresponding locations on the annual pan evaporation maps. The distinct advantages of the method must be kept in mind. Minimal meteorological requirements are only daily temperatures, latitude and elevation. A logical suggestion is to apply a standard "adjustment coefficient" of 0.75 (which is nearly identical to the accepted pan coefficient of 0.72) to the Penman/Linacre estimates. This would reduce the 200.4 mm August 1984 evaporation estimate to 150.3 mm, which is consistent with the other methods.

State-Wide Evaporation Estimates Using Climatic Normals and the Penman/Linacre Method

The problems encountered in extrapolation of meteorological data in mountainous terrain have already been discussed. This is most likely the reason for the lack of detail in the annual evaporation maps in the Rocky Mountains. The five degrees of latitude from top to bottom of Montana can account for considerable differences in the amount of radiation received annually. Free air lapse rates may allow estimation of mean temperature differences between a valley and an adjacent mountainside, but latitude can compensate for elevation in providing energy for evaporation. Thus, a high elevation site at the southern edge of the state can have far more evaporation than a low elevation site at the northern edge. Further, the indisputable differences in cloud cover and atmospheric moisture content from west to east across the mountains exert tremendous influence on radiation receipt and thus temperature and evaporation regimes.

In an attempt to see if any clear patterns of estimated annual evaporation exist in western Montana, climatological data for 39 NOAA weather stations operating from 1951 to 1980 in Montana west of 108 degrees longitude (U.S. Dept. of Commerce 1982) were used with the Penman/Linacre methodology. These stations with their latitude, longitude, elevation, estimated evaporation and "adjusted" evaporation (75% of the P/L estimate) appear in Table 4. Note that the adjusted annual evaporation estimates range from a low of 839 mm (33 inches) in West Glacier, to a high of 1105 mm (43.5 inches) in Big Timber. Figure 3 is a map of Montana indicating station location and unadjusted P/L evaporation estimates.

Latitude was found to be significantly correlated with estimated annual evaporation ($r = -.614$) in a correlation analysis. As expected, the inverse relationship indicates that evaporation decreases with the distance north. Elevation was also significantly correlated with evaporation ($r = .299$ at 95% confidence). However, note the sign of this "spurious" correlation, which indicates that evaporation increases with elevation; this illustrates the problems discussed in the beginning of this section and suggests that no empirical method for state-wide evaporation estimation will be found and that extrapolation of evaporation estimates should only be attempted over relatively short vertical and horizontal distances.

