Chapter Four

Evaporation in Seasonally Snow-Covered Catchments of the Sierra Nevada

by

Al Leydecker and John M. Melack

CHAPTER FOUR
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4. Chapter Four

4.1. An Introduction to Evaporation From Snow

Evaporation is the conversion of liquid water to vapor and its subsequent loss to the atmosphere; sublimation is the transformation from solid to vapor; and transpiration is the process by which plants take up soil moisture and release water vapor. In this report all are called "evaporation."

Evaporation in seasonally snow-covered mountain catchments is difficult to measure and the available methods all have limitations. Scaling up from point measurements in complex terrain is fraught with error, and solving the water balance equation is complicated by the large spatial variability of deposition when snow is the principal component of precipitation (Singh 1992).

The greatest uncertainty in calculating annual evaporation in high-altitude Sierra Nevada catchments lies in determining evaporation from snow. While evaporation during the short snow-free season is limited by quick-draining granitic soils and sparse vegetation, snow covers the ground for much of the year. Past estimates of evaporation from snow have varied from 1% to more than 80% of the total snowpack (Stewart 1982). Given this uncertainty and previous research indicating high evaporation in the Emerald catchment (Marks and Dozier 1992; Kattelmann and Elder 1991), we wanted to independently determine annual evaporation for the catchments in this study.

Agreement between evaporation calculated as the missing term in the water balance equation and evaporation estimated by an independent method would validate the accuracy of the outflow and precipitation measurements and increase confidence in the precision of the chemical mass balances. Most previous studies of high-altitude evaporation in the Sierra Nevada have concentrated on a single site or a single season. By examining a range of catchments and locations over the 4- to 5-year period of this study, an intensive analysis of evaporation would also provide insight into the contradictory conclusions of past research, especially concerning evaporation from snow.

4.1.1. Previous Studies

In 1934 Francois Matthes, referring to evaporation from snow in alpine areas of the Sierra Nevada, stated: "[I've] been impressed again and again by the fact that strongly sun-pitted snow fields above 12,000 feet waste away during the summer without contributing a drop of water to the streams below. . . . It follows that these upper snow fields are in no sense the sources of streams and are to be excluded from any estimates of runoff based on snow surveys." In an experiment designed to test Matthes' conclusions, Robert Sharp (1951) located an isolated snow bank on bedrock at 12,200 feet on the eastern side of the Sierra Nevada. After measuring its volume, he installed a weir to record snowmelt. He noted, "These measurements showed that approximately 99 percent of the wastage ran off through the weir, leaving about 1 percent for evaporation and other losses. . . . These results strongly suggest that evaporation occupies a relatively minor part in the ablation of snow in areas above 12,000 feet." It would be difficult to find two more diametrically opposed conclusions. The importance of evaporation from snow continues to be argued.

Energy balance studies on melting snow reflect this disagreement (Table IV-1). While many studies found evaporation to be low, others have concluded the opposite. About a third of the studies show a mass gain through condensation rather than evaporative loss. Measured evaporation varied widely, up to a maximum of 2.35 mm per day, and even the most recent studies reflect a wide disparity.

Stewart (1982) measured decreases in snow depth while monitoring snowmelt (using 7 buried snow lysimeters) over a 6-day period at the end of April 1981 on Mammoth Mountain (elevation 2930 m). He concluded that 25 to 40% of the snowpack decrease was due to evaporation, a loss of 3.3 to 6.7 mm per day. Analyzing periodic measurements of temperature and humidity and average wind speeds, he felt that meteorological patterns capable of producing this rate of loss existed at least 20% of the time.

At Emerald Lake in the central Sierra Nevada, Marks and Dozier (1992) used the mean-profile method to calculate evaporation from snow at two meteorological stations during 1986. They reported a total loss of 451 mm at the lake outlet and 537 mm on an exposed ridge (Table IV-2). Losses during snowmelt were 201 mm (9% of the snowpack) and 284 mm (13%), respectively. Continuing this study, Kattelmann and Elder (1991) reported total evaporation from snow as 18% in 1986 and 33% in 1987 (1987 was a drought year with a third of the 1986 snowpack and a short melt season). They concluded that "approximately 80 percent of the annual evaporative loss was from sublimation off of snow . . . at rates of 1 to 2 mm of water per day, with higher values on dry, windy days."

Anderson (1976), summarizing work done at the Central Sierra Snow Laboratory (CSSL) near Lake Tahoe, California, reached a contrasting conclusion. Using similar mean-profile methodology, he calculated average evaporation rates of about 2 mm per month during the accumulation season, and 4 mm per month during snowmelt (Table IV-3). While his study was conducted in forest clearings, where lower wind speeds account for some of the difference, his overall results were lower than those at Emerald Lake by a factor of more than 30.

"Pan studies" are experiments where snow is placed in a container of some type, weighed and set on or level with the snow surface. Over time, a careful account is kept of changes in mass; missing mass representing evaporative loss. Over the past 80 years, the numerous pan studies conducted in the Sierra Nevada have also shown a wide range of measured evaporation.

Sharp (1951) melted snow blocks under what he called "natural conditions," and reported that only 2.7% of the mass was lost to evaporation. West (1962), in one of the more careful experiments of this type, calculated average annual evaporation from snow for a three year period at CSSL at 13 mm in forested areas and 38 mm in small clearings. During 5 days in March he performed experiments in a large open area (a frozen lake surface ~2.4 km × 1.6 km) and found that daily evaporation varied from 0.3 to 1.6 mm. The total 5-day loss was equivalent to a monthly evaporation rate of 26 mm. Finally, using weighted averages for different terrain types, he estimated annual areal evaporation for a 10 km^{-2} catchment (elevation 2290 m) at 51 mm, or < 3% of the snowpack. Beaty (1975), working at 3600 m in the White Mountains (across the Owens Valley from

Sharp's site), reached a different conclusion: that 50 to 80% of the snow-pack was lost to evaporation.

Pan studies have to be viewed with reservation. Solar energy penetrates the snow-pack to some depth and container shape, size and material affect melt and evaporation. How the container was filled and whether it was set into or on top of the snow-surface (where it may generate increased turbulence) are important factors. For example, Beaty placed very small snow blocks (10 × 5 × 4 cm) on plastic sheets in hollowed out depressions in the snow-surface, collecting melt in ceramic coffee cups. The blocks melted in 2 to 3 days, at which time the melt-water was measured and weighed. Questions could be raised about the influence of the sun on plastic and the difference this may have made in the energy balance of the thin snow blocks, and how much melt-water had evaporated from plastic sheet and cup. This is not meant to single out Beaty's work for criticism (his is an enterprising and well-written paper), but to emphasize that all pan studies engender similar doubts.

4.1.2. Theoretical Considerations

4.1.2.1. Turbulent Transfer

Evaporation requires a vapor pressure difference between the surface and the overlying air, and wind generated turbulence. A third factor, atmospheric stability or instability, dampens or reinforces the upward flux of water vapor. In a stable atmosphere, upward motion (buoyancy), is restricted by warmer, less dense air overlying a colder, heavier layer. Instability results when the lower layer is warmer and less dense than the air above. Increases in either temperature or vapor pressure decrease air density, but the effect of a change in vapor pressure is very small and temperature is the primary determinant of buoyancy. High winds passing dry air over hot moist surfaces maximize evaporation.

Winter conditions in the Sierra Nevada are not usually conducive to high rates of evaporation. Mild days, with temperatures often above freezing, produce low vapor pressure differences: typical mid-winter values are 1 to 2 mb. At night, lower temperatures further diminish the small differences. In contrast, desert lakes can have summer daytime differences of 20 to 30 mb. With the snow-surface temperature usually lower than air temperatures, stable atmospheric conditions often prevail. Only high winds or unusual conditions could account for high evaporation.

During snowmelt, atmospheric vapor pressures increase with rising temperature while the snow-surface is constrained to the saturation vapor pressure at 0° C. Stable atmospheric stratification becomes the norm. As the season progresses, atmospheric vapor pressure may exceed that of the snow-surface, producing condensation. High evaporation at this time seems improbable.

4.1.2.2. The Energy Balance

Evaporation from snow is energy intensive, requiring 677 cal g^{-1} compared with the 80 cal g^{-1} needed for melt. High rates of evaporation can only be driven by large energy inputs onto the snow surface. The energy balance equation for the upper-most snowpack layer can be written as:

$$R_S + L_{in} - L_{out} + H - E = \Delta H_S + \text{Melt} \qquad (1)$$

where:

R_S is the short-wave solar radiation entering the surface,
L_{in} is long-wave (infrared) radiation directed at the surface,
L_{out} is long-wave radiation emitted from the snow surface,
H is sensible heat (sensible because it can be sensed by a change in temperature),
E is latent heat or evaporation (latent because it cannot be sensed by a change in temperature, only by a change in state, e.g., snow-to-vapor),
ΔH_S is the change in snow pack heat storage, and
Melt is the energy utilized for snowmelt.

All of the terms have units of energy flux per unit of time per unit area.

During winter, R_S, ΔH_S and Melt are either negligible or of much lower magnitude than the other terms and can be set to zero, simplifying Equation (1) to:

$$L_{in} - L_{out} + H = E \qquad (2)$$

The net long-wave radiation flux (L_{in} - L_{out}) is small, therefore, sensible heat provides most of the energy for winter evaporation from snow; most research shows that the winter sensible heat and evaporative fluxes are usually of similar magnitude but opposite sign (Bernier and Edwards 1990; Marks and Dozier 1992; Cline 1995). A flux of sensible heat requires a temperature difference (a surface temperature below air temperature if evaporation is to occur) and wind. Temperature differences between the air and snow-surface during the accumulation season are usually small, inferring low evaporation.

Another consequence of Equation (2) is that night-time snow-surface temperatures will typically be lower than the air temperature. On clear nights there is usually a net long-wave radiation loss from the snow-surface; a loss that has to be balanced by a sensible heat gain, i.e., snow-surface temperatures colder than the air. Any evaporation must be fueled by a further gain of sensible heat. The balance between decreases in evaporation and long-wave radiation and increases in sensible heat with decreasing surface temperature usually insures a snow-surface temperature significantly below that of even sub-zero air, as well as diminished evaporation during this time.

During snowmelt, when the snow-pack turns isothermal, ΔH_S can assumed to be zero and the net long-wave radiation can be considered negligible. Equation (1) can be reduced to:

$$R_S + H = \text{Melt} + E \qquad (3)$$

Both net solar radiation and the sensible heat flux become increasingly larger as the season progresses. If the incoming energy was divided equally between evaporation and melt, for every 1 mm of snow that evaporated 8.5 mm would melt (the ratio between the heats of sublimation and of fusion). Given spring melt rates between 10 and 20 mm per day, evaporation rates of over 2 mm per day are plausible. However, other considerations limit this possibility. Evaporation is a surface phenomenon, but the absorption of solar radiation takes place within a volume: radiation penetrates the snow to depths of up to a meter, with energy absorption exponentially decreasing with depth (Oke 1987). Less than

25% of the absorbed solar radiation is available for surface evaporation. And energy availability is not the sole criteria. In the absence of a favorable vapor pressure difference, all of the available energy will be used for melt. From an energy perspective, the potential for high evaporation from a melting snow-surface is limited.

4.1.2.3. Exceptional Evaporation

High rates of evaporation from snow can occur. Chinook winds in the eastern Rockies (caused by eastward-flowing air masses losing moisture while ascending the western-slope and then gaining heat by compression during the eastern-slope decent) travel at high speeds and are dry and comparatively warm. As such, they cause extremely high evaporation. Golding (1978) reported average potential rates, in Alberta, of 1.2 mm per day for the 19 Chinook days in 1975, and 2.0 mm per day for 20 days in 1976. His maximum calculated rate was 10.4 mm per day for a 4-day period in January 1976 (with an average recorded wind speed at an adjacent site of 21 m s^{-1}).

Blowing and tree captured snow also have extreme rates of evaporative loss. The literature contains many examples (see Schmidt and Gluns 1992): almost 20% of a avalanche deposit lost to evaporation while being re-distributed 600 m down-wind; an evaporative loss >50% for snow picked up by wind and re-deposited ~2 miles away, a scenario not unfamiliar during typical blizzard conditions; 33% of the seasonal snow-fall lost through evaporation from snow deposited on tree branches in a conifer forest.

These situations offer extraordinary opportunity for evaporation. Small flakes or particles suspended in the wind stream present a large surface area to the effects of turbulence and differential vapor pressure. Snow clumped on branches has similar characteristics: small volumes, with high ratios of surface area to mass, subject to wind and evaporation on all sides.

However, these examples are not typical of conditions in Sierran catchments. Hence, the Marks and Dozier (1992) results at Emerald Lake seem anomalously high and puzzling. The conditions that seem necessary for high evaporation, very high winds and/or dry air, do not appear to often exist in the Sierra Nevada. Longacre and Blaney (1962) analyzed 13 years of evaporation pan data collected in the west-side San Joaquin drainage. Table IV-4 summarizes results from the two lowest elevation lakes in the study, the only lakes with year-round records. The lowest monthly evaporation occurred in January and was less than half of the Marks and Dozier (1992) Emerald evaporation for the same month, i.e., the reported high-elevation evaporation from snow was double the measured low-elevation evaporation from water. The highest lake evaporation occurred in July, a time of high temperatures and low humidity. The Marks and Dozier estimate of evaporation from snow at Emerald in July was ~25% of these high rates, even though vapor pressure differences were more than an order of magnitude lower and often unfavorable. Winds along an open lake-shore should not have been substantially different from those occurring in the higher altitude catchment.

4.1.3. The Evaporation Study

Evaporation was estimated using four methods: (1) we selected an areal (area) evaporation model compatible with the available data and used it to estimate annual evaporation for each of the catchments; (2) meteorological data (air temperature, humidity and wind speed, measured at short intervals and then averaged and recorded for periods of up to an hour) from high-altitude sites located in or near the catchments were used to model point evaporation from snow with the mean-profile equations; (3) annual catchment evaporation was estimated using the evaporative concentration of conservative ions, i.e., chloride and, for some catchments, sulfate; and (4) the water balance equation was used to calculate annual evaporation for each catchment.

The evaporative concentration estimates and the water balances at Emerald and Spuller, where our most accurate measurements of precipitation and outflow were made, were used to judge the acceptability of the areal evaporation model. The model passed this test and model estimates of evaporation from snow were in general agreement with the meteorological station mean-profile results. Evaporation was estimated with the point and areal models at six of the catchments. Lost Lake was not included because of problems with the outflow record and the difficulty of accurately measuring the spring snowpack because of topography and the predominant influence of wind-blown deposition. We felt that no nearby meteorological station adequately represented conditions in this basin. However, the evaporative concentration results showed that evaporation at Lost was similar to losses in the other catchments.

The next part of this chapter presents the areal evaporation model and the evaporative concentration estimates; it includes an extensive discussion of the model, its strengths and weaknesses, and an evaluation of the model's overall acceptability. Following that, the mean-profile point model used at the meteorological stations is described and the results presented and analyzed. The last section summarizes our conclusions on evaporation from snow-covered catchments in the Sierra Nevada, and discusses some possible directions for future analysis and work.

4.2. The CRAE Model

4.2.1. Introduction

CRAE stands for "complementary relationship areal evaporation", and the model is based on Bouchet's (see Brutsaert and Stricker 1979; Morton 1983; Parlange 1992) concept that "potential evaporation" is dependent on the areal (the actual evaporation that occurs over an area) evaporation. (Potential evaporation is defined as evaporation occurring from a wet surface: a surface small enough to leave unchanged the characteristics of the over-passing air, but large enough to minimize the edge effect and other inaccuracies of un-corrected evaporation pan measurements.) This reverses the normal assumption of potential evaporation as the independent variable— and areal evaporation as some fraction of the potential evaporation based on moisture availability. According to Bouchet, if moisture is limiting, the energy that would have gone into evaporation is used instead to heat and dry the over-passing air, increasing the potential evaporation. Thus the relationship between the two is an inverse one: higher areal evaporation results in lower potential evaporation and visa versa.

The relationship is also complementary. At one extreme, when moisture is not limiting, the two are equal. As an area dries and areal evaporation decreases, this decrease is matched by an increase in potential evaporation. At the other extreme: when the areal evaporation is zero, potential evaporation becomes twice the wet-surface value. (Wet-surface evaporation is defined as evaporation occurring from an extensive landscape completely saturated or covered with water.) The validity of the complementary nature of this relationship is based on experimental evidence (Morton 1983; Morton 1991). CRAE model was formulated and designed by V. I. Morton, and as used in this study is documented in his 1983 paper; much of the empirical evidence used to justify the complementary relationship is given or referenced there. The final model is the culmination of a process begun in the early 1960s and documented in Morton's papers of 1965 1971 1974 1976 and 1978.

We were attracted to Morton's model for the following reasons:

1. It has been approximately validated for many regions of the world: one region of particular interest was northwestern Canada, where the evaporation regime is likewise dominated by lengthy seasonal snow cover.

2. The primary inputs are air temperature and dew point averaged over the time interval used in the model. This interval can be varied from 5-days to a month. Such data were available for Emerald and Spuller lakes and the relatively long time-step would enable us to fill data gaps using regression relationships with nearby meteorological stations. The simple input data requirement of the model would also allow us to estimate evaporation for catchments without meteorological data: by using standard temperature and dew-point lapse rates to transfer data between nearby sites of similar topography.

3. The model does not use wind speed as an input. Air temperature and humidity measured at a single point can be used to estimate conditions over a wide area, but wind speed cannot be "generalized" in a similar fashion; nor are there acceptable methods for transferring wind speed between stations or catchments. Thus a model that did not directly use wind speed had appeal.

4. There are no "tunable" parameters used in the model, and there is no calibration for local conditions. Thus the model is falsifiable, it either works or it does not. Its validation, without calibration, in many different environments gave us confidence that it could be applied to the catchments in this study.

5. No knowledge of the soil or plant community is required, and no measurement of soil moisture or temperature used; an attraction, since some of this type of data were unavailable.

4.2.2. CRAE Model Theory and Operation

In the CRAE model areal evaporation (E_{areal}) is equal to twice the wet-surface evaporation (E_{wet}) minus the potential evaporation (E_p):

$$E_{areal} = 2E_{wet} - E_p \qquad (4)$$

Potential evaporation, as used here, it is the same quantity expressed by the "Penman" equation:

$$E_P = [\Delta /(\Delta + \gamma)]\, R_{net} + [\, \gamma/(\Delta + \gamma)]\, E_a \qquad (5)$$

where:

Δ is the slope of the saturation vapor pressure vs. temperature curve,
γ is the psychometric constant,
R_{net} is the net radiation,

E_a represents the drying power of over-passing air, which Penman equated to a wind function multiplied by the difference between the saturation (e_a^*) and actual (e_a) vapor pressures, i.e., $E_a = f\{u\}[\, e_a^* - e_a \,]$.

In the model, wet surface evaporation is derived from the Priestley-Taylor equation (Priestley and Taylor 1972) which assumes that as a wet surface becomes more extensive, moisture in the air reaches a limiting value where no additional drying power remains:

$$E_{wet} = \alpha\, [\Delta /(\Delta + \gamma)]\, R_{net} \qquad (6)$$

In effect, the second term in Equation (5) drops out and alpha (α) can be considered a correction to the remaining term. Alpha is usually given a value of 1.26, experimentally determined over large wet or water surfaces. Morton increases α to 1.32 to account for "enhanced evaporation over typically rougher land areas."

Although the model uses the concepts expressed by Penman and Priestley-Taylor, the actual equations differ. Potential evaporation is calculated from the energy balance and vapor transport equations applied to a "suitably sized" wet area:

$$E_P = R_T - \gamma\, f_T (T_P - T) - 4\varepsilon\sigma\, (T)^3 (T_P - T) \qquad (7)$$

$$E_P = f_T (e_P^* - e_D) \qquad (8)$$

where:

R_T is the net radiation calculated with the assumption that the ground surface temperature is the same as that of the air,

f_T is the vapor transfer coefficient, consisting of an empirical constant modified to account for changes in vapor transport with elevation and atmospheric stability,

T is the air temperature,

T_P is the equilibrium temperature, that temperature at which equations (7) and (8) will give the same value for E_P,

e_P^* is the saturation vapor pressure at the equilibrium temperature,

e_D is the saturation vapor pressure at the dew point temperature, and

$\varepsilon\sigma$ is the product of the emissivity of the wet surface (ε) and the Stefan-Boltzmann constant.

Equation (7) calculates potential evaporation as the difference between net radiation and sensible heat loss. The second term is based on the usual assumption that the sensible heat and vapor transport coefficients are equal. The last term adjusts R_T for the correct wet-surface temperature, i.e., the equilibrium temperature, by calculating T_P as a two-term Taylor series expansion around the air temperature. Equation (8) is simply the vapor pressure difference between the air and the wet surface, multiplied by the transfer coefficient. These equations are iteratively solved for T_P, and the potential evaporation is calculated from this value using Equation (7). The model has no term for changes in subsurface heat storage. It assumes that over a sufficiently long period, 5-days to a month, short term changes will average out, and the overall magnitude of this term is insignificant in comparison with the other fluxes. Using similar reasoning, lag times and other short-term inputs are ignored.

Morton's version of the Priestley - Taylor equation is

$$E_{wet} = b_1 + b_2 [\Delta_P/(\Delta_P + \gamma)] R_{net} \qquad (9)$$

Coefficients b_1 and b_2 were empirically determined from measurements of potential evaporation during rainless periods at desert stations (recall that when areal evaporation is zero the potential evaporation is $2 \times E_{wet}$). The derivation of the constants was constrained so that the calculation of E_{wet} by either Equation (9) or (6) (assuming that $\alpha = 1.32$) would give the same answer. One remaining problem is the appropriate temperature at which to calculate Δ (γ is relatively insensitive to temperature changes). Since the equilibrium temperature (T_P) should not appreciably vary with the availability of water (and has been previously calculated in Equation (7)) it is used here. The net radiation includes the long-wave loss from a wet-surface at the equilibrium temperature. (Recently, Morton [1991] has suggested holding the relative humidity constant, somewhere between 82 and 85%, and using Equations (7) and (8) to calculate wet-surface evaporation.)

Computation in the model begins with the input of latitude, altitude and average annual precipitation for the catchment meteorological station. From this a pressure correction factor (the ratio of the average barometric pressure at the station to that at sea level) and the average snow-free clear-sky albedo are calculated. The inputs for each time interval are: (1) the average air temperature and (2) dew-point, (3) the ratio of "observed-to-maximum-possible sunshine duration" (the model calculates average incoming solar radiation but direct measurements can be used instead), (4) the fraction of snow-covered area, and (5) the number of days in the interval. In this study we used the "period," an interval of approximately a third of a month: defined as the first 10 days, the second 10 days, and the remaining days of each month. Although the model cannot be used for intervals of less than 5-days due to the inability to account for short term flux variations and lag times, daily output can be approximated by calculating both single and 5-day evaporation and then proportionally adjusting daily values to match the 5-day totals.

Clear-sky global radiation is calculated with the assumption that the basin is a flat plane at the given latitude. Morton used a difference of < 1 mb between the actual and saturation vapor pressures of the air to indicate seasonal snow cover. Seemingly designed for western Canada, this criteria did not apply in the more temperate conditions of the Sierra Nevada. Instead, a simple albedo model (Winther 1993) was substituted, along with the fraction of snow-covered area, to calculate average albedo during this time. The ratio of observed-to-maximum-possible sunshine duration is used to correct the clear-sky values of solar radiation, albedo and atmospheric long-wave radiation. After solving for potential and wet-surface evaporation, areal evaporation is calculated using the complementary relationship.

4.2.2.1. Meteorological Data for the Model

No meteorological data were collected at four of the catchments and data for the other two were incomplete. Nearby meteorological stations with relatively long and complete records were used to construct or supplement data for the model. The meteorological stations at Emerald and Spuller lakes record hourly data, and supplemental stations with similar instrumentation and recording intervals were sought to derive regression equations for meteorological variables between stations. Four stations with suitable data were available. Table IV-5 contains a brief description of each of the meteorological stations used in this study, along with station location, elevation, and the length and quality of the data record; Table IV-6 describes the meteorological instruments used at each station..

The catchments were divided into two geographical categories: the "eastside," consisting of Spuller, Ruby and Crystal lakes on the eastern side of the Sierra Nevada; and the "west-side," which included Emerald, Pear and Topaz lakes in the upper Marble Fork of the Kaweah drainage (the Tokopah Basin) of Sequoia National Park. Separate sets of regression equations were used to extend and repair meteorological records for the two catchments with meteorological stations. Records for the un-instrumented catchments were constructed using data from the geographically closest station, modified by standard lapse rates for temperature and dew-point. Gaps in the data records of the supplemental stations were repaired using regression relationships developed between stations.

4.2.2.1.1. The West-side Regression Equations

On the west-side, the supplemental data source was a NOAA meteorological station located at Wolverton. The station, installed mainly for the collection of wet and dry deposition data, is approximately 5 km west of Emerald Lake, at an altitude 600 m lower. It is located within a forest clearing bordering a large meadow. The station topography at Emerald is quite different: near the inlet of a glacial cirque lake in an area almost devoid of trees. This difference precluded the use of a simple lapse rate factor to account for the elevation change.

The time span chosen for comparing the two stations was September I 1991(the first or "I" period in September, i.e., the first 10 days, identified in Figures IV-1, IV-2 and IV-3 as period number 25) to December III 1993 (the third or "III" period in December, i.e., the last 11 days, identified in these figures as period number 108).

The equations, the square of the correlation coefficients (r^2) and the standard error of the estimates for the regressions between meteorological variables are shown in Table IV-7. Figure IV-1 illustrates the relationships between the two stations and compares actual with modeled temperature, dew-point and total incoming solar radiation at Emerald Lake. For temperature the r^2 value was 0.96. Surprisingly, the actual temperatures at the two sites were very close, the average lapse rate was 0.18° C/km as compared with standard values of 6 to 7° C/km (Running et al. 1987). Colder nights at Wolverton, in the valley bottom, and higher sunny-day temperatures in the Emerald cirque helped reduce the elevation differential.

The dew point r^2 was 0.92 with an average lapse rate between the two stations of 3.85° C /km. This was higher than the typical value of 1.5° C /km (Running et al. 1987) because of higher humidity and the increased frequency of night-time fog in Wolverton's forested valley environment. The relative humidity measurements showed little correlation ($r^2 = 0.27$).

The incident solar radiation r^2 was 0.92, but the probable error increases appreciably with increasing values (Figure IV-1f). Therefore, the very good correlation is probably an artifact of low magnitude winter values at both sites, reinforced by a greater similarity of winter conditions compared with those of summer. This effect becomes more pronounced for "S," the ratio of observed-to-maximum-possible sunshine duration

(Figures IV-2a and IV-2b). This parameter is a holdover from older efforts to quantify the intensity of solar radiation. The equation for the conversion between S and G, the now more common radiometer-measured incident solar radiation, is (Morton 1983)

$$S = 0.053G/(G_O - 0.47G) \qquad (10)$$

G_O is the clear sky incident solar radiation. The relationship of G_O vs. time for each station was derived from a smooth curve drawn along the upper boundary of the highest two or three daily radiometer totals for each period using all the years of record. Since G_O is a fixed value for a given period and station, a regressed value of S can be derived by either (1) directly comparing S values between the two stations, or (2) by calculating S from G using Equation (10). The second method resulted in a relatively poor r^2 of 0.38. Even though incident solar radiation correlates well between the two stations, the large standard error in the estimate is magnified in the calculation of S; direct correlation between S values produced better results ($r^2 = 0.64$). Although not as well correlated as temperature and dew-point, the relationship between stations was statistically significant and usable.

Since the Pear Lake catchment is adjacent and similar to the Emerald catchment, Emerald evaporation estimates were used for both. Topaz Lake is located 4.5 km to the northeast of Emerald Lake, at an elevation 415 m higher. Although the catchments are dissimilar in shape and aspect, they share similar sub-alpine characteristics and typical lapse rates chosen from the literature (Running et. al. 1987) were applied to the Emerald temperature and dew-point data for use at Topaz: the corrections were -2.70 and -0.52 °C, respectively. Incident solar radiation at Topaz was assumed to be the same as at Emerald.

4.2.2.1.2. The Eastside Regression Equations

The situation on the eastside was more complicated than that on the west. The record at Spuller Lake begins at about the same time as that at Emerald (September 1991) but is less complete due to operational difficulties. The other two lakes are located to the southeast of Spuller (Crystal, 45 km; Ruby, 74 km) and the latitudinal difference was expected to impact meteorological patterns. In addition, the Crystal catchment is located directly adjacent to Mammoth Pass. The pass is the lowest elevation gap in this area of the Sierra Nevada and beyond the pass lies a direct path to the west along the canyon of the upper San Joaquin: this route acts as a funnel for major westerly storm systems and results in the deposition of significantly more snow on Mammoth Mountain then in areas to the north or south. This was also expected to influence any correlation with the other catchments.

Three meteorological stations were used to supplement Spuller data:

1. Mammoth — Located on the flank of Mammoth Mt. in a level and open area. The meteorological record begins in September 1989 but has significant gaps. Measured parameters include temperature, relative humidity, wind speed and various measures of incoming and outgoing radiation. During the early years data was collected at 60 minute intervals (i.e., measurements are taken every 2 or 3 min. during the interval and then averaged and recorded). More recently the measurement interval had been shortened

to 15 minutes. The station is located 40 km southeast of Spuller Lake at an elevation 190 m lower. It is 5 km north of Crystal Lake at approximately the same elevation.

2. SNARL — Located at the Sierra Nevada Aquatic Research Laboratory, a research station maintained by UCSB in Long Valley. The record begins in October 1989. Temperature, relative humidity, wind speed and incoming solar radiation are recorded at 60 min. intervals and the record is essentially complete. The major problem with using this station is its topographic distinctiveness: Long Valley runs along the base of the Sierra Nevada and is part of the western edge of the high desert Basin and Range province extending eastward into Utah. The meteorological station is in a good location, flat and relatively open. It is approximately 2 km from the mouth of Convict Canyon and seems to be predominately affected by the up and down canyon meteorological pattern: large terminal moraines shield the site on the north and south. Thus measurements may be more representative of Convict Canyon than of the broader and more arid valley. Geographically, the station is 53 km southeast of Spuller Lake and 960 m lower (it is 18 km east of, and 770 m lower than, the Mammoth station).

3. Eastern Brook Lake (EBL) — Located just above the lake in a small clearing on a wooded slope, the station is surrounded by a tall stand of Red Fir. Established in the early 1980s, active maintenance ended before the records at the other sites began. Only in March 1993 was the station placed back into operation. Therefore, the data record useful to this study is limited to about 15 months. Counterbalancing these disadvantages is its geographic location in the same canyon (Rock Creek) as Ruby Lake, 3 km to the southwest and 260 m higher in elevation. Mammoth is 34 km to the northwest and 240 m lower, while SNARL is 21 km to the north and 1,010 m lower.

The regression equations, along with the respective r^2 values and standard errors are shown in Table IV-7. In most cases, the eastside meteorological variables were as statistically well correlated as those for the west-side. Again, regressions for S had lower r^2 values for reasons explained earlier. Overall, the eastside G and S regressions were better than those for Emerald. The poorest air temperature correlations were with SNARL due to a wider range (the low valley location had the coldest winter nights and warmest days) and an asymmetric seasonal variation (similar winter temperatures, with colder valley nights balanced by colder mountain days, and warmer valley summer temperatures). For dew-point, the Spuller data correlated better with SNARL, than with Mammoth. Although both the Spuller and SNARL meteorological stations are located at the mouth of valleys, a more likely explanation is atypically higher humidity at Mammoth due to the proximity of Mammoth Pass and its moisture laden winds.

Figures IV-2 and IV-3 show scatterplots for some of the eastside regressions, along with the variation in modeled and measured meteorological parameters over time. For data used in the model, gaps in the Spuller temperature record were filled by regressed Mammoth measurements or, if unavailable, with SNARL data. The opposite procedure was used for dew-point because of the lower standard error of the SNARL relationship. Mammoth data was used without change to estimate evaporation at Crystal. The EBL data set was extended with regressed Mammoth data and transferred to Ruby using temperature and dew-point corrections of -2.70 and -0.52 °C, respectively.

Snow-covered area depletion curves were developed from aerial photos (taken at the start of snowmelt each year and at subsequent times during 1993 and 1994), supplemented by subjective estimates during on-site visits. A series of three curves were derived for each catchment: for low, average and high water years.

4.2.3. Results and Discussion

Modeled evaporation for each of the catchments for water years 1990 through 1994 (e.g., water year 1990 begins on the first of October 1989, and ends on the last day of September 1990) are shown in Figures IV-4 and IV-5 (Figure IV-5.1 presents an alternate view of the same data). Evaporation is shown in units of millimeters (mm) of water equivalent lost during the period, i.e., the depth of water per unit area evaporated over the ~ 10-day interval of the period. Modeled evaporation had a general pattern of low values throughout winter and during the first part of the snowmelt season. Only as snow-covered area began to appreciably decrease, did evaporation rise. As the snowmelt season progressed, increased air temperatures and solar radiation accelerated melting and water availability, enhancing evaporation from warming snow-free surfaces. Past the evaporative peak, the lessened availability of free water from a diminishing pack and the exhaustion of soil moisture led to a rapid decrease. Large expanses of rock and course grained soils in the catchments limit the interval of moisture availability.

Comparing modeled results for five of the basins for each of two water years better illustrates these inferences (Figure IV-6). Both years have low winter values. High evaporation during this season of low vapor pressure gradients is possible only with high wind speeds. Since CRAE model assumes that the vapor transfer coefficient is independent of this parameter, the results are not unexpected.

In 1992, a drought year, snowmelt began at the end of March in most of the catchments. Past March, low evaporation continued for about a month (Figure IV-6a). At Ruby and Spuller lakes, where melt was delayed and prolonged due to higher elevation, the interval of low values continued well into May. The abrupt increase in evaporation after the first month or month and a half of snowmelt seems to coincide with a reduction in snow-covered area to 75 or 80%. From this time on, a steady reduction in snow-covered area decreased the average catchment albedo and drove the model's accelerating evaporation.

In 1993, late snow and rain storms complicated the situation, but even here the onset of rapid evaporation is significantly delayed (Figure IV-6b). At Emerald, the beginning of accelerated evaporation occurred around the middle of June, again at a time when snow-covered area was reduced to ~75%.

The overall patterns appear reasonable. Compared with 1992, evaporation in 1993 was lower and peaked later as the above average snowpack reduced total losses and delayed a significant reduction in snow-covered area until late in the summer. The higher-elevation basins, Ruby and Spuller, have lower evaporation compared with the lower-elevation Emerald and Crystal. In 1992, the evaporation rate at Emerald decreased after the middle of July, in agreement with an early disappearance of the snowpack at the end of June and the subsequent drying of basin soils. In 1993, the evaporative decrease occurs closer to the end of August, correlating with extended snowmelt and lower late-season temperatures. High evaporation at Topaz is most likely an artifact of the lapse rates used in transferring Emerald meteorological data to this basin. While there may be some physical justification for higher evaporation at Topaz, e.g., a higher percentage of soil cover and lake area and a more southerly aspect, the model responds only to meteorological data. Some combination of under-predicted temperature or over-predicted dew-point would decrease the computed potential evaporation and lead to an overestimate of areal evaporation.

Significant rainfall at Emerald produced surges in evaporation, and the evaporation rate followed temperature trends during the predominately snow-free season (Figure IV-7a). With snow cover, model evaporation was dominated by low energy input into the surface and was less sensitive to temperature and dew-point fluctuations. During most of the winter, potential, areal and wet-surface evaporation were roughly equal (Figures IV-7b and IV-7c), the snowpack providing a good approximation of an extensive wet-surface where potential equals wet-surface evaporation. During snowmelt, potential evaporation increased with higher temperatures, but with no corresponding increase in areal evaporation until significant ground surface was exposed. As the summer progressed, potential evaporation increased with the drying of the ground surface, reducing areal evaporation.

4.2.3.1. Evaluation of CRAE Modeled Evaporation

In evaluating the accuracy of the CRAE model's evaporation estimates we emphasize results from the two study catchments with meteorological stations and outflow weirs; these catchments, Emerald and Spuller lakes, also had the most intensive snow-water-equivalent measurements of the spring snowpack and offer the best data-sets for model evaluation.

We calculated evaporation from the snow surface at the Emerald and Spuller meteorological stations with the Brutsaert (1982) mean-profile equations. This method, the Brutsaert mean-profile model, is explained in detail in the next part of this chapter. The models differ in two fundamental ways. The mean-profile model is a point model, calculating evaporation at a point, while the CRAE model is an area model, estimating the average evaporation over a large area. The other difference is that the CRAE model estimates evaporation from all surfaces in the catchment, while the mean-profile model calculates only evaporation from the snowpack. Thus results can only be compared if the point evaporation measurements are assumed to approximate average evaporation, and only for periods when the basin is almost completely snow-covered. Given these caveats, the mean-profile results can be used to evaluate the CRAE model estimates of winter and early spring evaporation.

Evaporation estimates from both methods for winter months at Emerald are within the same low range (Figure IV-8a). As expected, results rapidly diverge as the snowmelt season progresses: evaporative loss from the snowpack diminishing to zero (often showing a net gain from condensation during the last month) while the CRAE model records increasing evaporation. The comparison at Spuller (Figure IV-8b), while it shows the same general agreement, is not as good. A possible explanation is atypical higher wind at the meteorological station; of the 8 sites analyzed with the mean-profile model, the Spuller station recorded the second highest average wind velocities. The Emerald and Spuller point evaporation values were, respectively, the lowest and highest calculated for alpine and sub-alpine sites in this study. Thus they arguably span the range of typical winter values.

4.2.3.1.1. Comparison of Model Estimates and Water Balance

The CRAE model estimates can also be evaluated by comparison with catchment water balances. The difference between annual stream outflow and precipitation can be attributed to changes in water storage and evaporation. Over a sufficient number of years the net change in storage can be considered zero, and evaporation becomes the sole loss. For water years 1990 through 1994 at Emerald, the average annual modeled evaporation differs by only 24 mm from that calculated from the water balance, an error equal to less than 2% of annual precipitation. Results for water years 1992 through 1994 at Spuller were within 1%.

Defining "water balance residual" as that quantity needed to balance water year input and output, the respective residuals for the 5 years (1990 through 1994) at Emerald were +11, +6, +19, -21 and +22 percent of annual precipitation (Figure IV-9a and see Chapter Two). A positive residual represents an excess of output over input: either underestimating snowpack and rain or overestimating outflow or evaporation. The 1993 snowpack was 240% of average; at the close of the water year, snow remained in the basin and water still flowed in the outlet. Hence, a large part of the negative residual in 1993 represents missing outflow. In the other years, which were all below average or drought years, streamflow stopped prior to the end of September. If we consider basin groundwater storage to be either limited or annually recharged by snowmelt, a sizable part of the positive error is probably due to overestimating evaporation since (1) rain is a minor input (< 10% of annual precipitation), (2) the snow surveys (involving hundreds of measurements over a relatively small catchment) should have accurately measured snow-water-equivalent, and (3) the outflow weir measurements were reasonably precise (±5 to 10%). The consistent positive error during these years reinforces this conclusion: the summation of errors for the other water balance components is unlikely to be similarly biased. Given the low estimate of evaporation from snow, the error must lie in over estimating evaporation from the ground surface and vegetation.

The water balance residuals for the same years at Spuller were respectively, +3, +19, -3, -22 and -8 percent (Figure IV-9b). During the first two years, as was the case with Emerald, the model may be overestimating evaporation given the positive water balance residuals. Spuller is a small catchment, yet the outlet flows year-round indicating substantial ground water storage. Much of the negative residual for 1993 is explained by carry-over of unmelted snow into 1994. The agreement between input and output for water years 1992 may be fortuitous and some exaggeration of summer evaporation is possible.

4.2.3.1.2. Comparison of Model Estimates with the Konstantinov Model

We compared the CRAE model with one based on different concepts: a model developed by A. I. Konstantinov (1966) which also avoids using wind speed as an input variable. Starting with the equations for turbulent mass transfer (the same methodology used in the mean-profile model), he empirically developed relationships for the change in temperature and vapor pressure between the surface and standard meteorological instrument heights over various time intervals. These differentials vary with both air temperature and the season, and are used with wind speed profiles selected on the basis of average air temperature and vapor pressure. The model, based on voluminous amounts of data from much of the northern hemisphere, can estimate evaporation for various time periods. After some favorable initial comparisons between Konstantinov's model, using one hour averaged meteorological data and the mean-profile results, we used it to estimate period (~10-day) evaporation at Emerald and Spuller.

A comparison between the Konstantinov and CRAE model evaporation estimates for water years 1993 and 1994 cannot be considered a success (Figure IV-10). Although both models agree on low evaporation from the winter snow surface and there is further agreement during the early snowmelt period, they differ radically in summer evaporative loss. The Konstantinov model estimates total water year evaporation as only a third to a half of the CRAE model totals. While this might initially be seen as confirmation that the CRAE model is over-predicting, a closer look at the Konstantinov results identifies some serious problems.

In 1993, the Konstantinov model had maximum evaporation occurring at the end of May; in 1994, maximum evaporation occurred at the beginning of May. At both of these times the snow cover was nearly complete and the mean-profile calculations showed evaporation from snow to be minimal. The model estimated evaporation as decreasing rapidly at the beginning of summer, when snowmelt and free water were abundantly available and warming temperatures should be maximizing evaporation. Evaporation in 1994 is less than that in 1993, whereas the reverse is more probable. The difference between models seems to lie in the relatively low summer temperatures and high humidity found in these mountain catchments. To CRAE model, this is evidence of low potential evaporation, thus high areal evaporation: to Konstantinov's model, a very small temperature and vapor pressure gradient, thus low turbulent transfer, i.e., low evaporation.

4.2.3.1.3. Comparison of Model Estimates with Evaporative Concentration of Solutes

If, over a period of years, ionic export from a catchment equals depositional input, we can infer that the ion is not a product of weathering and that its involvement in ion exchange or biologic processes has attained equilibrium. Ions exhibiting this behavior can be termed "conservative," and differences in volume weighted mean concentrations (VWM) between input and outflow over some specified time period can be used as a measure of evaporative concentration (Classen and Halm 1996). Defining "evaporative factor" (ef) as the fraction of total precipitation that is lost to evaporation:

$$ef = 1 - \frac{[VWM]_{in}}{[VWM]_{out}} \qquad (11)$$

Analysis of the chemical mass balances (4 to 6 years of data for each catchment) has shown that chloride was conservative for all of the study catchments and that sulfate was conservative for three (Figure IV-11a). Determining evaporation by chemical concentration should be robust for these catchments since estimates of the VWM concentrations of inflow and outflow can be obtained without an accurate determination of the mass balance: the methodology depends only on relative concentrations, not on relative volumes.

In the alpine and sub-alpine regions of the Sierra Nevada, ionic concentrations in precipitation are dominated by contributions from snow, the chemistry of which shows little variation over the spatial extent of these small basins (Melack et al. 1996). Sampling the spring snowpack just prior to the onset of melt gave good estimates of the annual input concentrations: the ionic contribution of the small amount of annual rainfall was minor, i.e., the annual input was insensitive to measurement errors in rainfall volume and concentration. While discharge was highly variable, the ionic concentrations of chloride and sulfate in discharge varied only slightly (concentrations varied by a factor typically less than 2, while discharge varied over 2 or 3 orders of magnitude); thus, the VWM outflow concentrations were not sensitive to errors in discharge measurement.

However, there were problems. Chloride concentrations in precipitation and discharge are low, in the range of 1 to 4 μeq/l, and the chloride determination had a standard error of ± 0.5 μeq/l. Given no bias, the large number of outflow samples, taken during high flow when most of the ionic mass was exported, reduced the probable error. Determining the input concentration, dependent on far fewer samples, was less precise.

The annual hydrologic cycle may have been influenced by "carryover" from previous years. Carryover is prevalent during high-snow years when residual snow or appreciable groundwater flow extends past the end of the water year. Summer and early fall rains, which typically produce little or no runoff, are also likely to result in the carryover of ions (Stoddard 1995). These problems are increased in our data since 1990-1992, severe drought years, were followed by an extremely wet year 1993. Although this raises questions about the validity of calculating evaporation for any one year, an average evaporation rate based on the analysis of a number of years of data minimizes the carryover problem and also reduces the possible error in the VWM inflow determination. Combining data from different catchments over a number of years should give a reasonable estimate of regional evaporation.

Another problem was that no accounting of snow-free dry deposition was available (spring snowpack samples include ions from dry deposition on snow). Since some studies comparing snow precipitation chemistry with snowpack chemistry in California and Colorado have shown that winter dry deposition is negligible (Williams and Melack 1991; Campbell et al. 1991), the assumption has heretofore been that summer dry deposition is also relatively unimportant. However, recent work at Loch Vale (Clow and Mast 1995) showed that summer dry deposition on granite was 10 to 20% greater than wet deposition from rain. The increase in dry deposition was attributed to greater airborne transport of particulates from dry summer soils, and more efficient scavenging of atmospheric gases and particulates by snow-free surfaces. If summer dry deposition in the alpine region of the Sierra Nevada is appreciable, failure to include it exaggerates the evaporation factor; the exaggeration is greater in drought years, less important for high water years.

Analysis of the annual VWM input and output concentrations of chloride and sulfate (sulfate was conservative only for Pear, Crystal and Topaz) for all catchments and years yields an average evaporation rate (ef × 100) for this region of the Sierra Nevada of 38 ± 5% (95% confidence interval) of annual precipitation (Figure IV-12a; $r^2 = 0.67$, $p < 0.0001$). The average rate as determined by the CRAE model for the same catchments and years was 37%. Limiting the comparison to the catchments of the Upper Marble Fork gave results of 35 ± 6% for chemical concentration (Figure IV-12b; $r^2 = 0.66$, $p < 0.0001$) vs. 36% for CRAE model. The y intercepts of the regressions were not significantly different from zero.