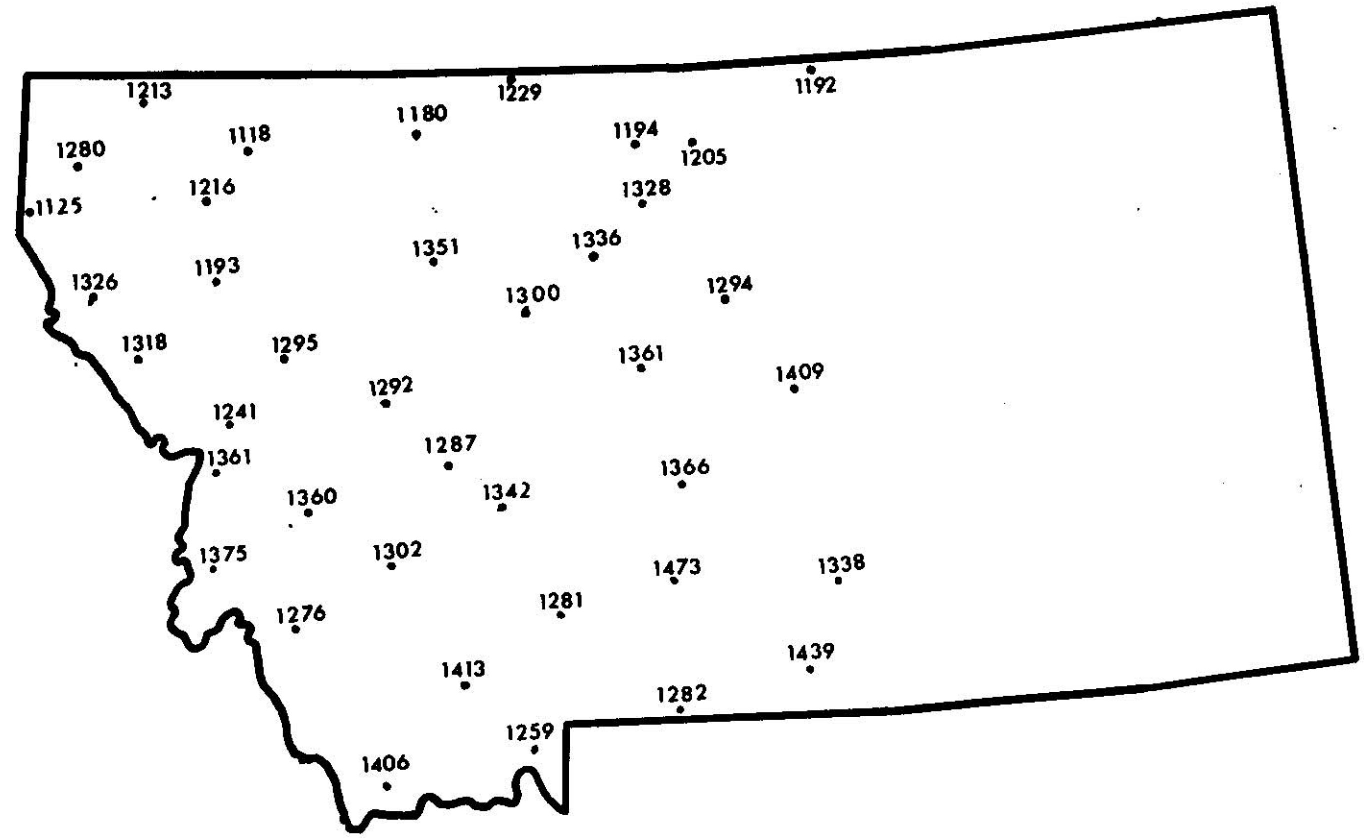
Adjusting Penman/Linacre for elevation

If absolutely necessary, the P/L evaporation estimates for the various stations can be adjusted to nearby locations at different elevations. The T_m variable in the Linacre equation already adjusts average daily or monthly temperature for a lapse rate of 6 degrees C. per 1000 meters. This is close enough to the average lapse rate of 5.5 degrees C. per 1000 meters cited by Finklin (1983) that it is not worth changing. Since dew point is

TABLE 4. Montana weather stations and average P/L estimated and average P/L adjusted pond evaporation rates for the period 1951-1980.

STATION LOCATION	STATION NUMBER	LATITUDE (deg.min)	LONGITUDE (deg.min)	ELEV. (ft)	Eo (mm/yr)	75%Eo (mm/yr)
Big Sandy	770	48.1	110.07	2700	1328	996
Big Timber	780	45.5	109.57	4100	1473	1105
Billings	807	45.48	108.32	3567	1338	1004
Bozeman	1050	45.49	110.53	5950	1281	961
Bridger	1102	45.18	108.55	3680	1439	1079
Butte	1318	45.57	112.3	5540	1302	977
Choteau	1737	47.49	112.1	3945	1351	1013
Cooke City	1995	45.01	109.56	7553	1282	962
Cut Bank	2173	48.36	112.22	3838	1180	885
Darby	2221	46.01	114.1	3880	1375	1031
Ft. Benton	3113	47.49	110.4	2636	1336	1002
Fortine	3139	48.47	114.54	3000	1213	910
Goldbutte	3617	48.59	111.24	3499	1229	922
Grassrange	3727	47.02	108.48	3480	1409	1057
Great Falls	3751	47.29	111.22	3662	1300	975
Harlowton	3939	46.26	109.5	4160	1366	1025
Havre	3996	48.33	109.46	2584	1205	904
Hebgen Dam	4038	44.52	111.2	6489	1259	944
Helena	4055	46.36	112	3828	1287	965
Heron	4084	48.05	116	2240	1125	844
Joplin	4512	48.35	110.47	3360	1194	896
Kalispell	4563	48.12	114.18	2971	1216	912
Libby	5015	48.24	115.32	2080	1280	960
Lima	5030	44.39	112.35	6275	1406	1055
Lincoln	5040	46.57	112.39	4540	1292	969
Missoula	5745	46.55	114.05	3190	1241	931
Philipsburg	6472	46.19	113.18	5270	1360	1020
Polson Kerr Dam	6640	47.41	114.14	2730	1193	895
Seeley Lake	7448	47.13	113.31	4100	1295	971
Stanford	7864	47.1	110.15	4308	1361	1021
Stevensville	7894	46.31	114.06	3370	1316	987
Superior	8043	47.12	114.53	2710	1318	989
Thompson Falls	8211	47.36	115.22	2380	1326	995
Townsend	8324	46.19	111.31	3833	1342	1007
Turner	8413	48.51	108.24	3045	1192	894
Virginia City	8597	45.18	111.57	5776	1413	1060
West Glacier	8809	48.3	113.59	3154	1118	839
Winifred	9033	47.33	109.23	3243	1294	971
Wisdom	9067	45.37	113.27	6060	1276	957

FIGURE 3. Locations of climatic stations and unadjusted P/L estimates of annual lake and pond evaporation.



an air mass property, it is reasonable to assume that absolute moisture content remains constant with elevation. Dew point temperature, however, changes with pressure, thus elevation. In the range of ambient temperatures from 0 to 20 degrees C., T_d has a lapse rate of about 2 degrees C per 1000 meters. The combined effect of the temperature changes with elevation is approximately a 20% reduction in daily, monthly or annual evaporation for every 1000 meter increase in elevation from a nearby reference station.

Adjusting Penman/Linacre for temperature departures from normal

If necessary, the normal monthly evaporation estimates for each station can be corrected for mean monthly temperature departures from normal. At a given elevation, estimated evaporation appears to change by approximately 0.5 mm per degree C. per day for deviation in mean monthly temperature. At a given elevation, estimated evaporation changes about 0.2 mm per degree C. per day for departure from normal in dew point temperature. Evaporation increases with an increase in average temperature and decreases with an increase in average dew point temperature.

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