The CRAE model average annual evaporation for each of the study catchments compares favorably with the evaporative concentration and water balance estimates (Figure IV-11b). Only the Crystal water balance and Topaz CRAE estimates seem anomalously high, probably due to ground water leakage in the first case and the previously mentioned lapse rate problem in the second.

4.2.3.2. Evaluation of the Reconstructed Meteorological Data

CRAE model evaporation estimates, based solely on modeled meteorological data, were compared with estimates using actual catchment data for water years 1993 and 1994 at Emerald and Spuller (Figure IV-13). At Emerald, the modeled data (data derived from the Wolverton regression equations) overestimated annual evaporation by about 7.5% ($r^2 = 0.86$), with most of the error occurring in June and July. At Spuller, using modeled values based on both SNARL and Mammoth, annual evaporation was overestimated by 10% ($r^2 = 0.93$) and 1% ($r^2 = 0.94$), respectively. We concluded that no appreciable error had been introduced by using the regression equations to repair and reconstruct records at these sites.

Standard lapse rates were used to transfer data to the Topaz and Ruby catchments. A temperature lapse rate presupposes a linear relationship between two points, with a slope of 1.0 and an intercept equal to the lapse rate times the elevation difference:

$$T_B = 1.00(T_A) + (\text{lapse rate in }^\circ\text{C/km})(\Delta \text{ elev. between A and B}) \qquad (12)$$

Measured air temperature lapse rates in this study varied considerably from the typically assumed -6 or -7 °C/km (Running et al. 1987): from + 0.35 (Wolverton-to-Emerald) to -12.01 °C/km (SNARL-to-Mammoth). The slope coefficients for equations based on Wolverton and SNARL were significantly different from 1.0. Dew-point lapse rates varied from -0.15 (SNARL-to-Mammoth) to -23.54 °C/km (Mammoth-to-Spuller), compared with standard values near -1.5 °C/km. Only the SNARL-to-Spuller and Mammoth-to-Spuller equations had slope coefficients of ~1.0. Lapse rates should be assumed only in the absence of viable alternatives. The probable overestimate of evaporation at Topaz is a cautionary example of using standard rates.

Ruby provides another example. In 1993, modeled evaporation slowly increased from April through the end of June, at a time when snow cover was nearly complete and actual evaporation must have been decreasing (Figure IV-6b). A low ridge separates the Eastern Brook meteorological station, used to provide data for Ruby, from Little Lakes Valley. The valley, a long broad glacial trough, is filled (as its name implies) with numerous lakes. It acts as a conduit for prevailing westerlies and is subject to high humidity and early and rapid snowmelt. The meteorological at Eastern Brook is warmer and more humid than at Ruby, and standard lapse rates probably minimized the difference, exaggerating evaporation.

A similar problem may be responsible for the error in the simulation using only regressed meteorological data at Emerald (Figure IV-13c). Most of the error was concentrated in precisely that interval, June and July, when Wolverton meteorological is no longer influenced by a melting snowpack, but Emerald's still is.

4.2.3.3. Basin-area and the Meteorological Data

Morton (1983) does not explicitly address the minimum watershed size appropriate for CRAE model. Ideally, the area should be large enough to develop temperature and humidity patterns that accurately characterize areal evaporation. Over-passing air is modified by changing surface conditions; change forces development of a transition zone reflecting a mixture of the characteristics of both the old and new surfaces. Only below this zone is the air "fully adjusted" to the new conditions. The fully adjusted layer grows at a rate of 1 m vertically for every 100 to 300 m along the surface (Oke 1987). Thus, over a kilometer of fetch may be required for meteorological instruments placed 4 to 7 m above the surface to fully reflect change.

Two questions determine if the meteorological data used in the model was appropriate. Does the air carry a signature representing catchment conditions when it reaches the meteorological station? If not, how close did the previously traversed terrain resemble the area of interest?

A possible problem at Emerald and Spuller is that the recorded temperature and humidity were developed externally and, more importantly, were not characteristic of the catchment. Failure to fully satisfy this criteria may explain the possible overestimation of summer evaporation. The prevailing winds at Emerald are westerly, up the Marble Fork from the extensive forested area around Wolverton. Moisture acquired over this area may not be totally lost by the time Emerald is reached; and higher humidity in the model results in higher areal evaporation. It is suggestive that the water balances for Spuller showed less evidence of overestimated evaporation. Winds at Spuller are mainly from the northwest, passing over similar alpine terrain prior to reaching the catchment. Even though the Emerald basin is larger, the temperature and humidity recorded Spuller may be more representative of catchment elevation and topography.

4.2.3.4. Effect of Wind-speed on Evaporation

Evaporation rates are usually proportional to wind speed, and CRAE model's use of a constant vapor transfer coefficient, independent of this variable, opens it to criticism. Morton (1983) offers 3 justifications: (1) the vapor transfer coefficient "increases with increases in both surface roughness and wind speed, and wind speeds tend to be lower in rough areas than in smooth areas"; (2) the vapor transfer coefficient "increases with increases in the instability of the atmosphere and this effect is more pronounced at low wind speeds than at high wind speeds"; and (3) routine wind measurements are more indicative of the peculiarities of instrument placement than a measure of average conditions, and using wind speed may introduce more error than assuming a constant coefficient.

Evaporation is directly proportional to the friction velocity, defined as the rate-of-change in wind speed with increasing height. Wind speeds typically increase with height, from zero (at some small distance just above the surface) to a maximum value at the top of the surface boundary layer. Above the boundary layer the surface exerts no influence. Rough terrain increases the height of the boundary layer and decreases the rate-of-change for a given wind speed. As the rate-of-change decreases so does evaporation. A simpler explanation is that a rougher surface exerts more "drag" or resistance on the over-passing air, lowering wind speed. Thus rougher terrain will decrease wind speed while increasing the vapor transport coefficient, maintaining consistent vapor transport.

Atmospheric instability enhances evaporation. It is typically caused by temperature stratification: colder air lying on top of warmer, more buoyant, air. Higher wind speeds decrease temperature stratification by turbulent mixing and are usually a precondition for a neutral atmosphere, one without buoyancy. Thus changes in stability will tend to reduce the effect of wind speed differences, again equalizing vapor transport: high instability, i.e., a high vapor transport coefficient, associated with low wind speeds; and higher wind speeds with a lower vapor transport coefficient. Morton's third assumption is un-arguably valid in mountain areas.

For a given surface, a constant vapor transfer coefficient implies constant wind speed. Although representing different wind environments, the wind speed distributions and mean values for five of the meteorological stations used in this study are similar (Figure IV-14). If the extremes of wind-swept or wind-sheltered locations are ignored (Wolverton, recording wind in a forest clearing at an elevation below the average tree height, and Mammoth on a wind-swept open ridge) the range of variation is considerably narrowed and using a constant vapor transfer coefficient seems defensible, especially for longer time periods and increasing catchment size. Whether the value of the vapor transfer coefficient used in the model is appropriate is another question: our results seem to indicate that it was.

4.2.3.5. Snowmelt Energy Balance

The CRAE model neglects sub-surface energy fluxes, assuming that over the computation period short-term fluctuations will average out, and that long-term changes will not be significant. Although usually valid when the sub-surface flux represents conductive transfer with bare ground, these assumptions are untrue during snowmelt. Typically, at least 70 to 80% of the energy received by a melting snowpack goes towards melt and not towards evaporation (Table IV-1). In geographical areas where snowmelt occupies only a short interval, incorrect modeling of the energy balance at this time may be unimportant. For the catchments in this study, however, the snowmelt season extends over two to five months, and the failure to account for this energy sink is troublesome.

Why then does the model produce realistic results during this interval? The answer is that the model almost never sees the surface as snow-covered during snowmelt. When the average air temperature is above zero, the wet-surface for which the model computes potential and wet-surface evaporation consists of liquid water; the only adjustment for the presence of snow is a decrease in net radiation from higher albedo. As net radiation is the principal determinant of wet-surface evaporation, E_{wet} remains low. In contrast, potential evaporation, dependent on sensible heat transfer, increases with increasing seasonal temperatures. As a consequence, areal evaporation ($E_{areal} = 2E_{wet} - E_p$) remains very low until the albedo appreciably increases.

The actual situation is different. Assume that potential evaporation can be estimated by a corrected evaporation pan measurement. Measurements during snowmelt, when average air temperatures were above zero, would give relatively high values of potential evaporation: higher than values calculated with the model since the solar radiation input would be greater, the water-surface albedo being far lower than the average catchment value used in the model. For a melting snowpack, the actual wet-surface evaporation would be very low: much lower than that calculated with the model since most of the net radiation is used for snowmelt and little is available for evaporation. Under these circumstances the areal evaporation would fall to near zero. A conclusion substantiated by the mean-profile results. The model therefore, is probably overestimating evaporation for this period; but since the difference between very low and near zero evaporation is small, its approximation of the situation remains realistic.

4.3. The Mean Profile Model

4.3.1. Introduction

Methods for calculating turbulent energy fluxes, using short term measurements of mean temperature, humidity and wind speed, are variously referred to as flux-profile, similarity, aerodynamic or mean-profile. We used equations from Brutsaert (1982) and the term "mean-profile."

4.3.2. The Mean-Profile Equations and Assumptions

The equations, for meteorological data from a single-level and including corrections for atmospheric stability, are:

$$u^* = kU \Big/ \left[\ln\left(\frac{Z_m - d_0}{z_0}\right) - \Psi_m\left(\frac{Z_m - d_0}{L}\right) \right] \tag{13}$$

$$H = \left(T_a - T_s\right) \rho c_p k u^* \Big/ \left[\ln\left(\frac{Z_t - d_0}{z_{0t}}\right) - \Psi_t\left(\frac{Z_t - d_0}{L}\right) \right] \tag{14}$$

$$E = \rho k u^* \left(q_a - q_s\right) \Big/ \left[\ln\left(\frac{Z_v - d_0}{z_{0v}}\right) - \Psi_v\left(\frac{Z_v - d_0}{L}\right) \right] \tag{15}$$

where:

E is evaporation (kg m^{-2} s^{-1}),
H is the sensible heat flux (W m^{-2}),
T_a and T_s are the air and snow-surface temperatures (°K),
U is wind speed (m s^{-1}),
Z is the meteorological instrument height above the surface (m) and the subscripts indicate the measurement, m for wind speed, v for water vapor and t for temperature,
c_p is the specific heat of air at constant pressure (1005 J kg^{-1} $\cdot$ $°K^{-1}$),
k is von Karman's dimensionless constant (0.4),
q_a and q_s are the specific humidity of the air and snow-surface,
u^* is the friction velocity (m s^{-1}),
z_0 is the momentum roughness length (m) and the subscripts t and v indicate roughness lengths for sensible heat and water vapor (m),
ρ is the density of the air (kg m^{-3}),
Ψ, the "psi" function, is the correction for atmospheric stability, and
m, t and v are the respective subscripts for momentum, sensible heat and water vapor.

Positive values signify a flux directed at the snow-surface. The value of L, the Obukhov stability length (m), is given by:

$$L = \frac{u^{*3}\rho}{gk\left(\frac{H}{T_a c_p} + 0.61E\right)} \qquad (16)$$

When surface irregularities are numerous and extensive, the base reference for measurement of Z lies between ground-level and the average height of the roughness elements (h_0); and the correction for this difference is called "displacement height" (d_0). Brutsaert (1982) recommends $d_0 = 0.66h_0$ as an approximation, but for open snow-surfaces d_0 was much smaller than the measurement error in Z, and this correction was ignored.

When L is infinitely large, the atmosphere is neutral and log-linear relationships for wind speed, temperature and vapor pressure extend throughout the lower surface sub-layer. L is positive for stable conditions, and a decreasing magnitude indicates increasing stability; the closer the value to zero the more effectively turbulence is dampened and the greater the flux reduction. Negative L indicates instability, and its magnitude might be envisioned as the height at which buoyant forces first begin to dominate over those generated by shear (Brutsaert,1982). Shear forces are at a maximum near the surface, and a decreasing magnitude of L indicates increasing dominance by strong buoyant forces.

The Ψ corrections for a stable atmosphere were taken from Imberger and Patterson (1990):

$$\begin{aligned}
&\Psi = -5(Z/L); && \text{for } Z/L < 0.5 \\
&\Psi = 0.5(Z/L)^{-2} - 4.25(Z/L)^{-1} - 7\ln(Z/L) - 0.825; && \text{for } Z/L \text{ from 0.5 to 10} \qquad (17) \\
&\Psi = \ln(Z/L) - 0.76(Z/L) - 12.093; && \text{for } Z/L > 10
\end{aligned}$$

These are the same equations given in Katul and Parlange (1992), corrected for continuity at the designated boundaries. Equal corrections are applied to all three fluxes, i.e., $\Psi_m = \Psi_t = \Psi_v$. Although some authors propose different functions for momentum and sensible heat or water vapor, we followed the recommendation of Brutsaert (1982): "that the differences between the functions in the stable case are less meaningful than those for instability, and until significant agreement and consensus develop, the assumption that the three are equal is conservative."

The corrections for an unstable atmosphere are from Brutsaert (1992). The Ψ function for specific heat and water vapor is different from the functions for momentum.

For sensible heat and water vapor:

$$\Psi_t\left(\tfrac{Z}{L}\right) \;\; or, \;\; \Psi_v\left(\tfrac{Z}{L}\right) = 1.2\ln\left[\frac{\left(0.33 + y^{0.78}\right)}{\left(0.33 + y_{Ov}^{0.78}\right)}\right] \qquad (18)$$

For momentum:

$$\Psi_m(y) = 0, \qquad \text{for } y < 0.0059 \qquad (19)$$

$$\Psi_m(y) = 1.47\ln\left[\frac{\left(0.28+y^{0.75}\right)}{\left(0.28+(0.0059+y_0)^{0.75}\right)}\right] - 1.29\left[y^{1/3} - \left(0.0059+y_o\right)^{1/3}\right],$$

$$\text{for } 0.0059 \le y \le 15.025 \qquad (20)$$

$$\Psi_m(y) = \Psi_m(15.025), \qquad \text{for } y > 15.025 \qquad (21)$$

where:

$y = - Z/L$, $y_0 = - z_0/L$ and z_0 is the respective roughness length for either sensible heat and water vapor or momentum.

Equations (13) through (16) are solved by iteration. The Ψ functions are initially set to zero, i.e., a neutral atmosphere is assumed, and the fluxes calculated with Equations (13), (14) and (15). The Obukhov length is computed using these values in Equation (16), and the Ψ corrections for the next iteration are calculated. When the difference between successive values of L reaches an acceptable limit, the iteration can stop. For certain combinations of meteorological parameters (stable conditions and light winds), the iterative procedure may fail to close, and a maximum value for the Ψ correction, or a limit on the number of iterations, can be imposed.

4.3.2.1. Roughness Length

The momentum roughness length (z_0) of snow varies from 0.0001 to 0.02 m (Moore 1983; Morris 1989). Generally, lengths around 0.001m are used for "smooth" snow, ~0.005 m for surfaces described as undulating, sun-cupped or hummocked; terrain obstacles or protruding vegetation further increase the value. With one exception, a roughness length of 0.005 m was used throughout the analysis.

The exception was at Wolverton where surrounding trees dictate a wind profile characterized by high z_0 and significant vertical displacement of the reference level from the ground (d_0). Roughness lengths for forested areas are typically within a range of 0.4 to 1.2 m (Brutsaert 1982; Parlange and Brutsaert 1989). At Wolverton, z_0 was set at 0.5 m and the displacement height calculated as $d_o = 4.9z_0$ (Brutsaert 1982).

Theoretical considerations indicate that roughness lengths for sensible heat and vapor should be one to two orders of magnitude smaller than for momentum (Moore 1983; Brutsaert 1982). As an experimental example, Pluss and Mazzoni (1994) calculated z_0 values of ~10^{-3} m (0.0019 m during the accumulation season, and 0.0044 m during snowmelt) compared with z_{0t} values ~10^{-6} m. In this study, the roughness lengths for sensible heat and water vapor were assumed to be a tenth of that for momentum ($0.1z_0$).

4.3.2.2. Snow-surface Temperature

Because the meteorological data were collected at a single-level above the surface, the second reference point for the solution of Equations (14) and (15) is the surface itself, or more precisely, a distance equal to the roughness length above the surface. In the absence of snow-surface temperature measurements, a simple model accounting for the observations that night-time snow-surface temperatures are 4 to 6 °C below the air temperature and that snow-surface temperatures will lag changes in air temperature (Bernier and Edwards 1990; Marks and Dozier 1992), was used:

1800 - 1945 hrs:	2 degrees below the air temperature
2000 - 0400 hrs:	4 degrees below the air temperature
0545 - 0600 hrs:	2 degrees below the air temperature
at all other times:	$T_{snow} = T_{snow(n-1)} + 0.5(T_{air(n)} - T_{snow(n-1)})$ (22) if $T_{snow} > 0^{\circ}$ C, set $T_{snow} = 0^{\circ}$ C

Modeled snow-surface temperatures were compared with data from the Central Sierra Snow Laboratory at Lake Tahoe (Figure IV-15a). As snowmelt progresses, snow-surface temperatures are constrained at zero an increasing percentage of the time, and the mean-profile results are less dependent on the surface temperature model (Figure IV-15c).

4.3.2.3. Specific Humidity

Specific humidity can be expressed as (Saucer 1955):

$$q = \frac{0.622 f_w e}{\left(p - 0.378 f_w e\right)} \qquad (23)$$

where f_w is a correction factor applied to the universal gas constant, and e and p are the vapor and barometric pressures (mb). Since f_w is near unity, and e is small compared with p, the equation can be simplified as

$$q = \frac{0.622e}{p} \qquad (24)$$

Barometric pressure was not recorded at any of the meteorological stations; instead, average sea-level pressure, corrected for altitude at each site, was used as a constant (Morton 1983). Atmospheric vapor pressure was calculated from the air temperature and relative humidity. Similarly, snow-surface vapor pressure was assumed to be the saturation vapor pressure for ice at the snow surface temperature. Saturation vapor pressure was calculated as (Saucer 1955):

$$e_{sat} = 6.11 \times 10^{\left(aT/_{T+b}\right)} \qquad (25)$$

where temperature is in ° C, and a and b are constants: for water vapor over water, a = 7.5 and b = 237.3; over ice, a = 9.5 and b = 365.5.

4.3.2.4. Density of Air

Since an average barometric air pressure was used, the equation for air density (Saucer 1955) was simplified to:

$$\rho = 0.3484\left(\frac{p}{T_a}\right) \qquad (26)$$

where ρ is the density of moist air (kg m^{-3}), and T_a the air temperature (°K).

4.3.3. Results and Discussion

The mean-profile estimates of evaporation at Emerald and Spuller lakes, SNARL and Mammoth Mountain are shown in Figures IV-16 and IV-17 (Figure IV-18 presents the same data in a different format). Each month was divided into 3 periods, i.e., the first

10 days, the second 10, and the remaining days, and period evaporation (in mm of water equivalent) is the sum of hourly calculated evaporation over each period (except at Mammoth Mountain where 15 minute data were summed). Gaps indicate missing meteorological data; proximity of maintenance personnel and climate severity were the key determinants of record completeness.

The general pattern was of higher evaporation during mid-winter, when vapor pressure differences were greatest, followed by continuously declining evaporation throughout snowmelt. Near the end of snowmelt, condensation of water vapor onto the snow-surface often matched evaporative loss and sometimes exceeded it. With the exposure of significant amounts of bare ground, atmospheric vapor pressures appreciably increased: a combined effect of high evaporation from moist snow-free areas, and increased air temperatures due to seasonal trends and sensible heat exchange. Higher atmospheric moisture content decreased the vapor pressure gradient over the remaining snow. Evaporation rates throughout April and into May were highest in 1993, when an above average snowpack maintained nearly complete snow coverage.

Condensation was prevalent at Mammoth Mountain in the late spring and summer. Although the persistence of snow into summer increases the probability of condensation, the primary cause was high humidity carried by prevailing westerlies. These winds are funneled through the lowest altitude pass in this region of the Sierra Nevada, and by May are warm enough to carry significant moisture. For example, vapor pressures in March and April of 1993 were normally distributed with median values ~3.7 mb; in May, the median vapor pressure increased by 50% to 5.58 mb, with a distribution skewed towards higher pressures. (The maximum vapor pressure exerted by a melting snow surface is 6.11 mb and the high median pressure indicates frequent condensation.) In contrast, median vapor pressure at Spuller in May was ~3 mb. Winds at Spuller are mostly down-canyon northwesterlies, relatively depleted in moisture after passing over the Sierra Nevada crest (~3800 m).

With the exception of April in both 1991 and 1993, monthly evaporation at SNARL (Figure IV-19a) varied between 5 and 20 mm. Higher April evaporation in 1991 was caused by two days of "Chinook" conditions: dry, high-speed winds. Average evaporation over the two days was approximately 4 mm per day. In April 1993, two separate Chinook events, averaging 3 mm per day, similarly increased evaporation.

Wind speed and atmospheric vapor pressure differences accounted for the variation in monthly evaporation between meteorological stations in 1993 (Figures IV-19b and IV-20b). Wolverton, with the lowest mean wind speed and highest mean vapor pressure, had the lowest evaporation. Trees surrounding this site dampen turbulence and maintain high humidity above the snow surface. At Emerald, higher wind speed and lower vapor pressure increased evaporation, but the total remained low, i.e., 61 mm, November through July. The other meteorological stations, exposed to higher winds, had losses two to three times higher: total evaporation at Mammoth, SNARL and Spuller was 115, 129, and 169 mm, respectively. Although wind speed at SNARL was less than half that at Mammoth, evaporation was higher due to the April Chinook intervals and lower mean vapor pressure. The highest evaporation occurred at Spuller, consistent with the driest air and the second highest wind speeds. Annual differences also reflected changes in mean wind speed and vapor pressure, e.g., at SNARL (Figure IV-20a).

Station mean wind speed in 1993 (December through May) varied from 0.74 to 4.6 m s^{-1}; mean wind speeds at Emerald, Spuller and SNARL were within ~1 m s^{-1} of each other. The Wolverton mean was about half, and Mammoth two and a half times, the Emerald average. Except for Wolverton, the frequency distributions of hourly wind speed (15 minutes at Mammoth) are similar (Figure IV-21a). Frequency distributions of temperature were also similar (Figure IV-21c); SNARL and Wolverton are slightly skewed towards higher temperatures due to lower elevation. Only at Wolverton and Spuller were the vapor pressure distributions noticeably different, with higher and lower values, respectively (Figure IV-21b).

Monthly and cumulative totals of wind speed, vapor pressure and wind direction at Emerald and Spuller are directly compared in Figure IV-22. The cumulative wind-speed modes are similar, but the Spuller distribution had a greater positive skew and higher mean. The highest wind speeds were recorded at Emerald (13.8 vs. 10.6 m·s^{-1}); up-canyon winds often have higher near-surface velocities due to compressed streamlines. Prevailing westerlies at Spuller are funneled downslope through the outlet gorge, oriented NW to SE. The Emerald cirque, oriented in the same direction, is subjected to both up- and down-canyon winds. Wind direction at Emerald shows greater variation; only northeasterly winds were rare, the canyon mouth facing away from this direction. The Spuller vapor pressure distribution was narrower, with a lower mean and pronounced low pressure peaks. The April and May distributions were noticeably different, broader and shallower at Emerald with a significant percentage > 6.11 mb. Almost no values indicating condensation are recorded at Spuller.

In 1993, meteorological data were collected at a ridge top site 2 km north of Emerald Lake (identified as the mini-catchments). In 1994, two other meteorological stations were installed, at 2.5 and 4 km from Emerald, on the high plateau to the east (identified as M1 and M3). Although these locations had higher wind speeds (~33% higher at M1 and the mini site, ~70% higher at M3) and lower vapor pressures than Emerald, monthly evaporation was similar (Figure IV-19c). The greatest differences (May 1993 and May and June 1994) resulted from higher Emerald vapor pressures rather than decreased wind speeds.

Excluding Wolverton, annual evaporation from snow fell within a range of 29 (SNARL 1992) to 166 mm (Spuller 1993). The lower estimate is from a shallow, low-elevation snowpack, the upper from a high-elevation location with abnormally deep snow; a site characterized by high winds and low vapor pressures. A regional estimate based on these results would lie between 80 to 100 mm of annual loss. Emerald (average annual evaporation of ~60 mm) and Mammoth and Spuller (~110 and ~140 mm) straddle the estimate.

During the drought years of 1990 through 1992, maximum snow accumulation varied from 0.6 to 0.9 m (water equivalent). The lower estimate of regional evaporation (80 mm, assuming a shortened snow-cover season) gives a snowpack loss of 9 to 13%. In the high snowfall year of 1993, peak accumulation varied from 1.7 to 2.3 m, and an evaporation estimate of 100 mm amounts to an average loss of 4 to 6%. During snowmelt, average evaporation at Mammoth, Emerald and Spuller was 5, 13 and 20 mm, respectively. An estimated average regional loss of 15 mm represents 2 to 3% of the snowpack during drought years, less than 1% for above normal accumulations. None of these estimates include losses that may occur during the re-deposition of wind blown snow (heavy snow density and the relative absence of major wind scour and depositional patterns in these catchments lead us to believe that these losses were minor).

4.3.3.1. Sensitivity of Evaporation to Model Assumptions

The sensitivity of calculated evaporation to the estimated parameters was evaluated with two data sets: Emerald from January through April of 1993, and Wolverton during January, March and April of 1993. A constant instrument height was assumed except when sensitivity to instrument height was specifically examined.

4.3.3.1.1. Roughness Length

Evaporation at Emerald was calculated for momentum roughness lengths from 0.0001 to 0.9 m (Figure IV-23a). At high values (> 0.1 m), increases in z_0 cause large increases in total evaporation, as the velocity and humidity profiles abruptly steepen. However, within the range of probable roughness lengths, 0.001 to 0.02 m, changes are small: evaporation doubling over this span.

At Wolverton, probable values of z_0 vary from 0.1 to 1.0m, and evaporation increases by 500% over this interval (Figure IV-23b). The effect of increasing z_0 is magnified by increases in displacement height, set at $4.9z_0$. Since a fixed anemometer height was assumed (7.7m), higher values of z_0 dramatically steepen the wind profile and increase calculated evaporation. Although the model is sensitive to the chosen value of z_0 at Wolverton, evaporation is so low as to make little overall difference. A doubling of z_0 would have increased evaporation from 4 to 7 mm.

We also assumed a ratio of 0.1 for z_{0t}/z_0 and z_{0v}/z_0. For the roughness length used in analyzing all but the Wolverton data (0.005 m), varying the ratio by an order of magnitude changed evaporation by ~12% (Figure IV-23c). For larger z_0 the change in evaporation becomes progressively greater: a 50 percent increase for z_0 of 0.1 m (Figure IV-23d). However, such large z_{0v} values are improbable. The Wolverton results were similar except for large values of z_0 which increased evaporation even further, by elevating the saturation vapor pressure unrealistically high above the snow (~6 m).

4.3.3.1.2. Wind Speed:

Wind speed is the critical variable; all of the fluxes are directly proportional to wind speed and the Obukhov length is proportional to its third power. Wind reduces the importance of atmospheric stability; with increasing wind speed, turbulent shear forces become dominant and the atmosphere approaches neutrality. A graph of evaporation vs. average wind speed for Emerald was developed by multiplying the hourly wind speeds by a variable "wind speed factor" (from 0.1 to 10) and recalculating evaporation (Figure IV-23e). Doubling the wind speed tripled the evaporation. The reliability of the evaporation results, therefore, rest heavily on anemometer accuracy.

Inaccurate, misplaced or improperly installed anemometers could lead to erroneous conclusions. We used 12 anemometers of 4 types, at times using 2 or 3 at an individual site. We regard our findings of low average wind speeds in this region as well founded. Average evaporation of 2 mm per day at Emerald (Kattelmann and Elder 1991) would have required a mean wind speed of 7.5 m s^{-1} over the four month data set (Figure IV-

23e). Values this high may occur, but rarely and in isolated locations, e.g., wind speed at Mammoth Mountain during the 1993 snow season exceeded 7.5 m s^{-1} only 20 percent of the time.

4.3.3.1.3. Instrument Height

Daily measurements of snow depth were made at Lodgepole Ranger Station, close to Wolverton. This record helped determine the relative snowpack fluctuation between major storms. When combined with measurements made during site-visits, it was used to reconstruct snow depth at each of the meteorological stations. Since instrument height errors did occur, Emerald data were used to evaluate their effect on calculated evaporation (Figure IV-23f). If instrument height is underestimated, the wind speed, temperature, and vapor profiles steepen, increasing the calculated fluxes. However, only when the instruments are close to the snow-surface will small errors in height produce large changes in evaporation. At Emerald, the meteorological instruments were 5 to 6 m above-ground and the maximum snow-depth varied from 1.5 to 3 m.; for this range, an instrument height error of ±1 m changed evaporation by approximately 20%.

4.3.3.1.4. Snow Temperature

Emerald evaporation calculated with the snow-surface temperature model of Marks et al. (1992; sTemp 2) was double the evaporation calculated using our model (sTemp 1). Marks et al. assumed that turbulent exchange takes place, not at the snow-surface, but 15 to 25 cm below the surface, i.e., within a porous upper layer. Their model was based on measurements within this layer. The difference between models is not about snow-surface temperature, per se, but air temperature at the snow-surface, and what snowpack strata corresponds with, and determines it.

Differences in calculated evaporation using the two models vary with wind speed and season. At very low wind speeds the difference was substantial, on the order of a few hundred percent, but declined to a constant 50% at wind speeds greater than 5 m s^{-1} (Figure IV-23e). In sTemp 2, snow temperature always lags air temperature, producing snow temperatures in excess of air temperature for a majority of the hourly accumulation season intervals. This has two effects: it produces neutral or unstable conditions much of the time, and it enhances vapor pressure differences during these periods.

During the 1993 snow accumulation season, sTemp1 produced stable conditions 82% of the time, compared with 47% for sTemp 2. Halving the occurrence of atmospheric stability accentuates evaporation at low wind speeds, when the role of instability is enhanced. For Wolverton, which exemplifies this situation, sTemp 2 increased evaporation six fold over that calculated with sTemp 1: from 4 to 22 mm. The instability effect and the large thermal lag used in the Marks et al. model maximized evaporation and set an upper limit on the probable snow-surface temperature error. During snowmelt, both models are similarly constrained to 0°C and there was little difference in calculated evaporation.

However, different snow temperature assumptions are responsible for only a small part of the large difference in calculated evaporation between the two studies at Emerald. Marks et al. (1992) report a total evaporation from November 1995 through April 1996 of 300 mm of water. In 1993, a similar snow year, our study showed a total of 55 mm for the same period. Using sTemp 2 increased the 1993 total to 110 mm. During the remainder of the snow season (May through July), when both temperature models would have estimated similar snow-surface temperatures, the differences between the studies are larger: 155 mm vs. 6 mm.

The principal disagreement lies in the measurement of wind speed. Marks et al. measured wind at two sites, described as "ridge" and "lake" locations. The respective means (and range), for December through May 1986, were 7.5 m s^{-1} (2.0 to 16.9 m s^{-1}), and 5.0 m s^{-1} (1.3 to 9.4 m s^{-1}). In contrast, mean wind speed measured in this study, for the same months in 1993, was 1.8 m s^{-1} (0 to 13.8 m s^{-1}). The wind speed differences account for the different evaporation results. Their average wind speeds seem anomalously high, based on the multiple sites and years of our study.

4.3.3.2. Mean-profile Assumptions and the Mountain Environment

The assumptions of steady-state flow and homogeneous surface conditions used to develop the mean-profile equations are violated in a mountain environment. The steady-state requirement is impossible to satisfy, but most authorities feel that ignoring it introduces little error (Moore 1983). Brutsaert 1982) recommends using averaging times of 20 to 60 minutes as a good compromise between reasonably constant atmospheric conditions and adequate sampling. The time interval used for averaging meteorological data in this study followed his recommendation: 60 minutes, except at Mammoth and the mini-catchments where a 15 minute interval was used. A test using 15 minute vs. 60 minute Mammoth data showed little difference in calculated evaporation. The second assumption is more critical. While snow reduces small scale differences by smoothing terrain features and covering vegetation, mountain surfaces are not homogenous.

A change in surface forces the development of a "transition zone" (Oke 1987), where the over-passing air reflects a mixture of both old and new surface characteristics. Below this zone the air is fully adjusted to the new conditions, above it, it reflects only the conditions of the old. The top of the transition zone develops at a rate of 1 m vertically for every 10 to 30 m of horizontal travel past the boundary, but the fully adjusted layer grows more slowly, 1 m vertically for every 100 to 300 m (Oke 1987); although with rougher surfaces and unstable conditions, vertical development may be more rapid. To insure fully adjusted conditions for instruments placed ~3 m above the snow surface, at least a few hundred meters are needed for an adequate fetch. Many of the measurements in this study were made under conditions of inadequate fetch.

Meteorological instruments in a transition zone record a spectrum of prior changes in surface and slope, and the principal mean-profile errors will lie in the assumptions of instrument height and surface roughness. We estimate the effect of the combined error on calculated evaporation as less than 20% (Figures IV-23a and IV-23f). That the measured meteorological variables represent values integrated over a range of terrain may even be an advantage, since the overall objective of the study is not the accurate calculation of point evaporation, but a broader spatial estimate.

4.3.3.3. Radiative Heating Above the Snow-surface

Halberstam and Schieldge (1981) performed extensive profile measurements over a melting snow surface in an open field near Lee Vining, California, in March 1978. They reported anomalous temperature profiles, with a persistent warm layer forming ~0.5 m above the snow pack on clear, calm days. Their interpretation was that under conditions of initial stability, evaporated water vapor was retained in the air layer just above the snow-surface and subsequently heated by absorbing solar radiation, directly from above and by reflection off of snow. As the layer became warmer than either the snow surface or the ambient air above, it re-radiated energy in the infrared from its upper surface towards the atmosphere and from its lower, towards the snow pack. Others have reported the same phenomenon (Male and Granger 1978).

These conditions would introduce error into a single-level calculation of the sensible heat flux. Typically during snowmelt, the single-level model portrays stable conditions and a near-logarithmic temperature profile. With a radiatively heated layer, the actual situation is a sensible heat flux reversal at the temperature maximum: stable air and a strong flux towards the surface below the layer, and unstable conditions and an upward flux above it. Thus a single-level model underestimates the sensible heat flux. Male and Granger (1978) suggest using a second temperature sensor at a distance of 10 to 30 cm above the snow to overcome this difficulty.

The effect of this problem on the water vapor flux is less clear. Increased turbulence above the temperature maximum enhances evaporative transport, yet moisture retention in the lowest air layer is necessary to maintain the maximum. Stable conditions below the temperature maximum and the maintenance of a high vapor pressure in this layer would indicate little additional vapor flux from the surface itself. Moore (1983) states that under the initial conditions necessary for the formation of this maximum, the turbulent fluxes are of minor importance to the overall energy balance and that, when they become important, on windy overcast days, this condition cannot develop.

At Emerald, during the later part of May and into June 1994, three temperature sensors were installed, the lower sensor placed near the 0.5 m elevation mentioned by Halberstam and Schieldge. During this period, late afternoon wind speeds were low (typically < 0.5 m s^{-1}), ideal for developing a radiation maximum. Night-time temperatures at the lowest sensor level were usually lower than temperatures recorded at the sensor above, as expected under stable atmospheric conditions. However, beginning just before noon, the reverse often occurred: the lower sensor recording warmer temperatures (Figure IV-24). Days when this pattern did not develop or developed differently were characterized by either much higher wind velocities or consistently low wind speeds throughout the day. The radiation maximum was neither as well developed nor as persistent as at Lee Vining (~1 vs. 3 to 4°C).

Instruments were intermittently maintained at multiple levels during snowmelt at Emerald in 1993 and at the mini-catchments in 1994; for these intervals, we separately calculated total evaporation and sensible heat transfer for each level (Table IV-8). The Emerald data show differences in the specific heat flux for different sensor heights, as predicted for a radiation maximum. However, the evaporation differences are small enough to have been caused by instrument variation and other errors. At the mini-catchments, a greater sensible heat flux was calculated for the higher sensor, the opposite of what was expected with a radiation maximum. Higher wind velocities at the site may be preventing its development. The small differences in calculated evaporation for different sensor levels argue for the relative insensitivity of our results to the radiation maximum problem.

4.3.3.4. Model Accuracy

The mean-profile calculations in this study cannot be independently verified. Combination methods such as Penman's add a radiation balance to the turbulent flux calculations, increasing both the probability of error and the reliance on estimated snow-surface temperature. Other authors report poor success with the Bowen ratio (Male and Granger 1978; McKay and Thurtell 1978). Calculating evaporation is problematic for Bowen ratios of -0.5 to -1.0: the expected range during snow accumulation when the water vapor and sensible heat fluxes are nearly identical but opposite in sign (Cline 1995; Marks and Dozier 1992). Net radiation is also required, dependent, as is the Bowen ratio, on estimated snow-surface temperatures. Eddy flux correlation can directly measure evaporation from the instantaneous fluctuations of both vertical air speed and water vapor density. However, the instrumentation is difficult to maintain and a large power requirement makes it unsuitable for remote locations.

Few evaluations of mean-profile results using sensors at a single-level are available. Pluss and Mazzoni (1994) compared single-level flux calculations during snowmelt with those obtained using sensors at four levels and found a "systematic overestimation." They felt that the results for sensible heat were better than for vapor. Comparisons between eddy correlation and the 4-level profile were generally good: somewhat overestimating larger and underestimating smaller fluxes. Some researchers have used lysimeters to verify the energy balance during snowmelt. However, single lysimeters are undependable and small amounts of evaporation cannot be distinguished within the standard error of even extensive arrays (Kattelmann 1995).

Snowmelt models, such as SNTHERM (Jordan 1991), calculate evaporation with single-level meteorological data using the energy balance to derive snow-surface temperature. Cline (1997) found that SNTHERM snowmelt evaporation closely matched his multi-level mean-profile calculations. However, besides meteorological data and long- and short-wave radiation, the model requires snow-profile data for initiation (layer-by-layer input of temperature, crystal size, and density). In our study, snow-profile data was only available (except at Mammoth) for the snowmelt season, when the snow-surface temperature is easily modeled.

We estimate that the probable error in winter (accumulation season) evaporation lies between 30 to 60%, while that for the snowmelt season is between 15 and 30% (Table IV-9). Given the small amounts of evaporation calculated at the study sites, even larger errors would not change the overall conclusions.

4.4. Conclusions

Only a small percentage of the settled snow-pack is lost to evaporation in the central Sierra Nevada. The typical evaporative loss from snow varies from 80 to 100 mm of water, approximately 6% of the average maximum accumulation. In drought years the loss is less, but represents a greater percentage of the maximum accumulation, 9 to 13%. During peak snow years the loss decreases to about 4%, although the extended length of the snow season somewhat increases the actual amount lost. Most of the loss occurs during the period of snow accumulation when vapor pressure differences are usually favorable for evaporation and conditions of atmospheric instability are often found. Even so, low vapor pressure differences between the air and the snow surface, the result of cold temperatures and the prior transit of the over-passing air over extensive snow-covered distances (which increase humidity and decrease the capacity to absorb additional moisture), limit total evaporation to relatively small quantities.

During snowmelt, stable atmospheric conditions and reduced vapor pressure differences, due to the higher vapor content of the warming spring air, reduce evaporation from snow to negligible amounts. Often near the end of this period, snow-pack evaporation losses are exceeded by gains from condensation. Total evaporation from the snow-pack during snowmelt is typically around 15 mm of water: 2 to 3% of the maximum accumulation during drought years, less than 1% for above normal snow years. Studies, such as ours, where the annual accounting and water balance analysis begins with the measurement of maximum accumulation at the start of spring snowmelt, can neglect evaporative loss from snow.

However, total or annual alpine and sub-alpine evaporation is neither negligible nor low. As snowmelt progresses, evaporation from saturated soils and free-water becomes appreciable. Wet soils and surface waters (standing pools and melt rivulets) are rapidly warmed by solar radiation, generating large vapor pressure differences and local atmospheric instability, enhancing evaporative loss. The heating of snow-free surfaces warms the over-passing air, increasing sensible heat transport to the remaining snow-pack, accelerating snow melt and further increasing free-water availability. Only as water availability diminishes with the gradual drying of the basin does the rate of evaporation decrease.

These conclusions are neither startling nor new. Hutchison (1963), in a snowmelt season study of evaporation from soil and snow surfaces conducted at 2740 m in the central Rockies, concluded: "Evaporation from wet soil surfaces greatly exceeded evaporation from nearby snow. This is readily explained by the much lower albedo of the soil surface, the consequent higher availability of energy, and the fact that the vapor pressure of the soil is not fixed by a temperature limit as is that of the snow. There is evidence that soil moisture evaporation tended to retard snow evaporation by increasing the atmospheric vapor pressure and thus depressing the gradient in vapor pressure between the temperature-limited snow surface and the ambient air. . . . As areas of bare wet soil increased and evaporation amounts from such surfaces increased, evaporation from snow decreased."

In a mid-July study near Spuller Lake, designed to test Matthes' conclusion of high evaporation at altitude, Kehrlein et. al. (1953) reported: "Examination of several patches of snow on Dana Plateau, near 11,500 ft elevation, showed runoff from each, including a large patch whose surface was 'sun-cupped,' with cups about 2 feet deep. Near this site, a small patch of snow 19″ × 30″ × 8″ deep was observed with no apparent runoff from it. The ground nearby was very dry. This patch of snow was removed from its position and placed on a nearby granite slab. Immediately there was runoff. Runoff continued for the half-hour of observation. Similar patches were noted in the vicinity. Evidently high infiltration capacity of the coarse granitic soil and glacial sand was as great as the rate of melt." After experimental work on Mt. Whitney, they concluded: "Overall consideration of the hydrologic water balance indicates that much water from melted snow does evaporate, but the vapor loss is predominantly from moist soil and from water surfaces rather than from the surface of the snow itself." We draw the same conclusion.

Complete analysis of data produced in the study will extend past the completion of this report. Additional time could be profitably spent in applying other evaporation models. Using the Bowen ratio and Penman equations to calculate accumulation season evaporation may prove feasible. At a minimum, seasonal Bowen ratio variations and the determination of potential evaporation (thus setting an upper bound) could provide additional insight. It may also be possible to extend these two methodologies into the snowmelt season, by using outflow volume and temperature to estimate energy inputs for snowmelt and by assuming that the meteorological data collected at the meteorological station represents average catchment values.

Calculation and evaluation of the energy balances for all the meteorological stations, comparing differences between sites and seasons, would be useful; particularly at Mammoth, where both incoming and outgoing solar and infrared radiation were monitored. Snow-surface temperatures, calculated from outgoing infrared radiation, could be used to evaluate the accuracy of the snow-surface temperature model used in the mean-profile equations. It may also be possible to use Mammoth radiation data to determine the "equilibrium temperature," i.e., the surface temperature at which evaporation calculated from the energy balance and evaporation calculated by turbulent transfer are equal. Diel variations in equilibrium temperature could decide which snow-surface temperature model, ours or the Marks et al. (1992) model, is more appropriate; or equilibrium temperatures could be used to develop an improved model. In-depth analysis of meteorological patterns at all of the sites may be warranted. Others (Male and Granger 1979; Cline 1997) have made efforts to relate point evaporation to synoptic meteorological patterns, and a similar attempt with this data set might be fruitful.

Obtaining meteorological data at two-or-more levels at Emerald and the mini-catchments failed due to temperature and humidity sensor drift, and our inability to provide the frequent re-calibration required. However, we may be able to calculate momentum roughness lengths using the multi-level wind speed data.

Using actual measurements of incoming solar and infrared radiation may improve CRAE model performance. Additional work could be done to improve the modeling of snow albedo, and more accurate models of clear sky radiation could be substituted for Morton's simple equations. In the CRAE model we used high quality hourly data to calculate period averages and to derive regression relationships between the stations. It would be instructive to test the feasibility of estimating alpine evaporation with CRAE model, relying solely on meteorological data from local National Meteorological Service reporting stations.

Brutsaert and Stricker (1979) and Parlange and Katul (1992) developed models based on the same complementary principle utilized by Morton (called Advection-Aridity in these papers). Their models use short time intervals (20 minutes to a few hours) and the mean-profile equation serves as the wind function in calculating potential evaporation. These concepts could be applied to meteorological data from this study and the results compared with the CRAE and mean-profile estimates. An advection-aridity model could only be used when the sub-surface energy flux is negligible, i.e., not during snowmelt. (Although as mentioned previously, if we assume that the Spuller and Emerald meteorological stations represent average catchment conditions, outflow discharge becomes a measure of snowmelt energy input and average catchment evaporation during snowmelt could be estimated.) Besides providing an additional check on study results, an advection-aridity model could estimate snow-free evaporation.

Aside from further data analysis, there are a number of questions raised by the study that can only be answered with additional field work and experimentation. The collection of multi-level meteorological data would allow evaporation estimates free of snow-surface assumptions. Thus providing an additional check on study results, as well as year-round point evaporation estimates. Since most evaporation is occurring from snow-free surfaces, experimental determination of a range of probable values would be beneficial. The need for re-calibration and frequent monitoring make Mammoth and SNARL the best locations for this work. We can again try collecting multi-level data at Emerald. If, in addition to instruments at the various levels, duplicate temperature-and-humidity sensors are installed at one of the levels, the second sensor can be moved to other levels during site visits, i.e., allowing calibration between the sensors. Since relative differences and not absolute values are important, the only necessary assumption is of linear instrument drift between visits.

The addition of a large snowmelt lysimeter at SNARL would provide a location for testing point evaporation models. Kattelmann (1995) has shown that a lysimeter of approximately 10 m^2 will give reasonably accurate snowmelt measurements for a level snow pack on level ground. SNARL meets that criteria, and has electrical power for keeping lines unfrozen, a resident manager and a restricted access site for construction.

Our acid rain research in Sequoia National Park has expanded to include questions of scale, i.e., what happens as we go from individual catchments to larger drainage basins. The installation of a gauging station on the Marble Fork of the Kaweah in the Tokopah Valley (just below the Emerald outlet confluence), allowed us to monitor river stage and water chemistry over the last three years. During this time we have also conducted basin-wide snow surveys at maximum accumulation. The snow survey and discharge data, combined with meteorological data from two new meteorological stations (M3 since 1995 and Topaz since 1996), will allow us to estimate evaporation for the larger Tokopah basin. Neither the Emerald basin nor the location of its meteorological station is typical of the Tokopah and the additional stations will provide better data for CRAE and other areal models, and better locations for point model evaporation estimates.

Lastly, along these same lines, more thought and work needs to be put into the development of appropriate models for snow-free alpine and sub-alpine evaporation. The key to this would seemingly lie in the determination of suitable algorithms to characterize the changing availability of moisture for the various catchment soil types and vegetation, or in the improvement of Morton's or some other advection-aridity model. That the latter approach appears to be more feasible, justifies some of the recommendations for continuing study made herein.

4.5. References

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Table IV-1. The relative contributions of snowmelt and evaporation from a melting snow pack. Data from Kuusisto (1986, Table 1) and Pluss and Mazzoni (1994, Table 4). Negative evaporation values and "C" indicate condensation. Conversion factors used: 1 watt m^{-2} = 0.259 mm day^{-1} of snowmelt = 0.0305 mm day^{-1} of evaporation from snow or ice = 0.0345 mm day^{-1} of condensation from water vapor.

Reference	Location	Elev. m	Time period	Snowmelt %	Snowmelt mm/day	Evaporation %	Evaporation mm/day
Miller (1955)	open field (California) 37°N			82		18	
Gold & Williams (1960)	open field (Canada) 45°N	100	Mar '59	26	7	74	2.35
Wendler (1967)	open field (Alaska) 67°N		Mar-Apr 1966	76		24	
Anderson (1968)	open field in mountains (California) 37°N		'47-51 snow seasons	100		C	
			Apr-May	100		C	
Treidl (1970)	open field (Michigan) 46°N		Jan 23 '69	100	15	C	-0.72
Dewalle & Meiman (1971)	forest opening (Colorado) 39°N	3260	Jun 1968	97	50	3	0.18
Fohn (1973)	Peyto Glacier (Canada)	2510	15 days in May	100	47	C	-0.50
de la Casiniere (1974)	open field in mountains (France) 46°N	3550	July 1968	85	16	15	0.33
	open field in mountains (Spain) 41°N	1860	Apr 1970	47	10	42	1.05
	Mt. Blanc, French Alps	3550	23 days in July	81	5	19	0.15
Weller & Holmgren (1974)	open field (Alaska) 71°N	10	Jun 1971	49		10	
Granger & Male (1974)	open field in prairies (Canada) 51°N		'74 melt	84	8	10	0.11
			'75 melt	70	5	29	0.24
			'76 melt	72	3	14	0.07
Martin (1975)	St. Sorlin Glacier (France)	2700	11 days in summer	93	14	7	0.11
Hendrie & Price (1979)	deciduous forest (Ontario) 46°N		Apr 1978	100	10	0	0
Kuusisto (1979)	open field (Finland) 51°N	60	melt season '68-'73	96	7	4	0.03
Harstveit (1981)	open field in mountains (Norway) 60°N	435	Apr-May '79-80	100	12	0	0
			cloudy days	100	23	C	-0.80
			clear days	76	7	24	0.26
Braun & Zuidema (1982)	small basin, 23% forest (Switzerland) 47°N	800	high melt days '77-'80	100	23	C	-0.80
Eaton & Wendler (1981)	open field (Alaska) 65°N		April 1980	32	3	68	0.75
Kuusisto (1982)	open field (Finland) 61°N		high melt days '59-'78	100	14	C	-0.04
	open field 67°N			77	15	23	0.53
Prowse & Owens	open field in mountains (New Zealand) 43°S	1500	Oct-Nov '76-'80	100		C	
			rainy days	100		C	
			days with high heat	100		C	
Funk (1984)	Rhonegletscher, Swiss Alps		summer	99	44	1	0.06
Moore & Owens (1984)	open field in mountains (New Zealand) 43°S	1450	melt season 1982	100	31	C	-1.03
Aguado (1985)	open field (Wisconsin) 3 sites 43-45°N		melt seasons '53-'54	57		12	
Vehvilainen (1986)	small basin, 82% forest (Finland) 64°N	120	melt seasons '71-'81	87	5	13	0.09
Harding (1986)	Finse (Norway)	1000	15 days in May	100	12	C	-0.01
Calanca & Heuberger (1990)	Urumqi Glacier #1, Tien Shan (China)	3900		81	16	19	.46
Marks & Dozier (1992)	Emerald Lake (California) 37°N	2800	May '86	43	11	57	1.65
			Jun '86	62	30	38	2.17
Pluss & Mazzoni (1994)	Swiss Alps	2600	18 days in May '92	97	10	3	0.03
Cline (1995)	Niwot Ridge (Colorado) 40°N	3520	melt season 1994	76	17	23	0.49

Table IV-2. Evaporation from snow at Emerald Lake in the winter of 1986-1987 (from Marks and Dozier, 1992).

Month	Evaporation (mm of water) At the Lake	At the Ridge	% increase at Ridge
Nov.	34	51	50
Dec.	73	25	-66
Jan.	64	61	-5
Feb.	36	54	50
Mar.	43	61	42
Apr.	46	83	80
May	49	65	33
June	66	72	11
July	40	64	60
Total	451	537	19

Table IV-3. Seasonal evaporation from snow at the Central Sierra Snow Laboratory, Lake Tahoe California (from Anderson, 1976).

	Evaporation (mm of water)		
Year	Accumulation Season	Snowmelt season	Total Evaporation
1968-1969	4.9	10.6	15.5
1969-1970	9.5	6.6	16.1
1970-1971	10.3	8.7	19.0
1971-1972	9.4	10.8	20.2
1972-1973	5.9	12.8	18.7
1973-1974	7.1	11.5	18.6
Average	7.9	10.2	18.0

Table IV-4. Average lake evaporation, from adjusted pan evaporation measurements, at Friant and Redinger lakes on the western slope of the Sierra Nevada (from Longacre and Blaney, 1962).

	Friant Lake	Redinger Lake
Elevation	125 m	436 m
Yearly evaporation	1722 mm	1564 mm
Max. month (July)	310 mm	279 mm
Min. month (Jan.)	28 mm	30 mm

Table IV-5. Meteorological stations used in the study. Numbers in brackets refer to instruments installed (as listed in Table IV-6) at each station.

Site and Location	Description
Eastern Brook 37° 26.22′ N, 118° 44.55′ E 3170 m	Located in a small clearing in a stand of old-growth fir and pine, 200 m up slope on the north side of Eastern Brook Lake. The instruments may be located below the lower boundary layer. The recent record begins in March of 1993. Other data was collected in the early 1980s. [2, 6, 8, 9 & 11]
Emerald 36° 35.97′ N, 118° 40.20′ E 2800 m	Located in a glacial cirque, behind (~80 m) and above (~20 m) the lake. The up-slope fetch (in the direction of the prevailing winds) is good, but the fetch in other directions is marginal. The record begins in Sept. 1991 and has few gaps. [1, 4, 6, 8, 9 & 11; 2 & 7 at two additional levels in spring 1994]
Mammoth 37° 38.48′ N, 119° 01.78′ E 2930 m	Located on a slightly sloping bench about half way up the flank of the mountain. The fetch is good in the direction of the prevailing wind and adequate for the other directions. The record begins in Sept. 1989 and is generally good, although there are a number of significant gaps. [2, 7, 7a, 8 (3 facing up, clear , red, thermal; 2 facing down, clear, red) & 10]
Mini-catchments 36° 36.67′ N, 118° 39.71′ E 2960 m	Located on a level and open ridge. The axis of the ridge is perpendicular to the direction of the predominant up-and-down canyon winds; the fetch is good in all directions with the exception of the southeast, where a small clump of trees is an obstruction. In operation during the spring of 1992 and 1993. The record is good for the snowmelt season of 1993. [2, 7, 7a, 8 (1 facing up; 1 facing down) & 10; 2 & 7 at an additional level in spring 1993]
M1 36° 36.55′ N, 118° 39.23′ E 3090 m	Located on a north-facing slope, next to an outlet stream below a small pond. The undulating, slightly sloping, terrain is open and free of vegetation and the fetch is good in all directions. A record exists for May and June of 1994. [2 & 7]
M3 36° 36.68′ N, 118° 38.50′ E 3250 m	Situated similar to M1 in a flat and open area , higher up on the same ridge, with good fetch in all directions. The record exists for April through June of 1993. [2 & 7]
SNARL 37° 36.79′ N, 118° 49.83′ E 2160 m	Located in Long Valley at the base of the eastern Sierra Nevada escarpment, adjacent to the outlet stream on a glacial out-wash plain ~2 km from the canyon mouth,. The record begins in Oct. 1989 and is essentially complete. The area is level and the fetch is good for up-and-down canyon winds, adequate in other directions (some low shrubs to the south, and buildings to the north). [3, 5, 8 & 10]
Spuller 37° 56.93′ N, 119° 16.91′ E 3120 m	Located next to the outlet stream, on a small flat in a narrow canyon ~100 m below the lake. The fetch for the prevailing wind direction (down canyon) is adequate, marginal for the other directions. The record begins in Oct. 1989, but little data are available for the early years. The latter record is better, but still has large gaps; only 1993 is substantially complete. [1, 4, 6, 8, 9 & 11]
Wolverton 36° 35.74′ N, 118° 44.16′ E 2190 m	Located in a small (~25 m) forest clearing adjacent to a large meadow; a few scattered tall trees amid younger growth form the surrounding stand. There are problems with fetch and instrument height, but given the openness of the stand, the tower is probably within the lower boundary layer. The record is excellent, providing almost continuous coverage from Nov. 1986. [2, 7, 7a, 8 & 11]

Table IV-6. Meteorological instruments used at the various meteorological stations.

Parameter	no.	Description
Air Temperature	1	Omnidata ES 060 Thermistor: range, -50 to +80 °C; precision, ±0.25 °C; response time, 90 s; mounted in a EA 130V radiation shield.
Air Temperature and Humidity	2	Vaisala ES 120 temperature and humidity sensor: range, -50 to +80 °C, 0 to 100%; precision, ±0.25 °C, ±0.3% ; response time, 5 s; mounted in a EA 130V radiation shield.
Air Temperature and Humidity	3	Vaisala HMP 113Y temperature and humidity sensor: range, -40 to +80 °C, 0 to 100%; precision, ±0.25 °C, ±0.2 to 0.3% ; response time, 5 s; mounted in a R. M. Young electrically aspirated radiation shield.
Humidity	4	Vaisala HMP 35A capacitance type sensor: range, 0 to 100%; response time, 5 s; precision, ±0.2 to 0.3% ; mounted in a EA 130V radiation shield.
Wind Speed and Direction	5	R. M. Young 05103 propeller and vane: range, 0 to 50 m s^{-1}; threshold sensitivity of propeller, 0.2 to 0.4 m s^{-1}, of vane, 0.7 m s^{-1}.
Wind Speed and Direction	6	Omnidata ES 050 propeller and vane: range, 0 to 40 m s^{-1}; threshold sensitivity of propeller, 0.4 m s^{-1}, of vane, 0.4 m s^{-1}.
Wind Speed	7	R. M. Young, Gill 3-cup anemometer: range, 0 to 50 m s^{-1}; threshold sensitivity, 0.5 m s^{-1}.
Wind Direction	7a	R. M. Young 12105, Gill Micro-vane; range, 355 °azimuth, threshold sensitivity, 0.4 m s^{-1}.
Solar Radiation	8	Eppley Precision Spectral Pyranometer: range, with clear glass, 0.285 to 2.80 µm; temperature dependence, ±1% for -20 to +40 °C; linearity, ±0.5% from 0 to 2800 w m^{-2}; response time, 1 s.
Infrared Radiation	9	Eppley Precision Infrared Radiometer (Pyrgeometer): range, 4 to 50 µm; temperature dependence, ±1% for -20 to +40 °C; linearity, ±1% from 0 to 700 W m^{-2}; response time, 2 s.
Precipitation	10	Weathertronics Model 6028A heated tipping bucket gauge with 'Alter' type shield: 12 in. orifice; sensitivity, 0.25 mm/tip; precision, 0.5% at 0.5 in hr^{-1}.
Precipitation	11	Qualimetrics 6011b tipping bucket gauge with 'Alter' type shield: 8 in. orifice; sensitivity, 0.25 mm/tip; precision, 0.5% at 0.5 in hr^{-1}.

Table IV-7. Regression equations developed between the for the various meteorological variables. The square of the regression coefficient (r^2) and the standard error of the estimate are given for each equation. T is air temperature; DP is dew point; TS is total solar, i.e., incoming short-wave radiation; and S is the ratio of observed-to maximum-possible sunshine duration. The stations are abbreviated as follows: Emerald Lake (eml), Mammoth Mountain (mm), Spuller Lake (spu), Wolverton (wolv), Eastern Brook Lake (ebl), and the Sierra Nevada Aquatic Research Laboratory (snarl).

West-side Regression Equations	**r^2**	**Standard Error**
Wolverton on Emerald Lake		
$T_{wolv} = 4.06 + 0.754T_{eml} + 0.33DP_{eml}$	0.98	0.88° C
$DP_{wolv} = 5.276 + 0.677DP_{eml}$	0.89	1.58° C
$TS_{wolv} = 16.233 + 0.943TS_{eml} + 2.867T_{eml}$	0.93	24.0 W m^{-2}
$S_{wolv} = 0.235 + 0.256S_{eml} + 0.034T_{eml} - 0.021DP_{eml}$	0.78	0.09
Emerald Lake on Wolverton		
$T_{eml} = -0.78 + 1.09T_{wolv} - 0.30DP_{wolv}$	0.96	1.27° C
$DP_{eml} = -9.438 + 0.921DP_{wolv} + 0.329T_{wolv}$	0.91	2.00° C
$TS_{elm} = 10.45 + 0.832TS_{wolv}$	0.92	22.6 W m^{-2}
$S_{eml} = 0.142 + 0.649S_{wolv} + 0.017T_{wolv} - 0.027DP_{wolv}$	0.64	0.11

Eastside Regression Equations	**r^2**	**Standard Error**
Mammoth Mt. on SNARL		
$T_{mm} = -0.85 + 0.787T_{snarl}$	0.87	2.41° C
$DP_{mm} = -2.22 + 0.806DP_{snarl} + 0.161T_{snarl}$	0.92	1.77° C
$TS_{mm} = -7.55 + 1.022TS_{snarl}$	0.95	20.8 W m^{-2}
$S_{mm} = -0.044 + 0.974S_{snarl} + 0.005T_{snarl}$	0.74	0.09
Spuller Lake on Mammoth Mt.		
$T_{spu} = -2.293 + 0.985T_{mm}$	0.96	1.24° C
$DP_{spu} = -6.247 + 0.201T_{mm} + 0.859DP_{mm}$	0.92	1.75° C
$TS_{spu} = -8.045 + 1.081TS_{mm} - 2.681T_{mm}$	0.95	16.9 W m^{-2}
$S_{spu} = 0.11 + 0.625S_{mm} - 0.012DP_{mm} + 0.0004TS_{mm}$	0.71	0.06
Spuller Lake on SNARL		
$T_{spu} = -3.242 + 0.782T_{snarl}$	0.89	2.05° C
$DP_{spu} = -6.878 + 0.208T_{snarl} + 0.883DP_{snarl}$	0.94	1.54° C
$TS_{spu} = 1.686 + 0.957TS_{snarl}$	0.96	17.3 W m^{-2}
$S_{spu} = 0.692 + 0.0043TS_{snarl} - 0.0035G_{o-spu}$	0.63	0.08
Eastern Brook Lake on SNARL		
$T_{ebl} = -2.684 + 0.833T_{snarl}$	0.91	1.83° C
$DP_{ebl} = -6.648 + 0.952DP_{snarl}$	0.89	2.08° C
$TS_{ebl} = -29.556 + 1.003TS_{snarl}$	0.95	20.3 W m^{-2}
$S_{ebl} = -0.044 + 1.291S_{snarl} - 0.0013TS_{snarl} + 0.013T_{snarl}$	0.54	0.10
Eastern Brook Lake on Mammoth Mt.		
$T_{ebl} = -1.31 + 1.008T_{mm}$	0.95	1.42° C
$DP_{ebl} = -5.846 + 0.878DP_{mm}$	0.91	1.82° C
$TS_{ebl} = -6.207 + 0.936TS_{mm}$	0.94	22.4 W m^{-2}
$S_{ebl} = 0.321 + 0.355S_{mm} + 0.017T_{mm} - 0.011DP_{mm}$	0.57	0.11

Table IV-8. Total evaporation and sensible heat fluxes for the given periods and locations. The fluxes were calculated for different sensor heights with the single-level mean-profile equations. Positive values signify a flux directed towards the snow surface.

Site	Period	Sensor height (m)	Evaporation (mm)	Sensible Heat (MJ m-
Emerald Lake	May 1994	4.15	-0.86	+2.16
	(16 days)	2.22	-1.77	+6.35
		0.67	-4.14	+17.21
Emerald Lake	June 1994	4.90	+0.06	+3.05
	(16 days)	2.62	-0.01	+6.18
		1.07	-0.47	+13.27
Mini-catchments	April 1993	2.50	-17.92	+30.85
	(23 days)	1.20	-12.26	+18.83
Mini-catchments	June 1993	4.34	-0.35	+3.52
	(4 days)	1.85	-0.10	+1.53

Table IV-9. Estimated errors in calculated evaporation (in %) attributed to the mean-profile assumptions. The combined error is calculated from the square root of the sum of the squares of the individual errors. Percentages are given for both reasonable and high estimates of the probable error.

	Accumulation Season (winter)		Snowmelt Season	
Item	Reasonable	High	Reasonable	High
Snow-surface temperature	25	50	5	10
Instrument height	10	20	5	10
Roughness length (z_0)	10	25	10	25
Roughness length ratio (z_{0v}/z_0)	6	6	6	6
Combined probable error	30	60	14	30

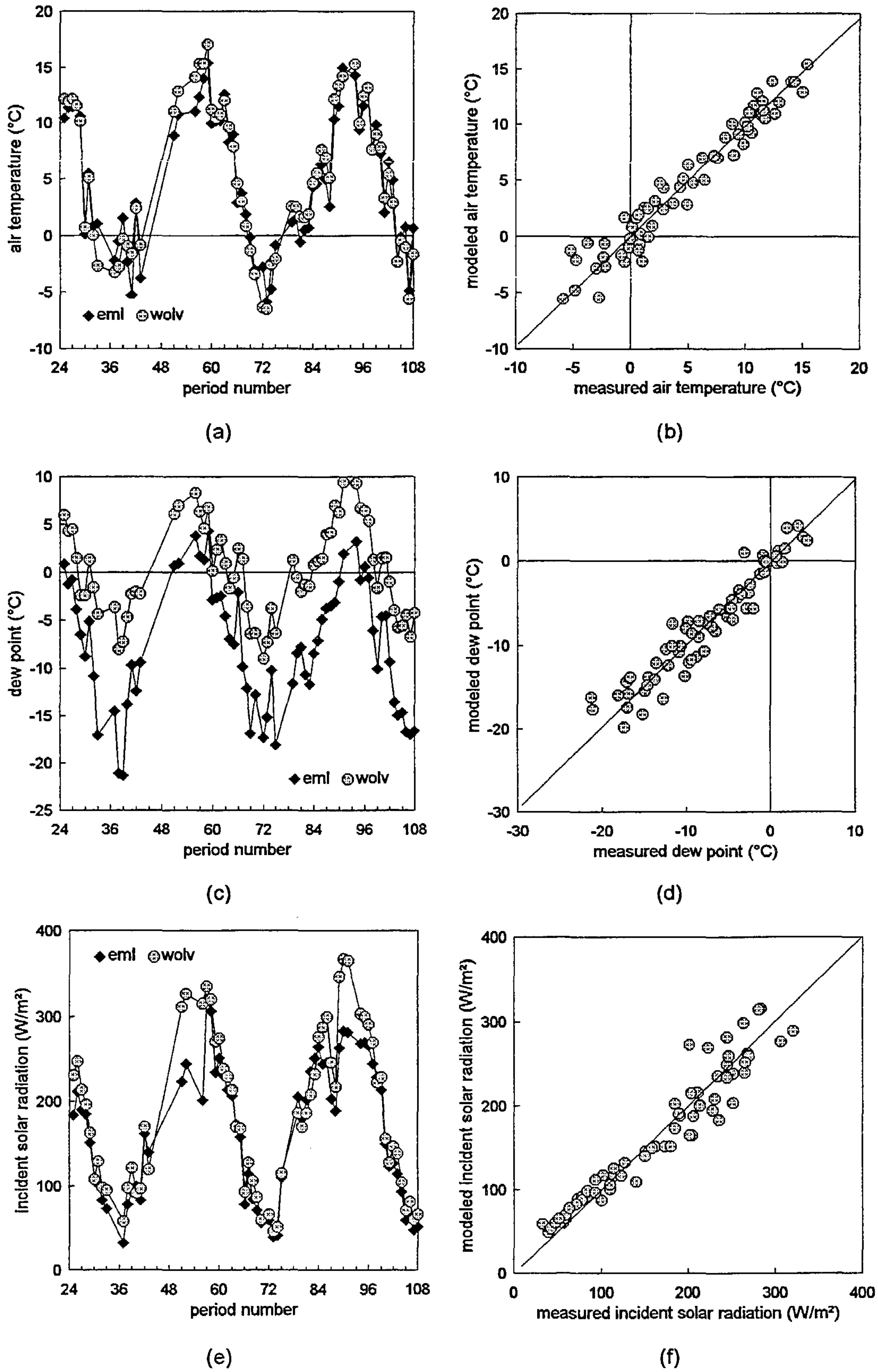


Figure IV-1: (a) Air temperature at Emerald Lake (eml) and Wolverton (wolv). (b) Modeled vs. measured air temperature at Emerald ($r^2 = 0.96$). (c) Dew point at Emerald and Wolverton. (d) Modeled vs. measured dew point at Emerald ($r^2 = 0.92$). (e) Incident solar radiation at Emerald and Wolverton. (f) Modeled vs. measured incident solar radiation at Emerald ($r^2 = 0.92$). All figures are based on period (~10 days) averages of data from Sept. 1 through 10, 1991 (period 25) to Dec. 21 through 31, 1993 (period 108). Modeled Emerald values are based on the regressions between Emerald and Wolverton.

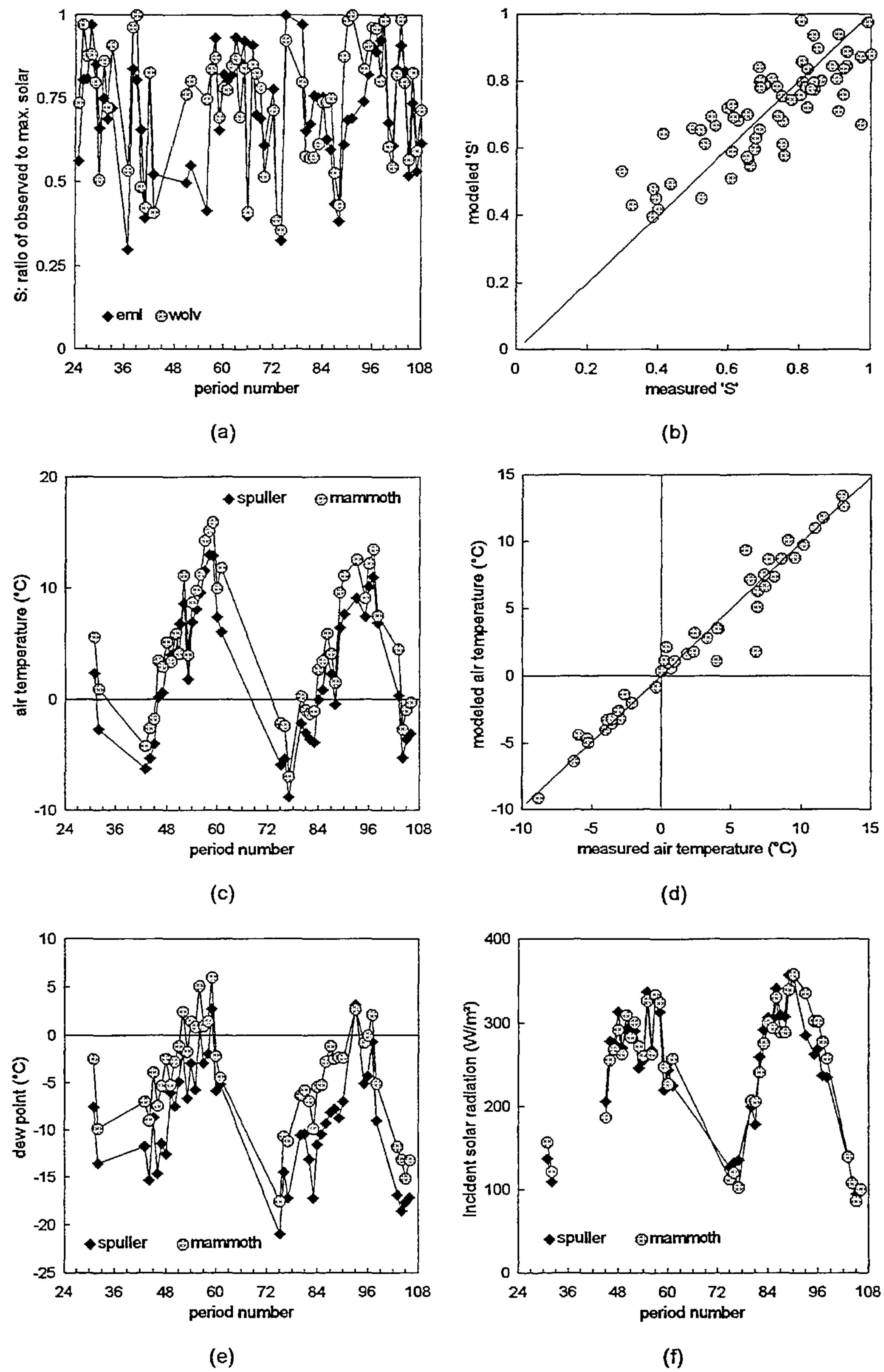


Figure IV-2: (a) S: the ratio of observed-to-maximum-possible sunshine duration at Emerald and Wolverton. (b) Modeled vs. measured S at Emerald ($r^2 = 0.64$). (c) Air temperature at Spuller Lake and Mammoth Mountain. (d) Modeled vs. measured air temperature at Spuller ($r^2 = 0.96$). (e) Dew point at Spuller and Mammoth. (f) Incident solar radiation at Spuller and Mammoth. All figures are based on period (~10 days) averages of data from period 25 to period 108 at Emerald and Wolverton, and from period 31 (Nov. 1 though 10, 1991) to period 106 (Dec. 1 through 10, 1993) at Spuller and Mammoth. Modeled Spuller air temperature in (d) is based on the regression between Spuller and Mammoth.

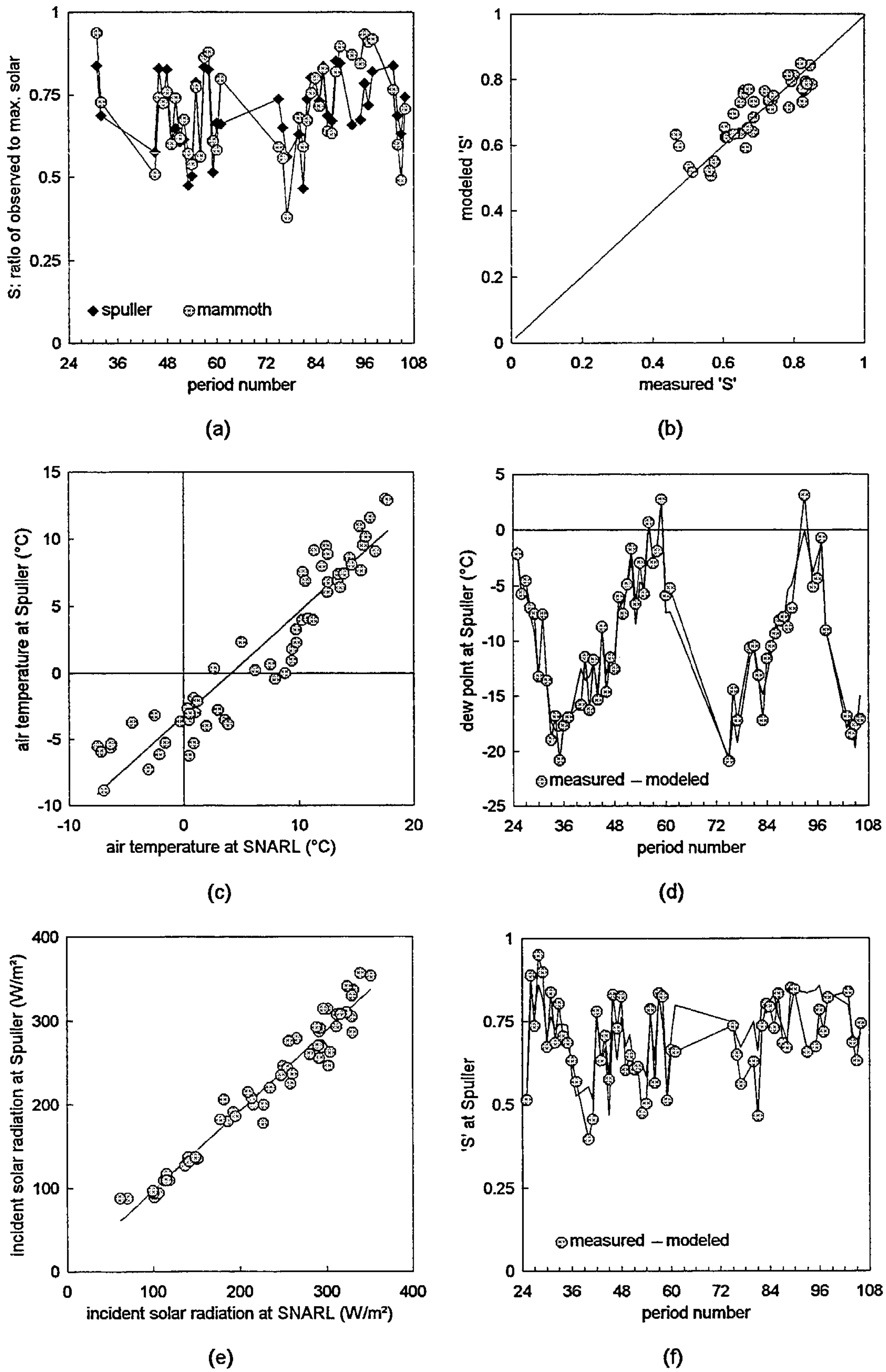
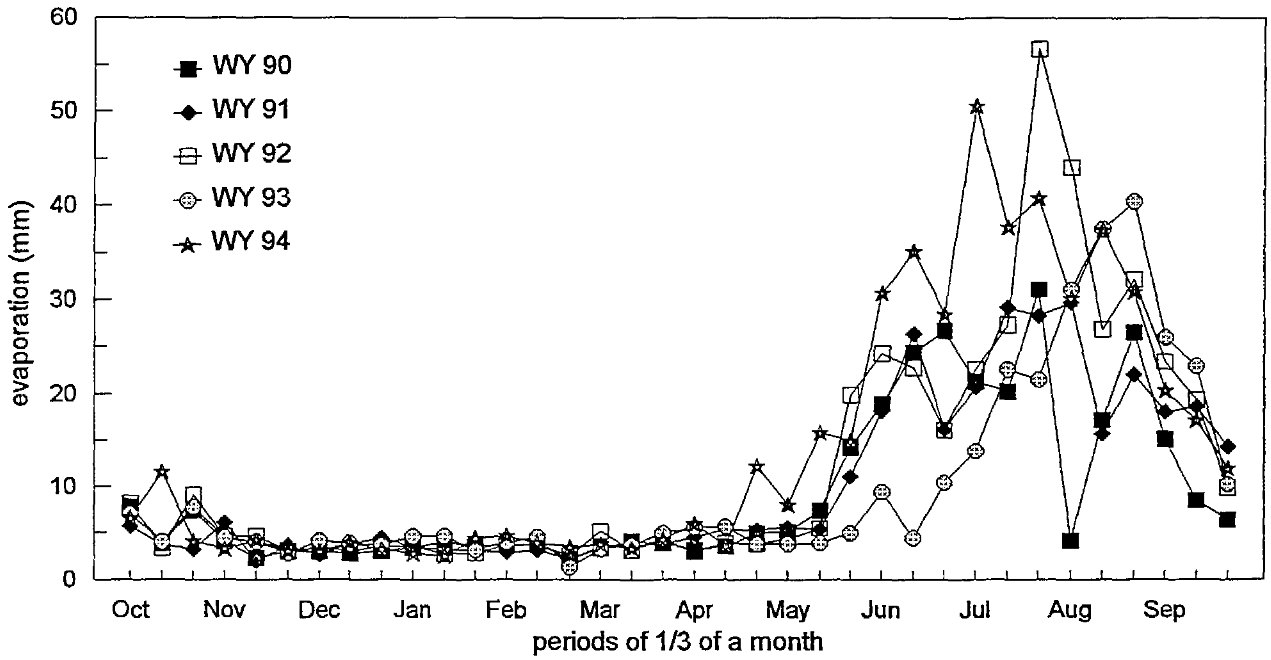
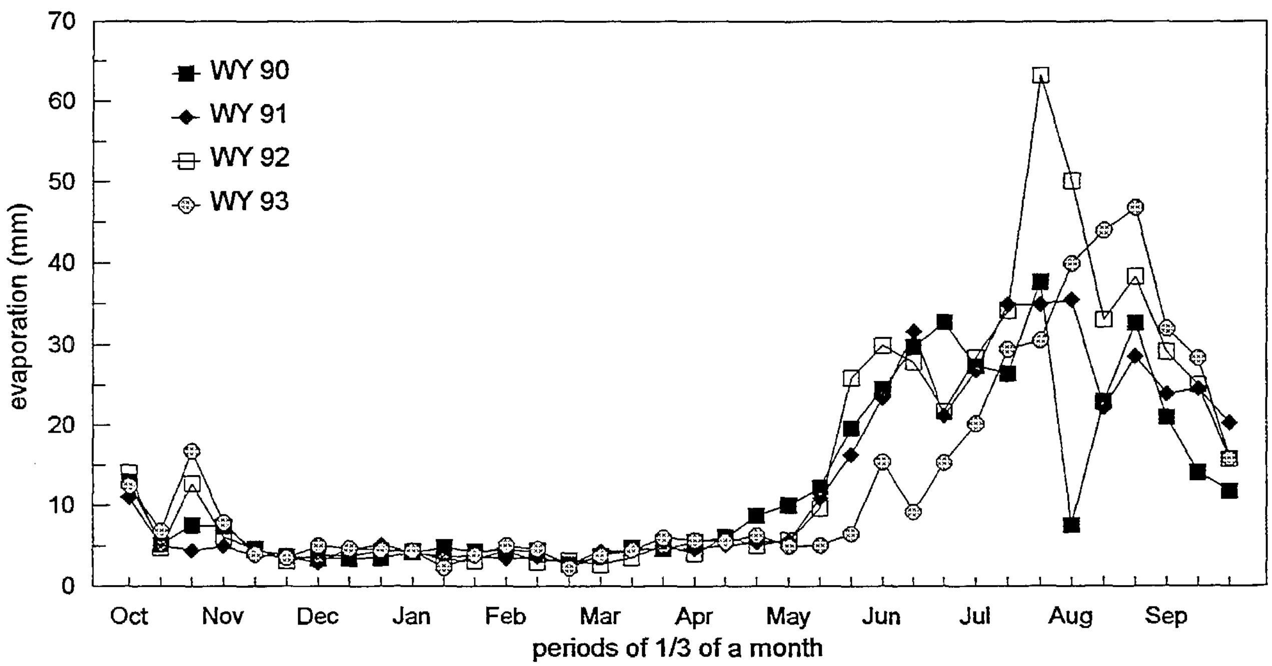


Figure IV-3: (a) S: the ratio of observed to maximum possible sunshine duration at Spuller Lake and Mammoth Mountain. (b) Modeled (on Mammoth) vs. measured S at Spuller ($r^2 = 0.71$). (c) Air temperature at Spuller vs. air temp. at SNARL ($r^2 = 0.89$). (d) Modeled (on SNARL) and measured dew point at Spuller ($r^2 = 0.94$). (e) Incident solar radiation at Spuller vs. incident solar at SNARL ($r^2 = 0.96$). All figures are based on period (~10 days) averages. Regression equations for the modeled values are shown in Table IV-5.

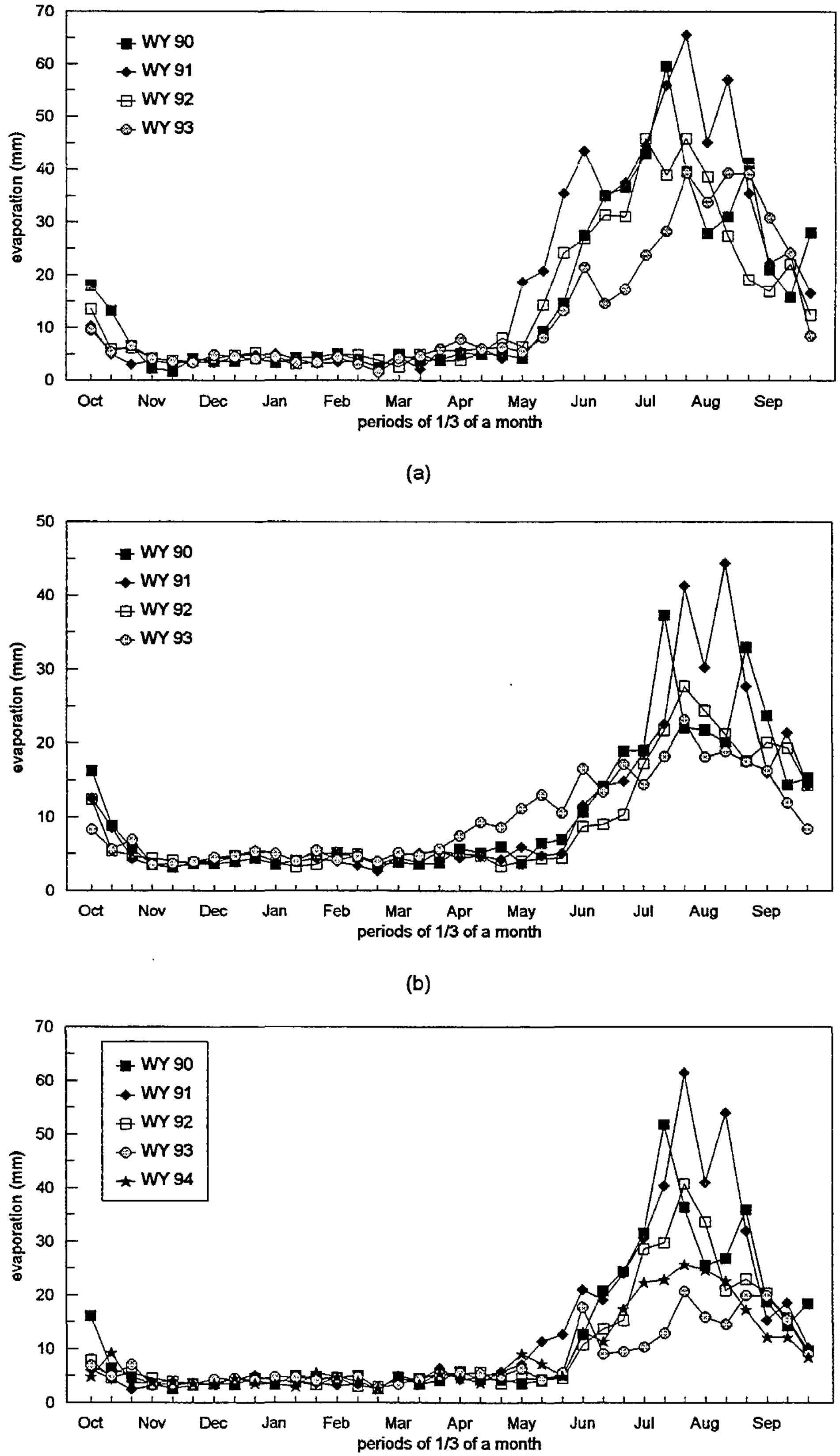


(a)



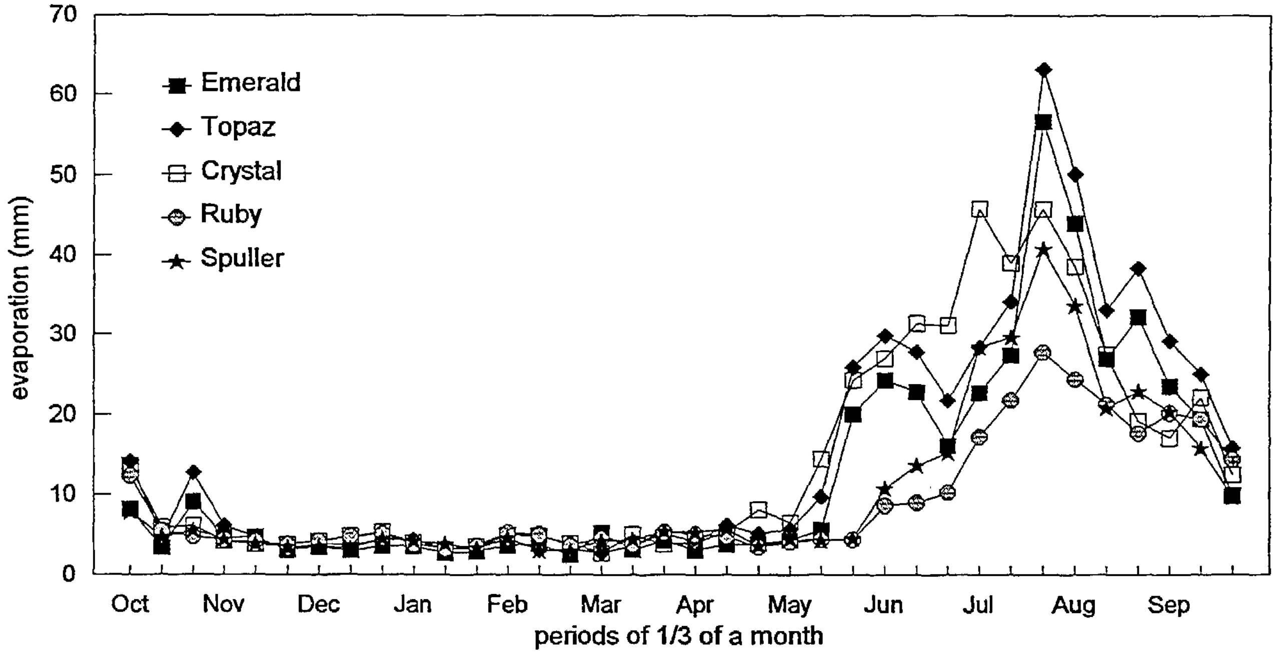
(b)

Figure IV-4: Evaporation in mm of water equivalent lost during the period (~10 day totals) calculated by the CRAE model for water years (WY) 1990 through 1994 for (a) the Emerald Lake catchment and through 1993 for (b) the Topaz Lake catchment. The water year begins on Oct. 1 and ends on the following Sept. 30. Evaporation at Pear Lake is identical to that at Emerald (same elevation and meteorological data).

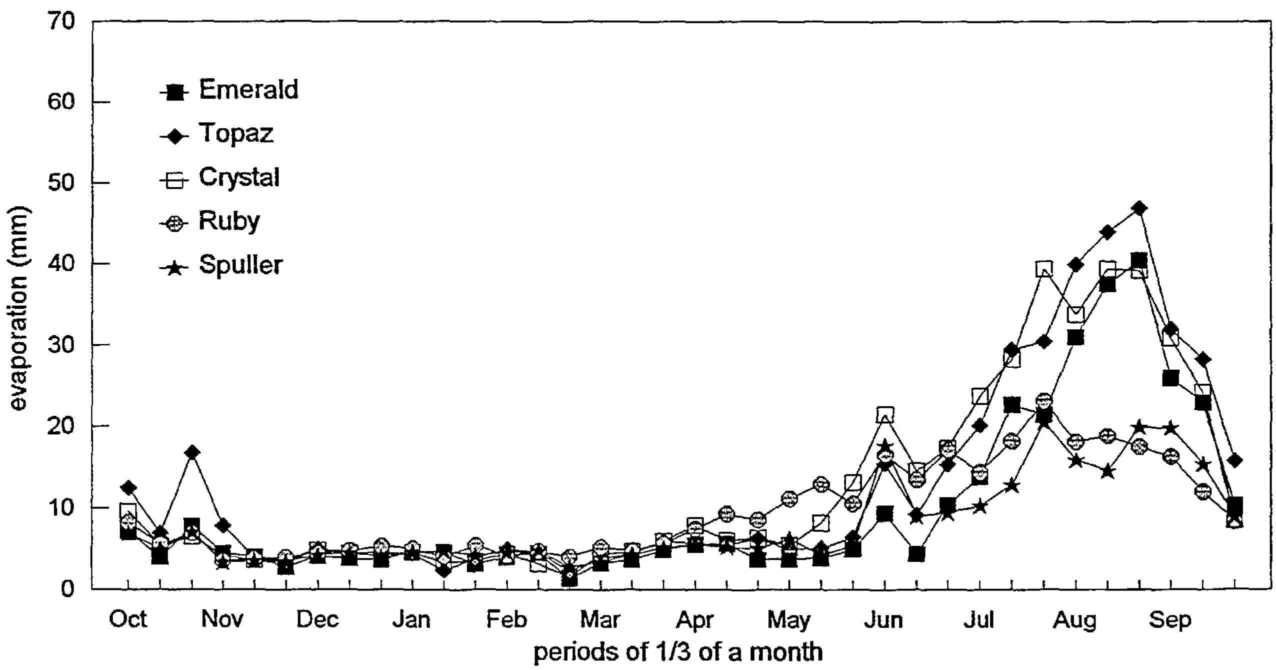


(c)

Figure IV-5: Evaporation in mm of water equivalent lost during the period (~10 day totals) calculated by the CRAE model for water years (WY) 1990 through 1993 for (a) the Crystal Lake catchment, (b) the Ruby Lake catchment and (c) through 1994 for the Spuller Lake catchment. The water year begins on Oct. 1 and ends on the following Sept. 30.

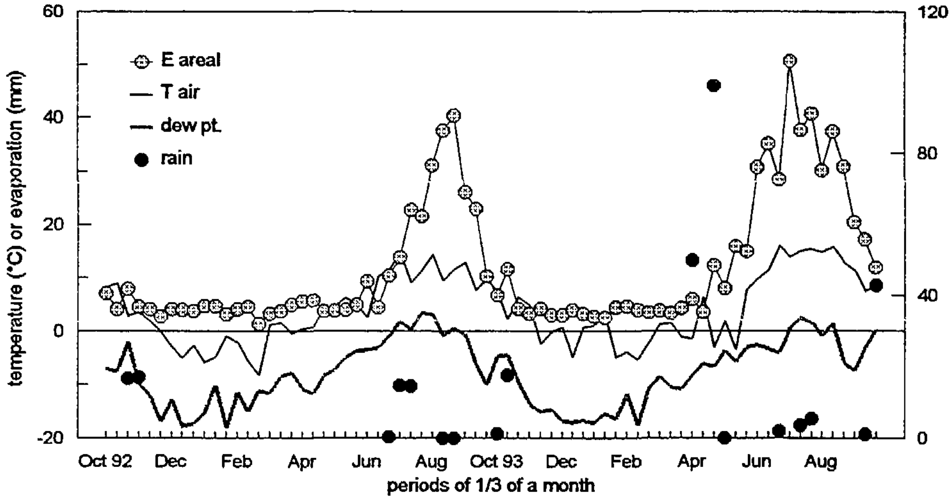


(a)

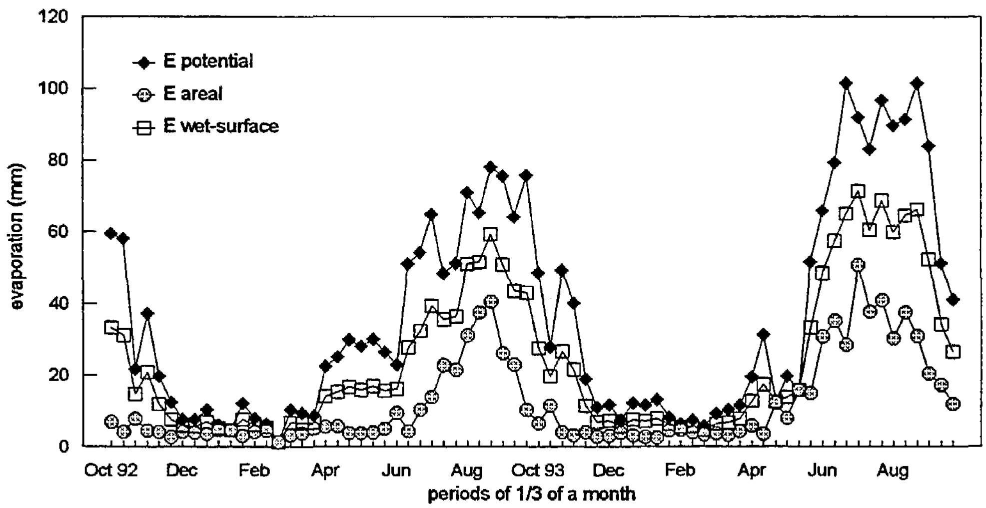


(b)

Figure IV-6: Evaporation in mm of water equivalent lost during the period (~10 day totals) calculated by the CRAE model for 5 study catchments (Pear is omitted as identical to Emerald) for water years (a) 1992, and (b) 1993.



(a)



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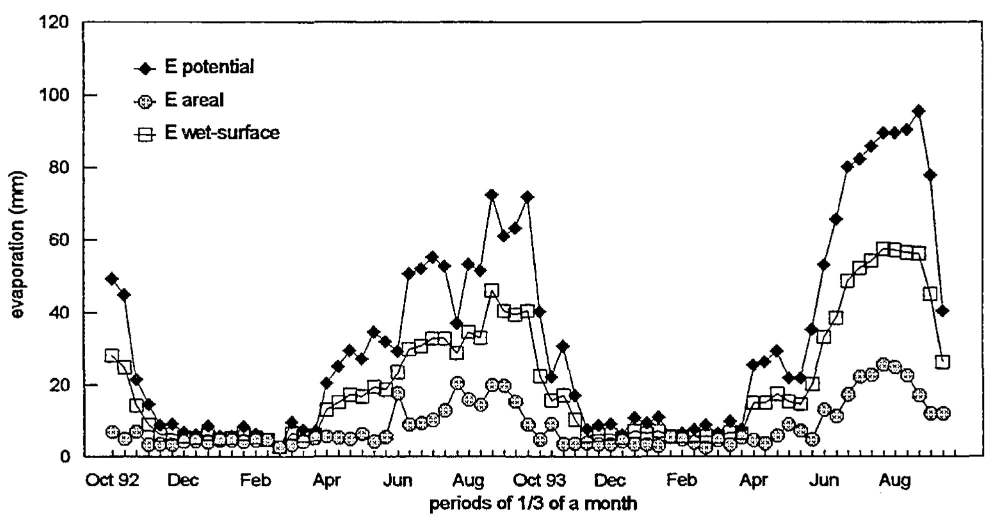
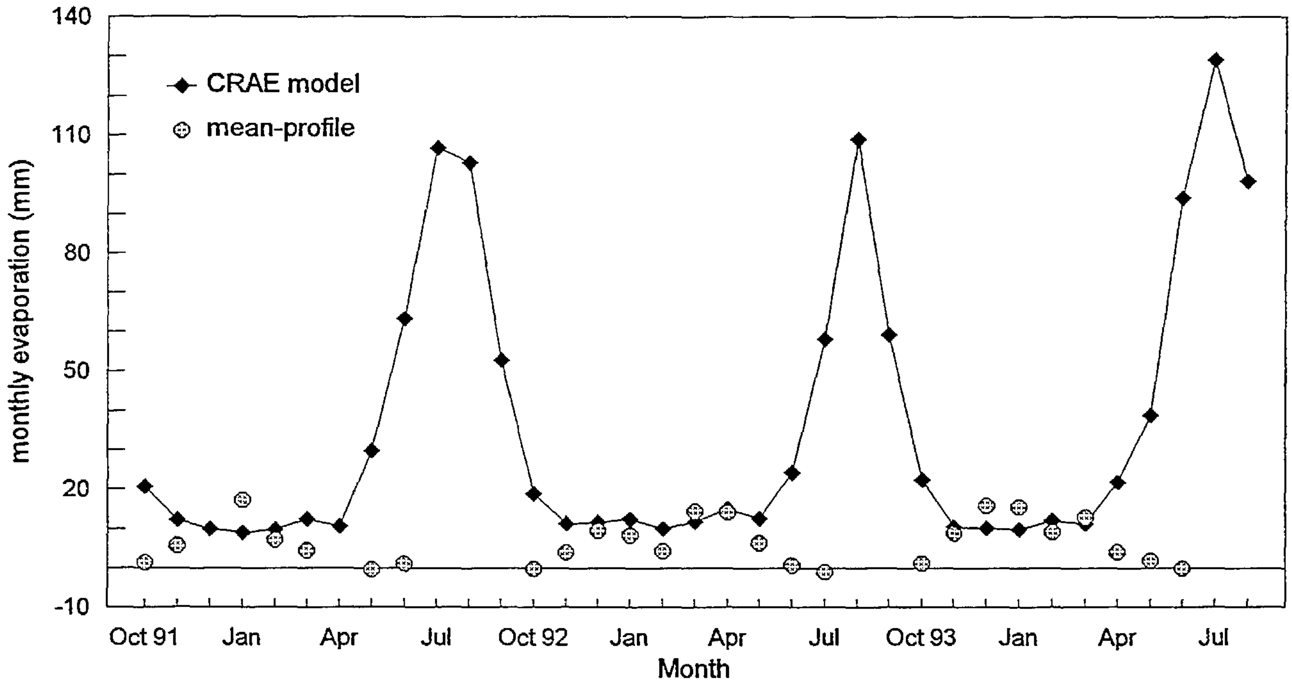
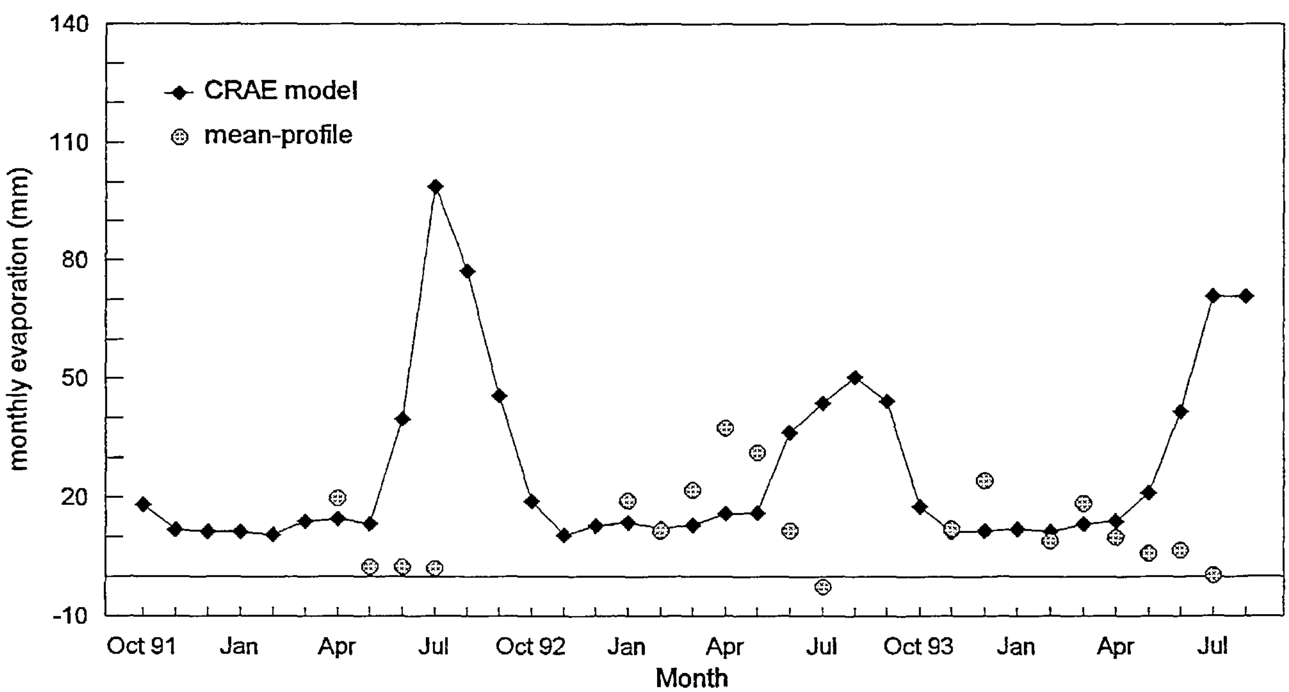


(c)

Figure IV-7: (a) Air temperature (Tair), dew point, rain and CRAE modeled evaporation (E areal) in mm of water equivalent lost during the period (~10 day totals) at Emerald Lake for water years 1993 and 1994. Period evaporation – potential, areal and wet-surface – as calculated by CRAE model for water years 1993 and 1994 at (b) Emerald Lake and (c) Spuller Lake.

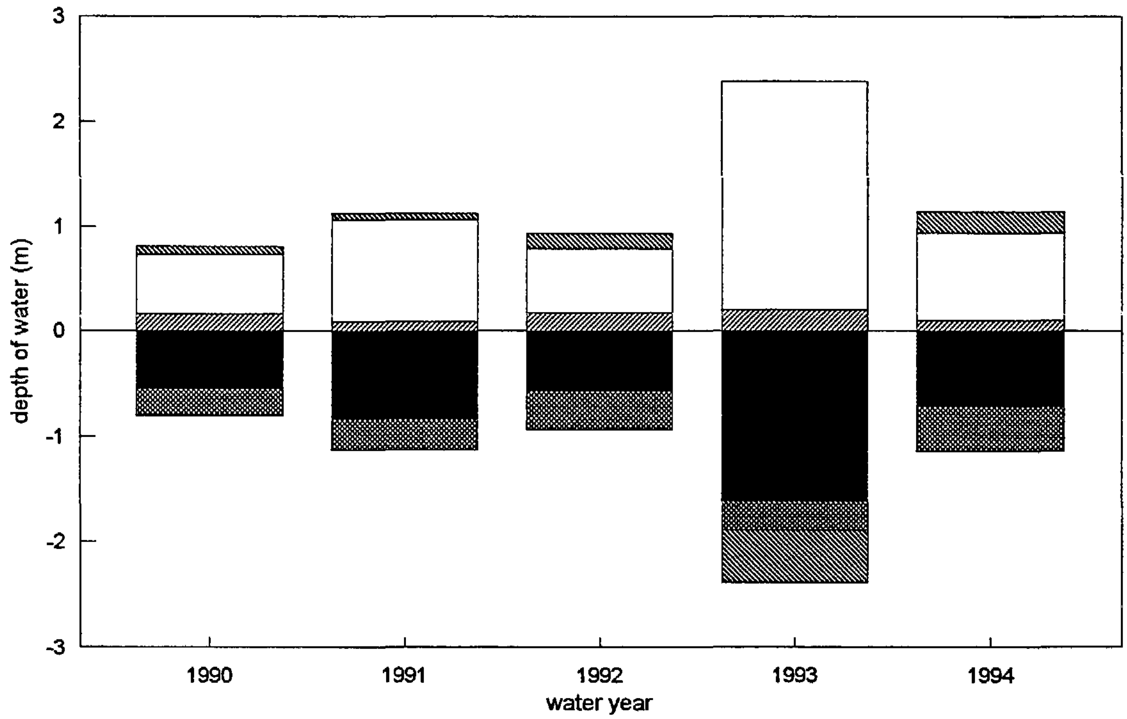


(a)



(b)

Figure IV-8: Monthly evaporation estimates from CRAE model (areal evaporation) and the mean-profile calculations for water years 1992, 1993 and 1994 at (a) Emerald Lake and (b) Spuller Lake. Mean-profile results are the hourly point model summations of evaporation from snow at the catchment weather stations; missing points indicate either the lack of meteorological data or the absence of snow.



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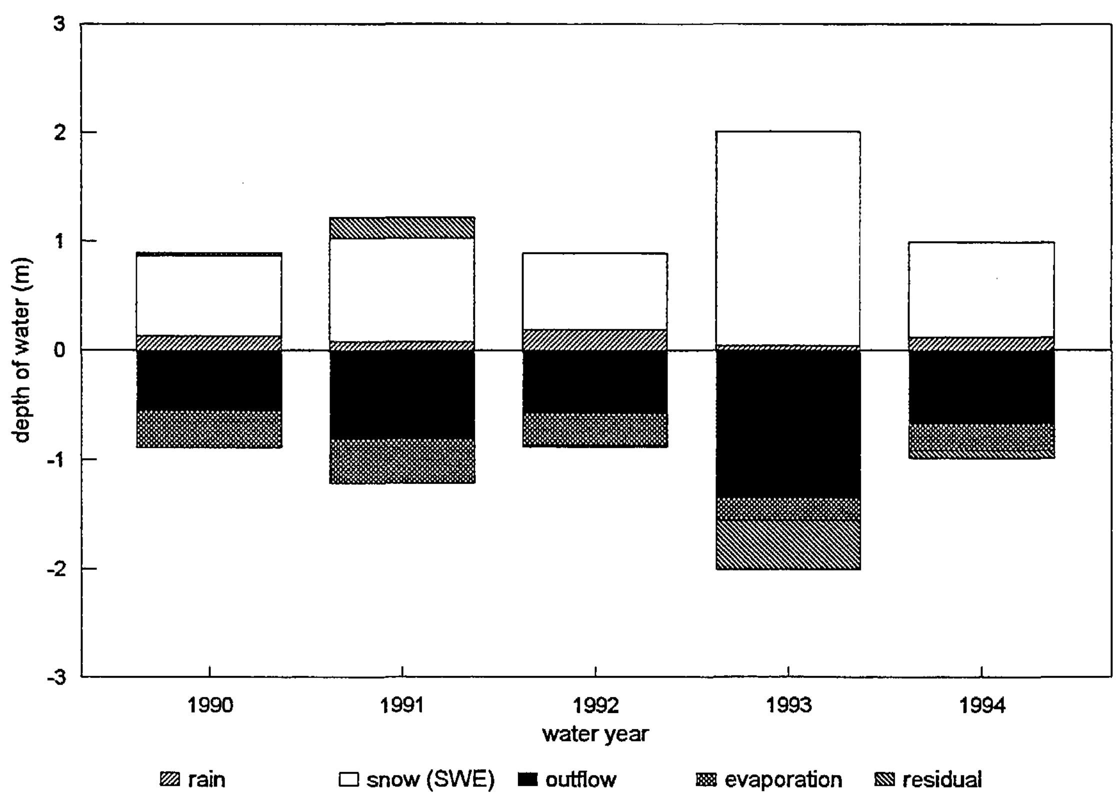
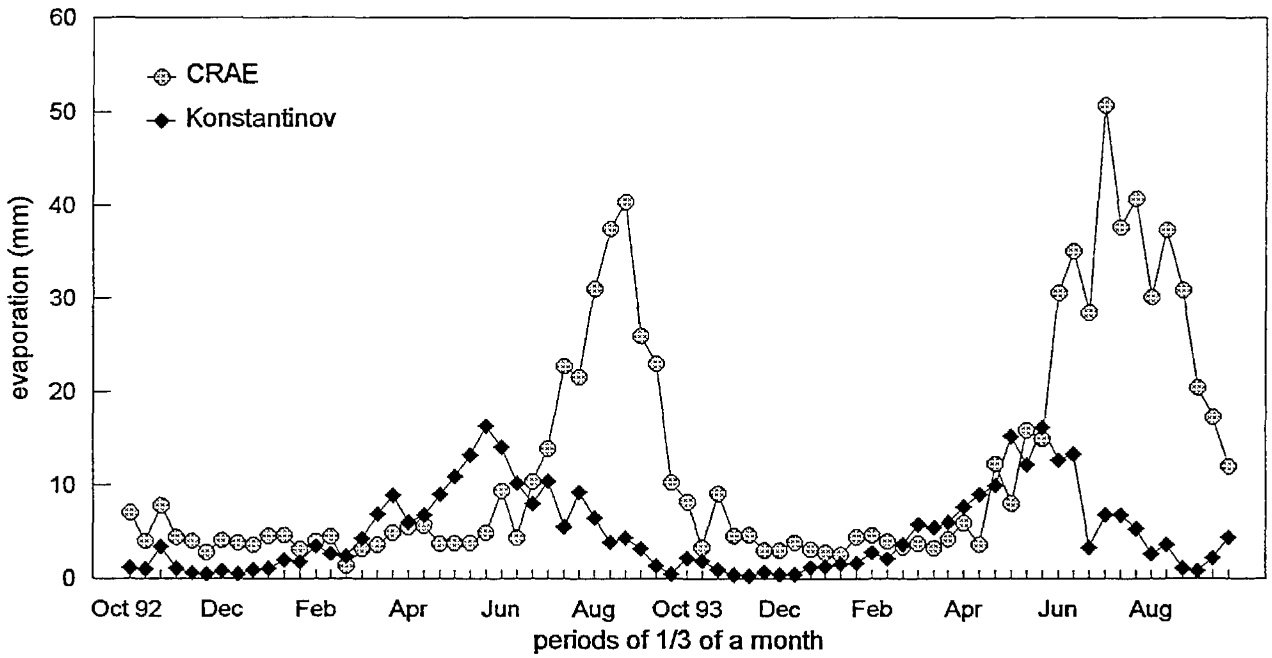
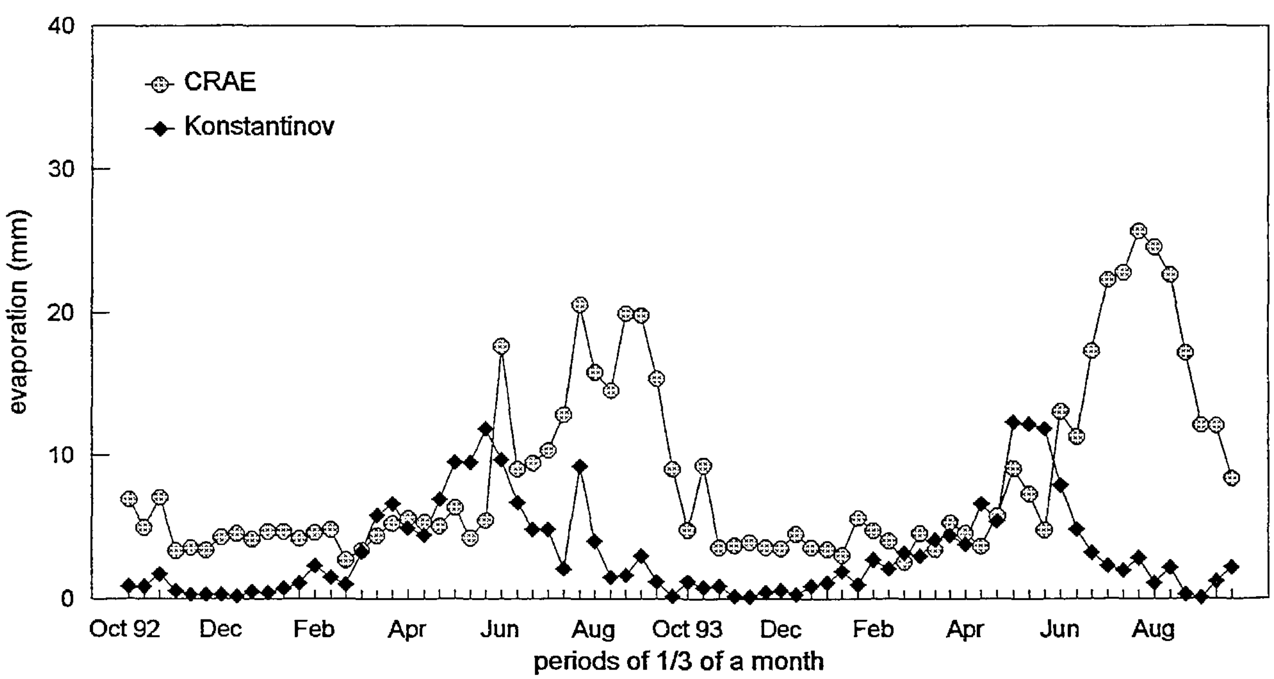


(b)

Figure IV-9: Annual water balances for the (a) Emerald Lake and (b) Spuller Lake catchments for water years 1990 through 1994. The water balance residual is the amount required to completely balance the annual input and output. A positive residual represents an excess of output over input: either an underestimate of precipitation or an overestimate of outflow or evaporation. Evaporation was estimated from CRAE model.



(a)



(b)

Figure IV-10: A comparison between the CRAE and Konstantinov evaporation models for water years 1993 and 1994 at the (a) Emerald Lake and (b) Spuller Lake catchments. Modeled evaporation is the total water equivalent lost over the ~10 day period.

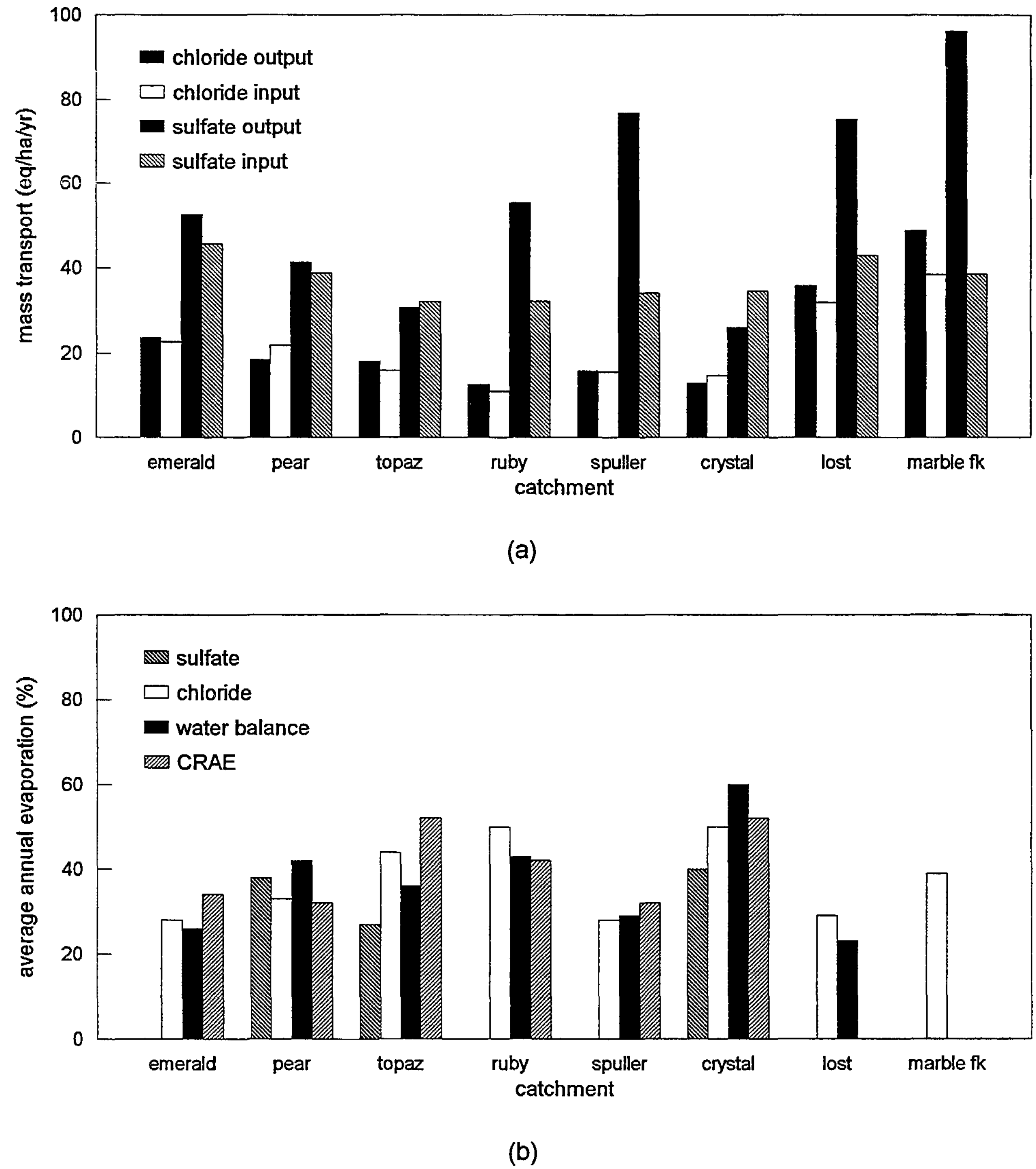
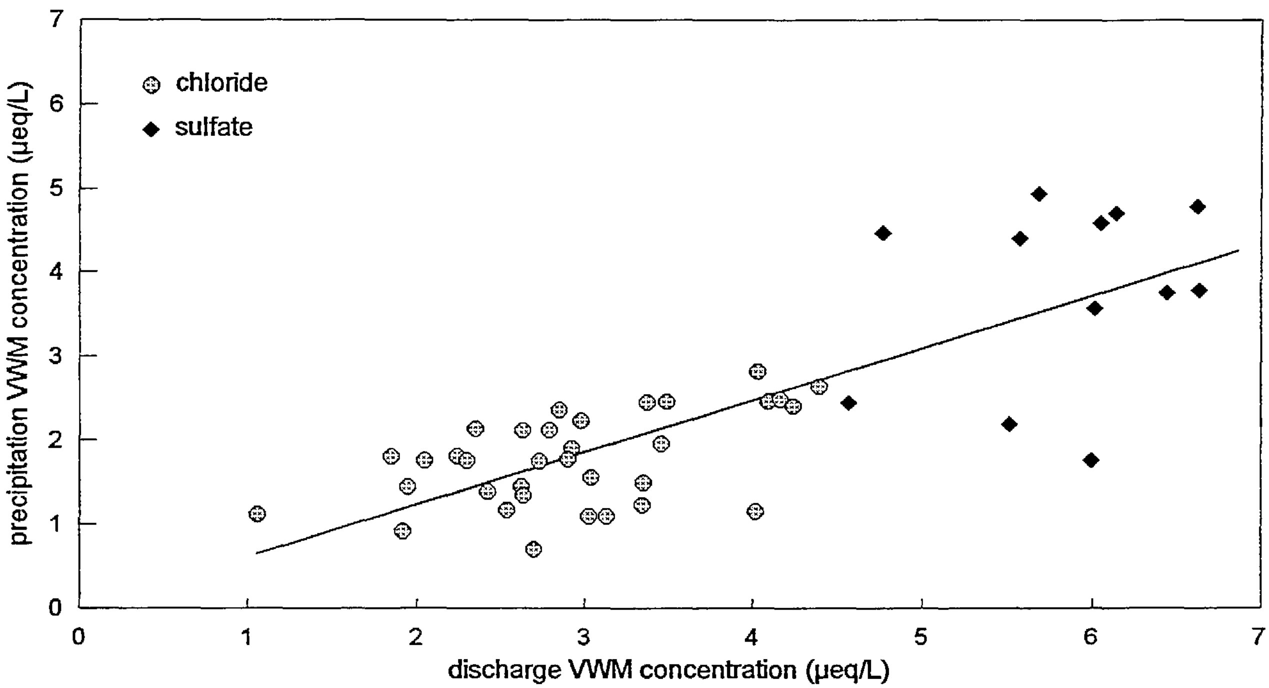
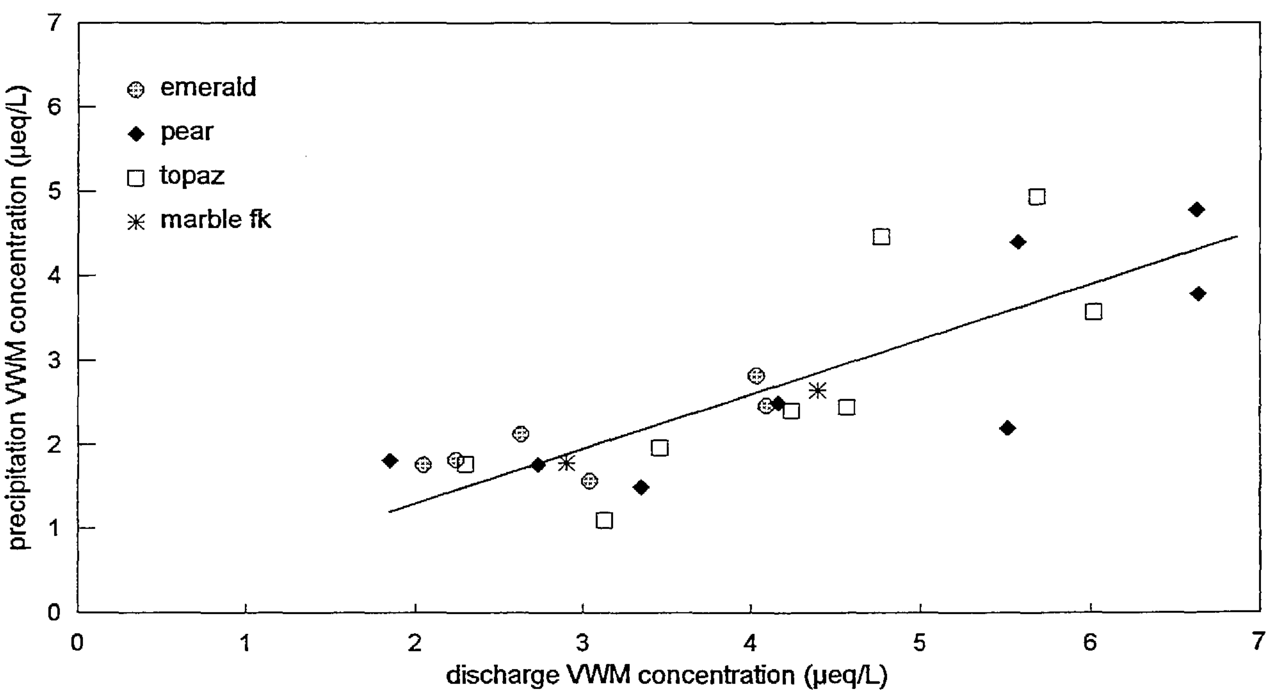


Figure IV-11: (a) The average annual study catchment mass balance for chloride and sulfate: all of the catchments appear conservative for chloride, and Crystal, Pear and Topaz appear conservative for sulfate. (b) Average annual evaporation, as percent of water year precipitation, determined by the evaporative concentration of sulfate and chloride, the water balance and CRAE model for the study catchments. Values represent 5-year averages for Emerald, Spuller and Ruby (1990-1994) and 4-year averages for the other catchments (1990-1993).

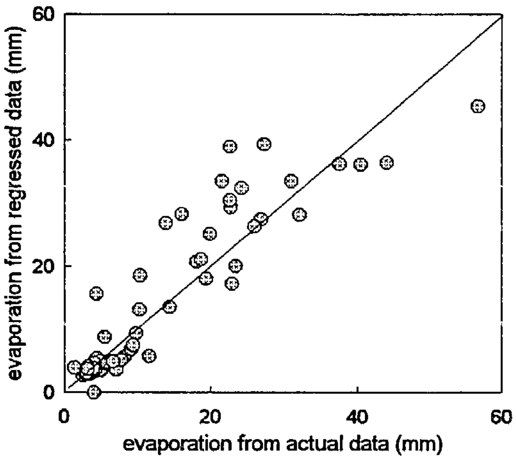


(a)



(b)

Figure IV-12: Annual (water year) precipitation vs. discharge VWM chloride and sulfate concentrations for (a) all study catchments and (b) the "westside" catchments of the upper Marble Fork; sulfate concentrations are shown only for catchments conservative for sulfate: Pear, Crystal and Topaz. The evaporation factor is equal to 1 minus the slope of the regression line.



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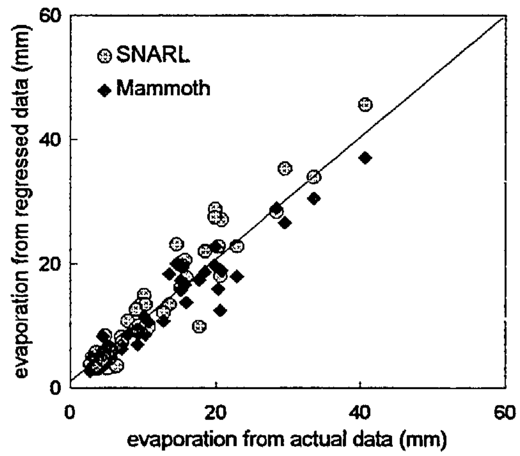

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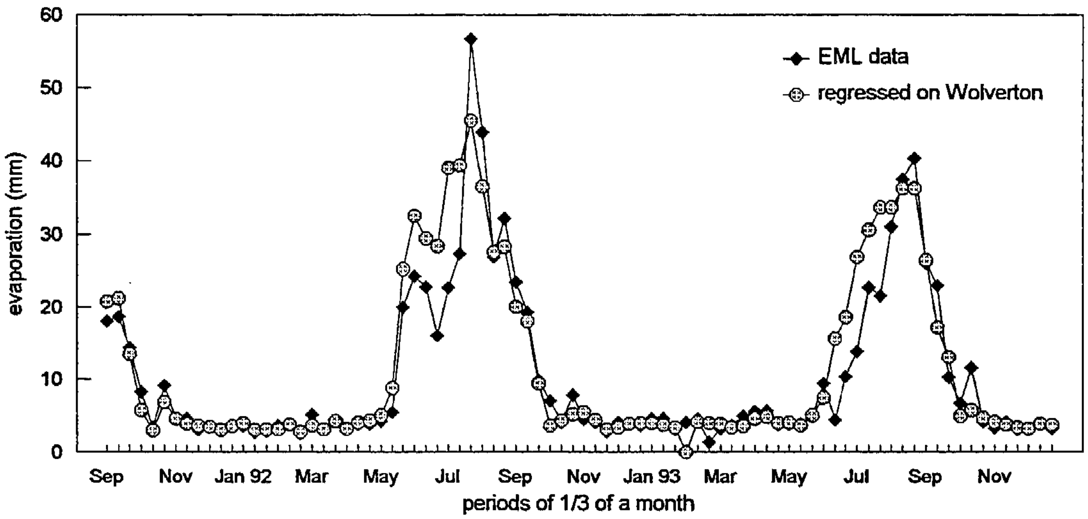

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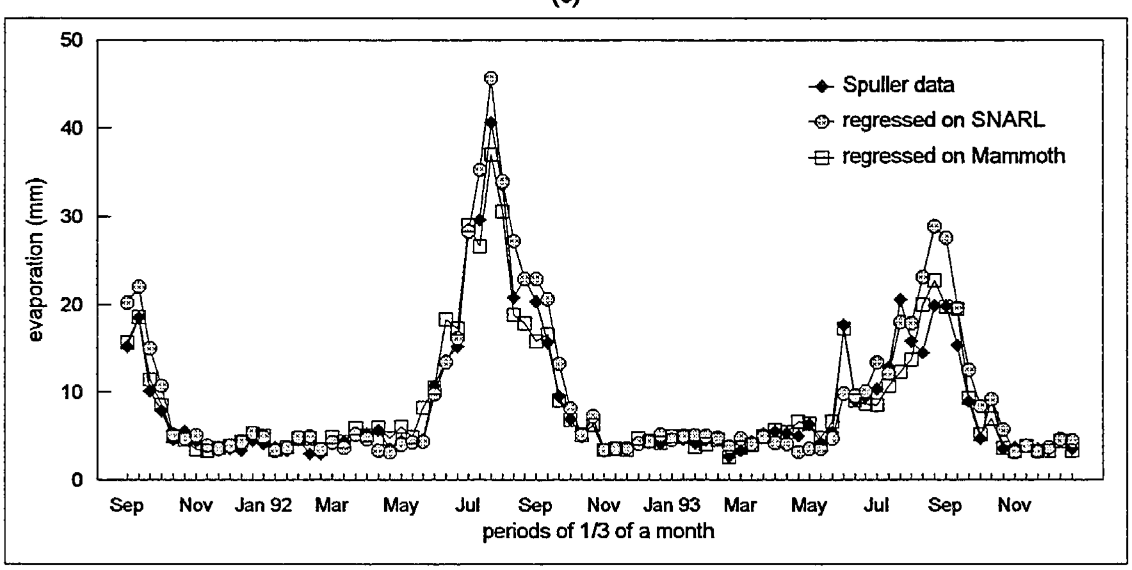


(d)

Figure IV-13: Comparison of CRAE evaporation calculated from measured vs. modeled meteorological data: (a) a scatterplot of period estimates from Sept. 1991 to Dec. 1993 at Emerald (r^2 = 0.86); (b) scatterplot of period estimates at Spuller; separate estimates are shown for meteorological data regressed on SNARL (r^2 = 0.93) and on Mammoth r^2 = 0.94); (c) period evaporation at Emerald; and (d) period evaporation at Spuller.

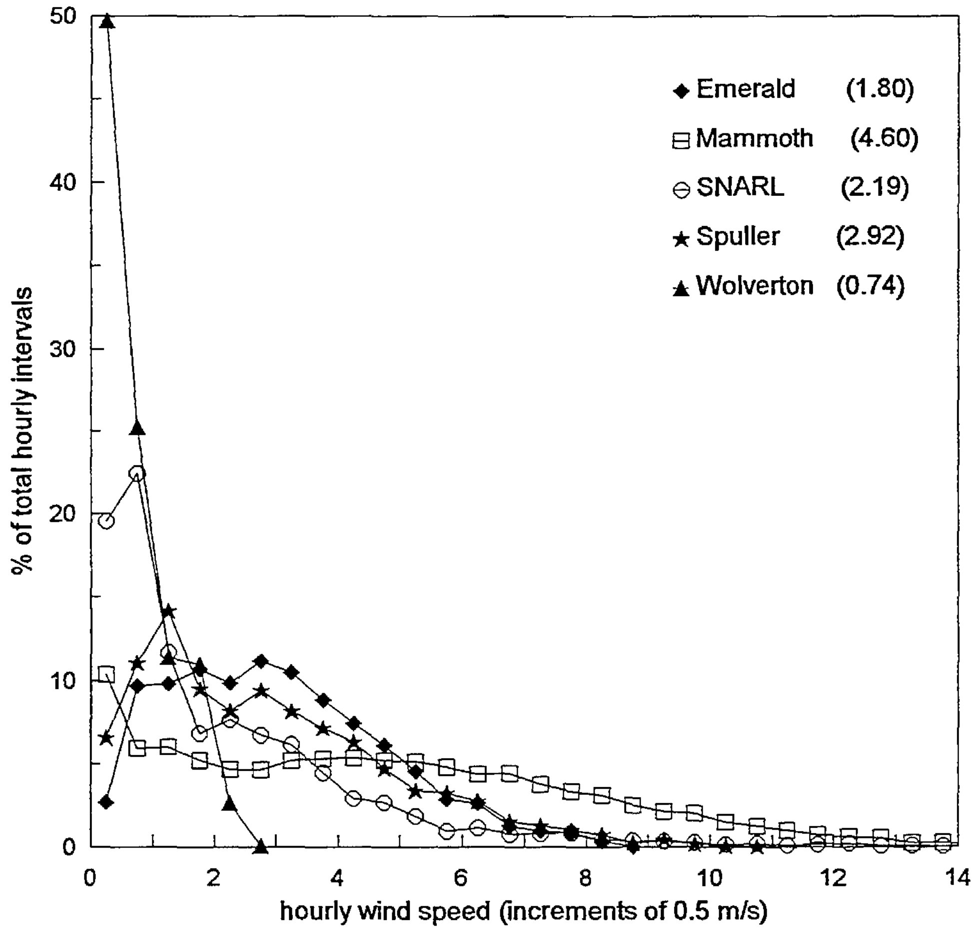


Figure IV-14: Seasonal distribution (December through May) of wind speed at 5 of the study weather stations during water year 1993; seasonal means for each site are shown on the legend in parenthesis.

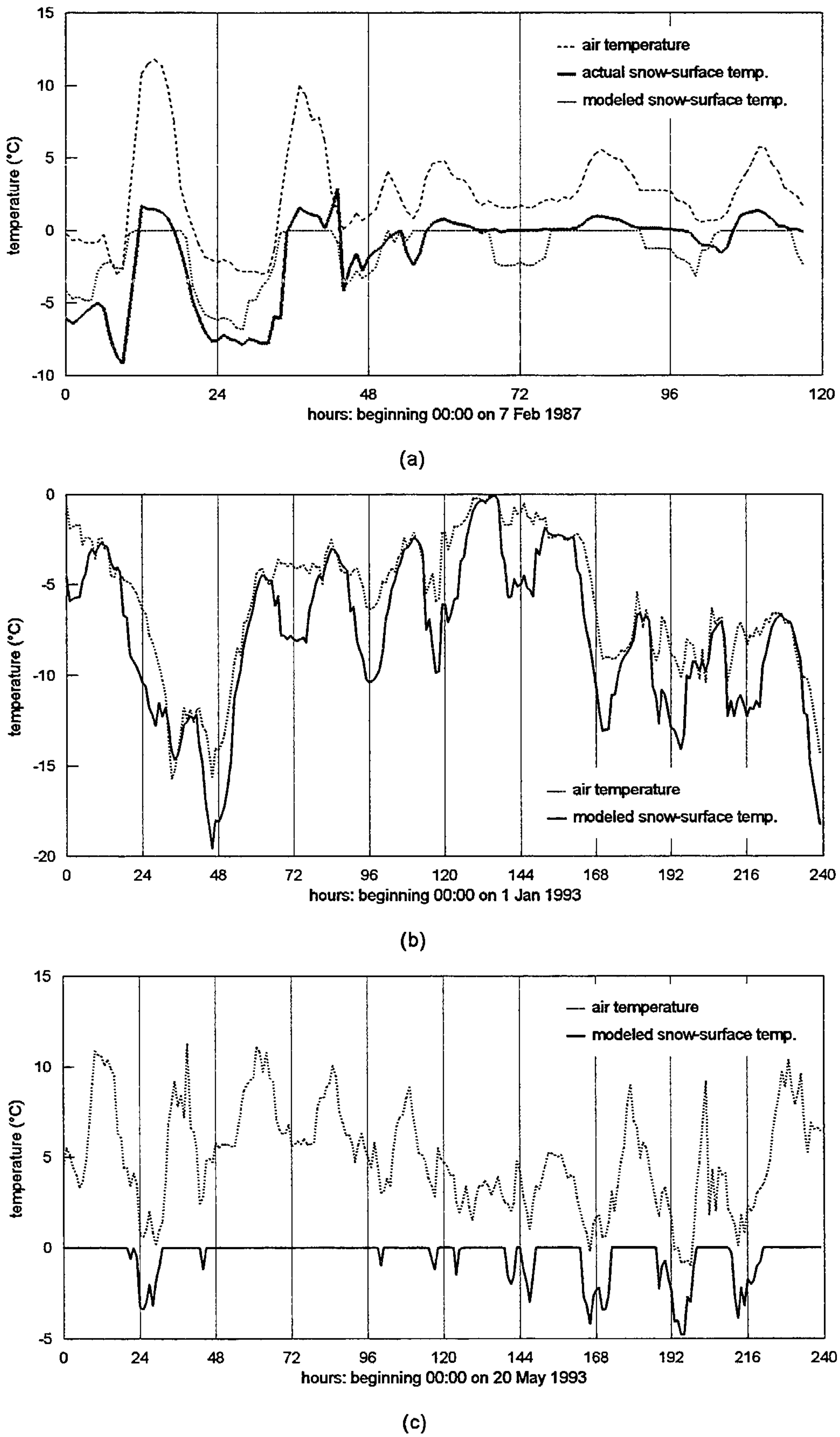


Figure IV-15: (a) Surface air temperature and observed and modeled snow-surface temperatures for 5 days in February 1987 at Tahoe CA (data from the CSSL). (b) Air and modeled snow-surface temperatures at Emerald Lake for 10 winter days in Janurary 1993. (c) Air and modeled snow-surface temperatures at Emerald Lake for 10 late spring days in May 1993.

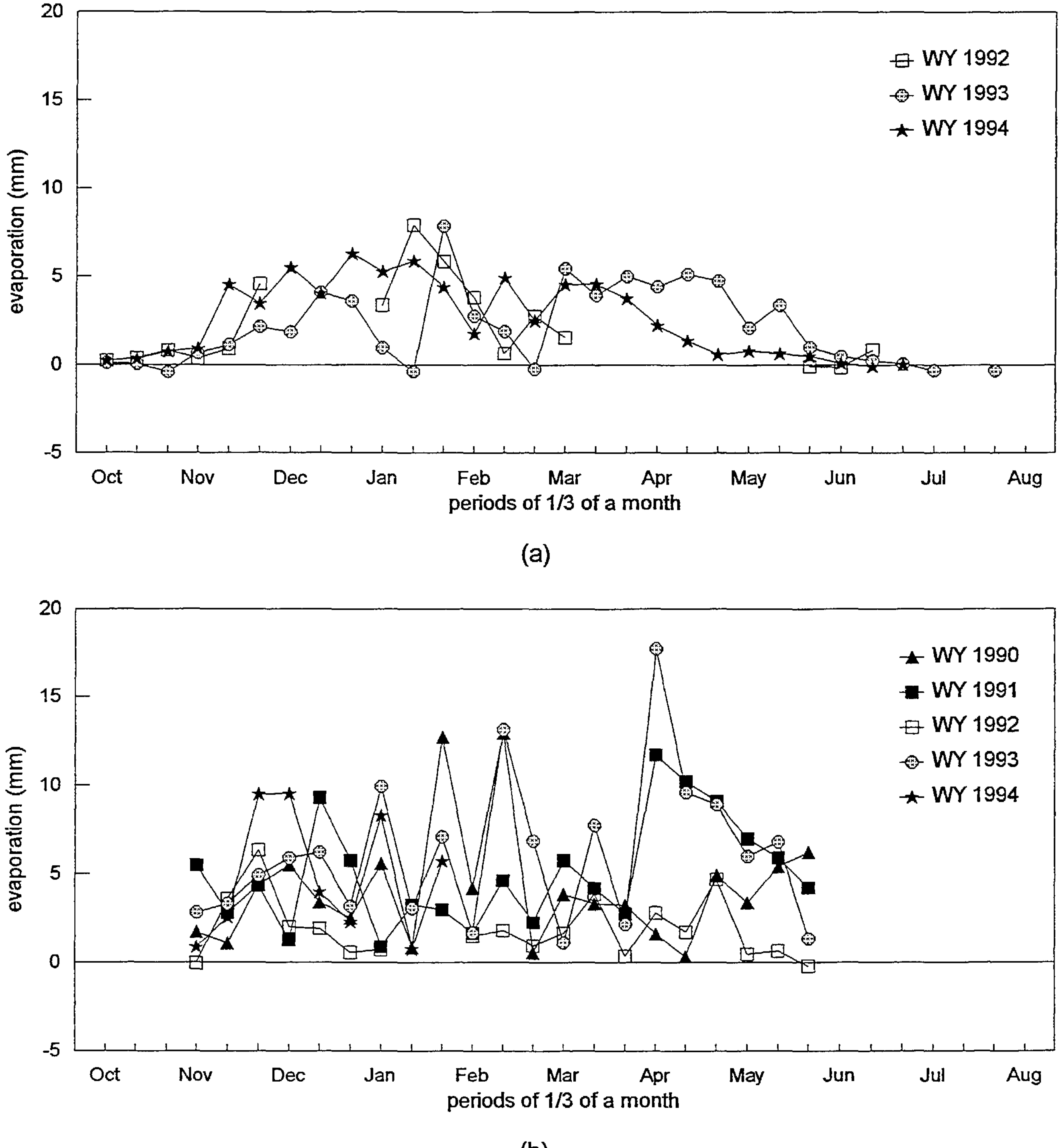
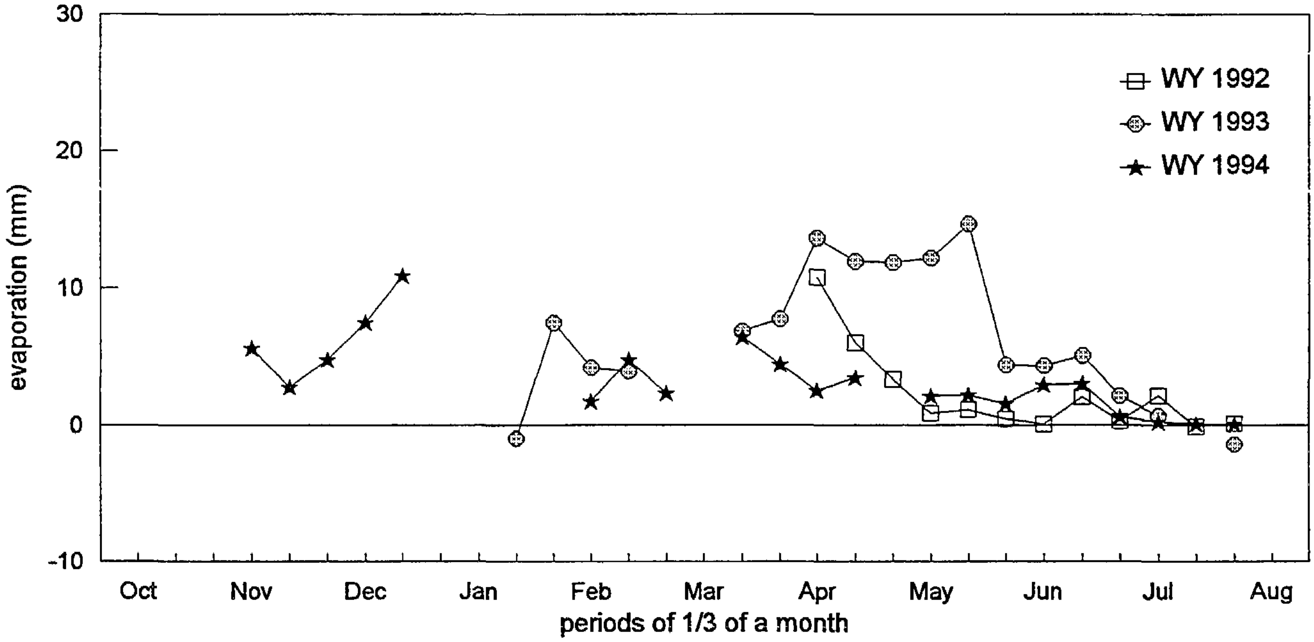


Figure IV-16: Period (~10 days) totals of evaporation from snow at (a) Emerald Lake and (b) SNARL. Points represent the sum, over each period, of hourly mean-profile estimates. Negative values indicate condensation.



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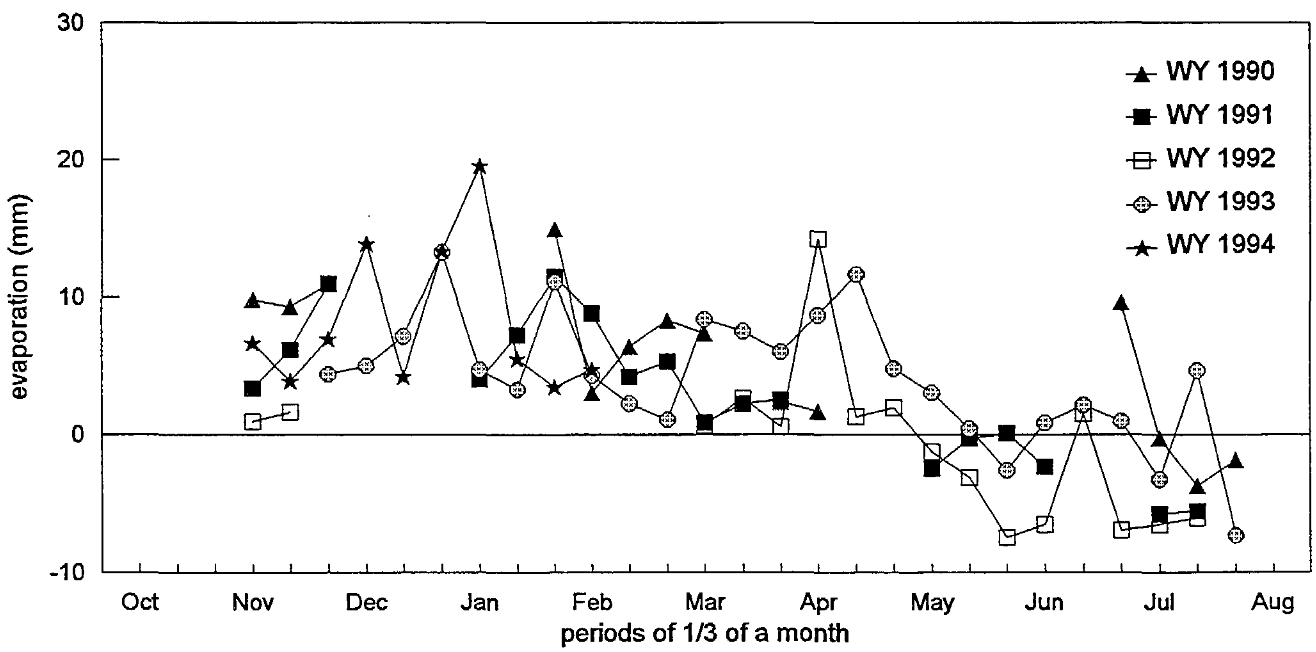
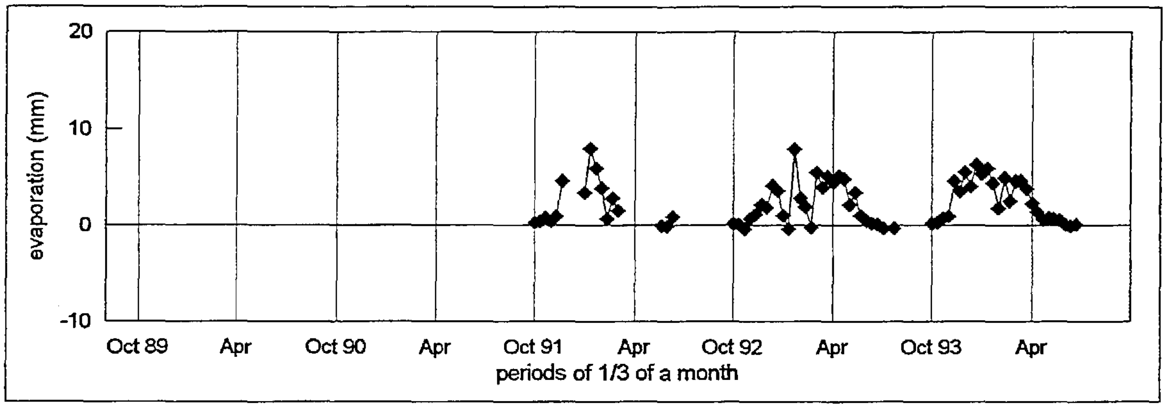


(b)

Figure IV-17: Period (~10 days) totals of evaporation from snow at (a) Spuller Lake and (b) Mammoth Mountain. Points represent the sum, over each period, of hourly mean-profile estimates at Spuller, 15 minute estimates at Mammoth. Negative values indicate condensation.



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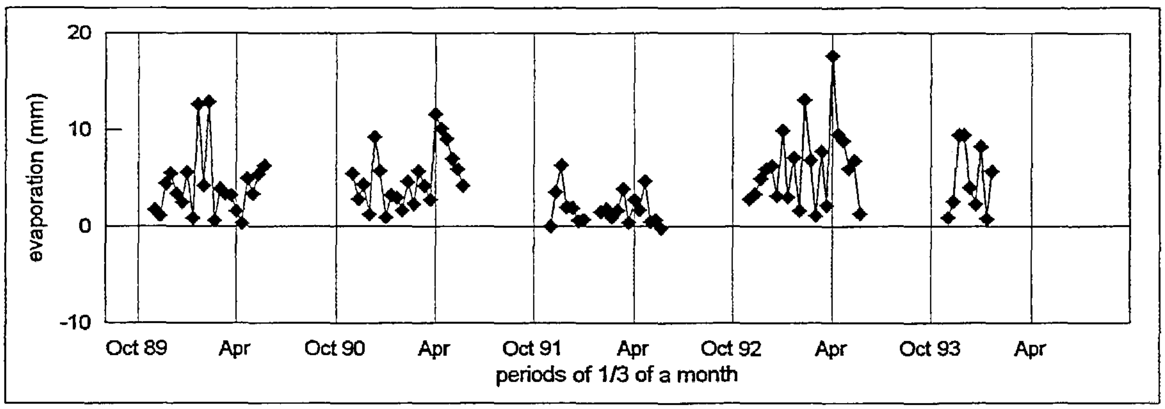

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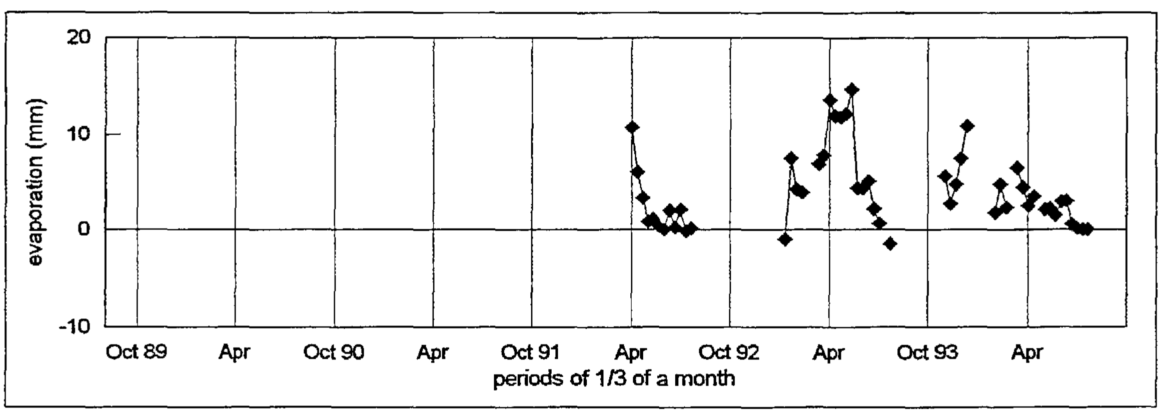

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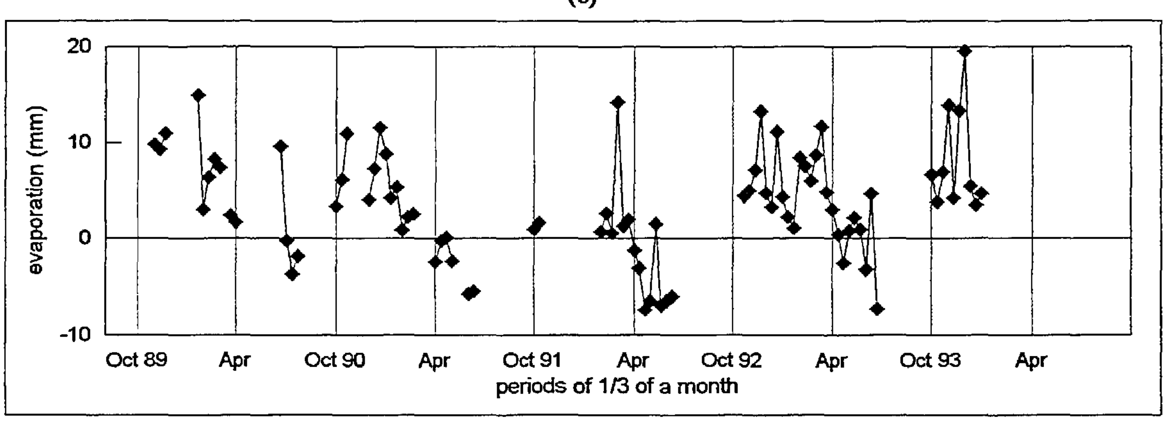
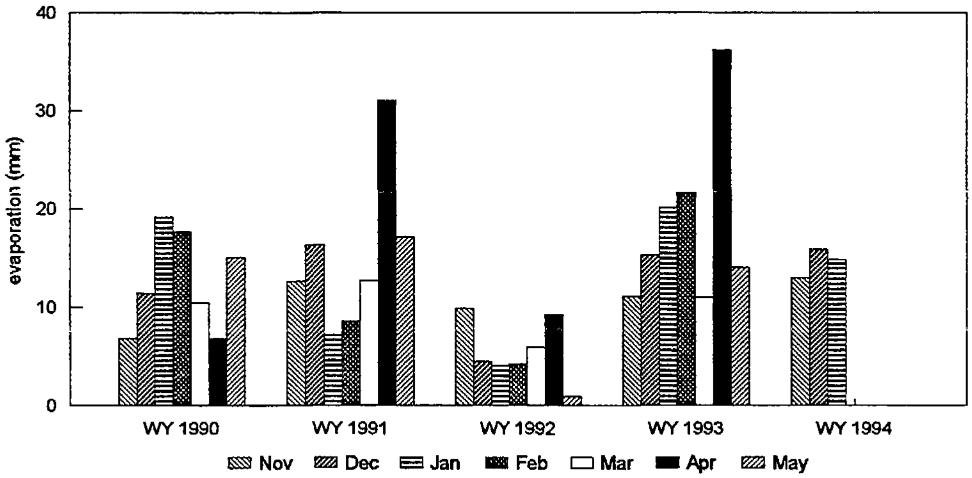
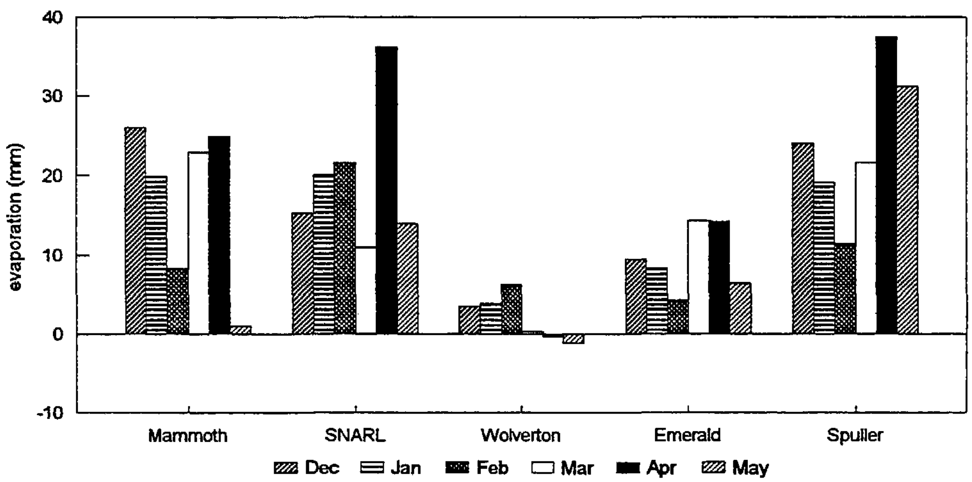


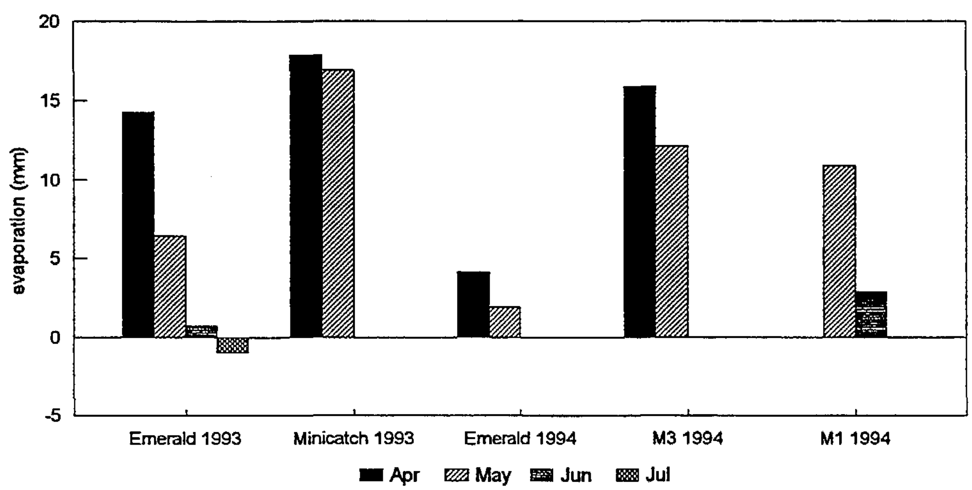
Figure IV-18: Period (~10 days) totals of evaporation from snow at (a) Emerald Lake, (b) SNARL, (c) Spuller Lake and (d) Mammoth Mountain. Points represent the sum, over each period, of hourly mean-profile estimates (15 minute estimates at Mammoth). Negative values indicate condensation.



(a)



(b)



(c)

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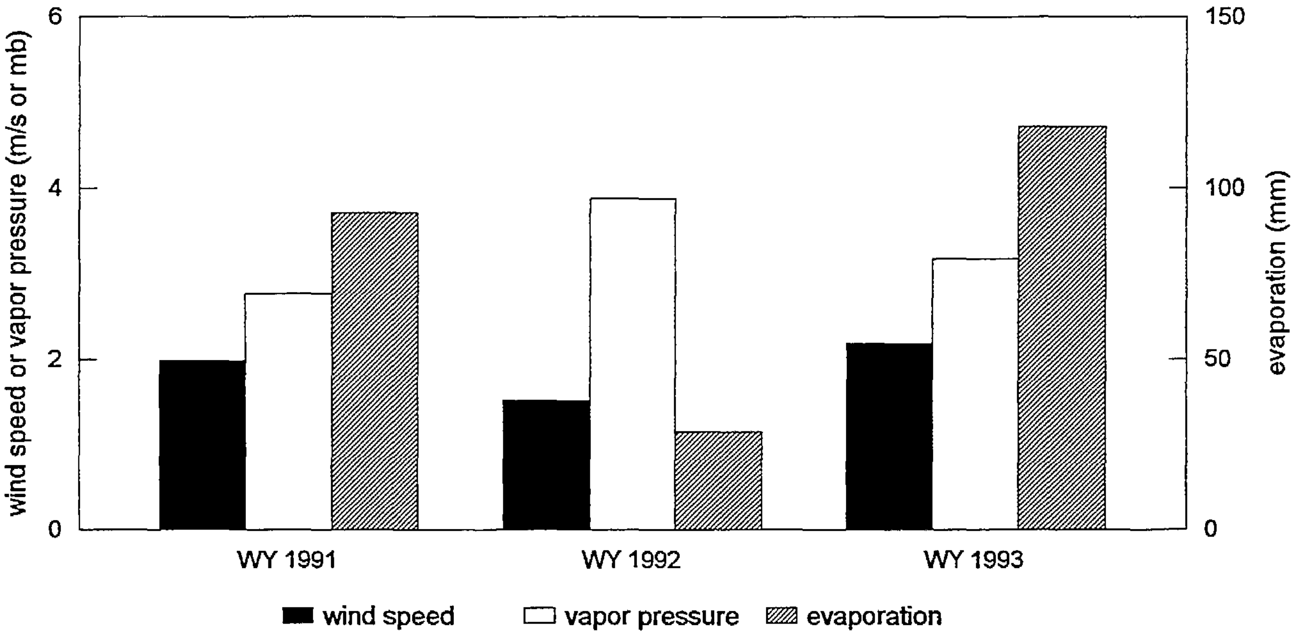
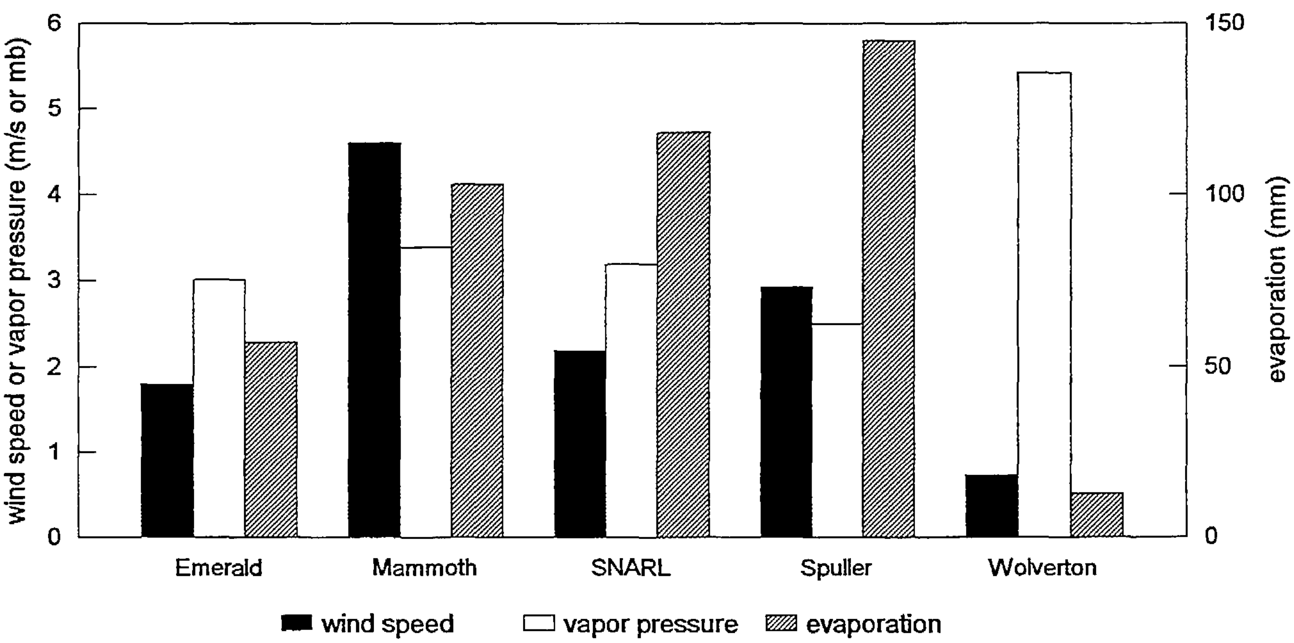


(a)



(b)

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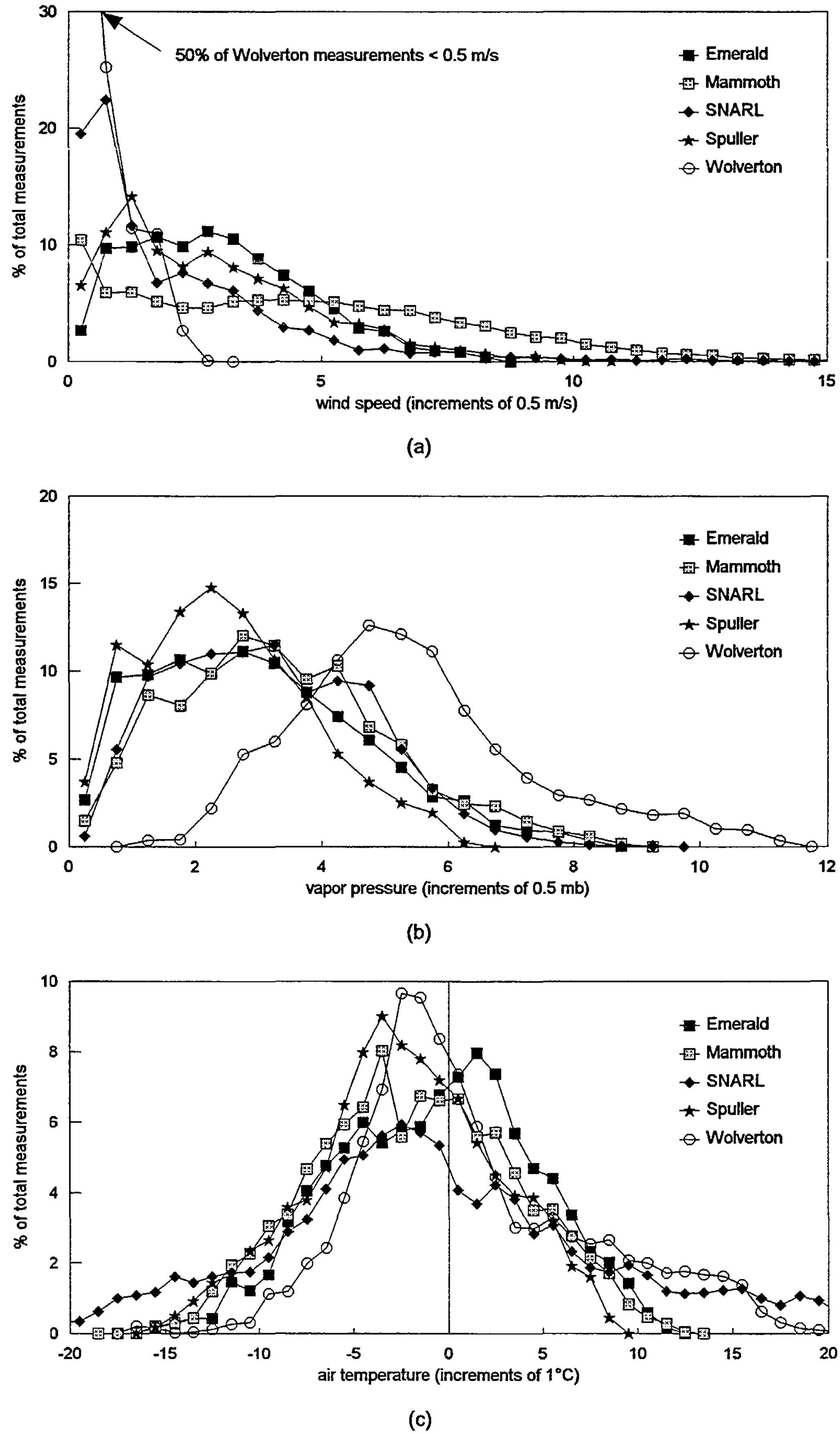

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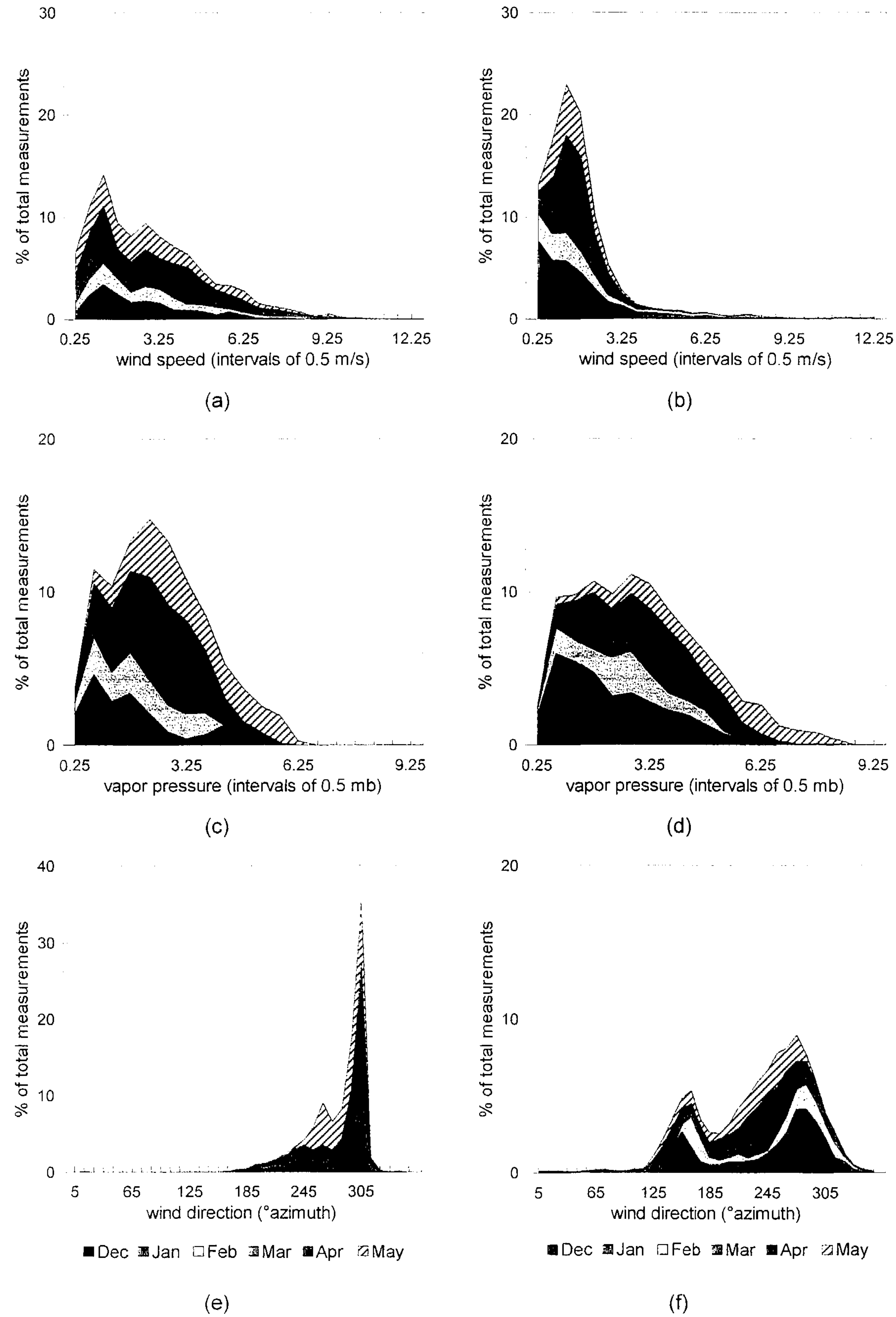

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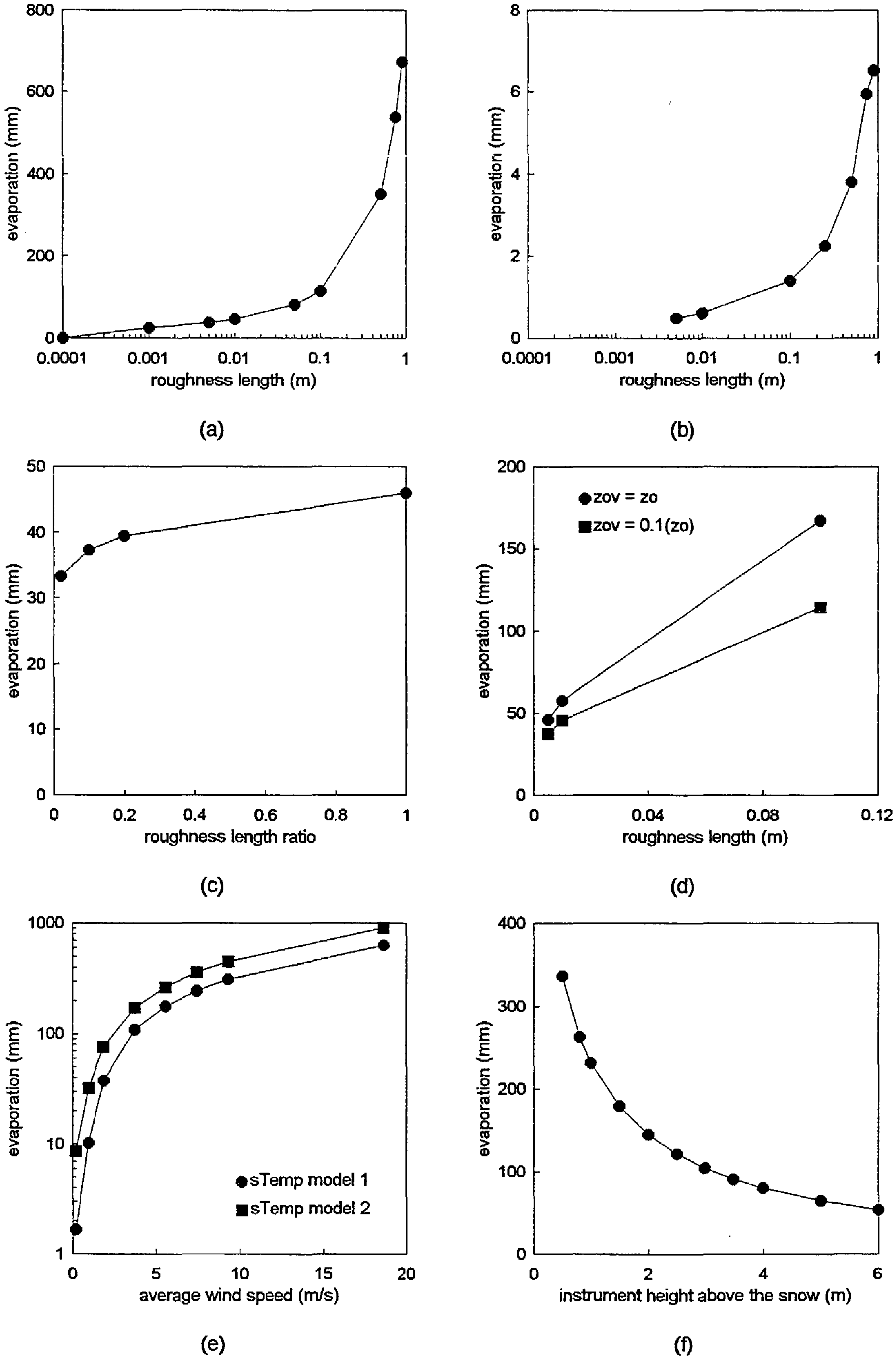

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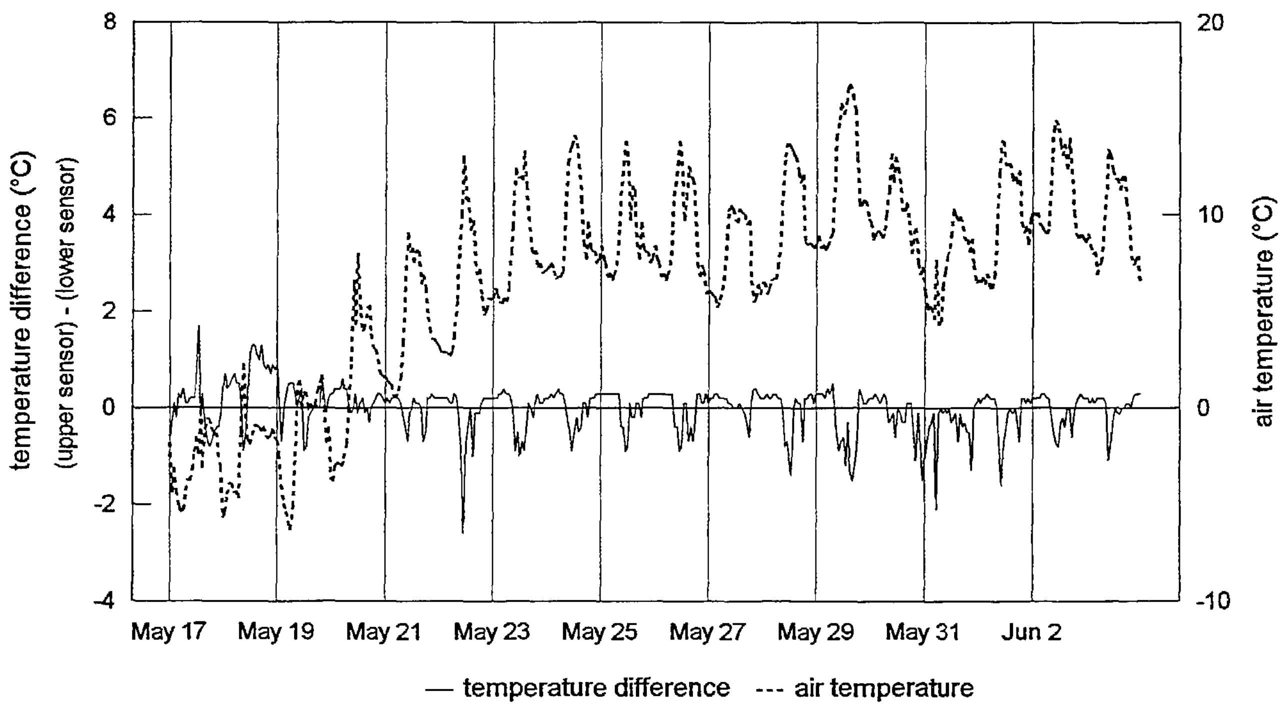

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Appendix One

Soil survey of Seasonally Snow-Covered Catchments of the Sierra Nevada

by

David Marrett

APPENDIX
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A. Appendix

A.1. Introduction

The core of any soil survey is made up of three parts : (1) soil maps, (2) mapping unit descriptions, (3) descriptions of the components named within mapping units. Soils and vegetation were described and mapped, during the summer of 1992, in the Pear, Topaz, Ruby, Crystal, Spuller and Lost lake basins; Emerald Lake was previously mapped by Huntington and Akeson (1987). These maps are presented in Chapter One of this report. Soil descriptions for the Mini Watersheds in Sequoia National Park are also included in this appendix. The only portions of the Marble Fork watershed mapped were the Pear Lake, Topaz Lake and Emerald Lake watersheds, which together make up 23% of the Marble Fork drainage. In contrast to more traditional classification schemes, an alternative approach, specifically designed to aid biogeochemical and hydrological investigations, was used to map the soils. The approach was a synthesis of old and new ideas in pedology, utilizing genetic soil groups, with no operational linkage to soil taxonomy *per se*. The genetic soil groups were locally defined based on their morphology, landscape setting and accepted theories in pedology, soil biogeochemistry and ecology. This appendix describes the soil classification approach and contains a detailed account of the soil characteristics of each watershed. The main objective of this project was to produce a high quality soil survey for these experimental watersheds to support current and future hydrologic and biogeochemical investigations.

A.2. Background and Methods

A.2.1. Approach

Soil groups were recognized based on their morphology, landscape setting, and accepted theories in pedology, soil biogeochemistry, and ecology. They are thought to represent "natural soil populations" or bodies of biogeochemically similar soil materials. Genetic soils are related to soil classification systems with the best available information, starting with simple field morphology.

An aspect of the soil survey is the omission of some types of soil properties, and soil-related environmental factors from the definitions of mapping units and named components. There were two classes of omissions: (1) internal soil morphologic properties that were impossible map in 1992, and (2) soil-related external environmental factors.

An operational decision was needed to make mapping and sampling in 1992 possible. The main internal properties were soil depth, texture and rock fragment contents. Colluvial stratification and bedrock contact geometry were too deep and too complex to accurately estimate and consistently map across all watersheds. Field notes were taken on these properties and some qualitative estimates of soil depth and rock fragment content are included in descriptive text sections. External soil-related factors such as slope gradient, aspect, microclimate, vegetation, and bedrock properties were omitted because these landscape attributes were assessed by others (see Chapter One). Certain surficial geologic features were also included in the soil inventory. Several types of non-soil surficial deposits (scree, talus, rubble and boulders) were recognized and distinguished in this study because their hydrologic properties are different from bedrock, and in some cases (scree) similar to soil, and most geologic maps do not accurately portray surficial deposits.

A.2.2. Soil Versus Non-soil

Numerous definitions of soil have been published. All of these definitions exclude some types of terrestrial surface materials such as massive bedrock, but other types of materials such as deep alluvial strata and exposed sandy river channels are classified variously. A number of similar terms including not-soil, nonsoil and non-soil have been used for the excluded materials. Differing perspectives regarding unusual surficial deposits have contributed to different estimates of proportions of soils versus non-soil materials at high-elevations in the Sierra Nevada.

An operational biogeochemical definition of soil has been adapted for the purposes of the 1992 soil inventory. In most respects this definition is identical to a pedologic definition of soil; it differs significantly from edaphic and engineering definitions. Soils are defined for this project as unconsolidated terrestrial mineral, organic, or mixed mineral and organic materials, showing evidence of pedogenic processes at least in the form of grain coatings on sand or finer particles. Deposits lacking sand or finer material are excluded. Pedologic definitions of soil require at least some evidence of *in situ* horizon development, the capability of supporting vascular land plants, and usually soil structure. These attributes are not absolutely required our definition. Our definition has been broadened so it can include recently deposited or disturbed, unstable or buried materials that were pedogenically altered in another setting, but do not now have clear horizons or significant vascular plant cover. These reworked soil materials are assumed to react with water more like stable soils, and less like raw geologic parent-materials. This modification was made to account for unstable, high-elevation stratified colluvial deposits.

A.2.3. Soil-survey Methods

A reconnaissance of the watersheds was completed and a preliminary mapping legend was developed. The mapping scheme was kept simple with as few named components and mapping units as possible. The mapping design was built around distinctions between soil and non-soil materials, and then around natural gradients within the soil continuum, especially moisture and redox gradients. Moisture gradients were considered the most important factor differentiating soil formation, and soil morphologic variation in association with moisture gradients was recognized as s mapable consistently across all watersheds. The importance of moisture and redox gradients is consistent with most cold soil and soil-ecosystem research (Marrett 1988; Rieger 1983; Burns and Tonkin 1982; Walker et al. 1980; Tedrow 1977, Billings 1976). In addition, these gradients within the soil have understandable biogeochemical and hydrologic importance.

Elevational gradients were recognized, but the gradual, subtle and often inconsistent relationships between observable soil morphology and elevation could not be unambiguously defined and incorporated into the basic mapping legend. Elevational gradients tended to be complexly mixed with variations in microclimatic and soil development. Some soil morphological variation in response to specific vegetation types were also noted, but these variations tended to be subtle and inconsistent. Rock fragment content and particle size properties could not be accurately and consistently mapped in 1992.

The consistent attempt to minimize the number of named components and mapping units was offset by detailed cartographic work. The observed soil variability was expressed with individual delineations separating small areas of relatively few mapping units. This can give the impression of complex soil maps when, in fact, most of an area is either one type of soil or one type of rock mixed in a variety of geometric patterns. The minimization of named components and mapping units was also partially offset by descriptions of how soils and soil-landscape patterns vary within each watershed (Section A.3.4).

The field mapping followed standard soil survey techniques (Soil Survey Staff 1951; 1993). Nearly every slope and delineation was either visited or viewed. However, this was not a detailed (order 1) soil survey in the strictest sense. Every soil line was not meticulously located on foot with multiple soil observations and notes taken flanking the line. The field maps were drawn with colored pencils on frosted mylar overlays of approximately 1:10,000 scale stereo-paired, low-elevation, true-color photographs. Frequent use was also made of topographic maps and false color images at 1:4,000. Only the topographic maps and 1:10,000 true color photographs were practical for field use.

In each of the six lake basins, two typical (or modal) soils in the lower watershed area were fully excavated, described in detail and sampled. One of these was always a well drained soil (Alpine or Volcanic Brown Soil) and the other was the most common of the wetter soil in that particular lower basin, usually a Moist Meadow Soil. In addition to the twelve fully described and sampled soils, many other soils were shallowly excavated and sampled in their upper parts. These shallow pits were usually arrayed in clusters along short and steep moisture gradients, or in contrasting vegetation types. Small pits were also sampled in unusual soil variants or landforms, and whenever possible in larger scale elevational gradients. In total, 235 soil samples were taken from 70 pits in the six lake basins. No soils were collected in the mini-watersheds.

The soil maps were made by transferring the 1:10,000 field sheets to mylar overlays of larger scale images. This process required interpretation without the opportunity to do field checking. In most cases, the final base map images were 1:4,000 scale color infrared images developed from high-elevation aerial photographs. Such images are less geometrically distorted than the low-elevation photographs, and contain less shadow. The Mini-watershed image was a high-elevation false color image at a scale of 1:2000, and the field mapping was done on an overlay of this same image. High-elevation false color photography was unavailable for the Lost Lake watershed, so the final map was an overlay of a low-elevation true color image at 1:2000. The Lost Lake map was field checked, found to be good, and only slightly modified during a visit in fall 1993. No other watershed was field checked.

A.3. Results

A.3.1. Explanation of the Overall Mapping Scheme

A.3.1.1. Genetic Soils and the Genetic Soil Group Scheme

A simple and original system was devised to define the minimum number of genetic soils needed to describe major soil variations. This system is shown schematically in Figure 1 and is explained in more detail below. Recognition of only seven genetic soils involved considerable lumping to emphasize general soil similarities, especially the hypothesized biogeochemical similarities. Each recognized soil is thought to be a group of genetically related and biogeochemically similar soil materials. The genetic soils do, however, contain considerable physical variability. Physical properties such as soil depth, texture and rock fragment composition were de-emphasized in the definitions of these seven genetic soils.

Figure 1 indicates important aspects and biases of our genetic soil classification scheme. The upper part divides surficial materials into soils and non-soil materials. Riparian Soils were mapped across all parent-materials and partitioned next. The other soils were split into pumice-dominated materials and all other parent-materials, and then subdivided along redox gradients in a parallel structure. The simple and general language used implies "fuzzy sets", intermediate forms and a recognition of the soil as a continuum.

The genetic soil names were intended to be as simple and connotative as possible. Most are based on traditional genetic names of similar soils such as Arctic Brown Soils, or Alpine Meadow Soils (Marrett 1988; Ugolini 1986; Tedrow 1977; Avery 1973; Kubiena 1953; Baldwin et al. 1938). Individual names are discussed below. Component terms in soil names such as alpine, moist meadow, wet meadow, and riparian, generally describe settings but are not meant to imply the soils are restricted in distribution to areas meeting botanical or hydrological definitions of these same terms. The soils were recognized primarily by their morphologic properties, and secondarily on their landscape setting. Although some of these terms imply vegetation, this system is fundamentally different than mapping systems used in other alpine and arctic ecosystems where landforms, soils and vegetation are mapped together in composite geomorphic-soil-vegetation mapping units (e.g., Walker et al. 1980).

The genetic soil scheme used in this report is not intended as an alternative soil classification system *per se*, but rather as a locally defined system needed to produce an easily understood soil survey. Like older official U.S. genetic soil classification systems (Thorp and Smith 1949; Baldwin et al. 1938; Marbut 1927), our scheme lacks some attributes of rigorous, comprehensive, modern soil classification system (Boul et al. 1989). It emphases natural gradients and natural relationships between soils, not rigorous subdivisions into numerous well defined, but narrow taxa. Cross correlations or classifications of genetic soils are done at varying levels of detail with the best information available as separate exercises (Table 1; Table 5). Adoption of a genetic soil scheme creates problems with correlation to Soil Taxonomy (1975 1992). The broadly defined genetic soils used in our soil survey cross more taxonomic boundaries than more narrowly defined genetic soils in smaller more intensive research areas (Marrett 1988). Rigorous classification with Soil Taxonomy generally requires some laboratory data, and in tephra derived soils requires considerable laboratory data. Such data were not available for our soil survey. Therefore, the cross-tabulations of genetic soils and taxa in Table 1 and Table 5 are provisional.

A.3.1.2. Mapping Unit Design and Names

The overall design of soil mapping units, like the genetic soil scheme, was intended to be as simple as possible while remaining scientifically valid. The general structure of the system is apparent in the Identification Legend (Table 2). Seven genetic soil types and seven types of non-soil materials were used as named components in a total of twenty nine mapping units, organized into four groups. The fourteen named components and the twenty nine mapping units are each described individually in later sections.

The four main groups of mapping units were defined on estimated areal proportions of soils versus non-soil materials. Predominantly soil units were used for areas estimated to have greater than 90 percent soil cover; soil dominated mixed units were used for areas estimated at 50 to 90 percent soil cover (with about 10 to 50 percent non-soil). Similarly the predominantly non-soil units were estimated to have over 90 percent non-soil cover and the non-soil dominated mixed units had about 50 to 90 percent non-soil cover. In field-mapping and subsequent cartographic efforts, high priority was given to keeping these four groups distinct. Accuracy with respect to proportions of soil cover was a primary objective in placing each line and defining every delineation. This scheme facilitated map generalization and rapid estimates of soil cover proportions.

Many key aspects of the mapping units are contained within the mapping unit names. Mapping unit names follow standard conventions such as listing components in the order of importance, but do not conform to all rules in conventional soil surveys (Soil Survey Staff 1951;1993). Thirteen of the twenty nine units contain one named component, fifteen units contain two named components, one unit has three named components, and five units also contain the word complex.

The word complex is used to indicate a intricately mixed pattern of contrasting soils and non-soil materials. These complexes could not be easily described with two or three component names. In some cases there were more than three important components, in other cases important components were unnamed in this survey. In most cases the main components could be recognized in the field from surface features, but the spatial scale of their mixing patterns was too fine for mapping separation. Proportions of most components could be estimated on a delineation by delineation basis, but these proportions varied considerably between delineations, and especially between watersheds. These complex mapping units have some properties of complexes, and some properties of undifferentiated units as these terms are officially defined and used in conventional soil surveys (Soil Survey Staff 1951;1993).

Mapping units with one named component and without the word complex are similar to consociation units as defined and used in conventional soil surveys (Soil

Survey Staff 1951;1993). Each delineation of these units should contain the named component in at least 75 percent, and more typically in about 85 percent of the area. Unnamed soils similar to the named soil component are often allowed to comprise a small part of the minimum 75 percent coverage. Strongly contrasting soils should not comprise more than about 15 percent of any delineation in one of these units. Similar rules apply to non-soil units.

Mapping units with two or three named components and without the word complex in their name are most similar to the officially defined complexes used in conventional soil surveys (Soil Survey Staff 1951;1993). The names of components appear in their general order of importance, but a few delineations of some units may not conform to this rule. As mentioned above, special attention was paid to accurate estimates of soils versus non-soil proportions and the least attention was paid to accuracy in mixtures of generally similar non-soil materials such as boulders, talus, rubble and scree.

Delineations of mapping units with two or three named components (but without the word complex) should have at least 75 percent and more typically 85 percent total coverage by the named components. The rules for inclusions are similar to those for single component units outlined above. Most inclusions in multi-name units are simply intermediate or mixed forms of the named components. In most cases these units are mixtures of soils and non-soil materials that can be simply distinguished in the field but cannot be cartographically separated at the scales used in this survey. While the proportions of contrasting components can usually be easily estimated within a delineation in the field, accurate assessments of these same properties from aerial photographs is difficult to impossible. Proportions of components may be similar in adjacent delineations but usually vary across watersheds, and are especially variable between watersheds.

A.3.2. Descriptions of Named Components

A.3.2.1. Major Genetic Soils

Alpine Brown Soils-- These well drained acidic soils are, by far, the most extensive group of soils observed. This name, adapted from Tedrow's Arctic Brown Soils (1977), relates these soils to several groups of similar, moderately developed genetic soils whose names contain the words or syllables brown, braun, brun, or camb (Dudal 1968; Duchaufour 1982; Canada Soil Survey Subcommittee 1978; Avery 1973; Kubiena 1953; Baldwin et al. 1938). 'Brown soils' are named for largely non-illuvial brown Bw subsoils, and to a lesser degree their darker brown A horizons, and transitional AB horizons. They are ambiguously related to specific official taxa (Soil Survey Staff 1975, 1992). Alpine Brown Soils have been mapped at all elevations in subalpine, treeline and true alpine settings. Their essential morphology is of the type abbreviated A-Bw-C, but several significant morphologic variations are also included in the diverse group (Table 5). The most typical A-Bw-C morphology consists of: (1) organic matter enriched, granular structured, dark brown surface horizons, (2) iron oxide stained, weak blocky structured, dark yellowish brown subsoils, and (3) massive pale brown substrata, sometimes enriched in secondary silica, and slightly silica cemented. These three master horizons usually contain identifiable gradations in properties which are recognized as subdivided and transitional horizons in detailed soil descriptions. By definition these well drained soils should not contain redox mottles, or other evidence of wetness at depths of less than 1 meter (Figure 1). Some of the more important of the many morphologic variations in Alpine Brown Soils are described below.

All Alpine Brown Soils are acidic (about pH 4 to 6), and all have coarse textures with intermediate to high rock fragment contents. Most textures were estimated to be sandy loams; loamy sands and loams were also common. Rock fragment content estimates varied from about 15 to 80% by volume with all sizes and most shapes of fragments well represented. The parent-materials of Alpine Brown Soils are mostly local alpine glacial drift and colluvium derived from granitic rocks, mixed with loess and minor amounts of volcanic ash. Residual rock weathering is locally important in several places. The Spuller watershed contains a large area of metavolcanic rocks, and other watersheds contain smaller areas of mineralized or non-granitic rocks. Alpine Brown Soils in these areas still contain significant components of granitic drift, but the parent-material mixture is more complex.

The most extensive and important morphologic variants of Alpine Brown Soils are the various shallow well drained soils. All of these soils have bedrock (R layers) within 1 meter of the surface, and most lack C horizons (A-Bw-R morphology). Many are shallower than 1 meter, and also lack B horizons (A-R type morphology). Shallow A-C-R type profiles are only common as inclusions at high-elevations, and in unstable colluvium. Bedrock contact geometry is complex in many small areas, and it is usually difficult to distinguish the larger deeper rock fragments from bedrock. Soil depth will require more study because it varies continuously from a few centimeters to over a meter without apparent "modes", and often with little surface expression.

A second important set of morphologic variations is related to elevational gradients. At high-elevations many areas of Alpine Brown Soils show minimal horizon development, and are associated with other, mostly unnamed, poorly developed genetic soils. These high-elevation areas are not well described in the current soil survey, and will require further study before they can be properly described and mapped. Unusually well developed 'forest soils' form the other end of the soil development-elevational gradient contained within Alpine Brown Soils. These soils have significant forest litter layers (O horizons) and weak to moderate morphologic evidence of strongly leached surface soil layers (horizons with some properties of E horizons). These leached horizons may be designated E, EA, AE, EB or BE. There is little or no clear morphological evidence of organic-metal complex accumulations in the upper B horizons (no clear evidence of spodic Bh or Bhs horizons). Upper B horizons may be designated as Bs to indicate apparent sesquioxide accumulation, without much organic accumulation. These well developed brown soils (also called "podzol-like") are usually relatively rich in volcanic ash in their upper parts, and are variable in depth and rock fragment content..

Other important types of morphologic variations within the Alpine Brown Soils include stratification of mostly weakly developed soils with non-soils in recent and unstable colluvium, with wide ranges of rock fragment contents. The current understandings of these variations in morphology are described as much as possible in individual mapping unit and watershed descriptions. With completion of current laboratory studies and further field study the Alpine Brown Soil group may be divided into three or four different genetic soils; it could also be subdivided into numerous phases (Table 5) to better express its variations. Since this soil is so extensive and subdivision would have advantages.

Moist Meadow Soils-- These poorly drained soils are not extensive, but may be important in watershed hydrology (e.g., storage of water) and biogeochemistry (e.g., decomposition of organic materials and low redox potentials). The name was adapted from the names Alpine Meadow Soils, Mountain Meadow Soils, and other similar terms used in older genetic soil classification systems (Tedrow 1977; Baldwin et al. 1938; Zakharov 1927). These soils have been mapped under a variety of plant communities, and the name is not meant to imply strict association of these soils with botanically defined moist meadows. They were mapped in middle to lower elevation areas in mostly subalpine or treeline settings. Their essential morphologic features are black A horizons and mottled subsoils (B or C horizons); many morphologic variations were observed. Many of these soils are strongly stratified with many A-C or A-Cg "sequa" or combinations in the upper 1 meter. Thick, black loamy A horizons over sandier Bw, Bg, Cg or C horizons are typical of the older, more stable and less stratified soils. Mottled subsoils are not usually gleyed. Where present gley is weak or confined to deeply buried layers. The subordinate "g" was used liberally for any indication of gley, not only for "strong gleying" as officially prescribed (Soil Survey Staff 1993; 1992). This technical adjustment was made to account for the cold, mostly oxidized saturated soil environments, and young soils; it may require some refinement. Most of these horizons were transitional between Bw and Bg horizons and would have been designated as mottled Bw horizons in well developed and stable valley soils. The A horizons of these soils appear to contain higher organic matter contents, and different organic matter composition than the Alpine Brown Soil A horizons: which are darker and feel siltier in the field.

Many Moist Meadow Soils are held together with dense medium to fine root networks. In older terminology these soils are "turfy". In addition to the stratified and unstratified variants these soils also vary in depth, texture and apparent redox status. By definition, these soils contain mottling or some other evidence of wetness and fluctuating redox in their upper meter, usually some of this evidence is within the upper 50 cm. By definition these soils do not contain evidence of prolonged chemical reduction, or major peat accumulation in their surface horizons (Figure 1). The morphologic variations of the Moist Meadow Soils are not well enough understood to describe in further detail in this report.

All Moist Meadow Soils are acidic (about pH 4 to 6), and most have medium to coarse textures with variable rock fragment contents. The A horizons, especially the thicker ones, tend to be loamier (silt loams, loams, fine and fine sandy loams) than the B and C horizons (more typically light sandy loams, loamy sands, and sands). These field estimated textures were done by feel. This method is influenced by organic matter content which makes A horizons feel siltier and/or loamier than deeper horizons with lower organic matter contents. Most Moist Meadow Soils contained relatively low rock fragment contents (about 0 to 15% by volume), but in certain settings, especially near outcrops, buried bedrock contacts, and talus rock fragment contents were variable (about 10 to 60%). The parent-materials of Moist Meadow Soils are mostly local alluvium derived from local sources of granitic glacial drift, colluvium, loess and volcanic ash. Residual rock weathering is apparently locally important in some of these soils. In parts of the Spuller watershed, and other non-granitic or mineralized rock areas parent-material mixtures are more complex. With laboratory analyses and perhaps further field studies the Moist Meadow Soil group could potentially be divided into two or more different genetic soils, or could be subdivided into phases (Table 5) to better express its variations. Since this soil is not extensive, such subdivisions would have few advantages unless they expressed known, contrasting biogeochemical properties.

Wet Meadow Soils-- These poorly soils are not extensive, but may be important in watershed hydrology and biogeochemistry. The name was adapted from the names Alpine Meadow Soils, Mountain Meadow Soils, and other similar terms used in older genetic soil classification systems (Tedrow 1977; Baldwin et al. 1938; Zakharov 1927). In older classification systems some of these soils would be categorized as wetter types of Meadow soils whereas most would be called Bog or Half Bog Soils. Wet Meadow Soils have been mapped under a variety of plant communities including wet meadows, marshes, willow thickets and bogs. They were only found in lower elevation areas of watersheds, usually in subalpine settings. The name is not meant to imply strict association of these soils with botanically defined wet meadows. The word meadow was also used in the name to emphasize similarities of the Moist and Wet Meadow Soils. Both types of Meadow Soils contain most of the features described above, but their hydrologic settings and surficial morphology differs. The essential morphologic features of the Wet Meadow Soils are accumulations of peat (or muck), gley, and low chroma colors in their upper parts indicative of prolonged saturation, oxygen depletion, and/or chemical reduction under current hydrologic conditions. Most of these soils are in stratified alluvial or lake-margin deposits with common buried O, A, and Cg horizons, and an occasional Bg horizon. Most profiles contain at least thin O horizons at some depth, some gley or low chroma colors, and some high chroma mottles. Some soils were composed completely of O horizons to the depth examined (about 60 cm). These various types of horizons and morphologic features were not investigated in enough detail to reveal fine scale spatial patterning. The most common types of layers were black loamy A horizons, sandy Cg horizons, and peaty Oi horizons. By definition, these soils have some evidence of peat accumulation or prolonged chemical reduction in their upper 50 cm, but they may also show some signs of fluctuating redox in their upper parts or in buried layers.

Wet Meadow Soils are thought to range from acid to neutral (about pH 5 to 7), and to experience significant coupled redox-acid base reactions. Most have medium to coarse textures with generally low but in places variable rock fragment contents. Organic layers are mostly fibrous (fibric) peat, or intermediate composition (hemic) mucky peats. Mineral layers vary in texture from silt loams, through loams, sandy loams and loamy sands to sands. As with the Moist Meadow Soils, rock fragment content is generally low (about 0 to 15% by volume) but can be locally high and variable (about 10 to 60%). Wet Meadow Soils are generally in deep, stratified, sandy, loamy and peat deposits, but shallow variants are locally important. The parent-materials of Wet Meadow Soils are much like Moist Meadow Soils except for the greater accumulations of peat in the wetter soils. These are mostly local alluvium from local sources of granitic glacial drift, colluvium, loess and volcanic ash. Residual (*in situ*)rock weathering may be locally important in some of these soils. In parts of the Spuller Watershed, and other non-granitic or mineralized rock areas, parent-material mixtures are more complex. With laboratory analyses and perhaps further field studies the Wet Meadow Soil group could be divided into different genetic soils, or subdivided into phases (Table 5) to better express its variations. Since this soil is not extensive, subdivisions would have few advantages unless they expressed known contrasting biogeochemical properties.

Riparian Soils-- These rocky, shallow, and wet but well oxidized soils occur extensively throughout most of the watersheds as many small delineations. Their proximity to flowing water makes them hydrologically and biogeochemically important. The name was coined in this study to describe their general setting in high energy stream corridors, seeps and similar settings where large volumes of cold, oxic water rapidly run over, under, around and through small volumes of unusual soils. Their distribution does not necessarily match botanical, or other definitions of riparian zones. They were mapped at all but the highest elevations in all types of settings.

They were recognized as wet soils contrasting with the two Meadow Soils mainly in depth, redox status and soil solution residence time. Their essential morphologic features include "turfy" dark brown to black A horizons, a total lack of B horizons, and shallow bedrock or fragmented rocky material. The main A horizon can often be subdivided and may be coarsely stratified. Transitional AC horizons, and sandy C materials are sometimes present below A horizons as small volumes of soil either in bedrock cracks, or in isolated pockets in otherwise fragmented rocky materials. No mottling, peat accumulations, other signs of either redox fluctuations or oxygen depletion have been observed in any of these soils. The A horizons of these wet Riparian Soils often look much like the dark A horizons of the Moist Meadow Soils, and not as similar to the A horizons of adjacent well drained Alpine Brown Soils. Subtle morphologic differences of these kinds are difficult to interpret without supporting laboratory data. The results from chemical analyses of A horizons are needed to clarify the biogeochemical relationships between Riparian Soils, Meadow Soils, and Brown Soils.

Riparian Soils are acidic (about pH 4 to 6), and range in texture from loamy surface layers to sandy subsoils in very rocky to rocky matrices. The upper A horizons are generally darker, and tend to be loamier (silt loams, loams, fine and very fine sandy loams) than the deeper A, AC and C horizons (more typically light sandy loams, loamy sands, and sands). These field estimated textures are influenced by organic matter content which makes upper A horizons feel siltier and loamier. The fragmented rocky components of these soils are most commonly colluvial or residual mechanically weathered bedrock. Alluvial stones and cobbles dominate a few delineations. Finer materials are fresh local alluvium, similar in specific composition to the surrounding soils. Riparian Soils were the only genetic soils mapped in both the granitic watersheds and in the tephra filled Crystal watershed. This was done on the assumption that the biogeochemical behavior of Riparian Soils is controlled by organics in A horizons, not weathering products in subsoils. The unusual biogeochemical properties of volcanic soils are usually attributed to tephra weathering products, not the fresh tephra itself (Boul et al. 1989). The finer soil materials in Riparian Soils are not in typical alluvial strata, but are in irregular pockets that appear to have been trapped and held in place by the plants that can colonize and survive in these high energy environments. Sedge dominated vegetation, willow thickets, and mixed herbaceous and shrub communities are most common. The current understandings of morphological variations are described as much as possible in individual mapping unit and watershed descriptions. Several morphologic variations of Riparian Soils were observed in the field, but none were geographically extensive. With laboratory analyses and perhaps further field studies the Riparian Soil group could be divided into different genetic soils, or could be subdivided into phases (Table 5) to better express its variations. Since this soil is not extensive, subdivisions of it would have few advantages unless they expressed known contrasting biogeochemical properties.

Volcanic Brown Soils-- These well drained to excessively drained pumice and volcanic-ash rich soils are extensive within the Crystal Lake watershed, and not found in any other watershed. The name was coined to emphasize the importance of tephra parent-materials, and the relationships to Alpine Brown Soils, Arctic Brown Soils (Tedrow 1977), and other types of "brown soils" as outlined above for Alpine Brown Soils. These soils have been mapped in subalpine and open treeline settings. Their essential morphology, like for the Alpine Brown Soils, is A-Bw-C, but several variations on this theme have been observed. One distinctive feature of these soils is their colluvial stratification due to shallow pumice grain flows over the soil surface; this stratification is common but not universal. In the colluvially stratified soils, surface soils may consist of several thin interbedded A, AC, and O horizons.

Subsoils show textural stratification, but their overall morphology is not as dramatically stratified as in surface horizons; the formation of weathered Bw horizons seems to cross and overprint most stratification. Stable pockets of older soils show a more typical A-Bw-C, or O-A-Bw-C horizonation more similar to most Alpine Brown Soils. In all cases observed the master horizons could be subdivided into many transitional and subordinate horizons with detailed examination of morphology. By definition these well drained soils should not contain redox mottles, or other evidence of wetness at depths of less than 1 meter.

Volcanic Brown Soils are coarse to very coarse grained and acidic (about pH 4 to 6). Field textures were mostly sandy loams, and pumice fragments ranging from about 1 to 10 mm in size comprised about 50% by volume of most soil layers examined. Larger, mostly colluvial, rock fragments comprised about an additional 10 to 30% of some soils. The parent-materials of most soils examined appeared to be at least 75% recent tephra, including both pumice fragments and sand to silt sized volcanic ash. These soils vary mostly in degree of stability, degree and type of internal stratification and depth to bedrock. These properties vary over short distances and cannot often be practically mapped. In many cases bedrock depth and relative stability can be estimated on-site for small areas. With laboratory studies and further field study the Volcanic Brown Soil group could be divided into different genetic soils, or numerous phases (Table 5), but it does not appear practical to attempt mapping these subdivisions.

Volcanic Moist Meadow Soils-- These moderately drained soils are only found as small areas in the Crystal Lake watershed. The name was coined by modifying the Moist Meadow Soils name to express the importance of tephra parent-materials. They were mapped under a variety of wet, subalpine plant communities, not just botanically defined moist meadows. Except for the tephra parent-materials and volcanic weathering products their morphology and drainage conditions are similar to the Moist Meadow Soils. Their essential morphologic features are black A horizons and stratified mottled subsoils (B or C horizons); several morphologic variations were observed. Most, if not all, of these soils are strongly stratified with many A-C, A-B or A-Cg "sequa" or combinations in the upper 1 meter. Thick, black loamy A horizons over sandier Bg, Cg or C horizons are typical of the older, more stable soils. Mottled subsoils are not usually gleyed, and the subordinate "g" was used liberally for any indication of gley, which may become problematic as described above for the Moist Meadow Soils. The A horizons of these soils appear to contain higher organic matter contents, and probably a different organic composition than the Volcanic Brown Soil A horizons. Organics affect the color and feel of these A horizons: they are darker and feel siltier than A horizons of well drained soils.

Like the other non-volcanic Moist Meadow Soils, many of these soils are held together with dense medium to fine root networks (i.e., "turfy"). These soils vary in depth, texture and apparent redox status. By definition these soils contain mottling or some other evidence of wetness and fluctuating redox in their upper meter, usually some of this evidence is within the upper 50 cm. By definition these soils do not contain evidence of prolonged chemical reduction, or major peat accumulation in their surface horizons. It is not currently understood how similar non-volcanic and Volcanic Moist Meadow Soils are.

Volcanic Moist Meadow Soils are acidic (about pH 4 to 6), and coarse to very coarse in texture. Organics and trapped silt make A horizons feel siltier and loamier than subsoils, but sandy loams are still most common. Subsoils are typically loamy sands to sands with high contents of small pumice fragments. Other than the abundant pumice fragments, other rock fragment contents are low to very low in these soils (about 0 to 20%). The parent-materials appear to be almost pure local alluvial deposits of reworked recent tephra. Subdivision of this group into different genetic soils, and /or phases is possible (Table 5), but does not appear justified given the small total area of soil involved, and the serious difficulties that would arise in attempting to map subdivisions or phases.

Volcanic Wet Meadow Soils-- These poorly drained to very poorly drained soils are not extensive; they are only found in one large wetland area adjacent to Crystal Lake. They are thought to be distinctive soils, important in watershed hydrology and biogeochemistry. The name was coined by modifying the Wet Meadow Soils name described above to indicate the importance of volcanic parent-materials. The name is not meant to imply strict association with botanically defined wet meadows; small areas of moss-bog, sedge marsh, willow thicket, and other wetland vegetation are mixed adjacent to Crystal Lake. These soils have variable morphologies due to strongly stratified recent alluvial deposition into the wetland complex. All soils are mottled, and show morphological evidence of fluctuating redox somewhere in their upper parts, but clear evidence of prolonged chemical reduction (gley or low chroma matrix colors) are rare. Very young pumice rich strata contain few signs of wetness. Peat has accumulated mostly as thin pockets under boggy moss mats or in sedge marsh areas. All profiles examined were stratified, with buried soil layers. The most common horizons were thin A, O, Cg and C horizons. These various types of horizons and morphologic features were not investigated in enough detail to reveal fine scale spatial patterning.

Volcanic Wet Meadow Soils are thought to range from acid to neutral (about pH 5 to 7), and to experience significant coupled redox-acid base reactions. They have coarse to very coarse textures due to the high proportions of pumice and sandy volcanic ash. Rock fragments other than pumice are probably present, but none were observed. These soils appeared to be in a deep alluvial (or delta) deposit filling in Crystal Lake. Shallow soils were not observed and are rare or absent. Although these variable soils could easily be subdivided into several genetic soils or many phases based on morphology (Table 5), such an effort does not seem justified at present.

A.3.2.2. Non-Soil Map Units

Boulders-- The standard definition of boulders as rock fragments greater than 60 cm in diameter or length (Soil Survey Staff 1993) is used in this report. Boulders may form colluvial deposits (like talus), or glacial drift deposits. Stones, especially large stones (30 to 60 cm) are commonly included with boulder deposits. Both colluvial and glacial boulders are common as scattered inclusions in several mapping units, and the term "rock" is sometimes used to refer to both rock outcrop and boulders.

Rock Outcrop-- Consolidated exposed bedrock of any composition is included in this term. Much of the exposed bedrock included in this term is jointed and/or fractured, some was mineralized and iron stained but no strongly chemically weathered rock (saprolite) was observed in the study.

Rock-covered Glaciers-- The two small glaciers mapped were completely covered with talus-like rocky mantles. The glacial ice itself was only observed in one place and it appeared to be alpine glacial ice. Some scientists may classify these features as rock glaciers, which form in different ways and contain much rockier cores than other types of alpine glaciers.

Rubble-- The term rubble is used for in-situ frost shattered rock fragment deposits with little or no evidence of colluvial or glacial displacement. These deposits are fragmented (fines do not fill most voids), and mostly composed of angular to subangular fragments in the cobble to stone size range (7.6 to 60 cm diameter).

Scree-- The term scree is used for loose and unstable sandy colluvial deposits. Most particles are individual uncoated quartz and feldspar grains in the medium sand to fine gravel size range (0.25 to 5 mm diameter). Scree deposits are often stratified with inclusions of soils and talus.

Talus-- This term has been used for colluvial rock fragment deposits. Most talus is fragmented (unfilled voids), and composed of angular fragments in the coarse gravel to stone size range (2 to 60 cm diameter). Talus deposits vary widely in relative age, physical stability, particle size distribution, and types of inclusions. Many deposits are stratified with varying compositions and inclusions. Common inclusions include scree layers, partially buried soils, and boulders.

Volcanic Scree-- This term was coined to emphasize the loose and unstable colluvial nature of some sandy to gravely tephra deposits. These deposits contain recent tephra ranging from fine sandy volcanic ash to gravel sized pumice fragments (0.1 to 10 mm diameter). Small subrounded pumice fragments dominate most strata.

A.3.3. Descriptions of Mapping Units

A.3.3.1. Predominantly Soil Units

All seven soil units were used for areas estimated to contain 90% or greater soil cover. Many divisions of mapping units were made and kept in the legend for the areas with such high proportions of soil. Most units are not extensive because it was rare to find areas large enough to map with less than 10% non-soil materials. Most of these areas have deep to very deep soils, or at least moderately deep soils in deeper surficial deposits .

Alpine Brown Soils (Br) -- These areas contain well drained to excessively drained soils formed in coarse granitic or mixed granitic dominated parent-materials. Most delineations are deep deposits of glacial till, or colluvially reworked till containing some stones and a few boulders. Most of the non-soil inclusions are boulders and isolated, small bedrock outcrops. The soils are dominantly the most stable, deep and well developed examples of Alpine Brown Soils. Small areas of intermediate to shallow examples of Alpine Brown Soils may be common in some delineations, especially adjacent to rock outcrops and near the edges of the delineation. This mapping unit is restricted to stable moderate gradient slopes, with little variation in drainage condition. Strong stratification or minimally developed soils have not been observed in this mapping unit. Areas with Alpine Brown Soils and little rock are rare and mostly small in area. Only ten delineations of this unit were mapped, five in the Topaz Lake watershed and the others in the Ruby and Spuller Lake watersheds (Table 3). In the Spuller Lake watershed the parent-material is of mixed lithology, and mineralized rock is near the Ruby Lake delineations. These geologic factors may affect soil chemistry, but at present there are no other known reasons to subdivide this unusual and fairly homogeneous unit.

Alpine Brown Soils-Meadow Complex (BC) -- These areas contain soils of variable drainage condition formed in coarse granitic or mixed granitic dominated parent-materials. Most delineations are in mixed surficial deposits including glacial till, local alluvium and colluvium. This complex unit was set up to describe intricately mixed contrasting soil areas dominated by moderately well drained to well drained soils, but also containing significant somewhat poorly drained areas, and occasionally even some small poorly drained areas, small streams or riparian areas. Many of these areas are dominantly deep to bedrock, but depth appears variable in some individual delineations, and moderate to shallow in some others. These depth differences are described to some degree in watershed descriptions (Section 3.4). Non-soil inclusions are higher in many delineations of this mapping unit than for other mapping units in this group. The inclusions of non-soil material are mostly bedrock, but boulders are present in some areas. Deep, stable and well developed moderately well drained Alpine Brown Soils are the most common soil. Next in importance are well drained Alpine Brown Soils, and unnamed soils intermediate between Alpine Brown Soils and Moist Meadow Soils. Moist Meadow Soils, and/or Riparian Soils are also usually present in small but important parts of many delineations. These wetter areas are too small to map out, but are probably hydrologically and biogeochemically critical. Moist Meadow Soils and some unnamed intermediate soils show evidence of fluctuating redox conditions in subsoils, other soils show no such evidence.

This mapping unit is as heterogeneous as any in this study, and the actual drainage properties and composition of soils in many delineations are poorly understood. It is a common mapping unit (57 delineations), but its general character and composition appear to vary considerably between the five watersheds where it has been mapped (Table 3). Most delineations are small with gentle to moderate slopes, many are associated with small intermittent stream channels. A number of small delineations at intermediate to high-elevation are surrounded by bedrock dominated areas (ROB, RO). These delineations have mostly shallow or intermediate depth soils. With more quantitative information this mapping unit could easily be subdivided into three or more units, or several phases. Such subdivisions may be well justified when more data become available.

Moist Meadow Soils (MM) -- These areas contain deep, moist to wet soils formed in coarse granitic or mixed granitic dominated parent-materials. Soils vary somewhat in drainage and redox conditions, but are mostly poorly drained with evidence of fluctuating redox in subsoils. Slopes are usually gentle, and local alluvium in lake or pond marginal deposits is the most common specific parent-material. Although the soils tend to have complex stratified morphologies with short range variability when examined in detail, most of these delineations show relatively little overall variation in slope, soil depth or general drainage condition. Soils gradations of poor drainage are the most common. This is an uncommon mapping unit with five mostly small delineations. This mapping unit appears similar in the four different watersheds where it was mapped (Table 3). At least one small delineation in the middle Ruby Lake watershed appears to contain some shallow soils.

Volcanic Brown Soils (VB) -- This area contains mostly well drained soils formed in deep tephra deposits over mostly volcanic rocks on the Mammoth Crest . The few inclusions are rock outcrops, large stones or boulders. The Volcanic Brown Soils appear typical for the genetic group, with some soils under groups of conifers displaying strongly leached mineral soils under a litter mat. The slope gradient is mostly moderate, but soils vary in stability and degree of stratification near their surfaces. Areas with Volcanic Brown Soils and so little rock appear rare and were only found on the gentle Mammoth Crest, not in the middle or lower Crystal Lake basin. The one delineation of this unit is large (Table 3).

Volcanic Moist Meadow Soils (VMM) -- These areas contain mostly deep, moist to wet soils formed in young pumice and volcanic ash dominated parent-materials. Soils vary somewhat in drainage and redox conditions, but are mostly moderately drained with evidence of fluctuating redox in subsoils. Slopes are gentle to moderate, and locally reworked tephra-rich alluvium and colluvium are the main specific parent-material. Bedrock is the main included non-soil material, but is generally absent. Soils contain complex morphologies composed of many recently deposited thin strata, similar in this respect to many other areas in the Crystal Lake basin. Very small areas of Volcanic Wet Meadow Soils may be included in some delineations. This unit differs from the Moist Meadow Soil unit in more than just composition of parent-materials. This unit contains more shallow soils, younger less developed soils, more moderately sloped areas, and some unnamed transitional Volcanic Moist Meadow to Riparian Soils. This unit has been mapped in five small delineations in the Crystal Lake watershed, and nowhere else (Table 3).

Volcanic Wet Meadow Soils (VWM) -- This one large area contains deep, wet soils formed in stratified alluvium mostly composed of recently deposited pumice, volcanic ash, and organic soil materials. Soils vary somewhat in drainage and redox conditions, but are mostly poorly drained with evidence of fluctuating redox and/or peat accumulation in surface soils; subsoils often contain some evidence of prolonged chemical reduction. The large delineation contains a few small somewhat poorly drained areas with soils similar to the Volcanic Moist Meadow Soils or unnamed intermediate soils. It also contains many small to medium sized poorly drained areas. The poorly drained areas have more peat accumulation, and evidence of prolonged chemical reduction at or near the surface. Detailed interpretation of redox and drainage condition from soil morphology is made difficult by the age and complex stratification of these young soils. This area is nearly level and appears to be a delta-like deposit essentially filling Crystal Lake from south to north. Although it contains short range variability there are few good reasons to subdivide this mapping unit.

Wet Meadow Soils (WM) -- These areas contain deep wet soils formed in stratified coarse granitic, mixed granitic dominated, or organic soil parent-materials. Soils vary somewhat in drainage and redox conditions, but are mostly poorly drained with evidence of fluctuating redox and peat accumulation in surface soils, and prolonged chemical reduction in subsoils. Slopes are nearly level lake margins. The specific parent-materials are local alluvium in deltas and lake marginal marsh deposits. Most areas appear to be inundated in the spring, but exposed with shallow ground water in the late summer. Some areas are poorly drained, and covered with water for most parts of most growing seasons. All areas have clear evidence of prolonged periods of anaerobic conditions and chemical reduction throughout most of the soil. Soils are stratified combinations of thin mineral and organic layers with a high degree of short range variability when examined in detail. This high spatial variability is part of the broad definition of Wet Meadow Soils. Delineations of this unit contain few if any inclusions of any other genetic soil defined in this report, but contain some intermediate soils grading toward Moist Meadow Soils. This unit was only mapped in one very small delineation adjacent to Lost Lake, and two medium to small delineations adjacent to Topaz Lake.

A.3.3.2. Soil Dominated Mixed Units

These are mixed soil and non-soil areas where soils were estimated to cover between about 50 to 90% of the land surface. They were estimated to have between about 10% and 49% non-soil cover including, but not limited to, bedrock. Many of these areas contain deep surficial deposits with little or no exposed bedrock. Since these areas are dominated by soils many subdivisions into mapping units were made, and kept in the legend. Unlike for the group of predominantly soil units described above (Section 3.3.1), many of the mapping units in this group are extensive. Soil depths and degrees of development are variable.

Alpine Brown Soils and Rock (BrR) -- These areas contain well drained to excessively drained soils formed in coarse granitic or mixed granitic dominated parent-materials. Most delineations are variable depth deposits of glacial till, or colluvially reworked till containing some stones and a few boulders. In most delineations the named Rock component is mostly rock outcrop. A few delineations are in deep bouldery till deposits with little rock outcrop. In these particular delineations the named Rock component refers mostly to boulders. The Alpine Brown Soils in this mapping unit are mostly stable and well developed, but are variable in depth. Deep and moderately deep soils are thought to be dominant in most delineations, but some areas contain mostly shallow or irregular depth versions of Alpine Brown Soils. This unit has been mapped in many delineations (63), in many different types of settings, in all six of the non-volcanic watersheds (Table 3). Slopes range in gradient from gentle to steep, elevation and degree of soil development also vary widely. In the lower Spuller Lake watershed the lithology is mixed. Although several variants (or phases; Table 1; Table 5) of Alpine Brown Soils comprise this unit, it contains few inclusions of any other named genetic soils. It also contains little variation in drainage, i.e., virtually no wet soils are included. When more quantitative information becomes available this unit could easily be subdivided into three or more units.

Alpine Brown Soils and Scree (BrS) -- These are unusual areas of deep, mostly excessively drained soils formed from granitic scree and wind deposited dust, mixed and interstratified with scree. These marginally stable stratified colluvial deposits are on moderate to steep slopes and appear to be deep to bedrock in most places. Most of these areas are relatively inaccessible and at very high-elevations, and have not been investigated in detail. Some areas contain stones and boulders at the surface, but most areas appear to be sandy scree deposits until they are closely examined. The areas investigated contain mostly an unusual variation of the Alpine Brown Soil mantled with about 5 to 10 cm of loose scree. They were sparsely vegetated, and interspersed with unvegetated areas of pure loose scree. The soils were surprisingly well developed in their upper parts. The relative proportions of soil and scree were difficult to estimate except by close field inspection. All of those areas examined were dominantly scree mantled soil. The delineation in the upper Spuller Lake watershed, and those north of Ruby Lake were examined; the delineations west and southwest of Ruby Lake were not visited. Weakly developed and stratified Alpine Brown Soils are the main component. These areas probably include a variety of soil-scree-bedrock combinations. They do not appear to contain any wet soils or other types of named genetic soils. The unit was only mapped in six delineations in two watersheds (Table 3), but most of these were relatively large delineations, and comprise a large proportion of the soil found at high-elevations in the two watersheds. This unit, more than any other one in the survey, requires further field inspection to verify mapping.

Alpine Brown Soils and Talus (BT) -- These areas contain well drained to excessively drained soils formed in coarse granitic or mixed lithology colluvial parent-materials. Many delineations have variable depths to bedrock, but on average most areas are deep. The shallower areas can usually be readily recognized in the field, but are difficult to interpret from aerial photographs. Mixed colluvial deposits including soil, talus, loamy textured colluvium and scree are common. In some areas the fines appear to have primarily filled in talus deposits, in other areas talus, scree and loamy reworked till seem to be interstratified. Colluvial boulders and large stones are common in some delineations. The Alpine Brown Soils are mostly deep to bedrock with unusually high rock fragment content and moderate to weak subsoil development. Shallowly developed young soils, and stratified mixtures of young soils are common. Soil age, stability and degree of development appear more variable in these colluvial deposits than in most glacial till deposits. Most slopes are moderate to steep talus slopes below cliffs or are confined within faults and joints. Very small areas of wet soils are included in this unit, usually around small seeps. Seeps are common, especially near the edges of delineations in the Pear Lake watershed. This unit was not examined in detail, but probably contains deep buried soil layers and complex stratigraphy in many places. The unit appears generally similar in the eleven mostly small delineations, where it was mapped in the Pear, Ruby and Spuller Lake watersheds (Table 3). At present there do not appear to be good reasons to subdivide this mapping unit.

Moist Meadow Soil Complex (MMC) -- These areas contain complex mixtures of soils and non-soil materials formed from mostly granitic alluvial parent-materials. The soils vary widely in drainage and redox conditions but are mostly somewhat poorly drained Moist Meadow Soils with evidence of fluctuating redox conditions in subsoils. Most delineations also contain some more poorly drained soils, some better drained soil areas and small rock outcrops. Soil depths appear variable in most delineations;

moderately deep and shallow soils dominate most delineations. Most delineations are small with complex bedrock controlled microtopography, short steep moisture gradients and fine scale mixtures of two or three genetic soils. In addition to Moist Meadow Soils, Alpine Brown Soils, Riparian Soils, Wet Meadow Soils and unnamed intermediate soils are all found within this mapping unit. Moist Meadow Soils, unnamed intermediate soils similar to Moist Meadow Soils, Wet Meadow Soils, and rock outcrops are the most common components. These variable complexes are generally similar in the many (24), mostly small delineations mapped in five watersheds (Table 3). Since so much of the soil variation within these areas is over short distances, subdivision into different mapping units would not be productive unless a much more intense mapping scale (about 1:1000) was used.

Rocky Riparian Soil Complex (RR) -- These areas contain mostly shallow, rocky and wet soils formed from granitic, mixed or volcanic parent-materials. Although the soils are wet, they show no evidence of either chemical reduction or fluctuating redox conditions. They are usually moist to saturated throughout the growing season, but the soil water appears to remain well oxidized as it moves quickly through the soil. Most areas could be described as seeps, stream corridors or intermittent snow melt channels. High flow rates through the soil seem critical in differentiating the wet oxidized Riparian Soils from the two Meadow Soils. The most important components in most delineations are: Riparian Soils, rock outcrops, stones, boulders, and talus. Riparian Soils are mostly shallow to very shallow, but similar unnamed moderately deep to deep soils form inclusions in some delineations of this mapping unit. Most of the many delineations of this unit are confined by bedrock structures. Unnamed soils intermediate between Riparian Soils and either Moist Meadow Soils or Alpine Brown Soils also form small inclusions in some delineations. This unit is generally similar between watersheds, although a number of variations have been noted. This was one of the most common mapping units used with a total of forty one delineations in all six lake watersheds (Table 3). Most delineations are small and elongated in shape, forming portions of drainage networks. One of several provisional subdivisions of this unit was kept in the legend (RRT below) and several others are possible. Quantitative field and laboratory data are needed before any more subdivisions can be properly evaluated. This unit represents small volumes of soil thought to play critical hydrologic and biogeochemical roles.

Rocky Riparian Soils and Talus Complex (RRT) -- These areas contain mostly rocky shallow wet soils over granitic talus or talus-rubble mixtures. This unit is similar to the Rocky Riparian Soil Complex (RR) in drainage and redox conditions; it is wet but remains well oxidized. The two units differ primarily in depth to bedrock; this unit has shallow soils formed in the upper part of deep talus, talus-rubble, or (rarely) alluvial stone deposits. These areas are not as clearly confined by bedrock structures as the other unit (RR). This unit can occur in many similar types of seep, stream corridor, and snow melt channel positions as the other Rocky Riparian Soil unit (RR), but occurs where bedrock has been deeply buried in fragmented materials (talus, rubble or alluvial stones). It is most often found on ledges or in cliff-base talus slopes where seepage and talus coincide. Like the other Riparian Soil unit (RR) this one occurs in many (42) mostly small delineations in several (4) watersheds, but most delineations (27) are in the mixed parent-materials of the Spuller Lake watershed (Table 3). These small volumes of soil are thought to play important hydrologic and biogeochemical roles, and therefore are worthy of detailed study.

Volcanic Brown Soils and Rock (VBR) -- These areas contain well drained to excessively drained soils formed in coarse pumice and volcanic ash (tephra) dominated parent-materials. Most delineations are variable depth deposits of colluvially reworked recent tephra around small rock outcrops. Glacial till, and granitic or mixed lithology colluvium are assumed to be buried under the recent tephra, but these materials have not been actually observed and described. The named Rock component is almost all rock outcrop, but colluvial stones and boulders from cliffs are also included. The rock itself may be either volcanic or granitic. Some delineations contain small areas of pure unstable pumice (volcanic scree) deposits. The Volcanic Brown Soils in this unit are variable in depth, but appear to be mostly deep and moderately deep. No distinctive morphologic variations of the Volcanic Brown Soils were observed to be characteristic of this particular mapping unit. Beyond depth variations and the fine scale variations in stratification characteristic of all Volcanic Brown Soils, little other genetic soil variation was observed. No other genetic soils or significant differences in drainage conditions were observed. This unit was mapped in fourteen delineations, including medium and large delineations, all in the Crystal Lake watershed (Table 3). The unit could be subdivided, but more quantitative field and laboratory data would be required to do so.

Volcanic Brown Soils and Scree (VBS) -- These areas contain mostly excessively drained soils formed in coarse pumice and volcanic ash (tephra) dominated parent-materials, mixed with unstable pumice or scree deposits. Most delineations are deep, stratified deposits of colluvially reworked recent tephra where soils have had time to form in most of the more stable areas. Bedrock appears to be deep to very deep in most places and both glacial till, and colluvium may be buried under the recent tephra in a few places. In addition to the scree component, small inclusions of talus, colluvial stones and small outcrops may be present. The Volcanic Brown Soils in this unit are somewhat variable in depth to bedrock, but deposits appear to be mostly deep to very deep. Although deposits are deep in most parts, they do not typically contain mostly deep, well developed soils. Shallow, weakly developed soils interstratified with non-soil pumice layers are common. Other distinctive morphologic variations of the Volcanic Brown Soils were not specifically observed to be characteristic of this mapping unit. Beyond some depth variations and the stratification described above, little other genetic soil variation was observed. No other genetic soils or significant differences in drainage conditions were observed. Most soils are excessively drained and some are well drained. This unit was mapped in nine small and medium sized delineations in the Crystal Lake watershed (Table 3). Given the complexity of colluvial stratification, this unit could easily be subdivided, but more quantitative field and laboratory data would be required to do so.

Volcanic Brown Soils and Talus (VBT) -- These areas contain mostly excessively drained volcanic soils, with a mixture of colluvial non-soil materials. The soils are formed in a mixture of coarse pumice, volcanic ash, and talus. Most of the non-soil material is talus, but volcanic scree (pumice) is an important unnamed component in a number of places. Most delineations are deep stratified colluvial deposits where the talus and volcanic strata are mixed in a variety of ways. For example, in some places recent talus deposits or individual fragments overlie volcanic soil and pumice strata, whereas more commonly soils seem to have formed where volcanic fines have filled voids in fragmented talus deposits. Depth to bedrock is variable, but is mostly deep. Some of these deposits may bury small areas of glacial till or other types of colluvial deposits. The named Talus component can be either of volcanic or granitic rock. The unnamed scree (pumice) inclusions were almost important enough to be named as a component, and in addition small rock outcrops may be present as non-soil inclusions. The Volcanic Brown Soils in this mapping unit are unusual in their high content of rock fragments. They are generally deeper, more stable and better developed than the soils in the Volcanic Brown Soils and Scree unit described above, but they are also interstratified with non-soil materials in places. Other distinctive morphologic variations of the Volcanic Brown Soils were not specifically observed in this mapping unit. Beyond the rock fragment content and stratification described above, little other genetic soil variation was observed. No other genetic soils or significant differences in drainage conditions were seen. This unit was mapped in five small and medium sized delineations in the Crystal Lake watershed (Table 3). Given the complexity of colluvial stratification, this unit could easily be subdivided, but more quantitative field and laboratory data would be required to do so.

Wet Meadow Soil Complex (WMC) -- These areas contain complex mixtures of soils and non-soil materials formed from mostly granitic parent-materials. The soils vary significantly in drainage and redox conditions but are mostly poorly drained Wet Meadow Soils with evidence of fluctuating redox conditions, anaerobic conditions or peat accumulation in surface soils and clear evidence of chemical reduction in subsoils. Most delineations also contain some poorly drained soils, some better drained soil areas and small rock outcrops. Soil depths appeared to be variable to highly variable in most delineations; moderately deep and shallow soils dominate most delineations. Most delineations are small with complex bedrock controlled microtopography, short steep moisture gradients and fine scale mixtures of at least two named genetic soils, and many unnamed variants of wet soils. In addition to the named Wet Meadow Soils, Moist Meadow Soils, unnamed intermediate meadow soils, and in some cases Riparian Soils were all found within this mapping unit. Shallow variations of meadow soils are common. Rock outcrops are the major non-soil inclusion. These variable complexes are generally similar between the two watersheds where they were mapped (Pear and Topaz Lake). Only five small delineations of this unit were mapped (Table 3), so subdivision cannot be justified.

A.3.3.3. Non-Soil Dominated Mixed Units

These are mixed non-soil and soil areas where the non-soil materials were estimated to cover between 50% and 90% of the land surface. Rock outcrop was named in two units and non-soil unconsolidated surficial materials were named in three. They were estimated to have between about 10% and 49% soil cover including large proportions of shallow and shallow soils in most cases. Most of these areas have relatively thin, but variable surficial deposits, including soils, over consolidated bedrock. Efforts made to minimize the number of mapping units in this group eliminated most fine distinctions between specific mixtures of non-soil materials, and some finer expressions of subtle soil variations.

Boulders and Alpine Brown Soils (BLB) -- These unusual middle to high-elevation areas contain pockets of coarse well drained soils in between granitic boulders and large stones. The depths to bedrock and deep compositions of these deposits appeared quite variable and were difficult to estimate. The genesis and evolution of these deposits are not well understood. Some areas appeared to be pure boulders at the surface, with "boulder-till" at greater depths. Other areas appeared to be pure deep boulder deposits that have only irregularly accumulated small pockets of soil. Although there was some evidence of colluvial and periglacial movements, most of these deposits appear essentially glacial. One small delineation near Ruby Lake appeared to be primarily colluvial. Some of the soil pockets may have been formed in loamy till, and others in mixtures of more recent wind and water deposited fines. Periglacial sorting of fines was not apparent. The Alpine Brown Soils observed in this unit were unusually high in rock fragments, variable in depths to boulders, and sometimes sat as isolated soil pockets above and surrounded by large unfilled voids between boulders. Otherwise they appeared generally similar to other moderately developed granitic and loess derived Alpine Brown Soils at these middle to high-elevations. Other genetic soils and other soil drainage conditions were not observed in this unit. The unit was only mapped in seven delineations in the Ruby and Spuller Lake watersheds (Table 3), but some of these are large delineations comprising much of the soil material at middle to high-elevations. The unit contains enough variability to consider its subdivision, but the operational problems of working with boulders would make mapping of any subdivisions difficult.

Rock Outcrop and Alpine Brown Soils (ROB) -- These common rock outcrop dominated areas contain a variety of fine scale non-soil/soil patterns. Most of the soils are shallow to moderately deep and well drained. The fine scale patterning, the proportional composition of soil and non-soil components, and soil depth patterns cannot be described in general terms across all the many delineations (137) of this unit in seven watersheds (Table 3). In most cases pattern, composition, and depth were determined by the specific bedrock structures, especially jointing patterns, within a delineation or across larger areas of several delineations. Joint orientation, for example, was a critical determinant of soil depth distributions, whereas joint spacing usually determined the proportions of soils versus bedrock. Slope gradient, other topographic factors, mechanical rock weathering patterns and a number of other site specific factors were also involved. Most of these combinations can be locally described and most of the key properties of interest can be estimated over small areas with field inspection. Alpine Brown Soils in various forms (or phases) are the main genetic soils in this unit, but some high-elevation delineations include unnamed wetter soils. The Alpine Brown Soils are variable in depth. They also vary to a lesser degree in drainage and degree of development in this mapping unit. Their drainage varies from moderately well drained, through well drained to excessively drained. The unit includes some deep and well developed soils, especially at lower elevations under conifers and many intermediate depth and development stage soils. Most soils are shallow, moderately deep or have variable rock contacts. At higher elevation this unit was used for many small delineations with a variety of poorly developed shallow soils not well described by any of the presently named genetic soils as discussed elsewhere. Given its variability, subdivision of this unit is desirable, but additional field and laboratory data is required first.

Rock Outcrop and Volcanic Brown Soils (RVB) -- These areas are dominantly small rock outcrops finely mixed with variable depth excessively drained volcanic soils. The rock may be volcanic or granitic, but the soils are composed mostly of colluvially reworked recent pumice and volcanic ash. Some delineations contain colluvial stones, and small areas of volcanic scree (loose pumice). The Volcanic Brown Soils in this unit are variable in depth, and appeared to be mostly shallow to moderately deep. No other distinctive morphologic variations of the Volcanic Brown Soils were observed to be characteristic of this particular mapping unit. No other genetic soils or significant differences in drainage conditions were observed. This was the most commonly mapped unit in the Crystal Lake watershed with twenty six mostly small to medium sized delineations, and was not mapped in any other basin (Table 3). With adequate quantitative data this unit could be subdivided, but such an effort doesn't appear justified.

Talus and Alpine Brown Soils (TB) -- These areas are relatively inactive and stable talus or mixed stability talus and scree slopes in which significant areas of coarse excessively drained soils were able to form. The deposits are mostly deep to bedrock, but the soils vary widely in depth and degree of development. Interstratified soil and non-soil materials are common. Scree is not presently a named component in this unit, but it is prevalent enough in some individual delineations to be named. These mixed scree areas were mapped in now defunct provisional mapping units naming both scree and talus. Most Alpine Brown Soils in this mapping unit are unusually high in rock fragments and only moderately developed; many are weakly developed without distinct B horizons (Table 5). Most of these soils appear to have formed in relatively stable and medium textured colluvial strata, but some look like small pockets of fines that filled into the voids of fragmented talus deposits. Small seeps were found along the edges of a few delineations, but no wet soils were found within delineations. This unit was not examined in great depth or detail, but probably contains deep buried soil layers in many places. The unit appeared generally similar in the thirty three delineations in four watersheds where it was mapped (Table 3). Detailed examination of these areas would allow this unit to be subdivided and when adequate quantitative data are available such subdivision may be justified. For example, separation of areas where soils totally lack B horizons can be justified.

Talus Rubble and Alpine Brown Soils (TRB) -- These areas consist mostly of mixtures of stable talus and frost shattered rubble with small pockets of finer soil materials. The soils are variable in depth and rock contact geometry, and are mostly well drained to excessively drained. Some of the angular rock fragments covering the surface are clearly colluvial talus, other rock fragments are residual frost shattered bedrock rubble, and much of this rocky debris has mixed, intermediate or unclear characteristics between these two named components. Talus is not necessarily any more prevalent than rubble in every delineation. Massive unfractured bedrock is rare and is only present in areas too small to map. All of the Alpine Brown Soils observed had unusually high contents of rock fragments and appeared to have formed from mixed granitic-derived soil and loess filling in voids in fragmented talus and rubble materials. There were indications of deep pockets of soil in bedrock cracks and the lower parts of talus and rubble below purely fragmented material. Talus may have buried soils in other places. These indications of deep soil pockets and the unusual fractured bedrock geometry (and not the nominal distinction between talus and rubble) were the main reasons this unit was split out from the stratified colluvial Talus and Alpine Brown Soil unit described above. Pockets of soil only in the upper parts of fragmented deposits with pure fragmented material below (the "Ranker Soils" of Kubiena (1953), Tedrow et al. 1977) may be present, but were not observed. Very little material resembling scree was observed. Soils were not examined in much detail within this unit, but no other genetic soils were indicated. The unit is restricted to seven delineations in one watershed (Lost Lake; Table 3). There are no apparent reasons to subdivide it.

A.3.3.4. Predominantly Non-Soil Units

These seven units are for areas estimated to contain less than 10% soil cover on an area basis. Most delineations contain much less than 10% soil cover. The included soils are usually shallow. One unit is for rock outcrop; most units describe surficial deposits of non-soil materials. These surficial deposits may function like soils hydrologically, and may also contain deeply buried soils. Efforts made to minimize the number of mapping units in this group eliminated most fine distinctions between mixtures of non-soil materials.

Boulders (BL) -- These areas are essentially covered with granitic boulders and large stones, with few, if any, small pockets of soil. One delineation in the Pear Lake watershed, and another adjacent to Ruby Lake are colluvial deposits, essentially coarse talus. The other delineations are middle to high-elevation glacial drift deposits, of unknown depth and genesis. These deposits have the general shape and distribution patterns of moraines, but also have minor colluvial and periglacial characteristics. In some places the boulders appeared to contain no soil or fine grained material at any depth; in other places there were indications of some bouldery-loamy glacial till under the upper 2 or 3 meters of boulders. The few observed soils included in these areas were shallow Alpine Brown Soils. Similar alpine and arctic bouldery deposits have been called by numerous names including: block fields, boulder fields, and felsenmeers. Several glacial, periglacial, colluvial, and weathering processes are involved in the formation of various different kinds of bouldery deposits. In several places streams can be heard running through the boulders at depths estimated at 2 to 5 meters. This unit was mapped in seven delineations in three watersheds (Table 3). There are no apparent reasons to subdivide it.

Rock covered Glaciers (RG) -- Two high-elevation areas of the Ruby Lake watershed apparently contain small alpine glaciers mantled with thick talus-like accumulations of rock fragments. Nearly pure glacial ice was observed adjacent to the north bergschrund (ice to rock gap) near the terminus of the larger, southern glacier. The surface of this glacier was observed from a nearby cliff, and no soils were apparent. The smaller glacier to the north was not examined in the field. It is possible that one or both of these glaciers might be better described as "rock glaciers" but resolving this question would probably require expert on-site investigation of the composition of the glaciers' cores. In addition to the main areas of glaciers both delineations also contain smaller areas of related surficial deposits including talus and permanent snow fields. These areas appeared to be essentially devoid of soils. Although surficially similar to talus, they were recognized in a special mapping unit for their unique hydrologic properties.

Rock Outcrop (RO) -- These areas are dominantly bedrock outcrops of any rock type, and with a variety of jointing and fracturing patterns. While soils occupy less than 10% of these areas, several types of non-soil materials in addition to consolidated bedrock are locally important inclusions in some delineations. For example 15 to 25% inclusions of shallow rubble or talus is common in some areas. Most delineations were drawn to keep soils inclusions to a minimum (less than 5%), and most included soils are shallow Alpine Brown Soils. Very small seepage areas with soils similar to Riparian Soils are also locally important inclusions. This unit was mapped in many (119) delineations in all watersheds. It could be subdivided, but any such subdivision should be done with a geologist.

Talus (T) -- These areas are mostly deep to bedrock colluvial rock fragment deposits, with the talus derived from any rock type. Most parts of most deposits are purely fragmented (large voids contain no fines). Talus deposits vary in age, physical stability, specific stratification and depth to bedrock. Some talus deposits are intermittently shallow. Small areas of buried soils, and larger areas of buried fine strata are likely under many talus deposits. Included soils are mostly rocky and poorly developed versions of Alpine Brown Soils; small seepage areas may contain Riparian Soils. Small areas of other types of non-soil materials were commonly included with talus delineations. These included boulders and large stones, small outcrops, scree and rubble. Talus was mapped in many delineations (34), including medium and large delineations, and was found in most watersheds (Table 3).

Talus and Rubble (TR) -- These areas are mixtures of relatively stable talus, frost shattered rubble, and intermediate forms of fragmented material. Very few areas contain any soil or fine grained material near the surface, but deep pockets of fines in bedrock cracks are probably common in places. The bedrock fracturing patterns and contact geometries between solid bedrock and overlying fragmented materials appeared complex. The few included soils are rocky and poorly developed Alpine Brown Soils. The basic rationale, and many of the descriptions of the Talus, Rubble and Alpine Brown Soils (TRB) unit given above also apply to this unit containing less soil. Unique bedrock contact geometry was more important than nominal distinctions between talus and rubble. It was mapped in twelve delineations in the Lost Lake watershed, and nowhere else (Table 3).

Talus and Scree (TS) -- These areas contain interstratified colluvial mixtures of fragmented materials (talus), and finer loose sandy materials (scree). This unit was only mapped in granitic rock. Most parts of most delineations are deep to bedrock, but depth is quite variable in some places. Depth relationships can usually be estimated in the field without excavation on an individual delineation basis. Some areas may contain more scree than talus since efforts made to minimize the number of mapping units eliminated most fine distinctions between mixtures of similar non-soil materials. These types of variation in composition were apparent in the field and in some cases on aerial photographs. In addition to the named components this unit includes other types of non-soil (bedrock, stones and boulders) and some small areas of soils. The soils are mostly rocky and poorly developed Alpine Brown Soils. Deeply buried soils are also likely in some of these areas. This unit was mapped in ten delineations in the Pear and Ruby Lake watersheds (Table 3). There is no apparent reason to subdivide this unit.

Volcanic Scree and Talus (VST) -- These areas contain mostly deep, unstable, interstratified colluvial mixtures of loose pumice (volcanic scree) and various types of rock fragments (talus). Depth relationships are variable in places, but depth can generally be estimated on an individual delineation basis in the field without excavation. Some areas contain more talus than volcanic scree since efforts made to minimize the number of mapping units eliminated most distinctions between mixtures of non-soil materials. These types of variation in composition were apparent both in the field and on some aerial photographs. Small areas of outcrops, stones, boulders, and Volcanic Brown Soils were included in mapping. Deeply buried weakly developed soils are probably present in some delineations. This unit was mapped in just three delineations, all in the Crystal Lake watershed (Table 3). Although it contains considerable stratification and variability there are few good reasons to subdivide it.

A.3.4. Individual Watershed Descriptions

A.3.4.1. Pear Lake

Overview -- The Pear Lake watershed is similar in several respects to most of the other watersheds in this study: It is a well defined geologically simple, recently glaciated granitic basin with minor volcanic ash accumulations in places. No unique mapping units were needed to describe the soil patterns in this basin, and ten of the fourteen mapping units used in the Pear Lake basin were common units also used in at least three other basins (Table 3). The Pear Lake area is similar to the Ruby Lake area in that both contain large relatively uniform areas of rock and mixed areas dominated by non-soil materials. The soil maps reflect these patterns with large delineations, especially at higher elevations. These patterns are largely controlled by the glacial topography and bedrock structure, reflecting high relief and massive granitic bodies with widely spaced jointing patterns and little frost shattered rubble. The Pear Lake watershed appeared unique in having pure moderate to low relief rock outcrop areas (much less than 5% soil) in the middle basin due to glacial scouring of essentially unjointed rock masses. The unusual patterns of seeps and soils in these massive bedrock areas are described below.

Topography and Physical Geography -- The Pear Lake watershed is intermediate in size and elevation and relatively steep with a wide elevation range when compared to the other watersheds (Table 4). This 136 ha north facing watershed is similar in size to the Topaz, Spuller, and Crystal Lake watersheds. The lake elevation of about 2,904m is similar to the elevation of Crystal Lake, but lower than most of the other lakes. However this steep watershed rises about 471m in a distance of only 1.2 km, and has an upper elevation of about 3,375 m.

Glacial erosion has given this watershed a distinctly U-shaped and stepped topography with an overall trough or bowl shape comprised of several smaller bowls or cirque basins. In this watershed there are two relatively large and distinct sub-basins: a lower elevation northern sub-basin around the lake itself, and an intermediate elevation (3,110 to 3,170 ; 10,200 to 10,400 ft) sub-basin to the southeast of the lake. In addition there are smaller cirque basins (under the large high-elevation cliffs), large bedrock benches, steep talus and boulder fields, and a complex variety of other types of small slopes. Since this steep watershed faces north and contains large north and east facing cliffs, the shadowed middle and upper parts of it are relatively cold for this elevation. Late snow melt also contributes to these areas having more of a short growing season-alpine character than most other slopes at similar elevations. This alpine character is reflected in minimally developed soils. The overall steepness of this watershed contributes to the relatively high proportions of colluvial deposits (such as talus), and soils formed in colluvial deposits.

Bedrock and Surficial Geology -- The following description of bedrock geology was primarily taken from a 1:62,500 scale geologic map of the Triple Divide Peak Quadrangle (Moore and Sisson 1987). Such intermediate scale maps express little specific detail about rock composition, and little or nothing about surficial deposits. Three types of granitic rock were mapped in the Pear Lake watershed. At least three quarters of the basin was mapped as the Granite of Lodgepole Campground, a coarse grained true granite containing sparse variable mafic inclusions. This was the only true granite mapped in any of the watersheds in this study, and also was among the coarsest and most felsic rocks mapped in any of the study areas. The eastern and southeastern parts of the basin were mapped as two types of granodiorite in the Mitchell Intrusive Suite: (1) Mitchell Peak Granodiorite, fine facies, and (2) Granodiorite of Castle Creek. The Mitchell Peak rock forms much of the high eastern basin divide including the unnamed 11,328 ft peak and its flanks. The Castle Creek unit forms a narrow band between the Mitchell Peak rock and the Lodgepole Campground Granite. This band crosses the northern and southeastern basin divides. The Mitchell Peak unit was described as a fine grained porphyritic granodiorite with abundant mafic inclusions and K-feldspar, hornblende, biotite, and plagioclase phenocrysts. The Castle Creek units were described as a medium grained, equigranular, and hornblende-rich granodiorite. Both granodiorite units were described as including areas of the other unit or as being marginal to the other unit. Similar granodiorites were mapped in several other watersheds.

Granitic rocks vary somewhat in their resistance to chemical weathering based on differences in texture, mineralogy and bulk elemental chemistry. Factors which increase resistance are coarse textures, high quartz, K-feldspar and Si contents. Finer textures, higher plagioclase feldspar and dark mineral contents, higher Ca, Mg, and Fe contents all make granitic rocks more susceptible to chemical weathering. Therefore the coarse, light colored true granites should be more resistant than the darker and finer grained granodiorites in the Pear Lake watershed. In theory, soils derived from the *in situ* weathering of such contrasting granitic rocks would also vary in a predictable ways. These potential differences may or may not be observable in the Pear Lake basin because most soils form in mixed transported parent-materials and other soil forming factors also vary considerably across the watershed.

The most significant surficial geology features of this basin are small areas not expressed on the geology map. The prominent southwest to northeast trending fault which cuts across the Pear and Emerald Lake watersheds contains unusual soil materials possibly including fault gouge clays. Just north of Alta Peak is a small area of deep glacial till in a moraine complex. Tephra deposits are thin and discontinuous, and were not mentioned or described by Moore and Sisson (1987) on the geologic map, but were mentioned by Huntington and Akeson (1987) in the soil survey of Sequoia Park.

The geologic parent-materials for Pear Lake basin soils are mostly locally derived glacial till, reworked till and colluvium with locally significant accumulations of tephra, loess and stream alluvium. Post glacial mechanical weathering, rock fall and other forms of colluvial movement have been important recent geomorphic processes in this steep basin.

Landscape, Soil and Vegetation Patterns -- The intermediate scale patterns of topography, landforms, drainage, soils and vegetation in this watershed can be described in terms of three areas. The lower elevation northern basin around the lake outflow contains coniferous subalpine forest and open coniferous woodland interspersed with rock outcrop. Most of these soils are well drained to excessively drained; wetlands and soils with intermediate drainage conditions are mostly limited to small meadow and marsh areas immediately adjacent to the lake or outflow stream. Most of these moist to wet areas were too small to be mapped out in the soil survey. The setting and patterns around other parts of the lake contrast with the outflow area. Massive bedrock outcrops, toes of talus chutes and a few small wet delta-like areas control soil and vegetation patterns. These areas contain only a few scattered trees, perhaps due in part to talus-soil instability and avalanches. Three of the wetter areas were large enough to map out, the others areas were too small. Soils in talus chutes tend to be well drained to excessively drained, but small seepage areas are common especially near the margins of these areas.

Most of the Pear Lake watershed can be described as a mixture of large unvegetated rockland areas with scattered conifers, confined wet riparian zones, scattered shrub thickets and small isolated meadows of varying drainage soil conditions. The overall pattern is strongly controlled by well defined geologic structures (joint patterns and faults) and the distinctly stepped glacial topography. The rockland areas (rock outcrop, talus, and boulders) and non-soil dominated mixed units (mostly the Rock Outcrop and Alpine Brown (ROB)), and Talus and Alpine Brown (TB) units) form a broad scale pattern shown on the soil map with large delineations. Networks of mostly small elongate riparian zones and meadows break up the coarse scale pattern of rockland and non-soil dominated mixtures in several middle elevation parts of the basin. Most of these small riparian and meadow areas contain variable soils. These soils are mostly shallow but vary in depth, drainage condition and morphology within most individual delineations.

The high-elevation areas of the Pear Lake watershed have a relatively simple and coarse scale pattern dominated by large cliffs (rock outcrop), adjacent large talus slopes, and mixed units dominated by rock outcrop or talus. Most of this area is essentially unvegetated rockland, but small areas of dry to moist alpine meadow vegetation are common. A few scattered trees and shrubby areas are also present. Some of the simplicity in soil mapping at the highest elevations reflects decreased soil development and differentiation along moisture gradients. Only a few small areas of riparian or meadow soils were mapped in the highest elevation areas of this basin, because most areas lacked clearly apparent riparian zones or wetter meadows at a mappable scale. Even in the best differentiated areas, soil morphology was difficult to distinguish. Since the soils themselves in well drained areas, moister meadows and riparian zones at the highest elevations could not always be distinguished they were mapped together, usually in the Rock Outcrop and Alpine Brown (ROB) unit. Soils in these higher elevation areas on average are shallower than at lower elevations, but the actual depth distributions of soils are complex and not well understood.

Unusual Soils or Soil, Non-Soil and Water Patterns -- The Pear Lake watershed contains a unique pattern of shallow, Riparian Soils on ledges in association with seepage. These soils appeared to have lower rock fragment content, higher organic matter content, and other morphologic differences from typical Riparian Soils. Unlike the more common steep linear riparian zones these drainage systems involve isolated areas of Riparian Soils on moderately sloping ledges connected by seepage over and within the rock outcrops. Often the seepage is over relatively wide areas of darkly stained massive bedrock. Some of these areas were clearly visible above the south end of the lake. It is not known how the biogeochemical properties of these soils and waters may compare to those in more typical riparian zones, but results of laboratory analyses should help clarify the problem.

Layers and pockets of fine sand sized volcanic ash or tephra were found in the Pear Lake watershed. The pockets are usually near the mineral soil surface and often associated with strongly leached stable well drained soils under conifers ("podzol-like" soils). A similar soil was reported from the adjacent Emerald Lake watershed (Huntington and Akeson 1987), and was also observed during this study in the Ruby Lake basin. At present these leached well drained soils have been included with Alpine Brown Soils (Table 5).

A distinct layer of sandy tephra was sampled from deep in a stratified meadow soil adjacent to Pear Lake. This deep layer was one of about fifteen thin strata identified where a toeslope interfingered with a lake margin marsh. The age and source of this tephra, its mineralogical and chemical composition, and its exact distribution in the basins are all unknown. It was probably a widespread thin ash deposit that was subsequently eroded off rock outcrops and more open areas forming alluvial depositional strata in toeslopes, flatter meadows, ponds and lakes. It appears tephra was likely to be trapped and stabilized under conifers, where it also was likely to show the effects of organic acid leaching (podzolization).

Adjacent to two Talus (T) delineations and a Talus and Scree (TS) delineation below Alta Peak in the southwest corner of the basin an unusual delineation of Alpine Brown Soils and Rock (BrR) was mapped. The rock component described in this glacial moraine was large stones and boulders (not bedrock). The depth to bedrock in this area appeared to be at least two to three meters. The deep but rocky morainal soil pattern in this delineation was similar to a delineation adjacent to Topaz Lake and was considered for a separate mapping unit to clearly indicate the differences from mixed Alpine Brown Soil and bedrock delineations of BrR. However these deep morainal situations were too rare. This delineation of BrR is at a higher elevation and appears to contain colder drier soils than the soils adjacent to Topaz Lake.

A.3.4.2. Topaz Lake

Overview -- The Topaz Lake watershed has soils similar to most of the other watersheds, but it is also an unusual area in some respects. It lacks the distinct basin topography composed of steep cirques, large cliffs, bedrock steps and a deep central trough. Instead it is more a part of a high-elevation moderate relief glaciated plateau area bordering a major divide, with steep areas just outside the watershed. The lower basin is part of a plateau which extends to the south southeast across a complex and indistinct watershed boundary. The other basin boundaries are more distinct and often near transitions from moderate topography within the Topaz watershed to steep topography just outside it. The Topaz watershed is similar to most of the other study areas in having simple granitic lithology and essentially the same major genetic soils. No unique soil mapping units were needed to describe soil patterns in this watershed and only ten mapping units were used in it (Table 3). Of these ten, seven were common mapping units used in at least three other basins. The two unusual mapping units named for Wet Meadow Soils (WM and WMC), and the unusual relatively pure areas of Alpine Brown Soils (Br) were more important in the Topaz watershed than in any other watersheds. Although the range in soils through most of this basin is not great the intricate mixtures of similar soil and non-soil patterns were mappable at 1:4,000. This resulted in a complex soil map mostly composed of only four or five mapping units with one large delineation, a few other moderate delineations, and many small ones. These patterns of soils and non-soils were determined mostly by geologic structures (faults and closely spaced joint systems), and glacial erosion.

Topography and Physical Geography -- The Topaz Lake watershed is intermediate in size and elevation, but has relatively low relief and a narrow elevational range when compared to the other watersheds (Table 4). The Topaz Lake watershed at 178 ha, is larger than most, but is generally similar in size to the Pear, Spuller, and Crystal Lake watersheds. The outlet to the basin is relatively high (3,218m) but this watershed only rises 275 m (800 ft) in 1.2 km to a high unnamed ridge at 3,493m. The watershed drains west and faces generally southwest.

The overall topographic pattern of the Topaz Lake watershed is not itself a glacial trough, but is part of a high plateau or tableland bounded by higher ridges and steep cliffs dropping down to surrounding lower glacial troughs. The high ridge which forms the northern and northeastern Topaz watershed divides is also the Kings-Kaweah Divide. No distinct topographic break defines the southeastern to southern Topaz watershed divide--the Tableland Plateau continues for about another kilometer with indistinct drainage or southern drainage into Buck Creek. Surface drainage patterns are difficult to determine on the tablelands, and there are indications that a large proportion of the water percolates down into rock joints.

The Topaz Lake watershed contains a large low relief lower sub-basin around the lake. This is surrounded by broken and complex terrain, but with generally moderate overall relief. This complex and broken relief contains many small cliffs and flat ledges cut by deep soil filled joints and faults. In parts of the northern and northeastern basin smooth meadow areas with gentle to moderate relief are common. A small area of steep mountain summit ridge forms the northern boundary.

Moderate relief in this basin contributes to relatively low proportions of colluvial deposits, and few soils formed in colluvium. The combination of moderate relief and generally southwest facing slopes give some high-elevation areas of the northern watershed a longer growing season and less of an alpine-character than is typical for most other slopes at similar elevations. These areas contain relatively well developed soils, and alpine meadows.

Bedrock and Surficial Geology -- The bedrock geology descriptions that follow were adapted from Triple Divide Peak geology map discussed above for Pear Lake (Moore and Sisson 1987). The entire Topaz Lake watershed was mapped within a large delineation of Mitchell Peak Granodiorite, fine grained facies. Two small islands of the Mitchell Peak Granodiorite, coarse grained facies were mapped in the northwest part of the Topaz Lake watershed. The small islands of slightly contrasting rock are unlikely to affect soil properties. These and similar types of granodiorite were mapped in several other watersheds.

Surficial geology was not expressed on the geology map. This watershed contains relatively large and pure glacial till, alluvial and lake delta deposits in the lower basin around the lake. The higher elevation smooth meadow areas appeared to also contain deep, pure till deposits but this area needs more investigation. Above the lower basin most of the soils formed in glacial till deposited in between rock outcrops, or combinations of these till materials with *in situ* residual bedrock soils, and small pockets of colluvium. When compared to the other watersheds this one contains more residual materials, and far less colluvial materials. In addition tephra, loess, and stream alluvium are locally important.

Landscape, Soil and Vegetation Patterns -- The intermediate scale patterns of topography, landforms, drainage, soils and vegetation in this watershed can be generally described in terms of three areas. Topaz Lake is mostly surrounded by low relief meadows and marshes of varying drainage conditions. Some of these areas are quite complex, containing rock outcrop and mixtures of several soils in close association, whereas other nearly level slopes are relatively pure areas of the same genetic soil lacking rock outcrop. The large pure areas were formed in glacial till, alluvium, delta and lake deposits. These areas appeared as large and uniform as any in the 1992 study. Small and scattered conifers are present in some rock outcrop dominated areas near the lake outflow, shrubs are mostly willow thickets adjacent to meadows or riparian corridors. Dry grassy meadows are most common on well drained soils.

Surrounding this low relief area the great majority of the basin is a complex moderate relief landscape composed of many small rock outcrops, soil covered ledges and soil filled faults, joints and cracks. Careful inspections of many areas consistently revealed much more soil cover than was apparent on aerial photographs. Most of the observed soils were well drained Alpine Brown Soils in association with dry meadow or sparse shrubby vegetation, but complexes including wetter soils were also common. Soil depth appears variable with many shallow soil areas especially on ledges. Patterns and proportions of soil versus outcrop were variable. Soils contained high proportions of rock fragments in places, and mapped rock outcrops included some rubble but no significant areas of talus or colluvial boulder deposits were observed. Rock outcrops contain many small cliffs, ledges and flattened summit areas, and a few medium sized cliffs. Larger faults and joints have channeled most surface water into well defined linear rocky riparian zones and long narrow moist or mixed drainage class meadows. Conifers were mostly limited in distribution to a mid-slope band of scattered medium sized trees growing out of rocky soils near the middle of the basin. Their distribution did not seem to be related to any specific or unusual type of soil, rock, or drainage features. Microclimatic factors may have been important in this unusual vegetation pattern.

Portions of the higher elevation northern and northeastern basin have different general patterns of soils and vegetation from those described above. In some of these areas the intricate mixture of small jointed rock outcrops and soils characteristic of most of the basin is replaced by broad smooth meadows containing isolated rocks and rocky areas. The drainage condition of these meadow soils was not always apparent, but most appeared moderately well drained to somewhat poorly drained. The dominant soil could be intermediate between Alpine Brown Soils and Moist Meadow Soils. Botanical information and more detailed soil investigations are needed. Soils in these smooth meadow areas appeared mostly deep to deep, but the soils were coarse textured and contained high proportions of rock fragments in places. Some of these meadows appeared unusually productive given their high-elevation.

Unusual Soils or Soil, Non-Soil and Water Patterns -- Few unusual soils or unique soil patterns were found in this watershed partly because relatively little time was spent in it. For example, no tephra was found, although it is probably present. This was the first watershed mapped, so the legend was not fully developed. The relatively extensive areas of marsh and Wet Meadow Soil around the lake were unusual in this study, and so were the relatively pure areas of Alpine Brown Soils nearby the lake. These were apparently in deep glacial moraine deposits. The lake margins also had relatively large areas of Moist Meadow Soils, although this was not as unusual in this study.

The intricate pattern of rock outcrop and soil throughout most of the basin is unusual in its complexity but not in its component parts. The high-elevation, smooth meadow areas appeared unusual, and need more investigation. The highest elevation soils observed in this watershed (about 3,380 m or 11,100 ft) appeared stable, well developed, and well differentiated along moisture gradients, unlike those observed at similar elevations in the nearby Pear Lake area. These soil differences appeared to be related to differences in slope gradient, aspect, soil temperature and the length of the growing.

A.3.4.3. Mini Watersheds

Overview -- The Mini-watersheds are a pair of very small, low-relief, experimental drainage basins near the Pear Lake watershed. One watershed was well defined, but the other had an indistinct eastern boundary. They were mapped entirely with the three most common mapping units in the study, and no distinctive soils or soil patterns were found in them. Only one genetic soil, Alpine Brown Soils, was recognized in these basins. The three mapping units used simply reflect different proportions of soil and rock (Table 3). Some of the rock outcrop in one of the basins is composed partly of large detached granitic slabs. Such slabs are common in many parts of the Sierra Nevada, but were not common in any of the lake basins in the 1992 study. Since they were not extensive enough to name mapping units for, large slabs were treated as inclusions in rock outcrops.

Topography and Physical Geography -- The Mini-watersheds are by far the smallest areas mapped, with the lowest relief. They are intermediate in elevation, and similar in most respects to much of the Topaz Lake watershed and some eastern parts of the Pear Lake watershed (Table 4). Low relief contributed to low proportions of colluvial materials.

Bedrock and Surficial Geology -- These small areas were mapped entirely within the same large delineation of Mitchell Peak Granodiorite (fine grained facies) as the Topaz Lake watershed (Moore and Sisson 1987). The pattern of bedrock and thin surficial deposits is similar to the outcrops and ledges north and northwest of Topaz Lake. Most of the soil seemed to be locally derived residuum from *in situ* bedrock weathering but minor inclusions of till, loess, and tephra were probably also present as minor components.

Landscape, Soil and Vegetation Patterns -- Two small watersheds were chosen and named the soil watershed and the rock watershed. Both contain significant proportions of both soil and rock. The larger rock watershed has a higher proportion of rock and lower proportion of soil than the soil watershed, and also has an indistinct eastern boundary. The difference in the proportions of rock and soil between the two watersheds appeared great if the questionable eastern portion of the rock watershed was not included in comparisons. However, if the eastern area of the rock watershed was included in the total analysis, the absolute areas and volumes of soils in the two watersheds appeared to be similar. Both watersheds had soil dominated areas near their lowest or middle parts surrounded by rock dominated areas (ROB) and rock outcrop. Both watersheds appeared to contain relatively high proportions of shallow soils. No significant differences in soil drainage condition were indicated--all soils appeared well drained. Scattered small conifers and some shrubs were present in a matrix of mostly dry meadow vegetation and unvegetated rockland.

Unusual Soils or Soil, Non-Soil and Water Patterns -- The slabby parts of the rock outcrop in the southwest parts of the rock watershed are unusual. No other unusual soils, non-soil materials or patterns were noted, but little time was spent in these areas and no real soil pits were dug, described, or sampled.

A.3.4.4. Ruby Lake

Overview -- The Ruby Lake watershed is by far the largest and highest elevation watershed, and in many respects the most complex. However, it is similar to most of the other watersheds in several important ways: it is a well defined, and recently glaciated subalpine to alpine granitic basin with minor volcanic ash accumulations in places. The genetic soils, and soil patterns within the common mapping units are similar to those in most of the other basins, therefore the standard descriptions of soils and units apply well to most of the Ruby Lake area. Only one unique mapping unit was needed in this watershed, and eleven of the seventeen mapping units used in the Ruby Lake area were common units also used in at least three other watersheds (Table 3). The one unique mapping unit at Ruby Lake is Rock covered Glaciers (RG), this and most of the other unusual mapping units are also typically high-elevation units. The Ruby Lake and the Spuller Lake watersheds are especially similar; these two areas share fifteen mapping units.

Topography and Physical Geography -- The Ruby Lake watershed is by far the largest, highest, and steepest watershed, with the greatest range in elevation (Table 4). It is a long, relatively narrow and generally north-south oriented watershed. It runs from a high southern peak and glacier area northward through a distinctly stepped, U-shaped glacial trough down to Ruby Lake, which has an eastern outflow stream. A smaller side valley runs from Mono Pass south into Ruby Lake. The ridges surrounding this watershed are unusually high and steep, especially the western and southern ridges and peaks. The watershed has an area of 441 ha, an outlet elevation of 3,390m and vertical relief of 812m. The highest points in the watershed are the named peaks around its southern end: Mt. Abbot and Mt. Mills. In addition most of the unnamed peaks and ridges surrounding this watershed are also high; the watershed divide mostly ranges from 3,600 to 4,000m. The total elevation range of the watershed from Ruby Lake to Mt. Abbot is 787m in a distance of about 3.35 km. Many steeper gradients, including large, nearly vertical cliffs are characteristic of this watershed. For example an unnamed 4,000m peak is only about 0.9 km west of Ruby Lake, and almost half of this (610m) relief is in one large, nearly vertical cliff.

The size and complexity of this watershed led to its division into six areas for descriptions of overall landform, soil and vegetation patterns (see below). The alpine temperature regime contributed to high mechanical weathering rates. Large volumes of bedrock have been broken down into rock fragments of all sizes including sand. Extreme slope gradients accelerate these processes as the debris falls and slides away from the bedrock into large talus, scree, and mixed colluvial deposits. Such weathering and slope processes have affected most of the soils in this watershed since latest Pleistocene deglaciaton. High-elevations, cold air drainage, short growing seasons on cold, shadowed slopes also affect most soils in this watershed. Large parts of the Ruby Lake watershed have a set of high-elevation, alpine soil characteristics only found in the highest and coldest parts of two other watersheds.

Bedrock and Surficial Geology -- The Mt. Abbot Quadrangle geologic map (Lockwood and Lydon 1975) showed significant geologic complexity in and near the Ruby Lake area. One major fault, a mineralized contact between at least two granitic plutons, three types of granitic rocks, two mineralized rock units, and three types of surficial materials as well as several other features were all mapped in this watershed. Areas just east of the watershed were mapped with a number of additional features. The northern, western and southwestern parts of the Ruby Lake watershed were mostly mapped as part of a young pluton of quartz monzonite which also occupies about a third of the entire (15') quadrangle and extends into two other adjacent quadrangles. This large delineation of the Quartz Monzonite of Mono Recesses appeared geologically simple on this map; it contained surficial deposits, but few areas of other rock types and few geologic structures. The eastern and central areas of the Ruby Lake watershed appeared more complex, and seemed to represent the western, mineralized margin of an older mostly granodiorite pluton dominated by the Granodiorite of Chickenfoot Lake. This granodiorite was shown surrounding two types of mineralized rock and the Quartz Monzonite of Ruby Lake within the Ruby Lake watershed. This eastern granodiorite-dominated pluton area was also hatched with red lines indicating "zones in which felsic dikes are uncommonly abundant". Three other types of rocks were mapped just east of the watershed, apparently part of this same pluton. Surficial deposits mapped within the watershed included Talus, Glacial Moraine and Fill, and a small area of Olivine Trachybasalt (Tt) also described as a dacite dike. The geologic map showed alpine glaciers in two adjacent watersheds, but none within the Ruby Lake watershed.

The Quartz Monzonite of Mono Recesses was described as typically coarse grained and strongly porphyritic, with minor hornblende and sphene. The Granodiorite of Chickenfoot Lake was described as medium grained and slightly porphyritic in places. It was also described as being strongly sheared, and cut by many mafic dikes in its southern part. The Ruby Lake watershed probably falls somewhat north of the southern part described. The Quartz Monzonite of Ruby Lake has been mapped only in a small area right around the eastern and southern parts of Ruby Lake, where it is shown surrounding a smaller area of mineralized rock. It was described as fine to medium grained biotite quartz monzonite; the mineralized rock was described as masses of quartz-grossularite-epidote-clinopyroxene rocks. A larger body of mineralized rock was shown crossing the eastern basin divide and covering an area south of Mills Lake in the middle Ruby Lake basin. This rock was described as areas of oxidized and rust-stained granitic rock with altered feldspars and common sulfide minerals. This is the same area where iron oxide rich variations of Alpine Brown Soils were observed. The short report accompanying Lockwood and Lydon's 1975 geologic map contained a number of good references including one on the chemical and mineralogical composition of the major rock units (Lockwood 1975) and others on landscape evolution and glacial geology (Bateman and Wahrhaftig 1966; Birman 1964). The possible effects of these rock formations on soil formation within the Ruby Lake watershed may be discussed elsewhere.

The surficial geology of this relatively large, high-elevation and steep basin appeared complex. A number of different types of rock and types of glacial drift contributed soil parent-materials in this geomorphically complex landscape. Since latest Pleistocene deglaciation there appears to have been a great deal of mechanical weathering and colluvial redistribution of surficial materials (including drift and soils) within the watershed. There have also been alluvial and aeolian processes at work. Significant glacial advances and retreats in the middle basin-Mills Lake area are reported to have taken place during the Holocene "little ice ages" (Birman 1964). Tephra was not shown on the geologic maps, nor was it mentioned in any of the reports. It appeared to be locally important as a soil parent-material in at least parts of the lower basin. It was most apparent under large conifers and in stratified meadow soils. Minor loess deposits were apparent throughout the basin, and much of the high-elevation soil material appears to have been wind deposited.

These recent surficial processes, combined with glacial processes and bedrock variations have created a complex pattern of soil parent-materials in the Ruby Lake watershed. The recent processes appeared to have been active enough to have deposited or at least reworked most soil materials in this watershed during the middle to late Holocene. Most Pleistocene glacial drift appeared to have been recently eroded, buried, or colluvially reworked.

Landscape, Soil and Vegetation Patterns -- The intermediate scale patterns of topography, landforms, drainage, soils and vegetation would be difficult to describe for this watershed as a whole. The Ruby basin contains at least five or six distinct areas including small sub-basins. The six areas used for description in this report are: (1) the lower basin around Ruby Lake; (2) the northern slopes and side valley area from Mono Pass to the lower basin; (3) the middle basin around Mills Lake and numerous smaller ponds (3,475-3,600m); (4) the upper basin southwest and south of the middle basin (3,600-3810m) ; (5) the high western and southern cliff and mountain ridge system running south from just southwest of Mono Pass to just past the summit of Mt. Abbot; and (6) the eastern basin ridge system running from an unnamed (3,932m) peak northeast of Mt. Abbot north-northeast to an unnamed (3,628m) peak southeast of Ruby Lake.

The lower basin around Ruby Lake itself contained the greatest observed variability in soils and soil forming factors. The northeast parts of this area were relatively simple, whereas the southwestern parts were complex. The simpler areas were mapped as medium to large delineations of Talus (T), mixed Talus and Alpine Brown Soils units (TB and BT), and the most common Alpine Brown Soils and Rock units (BrR, ROB, and RO). The Alpine Brown Soils observed in all of these units had a strongly colluvial character, and were well drained to excessively drained. Soil depths were variable and not well understood, but many soils appeared to be both deep and rocky. Some of the soils north of the lake appeared unusually warm and dry given their elevation. The one unusual mapping unit (BrS) in this area was used for a deposit of loose unstable sandy soils interstratified with scree north of the lake. This mapping unit (BrS) was normally mapped at higher elevations in more alpine settings. The area north of the lake had mostly dry subalpine woodland vegetation, whereas southeast of the lake the subalpine vegetation appeared somewhat more mesic. The smaller delineations of wetter soils (RR, BC, and MMC) were mostly similar to their standard descriptions, with one exception. A prominent group of springs along the north side of the lake near an island displayed unusual soil characteristics. The fine scale patterns of soil variation within these spring complexes could not be adequately mapped or described in separate mapping units. The complex contained small pockets of peat, reduced mineral soil, and on one visit contained warm water. Although mapped as Riparian Soils (RR) much of the complex fits no recognized genetic soils.

The southwestern part of the lower Ruby Lake basin contains most of the major inflows, in a complex soil-landscape pattern. Near the lake these patterns were mapped with many small delineations of a number of different mapping units, and even this map was a simplification. This area was so complex that it would have included far more delineations and mapping units if the map had been at a larger scale with a larger legend. Bedrock, talus and colluvial boulders were the main non-soil components seen in this mixture with Riparian Soils, Moist Meadow Soils, Alpine Brown Soils and several unnamed included soils. Wet Meadow Soils were only a minor inclusion in some areas, so although much of this area was wet it was also well aerated. Soil depth patterns and stratification patterns within colluvial deposits appeared to be complex; they were not fully investigated and could not be well described. Shallow soils, and areas of variable soil depth appeared common. Several different soils in this area were sampled, and many field notes were taken. The small bouldery delineations (BLB, BL) in this particular area appeared to be colluvial. The vegetation patterns in the complex soil landscape around the southwest parts of Ruby Lake were complex mixtures of subalpine meadows of several types, shrub communities and scattered conifers. Above this complex area near the lake were much simpler and less distinctive Talus, bedrock and Alpine Brown Soil dominated slopes. These were mapped with medium to large delineations of common mapping units (T, TB, ROB, RO). The vegetation on these higher slopes, like the soils, was simpler. It appeared to be a fairly typical mixture of treeline shrub, meadow and conifer communities.

The northern side valley-Mono Pass area had some distinctive soil features. The soils were mostly mapped with a few large delineations. This map accurately reflected a broad, large scale and relatively simple soil landscape. What is not as apparent and as accurately reflected on the map is the complex colluvial stratification contained in the large delineations of the Alpine Brown and Scree (BrS) unit. Contrasting subsurface stratification patterns were observed in these areas, but were impractical to map because they had little or no clear surface expression. The medium sized delineation of Boulders and Alpine Brown Soils (BLB) in this area is colluvial. Where vegetated, this northern side valley area had mostly dry alpine meadows or dry treeline vegetation.

The middle sub-basin of the Ruby Lake watershed is a long narrow valley area between 3,475 and 3,600m generally around Mills Lake and a string of small ponds. It is a mixture of broad relatively simple rocky landforms, one complex soil area near Mills Lake and several smaller somewhat complex soil areas. The observed soils were not well developed or differentiated in most of the area, only Alpine Brown Soils and Riparian Soils were well represented. One small area with Moist Meadow Soil was mapped at the northern end of this area, near the major topographic break, down to the lower Ruby Lake basin; no other Meadow Soils were observed above this small area. Most of the middle basin was mapped with medium to large delineations dominated by glacial boulder deposits, talus or bedrock with some Alpine Brown Soils (BLB, T, TB, ROB, RO) included. One relatively large, complex area was mapped southwest of Mills Lake. The wetter portions of this small area were mapped as Alpine Brown Soil-Meadow Complexes (BC) or Riparian Soil units (RR, RRT). The better drained areas were mostly mapped as mixtures of Alpine Brown Soils, Rock and Talus (Br, BrR, ROB, TB).

Several variations of Alpine Brown Soils were observed in this same small area southwest of Mills Lake. Some soils just west of Mills Lake were unexpectedly deep and well developed. Other soils, especially higher elevation ones (about 3,600 m) to the southwest were examples of the high-elevation poorly-developed and poorly-differentiated soils. The most distinctive geologic and soil feature of this area near Mills Lake was a band of mineralized reddish rock, reddish scree and reddish brown well drained soils (now lumped with Alpine Brown Soils). These iron oxide rich soils were initially field mapped as a special ("Iron-Brown") genetic soil but due to minor geographic extent and unknown chemical properties these were later lumped with Alpine Brown Soils. They have been sampled in several places. When chemical data become available their status can be reevaluated (Table 5). A number of smaller soil dominated and sometimes wetter areas were concentrated around ponds and drainageways north of Mills Lake. The vegetation of the middle basin could be described as a middle elevation treeline transition zone. Scattered small conifers west of Mills Lake, and a variety of shrub and herbaceous meadow types were observed.

Above the Mills Lake or middle sub-basin area is a high-elevation upper valley, which is essentially a fourth sub-basin area within the Ruby Lake watershed. This upper basin ranges in elevation from about 3,600 to 3,810m and has a distinctly alpine character. The soils were mapped mostly in large delineations of Rock covered Glacier (RG), Talus (T), Rock Outcrop (RO), and Rock Outcrop and Alpine Brown

Soils (ROB), and one small Alpine Brown Soil and Rock (BrR) delineation. This simple soil map accurately reflected a broad, simple and largely unvegetated landscape pattern in the upper basin. The only soil complexity observed was in those two areas mapped with soils (ROB, BrR). These contain fine scale variations among subtle variations of high-elevation soils and non-soil materials, mostly bedrock (Table 5). Soils appeared shallow in the ROB delineation, and on average slightly deeper in the BrR delineation. Well developed and well differentiated Alpine Brown Soils and Riparian Soils may be present, but were not observed. This area was only briefly visited and sampled in the general area surrounding the BrR delineation. The observed vegetation, where present, was distinctly alpine with little species diversity but some subtle variations indicative of moisture gradients, and differences in snow-free growing season lengths.

The high mountain ridge and western watershed divide running from near Mono Pass in the north to Mt. Abbot in the south was the only large part of the Ruby Lake basin never visited on foot. The essentially non-soil areas are more reliable mapped than the Alpine Brown and Scree (BrS), and Talus and Alpine Brown Soils (TB) delineations. In spite of these problems, it is unlikely that detailed on-site investigations would greatly alter the broad-scale soil map patterns. Such investigations would primarily clarify the composition of soils within some problematic delineations. The area appeared mostly unvegetated, but probably contains small areas of high-elevation dry alpine vegetation and perhaps riparian areas.

The eastern watershed divide was visited and investigated along most of its length. This divide is a distinct ridge composed of three unnamed peaks and three saddles. The soils were mapped mostly as large delineations of Rock Outcrop, Boulder or Talus dominated areas (RO, ROB, BLB, T, TB). The soil map accurately reflected a simple, broad-scale pattern and most of the delineations contained soil landscape-patterns typical for their mapping units. It contained scattered conifers in areas, but also had unvegetated rocklands and small areas of dry meadow vegetation. The most distinctive area was the band of reddish mineralized rock and reddish brown soils east of Mills Lake. These were the same iron oxide rich or "iron-brown" variations of Alpine Brown Soils discussed above for the middle basin. They were lumped with other Alpine Brown Soils and not explicitly expressed on the soil map.

Unusual Soils or Soil, Non-Soil and Water Patterns -- Many of the numerous small areas of unusual soils and soils-related patterns were discussed above for the different parts of this watershed. The most distinctive of these are outlined below. High-elevation stratified colluvial combinations of Alpine Brown Soils, Scree, Talus and Boulders (BrS, TS, BrR, BLB) were one unusual and poorly understood group of materials. Another middle to high-elevation delineation of glacial Boulders and Alpine Brown Soils (BLB) in the middle basin appeared unusual and was poorly understood. Soils around the spring complex just above and north of Ruby Lake were a unique mixture of materials. The reddish iron oxide stained rocks, scree, and soils around Mills Lake were also unique in this study. All of these and other unusual or unique areas were sampled to determine how unusual morphology and landscape settings may be related to biogeochemical properties.

A.3.4.5. Spuller Lake

Overview -- The Spuller Lake watershed is similar in several important respects to most of the other watersheds in this study, but it also contains more geologic variation than most. The high-elevation granitic upper slopes are most similar to the other basins, whereas the middle and lower Spuller basin was formed in metamorphosed volcanic rock with mixed soil parent-materials. Like many of the other basins Spuller appeared to contain minor deposits of recent volcanic ash (tephra). Spuller is a medium sized, steep, well defined, high-elevation watershed most similar in these respects to the Pear Lake watershed, and generally similar to both the Crystal Lake and Topaz Lake watersheds. Its soil patterns, especially those in the highest elevation areas, were most similar to the larger Ruby Lake watershed. The large area of complex bedrock in the middle and lower basin was not recognized in special soils or mapping units; in fact, no unique mapping units were used in this basin. Ten of the fifteen mapping units used in the Spuller Lake watershed were common units also used in three or more other watersheds (Table 3). Most of the less common mapping units used were high-elevation units involving boulders or scree, and all fifteen mapping units used in the Spuller Lake watershed were also used in the Ruby Lake watershed (Table 3).

Topography and Physical Geography -- The Spuller Lake watershed is intermediate in size and relatively high and steep with a wide elevation range when compared to the other basins (Table 4). It is about 97 ha in area, with an elevation range from 3,131m to 3,668m and contains a number of distinct sub-basins and slopes. The Spuller Lake watershed is a distinctly stepped alpine basin that can be divided into a distinct lower basin, a distinct middle basin and broad high-elevation slope areas. It lacks definite south or southeast watershed boundaries. The upper slopes face mostly eastward, the middle basin is a deep southwest to northeast trending trough, and the lower basin is an east draining and east to northeast-facing bowl or half bowl. These three areas are described in more detail below. The western watershed divide is high as is the indistinct southwestern divide. The east and north divides are lower and were not effective divides for glacial ice--the glacial flow patterns (generally eastward from the upper slopes, and north in the middle and lower basins) were quite different from the current sinuous main drainage pathways. Most slopes face east to northeast. The steepness of the watershed contributes to the prevalence of colluvial soils and non-soil colluvial deposits. Many slopes are relatively cold for their elevations due to high-elevation, and northeast aspect upper slopes which shadow and provide cold air drainage to many of the lower slopes. These factors contributed to the distinctively alpine character of soils, vegetation and landforms on the upper and middle slopes of this watershed.

Bedrock and Surficial Geology -- The following description of bedrock geology was adapted from a 1:62,500 scale map (Bateman, et al. 1983), and field observations by Aaron Brown in 1990. The bedrock of this basin has been mapped in two medium sized delineations: (1) the middle and lower basins were mapped as metamorphosed Tuffaceous Lake Beds and (2) the upper slopes were mapped as Granodiorite of the

Kuna Crest. Bateman et al. described the Tuffaceous Lake Beds as thinly bedded, fine grained and mostly of volcanic sediments. The most common minerals reported in these metavolcanic rocks were plagioclase, quartz, biotite, hornblende, and opaque minerals. Aaron Brown described the properties of this rock-unit within the Spuller watershed as varied with many vertically oriented layers running north-south. Some layers looked like schist, some like pure quartz, others like quartz with biotite inclusions and others were tuffaceous. Brown also found some pyrite near the old mines west of Spuller Lake--evidence of mineralized zones. Brown didn't describe the Granodiorite of the Kuna Crest, but Bateman et al. described it as dark colored, medium grained hornblende-biotite granodiorite. This granodiorite unit was similar to rock-units mapped in several other watersheds, while the metavolcanic rock-unit was unique in this study.

Surficial deposits were not shown on the geologic map, but Brown mentioned some important surficial features. These included the probable split paths of glacial ice, the presence of tephra, and the proximity of Mono-Inyo volcanic chain. Past glacial ice pathways did not simply follow current drainage patterns from the upper slopes to the Spuller Lake outlet; the current drainage pathway from the highest slopes to the lake outflow is an S-shaped curve. Much of the ice from the highest slopes appeared to have continued east, straight toward the Fantail Lake area across the current eastern ridge of the Spuller Lake watershed. Some of the ice turned north, down into the middle and lower Spuller basins, but there was evidence that ice in the lower basin continued north toward the Maul Lake area and did not turn east down the current Spuller Lake outlet creek valley. These glacial ice flow patterns affected glacial drift depth and composition and therefore soil patterns in the watershed.

The complicated bedrock and surficial geology of this basin contributed to complex mixtures of soil parent-materials. Only the highest elevation soils appeared to be truly granitic. The middle and lower basin, however, contained substantial granitic components in their glacial till and alluvial deposits overlying metavolcanic rocks. Tephra has probably been deposited in significant quantities over most of the basin, but its subsequent erosion, redistribution and mixing with other soil materials is not understood at this time.

Landscape, Soil and Vegetation Patterns -- The intermediate scale patterns of topography, landforms, drainage, soils and vegetation in this basin can be generally described in terms of a lower basin, a middle basin and high-elevation slope areas. The lower basin is essentially a bowl of cliffs, talus and boulder slopes, riparian areas, meadows and open subalpine woodlands surrounding Spuller Lake itself. The soil map for this area is complex, but it captured most of the soil variation with relatively distinct and pure delineations. Although the influences of metavolcanic rocks on soils was not explicitly recognized in the legend, otherwise the mapping units appeared to describe the actual soil patterns of the lower Spuller Lake basin well. Some of the isolated smaller RRT and ROB delineations in the cliffs west of the lake were not field verified.

The middle Spuller basin is a deep, U-shaped glacial trough, or elongated bowl trending from southwest to north in a slight curve. A small pond, small meadows, riparian areas and mostly large boulder fields occupy its gently sloped center. Talus and boulder covered slopes as well as riparian areas, and combination bedrock and soil slopes surround the central basin. Very few conifers are present in either of these central areas; the vegetation, where present, is mostly riparian, low shrub or a variety of meadow types. The middle basin drains north and is partially surrounded and shadowed by large cliffs and steep boulder fields to the west and southwest. These cliffs and boulder fields are cold, shadowed slopes with an alpine character. The northeastern to southern edges of the middle basin are a more complicated mixture of boulders, bedrock and soil-covered slopes. These upper slopes and eastern edges of the basin contain small stands of conifers, riparian areas and a variety of meadow types. Although they are higher than the central middle basin, they appeared warmer with more of a mixed subalpine woodland and meadow character similar to the lower basin.

Like in the lower Spuller Lake basin, the overall patterns of soil and non-soil materials in the middle basin were well expressed on the soil map. In the middle basin the soil map was composed of a few large and many small delineations. Most of the observed soil and landform patterns were at mappable scales, and most of the delineations were relatively distinct and appeared pure in composition. These areas were also well described by the standard descriptions of genetic soils and mapping units.

The middle and lower Spuller basins are both bounded on the southwest and west by a large steep complex of cliffs. This cliff complex trends across the entire basin from south-southeast to north-northwest between elevations of about 3,170 and 3,292m. The upper parts of this cliff complex and all areas above it to the south and west (or all areas above about 3,230m comprise what is being described here as the Spuller upper slope areas.

These upper slopes contain a variety of materials and landforms, but all areas share a distinct high-elevation and strongly alpine character. Vascular vegetation consisted of only a few species of low-stature alpine plants, and the soils were minimally developed but extensive and deep in places. The areal extent of soils and soil-like scree on the highest elevation slopes (above 3, 410m) was unexpected. Alpine Brown Soils and Riparian Soils were both mapped on these upper slopes.

Unusual Soils or Soil, Non-Soil and Water Patterns -- Metavolcanic bedrock in the middle and lower basins may distinguish some soils from granitic soils. These soil have been sampled, and chemical and mineralogical data should help clarify these potential differences. It is important to note, however, that soil morphology associated with metavolcanic rock was similar to that in corresponding landscape positions of watersheds with only granitic bedrock.

The lower basin contained soils under groups of conifers that appeared leached in their upper parts. This was the leached phase of "podzol-like" Alpine Brown Soil that showed evolution toward Spodosols or Podzols (Table 5). Although unusual and rarely reported, similar leached soils were found in four other watersheds in this study. Unusual soils were not observed in the middle basin, but few pits were excavated there.

The upper slopes of the Spuller watershed had problematic high-elevation variants of the Alpine Brown Soils and Riparian Soils as discussed elsewhere. The highest slope had a large area of an unusual high-elevation scree-mantled Alpine Brown Soil (BrS) which was also found in the Ruby Lake watershed. Other unusual mapping units in the middle and upper Spuller basin were dominated by Boulders. These were the Boulders (BL) and the Boulders and Alpine Brown Soils (BLB) units. These boulder dominated areas appeared to be primarily an unusual type of glacial drift (not colluvial). Some parts of these bouldery areas had mixed drift and colluvial character. The bouldery areas in the middle and upper Spuller watershed looked similar to the large bouldery areas in the middle Ruby Lake watershed.

A.3.4.6. Crystal Lake

Overview -- The Crystal Lake watershed had, by far, the most distinctive soil patterns in this study. Calling this watershed the "pumice bowl" would sum up many of its unique properties; it contains large areas of unstable pumice, and mixtures of pumice, talus, rock and soils with varying degrees of stability. All soils examined in this watershed were predominantly composed of recently deposited strata of pumice or combinations of pumice, volcanic ash and rock fragments. As such, the Crystal Lake area could have been mapped with its own unique legend. However, one of the objectives of this study was to use a common legend in all of the watersheds, so this option was eliminated. As a compromise the Crystal Lake watershed was mapped with a combination of mostly unique genetic soils and unique mapping units with the more common soils, non-soil materials and mapping units used whenever possible. The unique volcanic soils and volcanic mapping units were set up to be as analogous or parallel as possible to their non-volcanic counterparts (Figure 1, Tables 1 and 2). The terminology was also simplified and standardized as much as possible. The word volcanic was simply added as an adjective in most cases to modify genetic soil names. In one case 'volcanic' was added to a non-soil material name (scree) instead of using the more technical term (pumice). Three of the four genetic soils, one of the three named non-soil materials, and eight of the eleven mapping units used in the Crystal Lake watershed were unique to this areas (Tables 1, 2 and 3).

Topography and Physical Geography -- The Crystal Lake watershed is a topographically distinct basin, intermediate in size and overall relief, with slightly lower elevations and a slightly lower elevation range than most of the other watersheds (Table 4). In size (135 ha) is similar to the Pear, Topaz, and Spuller Lake watersheds and has an elevation range, from 2,951 to 3,244m. Although significantly higher than the Lost Lake area, it is on average lower in elevation than all the other watersheds. This basin has an unusual combination of contrasting slopes with many distinct medium to large vertical cliffs but also with many areas of smooth, gentle to moderate slopes covered with pumice and pumice-rich soils. Like several of the other watersheds, this one contains distinct lower and upper sub-basins, with a variety of other distinctive slopes. The Crystal Lake watershed contains two large cliff systems, one associated with the granitic Crystal Crag, and another set of mixed granitic-andesitic cliffs ringing the lower basin just below the Mammoth Crest. The Mammoth Crest itself contains large, gentle, relatively high-elevation slopes draining through the cliffs into the lower Crystal Lake watershed. The watershed has a relatively simple overall shape, both facing and draining northeast. Near-vertical cliffs contribute talus, but most of the unstable, colluvial character of soils and non-soil materials in this watershed are due to the physical properties of pumice. Soils, vegetation and landforms appeared mostly subalpine in character due to the relatively low elevation of the catchment. The higher parts of this watershed appeared to have more of an intermediate treeline character but no, truly alpine appearing areas were observed. The potentially cold and shadowed north and east facing slopes in this watershed were not greatly different than other slopes with south and west aspects.

Bedrock and Surficial Geology -- Information on bedrock geology that follows was adapted from an old 1:62,500 scale map of the Devils Postpile Quadrangle (Huber and Rinehart 1965) and from Aaron Brown's 1990 observations. The whole Crystal Lake watershed was mapped within a large delineation of "Rocks similar to the Cathedral Peak Granite", but the small inclusions of the Andesite of Deadman Pass were also shown near the Mammoth Crest on the western watershed boundary. The map gives the false impression of a simple, granite-dominated geology. Problems include broadly defined rock units and no expression of surficial geology. The "Rocks similar to the Cathedral Peak Granite" are described as ranging in composition from granodiorite to alaskite, averaging as mafic quartz monzonite. This broad description includes true granites, and perhaps two or more other types of felsic pluton rocks as defined in the modern international rock classification system, which was first published after this particular geologic map. The Andesite of Deadman Pass was described on the published map as primarily a series of interbedded, and commonly vesicular andesitic flows, cinders and rubble and also some scattered remnants of other andesitic rock. Based on field observations, Brown described the volcanic rock as patches of a reddish andesite cap rock, but appropriately spent more time describing the recently deposited tephra.

Brown described the tephra as deep (up to 1 meter), distributed throughout the basin, mostly gray, ranging in size up to 2 cm in diameter, and probably from an eruption or eruptions within tens of kilometers. The geologic map did not show tephra in this basin but had a large area of Quaternary Rhyolitic Pumice mapped 7 km to the north. This pumice unit was: "Mapped only where it constitutes a blanket up to several tens of feet in thickness in the rather broad and flat area in the northeastern part of the quadrangle" (Huber and Rinehart 1965). This same type of pumice was then described as being locally a major constituent elsewhere outside the one large delineation mapped. The rhyolite pumice was related to both the recent, southern Mono Crater Rhyolitic Domes and also to the older Quartz Latite of Mammoth Mountain. The tephra materials observed during the soil survey where mostly light gray, rhyolitic ash and pumice about 1 to 8 mm in diameter but they ranged in size and composition from glassy fine sand sized ash to medium gravel sized mostly subrounded pumice fragments. The maximum depth of this pumice deposit was estimated to be at least 3 to 5 m in some areas.

Volcanic bedrock and tephra are more susceptible to chemical weathering than granitic rocks. This susceptibility is most pronounced in fine grained glassy ash, and less apparent in massive andesite rock. Upon weathering, glassy volcanics release high concentrations of soluble silica which in turn form distinctive secondary weathering products. The unusual physical, chemical and mineralogical properties of volcanic soils are mostly related to the weathering products, not the primary tephra. Volcanic soils are also classified on the unusual properties imparted by weathering products. Although tephra was the dominant parent-material seen throughout the Crystal Lake area, much of it appeared to be recently deposited, coarse, and weakly weathered.

Landscape, Soil and Vegetation Patterns -- The intermediate scale patterns of topography, landforms, drainage, soils and vegetation in the Crystal Lake watershed can be generally described as follows. Just south of the lake and at essentially the same elevation is a large flat wetland composed of recently deposited tephra (mostly stream deposited pumice). This flat delta contains a fine-scale mosaic of contrasting soils and wetland plant communities. Willow thickets, sedge marshes, and moss and shrub bog with associated soils were all sampled in this wetland. Surrounding the lake and its delta are mostly forested slopes, generally moderate in slope gradient and containing many individual small to medium sized rock outcrop areas. Between the rocky areas in the lower basin are smoother slopes with clear evidence of many shallow pumice grain flows over the soil surfaces. The large forested and soil-dominated slopes that flank the lake had high short range (1 to 10 m) soil variability, but similar, almost uniform soil-landscape mosaics over longer distances (50 to 200m). Narrow riparian areas and less stable mixed soil and scree-like chutes cross these lower slopes.

Talus and scree-like areas become more prominent southwest of the delta wetland and beneath the large cliffs that ring the lower basin. These unstable, colluvial-deposits contained variable proportions of soils, but generally lacked large conifers. Adjacent cliffs and rocky ledges also had variable soil proportions, but their soils appeared more stable and therefore more favorable to conifer establishment and growth. The small upper basin and adjacent areas between the upper and lower basin contained complex mixtures of soils and non-soil materials. In addition to all of the components described above, these upper areas had a variety of unusual, small moist-meadows, seepage and riparian areas not well described by any soil mapping unit. Many of these wetter areas were covered with willow thickets. The smooth and gentle slopes of the Mammoth Crest itself appeared to contain mostly deep, somewhat unstable pumice rich soils with sparse vegetation. The isolated stands of conifers on the crest were usually associated with rock outcrops or other slopes with unusually stable soils.

Unusual Soils or Soil, Non-Soil and Water Patterns -- Most of the unusual soils and soil patterns in this basin were recognized with unique genetic soils and unique mapping units. The standard descriptions of these soils and units contain most of the relevant information about their unusual properties. What follows is mostly a description of the unusual properties of the one non-unique soil, the common non-soil materials and mapping units as they were found within the Crystal Lake watershed.

There were two reasons for mapping Riparian Soils in the Crystal Lake area instead of keeping the unique Volcanic Riparian Soil which appeared in earlier working legends. The first reason was simply the small area of soils represented. The second reason was that the physical and biogeochemical properties of these shallow, relatively unweathered but organic matter rich soils were probably not strongly influenced by tephra weathering products.

Loose, unstable colluvial pumice deposits were considered important and distinctive enough landforms and non-soil materials to be worthy of specific recognition in unique mapping units. The term 'volcanic scree' was coined for these pumice deposits to help relate them to the more common types of Sierra Nevada loose granitic scree. On the other hand volcanic bedrock, and volcanic rock talus were not considered distinctive enough to specifically recognize in the soil mapping legend. It was assumed that surficial deposits including tephra should be mapped with soils).

A.3.4.7. Lost Lake

Overview -- Although it is a well defined granitic basin with genetic soil types chemically and mineralogically similar to most of the other watersheds, in many respects the Lost Lake watershed is distinctive. It is by far the smallest, lowest elevation, and most northern lake basin studied. The Lost Lake watershed has more of a subalpine character than any of the other watersheds. It is similar to most of the others in having a soil and non-soil pattern strongly influenced by recent glaciation and geologic structures such as faults and joints. However, only in the Lost Lake watershed has frost shattering of bedrock played the a dominant role in controlling soil and non-soil patterns. The soil properties themselves are also strongly influenced by frost processes, and subsequent redistribution of fines. In particular, the watershed contains extensive areas of rubble, soils formed by fines filling into the open voids of rubble, and mixtures of these with talus and rock outcrop. Two unique mapping units (TRB and TR) were designed to describe areas with high proportions of rubble, and several other unique mapping units were considered but rejected due to the small areas involved. Eight of the eleven mapping units used in this basin were common to at least three other basins (Table 3), but most of these had unique characteristics within the Lost Lake watershed as described below.

The overall relief of this small basin is moderate, but the microtopography is complex. This is reflected in a complex soil map that was, in fact, significantly simplified from the field map sheet for three main reasons: (1) to help legibility, (2) to allow the 1:2000 scale maps to be reduced to the standard 1:4000 scale if desired and (3) because several specific mapping units reflecting variation within the Lost Lake watershed were eliminated from the legend in favor of more general and mixed units. In order to properly portray the soil landscape patterns observed in the field an expanded and more specifically adapted legend and a remapping effort at a scale of about 1:1000 would be needed. Nevertheless, the Lost Lake area was the most thoroughly investigated lake basin in this study. It was traversed more completely than any of the other areas and some field verification and mapping modifications were possible in 1993. This was largely opportunistic, and not specifically by design. Field sampling in 1993 took less time than anticipated allowing field verification and minor adjustments to the soil maps produced in 1992. Small watershed size, large areas of non-soil, easy access and moderate topography all contributed to a relatively efficient and intense investigation of the soils.

Topography and Physical Geography -- When compared to the other five lake basins, the Lost Lake watershed is small (only about 25 ha) and low in elevation with moderate relief (Table 4). The elevation range is low, only about 160m from 2,475 to 2,635m. Nonetheless, this watershed does include a number of medium to small cliffs and associated talus deposits. The low relief contributed to the prevalence of more or less *in situ* rubble and intermediate rubble to talus bodies, in contrast to the purely colluvial talus more common in other study areas. The relatively low elevation of the basin gave it a generally subalpine character. The dominantly north aspect of this watershed affects snowmelt patterns and therefore limits growing season length and soil temperatures relative to warmer-aspect slopes at similar elevations nearby. Parts of the middle basin appeared to have soil and vegetation patterns strongly controlled by late snowmelt.

Bedrock and Surficial Geology -- The following surficial and bedrock geology information was adapted mostly from a 1:62,500 geology map, and related report (Loomis 1983;1981). Loomis (1981) discussed landscape evolution, glacial and other surficial geology in more detail than the geologic references cited for the other watersheds. It reviewed some classic literature on these subjects (Birkeland 1963; McAllister 1936; Blackwelder 1931). The Lost Lake watershed was all mapped within a medium sized delineation of Keiths Dome Quartz Monzonite. A small and narrow northwest to southeast trending metavolcanic tuff-breccia area was also mapped crossing the upper basin. The Keiths Dome unit was described as being an unusually mafic quartz monzonite with 16% dark minerals. The area was described as being somewhat heterogeneous with respect to quartz and mafic mineral contents, and tables in Loomis (1981) gave chemical and mineralogical data for three samples of this rock. Aaron Brown looked at the rocks within the Lost Lake watershed in 1990, and reported that the quartz monzonite looked fairly uniform. Brown didn't see the tuff breccia inclusion or any of the aplite dikes that were mapped. Loomis reported the Keiths Dome unit was intruded by basalt dikes and all adjoining granitic bodies. The metavolcanic tuff-breccias were described in some detail as variable rocks with basaltic or pyroxene andesite composition (Loomis 1981).

Loomis described the Keiths Dome area (presumably including the upper parts of the Lost Lake watershed) as a remnant of the lowest of three Tertiary erosional surfaces now being modified by glacial and other forms of erosion. He reviewed and interpreted earlier glacial geology (McAllister 1936) and described large late Pleistocene glaciers moving east from the Desolation Valley down into the Echo Lake area with "fingers" of ice diverted northward through the col between Echo Peak and Keiths Dome. He described only the highest peaks and some high ridge-crests as remaining above glaciers (nunataks), but most of the area above the elevation of Lake Tahoe as being buried by ice. The entire Lost Lake watershed was undoubtedly scoured during the late Pleistocene by glaciers flowing north to northeast over the shoulder of Keiths Dome. However, frost shattering of bedrock also suggests long periods of subaerial exposure to a harsh, periglacial weathering environment. This degree of frost shattering was not apparent in similar rock around upper Echo Lake. The Lost Lake watershed has apparently been deglaciated longer than the nearby

Desolation Valley to Echo Lake area. This is consistent with Loomis' description of a main ice lobe moving east down the Echo Lake valley with fingers of ice peeling off to the north only when the main lobe was at its thickest.

Frost shattered of rock is by far the most distinctive geologic feature of this basin with respect to soil formation. Although long exposure to periglacial climates in the late Pleistocene seems to be the best explanation of this phenomena, climatic rock properties, and other geomorphic processes also contributed. The unusual example of quartz monzonite in this area contains more mafic minerals than most granitics and may be relatively susceptible to chemical weathering.

Landscape, Soil and Vegetation Patterns -- The intermediate scale patterns of topography, landforms, drainage, soils and vegetation in this watershed can be generally described in terms of three areas within the watershed. The Lost Lake watershed contains a lower basin around the lake, a complex middle basin area, and an upper area that is largely a rounded wind swept, rockland ridge. The lower basin and much of the middle basin area have a distinctly subalpine character. Other parts of the middle basin are unvegetated rockland difficult to associate with any mountain zone; similar looking rocklands can be found anywhere from the lower montane to alpine zones. The middle watershed also contains a rocky southern basin and riparian complex where deep snow accumulation and late melting appeared to control soil and plant community development. In parts of the middle basin and most of the upper watershed, treeline along ridges appeared to be controlled by wind exposure. Conifers are present, at least as scattered individuals, throughout the watershed.

The lower basin around the lake is a mixture of small marshes and meadows, well defined riparian zones, coniferous subalpine forest, open rocky woodlands and rocklands. The overall pattern is controlled by glacial topography, rock structures, and drainage from the middle basin. Contrasting soils and non-soil materials are mixed at fine spatial scales. Well drained soils are variable in depth and rock fragment content, but on average appear deep and high in fragments. Riparian soils are shallower, sometimes to bedrock and sometimes to fragmented material (talus or rubble). Meadow soils appear relatively deep in most places with variable rock fragment contents. Unusual soil properties and patterns are described below.

The lower basin is surrounded by relatively steep areas of small to medium sized cliffs, mixed rock outcrop areas, associated talus, talus-rubble mixtures and riparian zones. The riparian zones crossing these steeper areas and in complex bench areas just above the steep slopes were in themselves complex. Most contained closely mixed combinations of contrasting soils in banded linear patterns. The mapping scale and common legend used in this study did not allow for a good expression of soil properties and soil patterns in these areas.

The middle basin part of this watershed contains some relatively simple areas shown on the soil map as large delineations of Talus and Rubble (TR) or Rock Outcrop (RO), and some areas of intermediate complexity shown as medium to large delineations of Talus, Rubble and Alpine Brown Soils (TRB) or Rock Outcrop and Alpine Brown Soils (ROB). The middle basin also contains a relatively large and complex riparian and rockland area south and southeast of the central rock outcrop area. There is also a smaller complex area of riparian and meadow soils mixed with rocklands in the northwestern part of the central basin. The unique fine-scale patterns of soil and non-soil variation were only marginally described by the common legend, and only marginally resolved on the soil map. The simple to intermediate complexity areas of the middle basin had large unvegetated rockland areas, with patches and pockets of mostly dry meadow vegetation and scattered conifers. In the more complex and wetter areas small linear groups of conifers with ericaeous shrub understories and fringing areas were usually found on the better drained Alpine Brown Soils. Mixtures of shrubs and herbaceous meadow vegetation, often sedge dominated, dominated intermediate drainage and wetter soils in these complexes.

The upper parts of the Lost Lake watershed appeared as essentially wind swept, rounded rockland ridges, containing scattered conifers and some areas with significant inclusions of deep well drained soils. Close inspection of the area revealed that the prominent conifers were often in rocky areas, whereas the areas with high proportions of soil cover often had only sparse dry meadow or shrub cover. The relatively large and healthy conifers growing out of rubble and cracked bedrock indicated deep pockets of soil hidden under some apparent rockland surfaces. It was considered likely that such deep pockets of fines were more extensive than the few conifers able to reach and exploit them.

Unusual Soils or Soil, Non-Soil and Water Patterns -- Unlike in most other watersheds, little of the Alpine Brown Soil volume of the Lost Lake watershed appeared to have formed directly in glacial till or loamy colluvial deposits. Instead, most of these soils appeared to have formed as fine grained materials filled into bedrock cracks, rubble, or talus. Wind and water deposition appeared to have contributed to this "in-filling" process. The physical properties of these soils were quite distinctive. Alpine Brown Soils in the Lost Lake basin had higher proportions of coarse fragments, especially large angular fragments, than was typical for this same genetic soil in the other watersheds. Bedrock depths and soil/bedrock contact geometry were also irregular at Lost Lake.

In spite of these obvious physical differences, field morphology did not indicate any unique or unusual mineralogy or biogeochemical properties of the Alpine Brown Soils in this watershed. In these respects they appeared similar to the lower elevation soils of most other watersheds. The term 'Alpine' is not an ideal descriptor for these subalpine forest soils, and it may be more appropriate to subdivide them along a elevational and/or developmental gradient (Table 5). The typically subalpine and alpine variants were kept together because morphological differences in the mineral soils were subtle and inconsistent, and forest litter layers (O horizons) were variable over short distances.

Most of the wetter genetic soils in the Lost Lake watershed formed in coarse alluvium, delta and lake deposits. Some also formed in rubble or talus and perhaps a few formed in loamy colluvium. Some of the lake and delta deposits were fine to medium textured, but most of the wetter soils (Riparian Soils and Moist Meadow Soils) became increasingly coarse and rocky with distance from the lake. Several processes of fines filling into rocky materials, including sediment capture in turfy vegetation, appeared to play important roles in the formation of these soils. Beyond physical differences, the wetter soils, like their better drained counterparts, appeared to be mineralogically and biogeochemically similar in this watershed to those in most of other study areas.

The Lost Lake watershed was most distinctive in its fine scale mixtures of contrasting soils and non-soil materials. These included relatively simple mixtures of well drained Alpine Brown Soils with different types of rockland (bedrock, rubble or talus) as well as far more complex mixtures including contrasting soils.

Many of the wetter areas on the bench level just above the lake and in the middle basin contained fine scale mixtures of narrow bands of contrasting soils varying in drainage condition. These were typically linear areas of moderate slope developed in faults or joints with combinations of Alpine Brown Soils, Moist Meadow Soils, Riparian Soils, and some unnamed variations of these types. In some cases the contrasting soil pattern was mappable at 1:2000, but in most cases the components were too closely mixed. A number of small delineations had distinct bands with characteristics of the following mapping units: Rocky Riparian Soils (RR), Alpine Brown-Meadow Soil Complex (BC), and Alpine Brown Soils and Rock (BrR). These soil bands were on the order of two to five meters wide and 10 to 30 meters long, arranged with a central riparian area, surrounded in most areas with long narrow meadows. The well drained soils, when present, were usually in higher forested areas around the edges and transitional to adjacent rockland areas. Most of these areas were mapped as either Rocky Riparian Soil Complexes (RR), or Alpine Brown Soil-Meadow Complexes (BC) based on estimates of composition, but neither properly described the patterns as explained below.

In the two complex areas of the middle basin most of the Rocky Riparian Soil Complex delineations (RR) contained many small areas of shallow or minimally developed Alpine Brown Soils and some Moist Meadow Soils. Conversely, Alpine Brown Soil-Meadow Complex (BC) delineations contained more Riparian Soils, and more shallow poorly developed Alpine Brown Soils and often more rock than typical for the mapping unit. Consequently some of the areas mapped as Rocky Riparian Soil Complex (RR) and those mapped as Alpine Brown Soil-Meadow Complex (BC) in the middle Lost Lake basin were fairly similar mixtures of soils and rock. These two mapping units were far more distinct in soil composition in the other lake basins.

One relatively large riparian complex west of the lake appeared to be essentially a steep moist meadow and riparian system developed over steep bedrock, talus, rubble and coarse alluvium. It was a mixture of Riparian Soils in Talus, Talus, and a mixed rocky meadow soil complex. This rocky meadow area was mapped in the Alpine Brown-Meadow Complex (BC) unit but was really a unique area not well described by any unit in the current legend. It appeared more like a meadow equivalent of the RRT unit.

A.4. Discussion

The main strengths of the soils maps are: (1) a simple, systematic mapping legend was used, (2) scientifically conservative treatment of difficult to determine soil properties (such as depth to bedrock) was employed, and (3) detailed, and consistent maps of major soil and non-soil areal distributions were produced. These attributes make the current product suitable for calculations of soil-landscape properties on an area basis, and for a number of biogeochemical analyses. However, they are poorly suited for volumetric calculations of soils and surficial deposits without further field measurements.

Future research needs include: (1) field checking of soil maps in areas either never visited (e.g., the western ridge of the Ruby Lake watershed), or quickly traversed (much of the Topaz Lake watershed), (2) quantitative field studies of soil depth, texture and rock fragment contents, and (3) some systematic transect studies of major slopes to quantitatively assess actual mapping unit composition. All of the above studies can be done as simple field studies with a minimum of equipment and without the need to make major changes in the existing soil maps or soil legend.

Although these field studies are theoretically simple obtaining adequate sampling density will be time consuming. Soil depths could be directly measured or reasonably estimated without special equipment at some locations in most landscape positions, however, accurate, comprehensive and quantitative depth to bedrock studies would require sophisticated geophysical techniques, specially trained technicians, and field verification of geophysical data.

I suggest that a long-term goal of developing purely quantitative volumetric soil-landscape models be adopted. A stratigraphic approach appears practical at small watershed scales. A variety of maps could be generated from such models, but the models themselves would not rely on any defined soil types or mapping units. They would rely on continuous three dimensional predictions of soil properties organized around measurable soil layers. These models could logically be divided into physical and biogeochemical modules. Quantitative volumetric models are conceptually attractive, but would be intractable. They should be initially developed and tested on small areas (tens of hectares) before scaling up to small watersheds such as those in this study.

Another potential area for future research involves geographically extensive studies of soil-landscapes throughout the Sierra Nevada. These studies could help extrapolate research results from these small watersheds over larger areas for calculations in regional to global-scale models. The mapping legend used in this report, for example, could be used in belt transects to efficiently evaluate large areas. A study of this kind could rapidly assess the general applicability of important results from this study such as the high proportions of soils found in some high-elevation settings.

A.5. Summary

This report and the accompanying maps in Chapter One are an integrated soil survey of seven watersheds in the Sierra Nevada. The approach taken was original and designed for use by researchers from a variety of disciplines with particular emphasis on hydrology and biogeochemistry. This soil survey is essentially a semi-quantitative model of soil-landscape patterns on an areal basis, designed to be periodically improved with additions of quantitative data. Eventually it may be transformed into quantitative volumetric models of soil physical and biogeochemical properties at small watershed scales.

This soil survey used few soils and mapping units to emphasize major relationships in the soil-landscape continuum. The soil maps contain great cartographic detail in places. This detail was used in complex soil-landscapes to express some of the observed soil variability that otherwise would have been lost with less intensive mapping of a small legend. Maps of these complex areas can be simplified for specific users. The system was specifically designed to facilitate map generalizations and other cartographic transformations.

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Table 1. Named Genetic Soils, their major variants and horizon sequences, and probable major taxa

Genetic Soil	Major Variants	Horizonation	Major Taxa (Subgroups)
Alpine Brown Soils	Deep, Stable	A-Bw-C; A-Bw-C-R O-A-Bw-C; O-EA-Bs-C	Typic Cryumbrepts Entic Cryumbrepts Dystric Cryochrepts
	Deep , Stratified	A-Bw-2A; A-AB-2A	Entic Cryumbrepts
	Shallow or Variable	A-R, A-Bw-R	Lithic Cryumbrepts Lithic Cyrorthents
Moist Meadow Soils	Deep, Stratified	A-Bw-Cg; A-C-2A-2Cg	Aquic Cryumbrepts Oxyaquic Cryumbrepts Aquic Cryofluvents
	Shallow or Variable	A-R; A-Bw-R; A-Cg-R	Lithic Cryumbrepts Lithic Cryorthents
Wet Meadow Soils	Deep, Stratified	A-Cg-2A-2Cg; O-A-Cg	Humic Cryaquepts. Typic Cryaquents
	Shallow or Variable	A-Cg-R; A-R; O-A-R	Histic Cryaquept Lithic Cryaquept Lithic Cryaquent
Riparian Soils	Bedrock	A-R; A-C-R	Lithic Cryumbrepts Ruptic-Lithic Cryumbrept
	Talus	A-C	Oxyaquic Cryumbrepts Oxyaquic Cryorthent
	Volcanic	A-R; A-C-R	Vitrandic Cryumbrepts Vitrandic Cryorthent
Volcanic Brown Soils	Deep, Stable	O-A-Bw-C; A-Bw-C	Andisols or intergrades to Andisols, Laboratory data needed even for preliminary assessment Vitrandic intergrade subgroups appear most likely
	Deep , Stratified	O-A-C-2A-2Bw-2C	
	Shallow or Variable	O-A-Bw-R; A-Bw-R	
Volcanic Moist Meadow Soils	Deep , Stratified	A-Bw-Cg; A-C-A-Cg	
	Shallow or Variable	A-C-R; A-R	
Volcanic Wet Meadow Soils	Deep , Stratified	O-A-Cg; A-Cg-2A-2Cg	

Table 2. Identification legend, alphabetically by map symbol, and in groups.

Alphabetical by Symbol

Map symbol	Mapping Unit Name
BC	Alpine Brown Soil-Meadow Complex
BL	Boulders
BLB	Boulders and Alpine Brown Soils
Br	Alpine Brown Soils
BrR	Alpine Brown Soils and Rock
BrS	Alpine Brown Soils and Scree
BT	Alpine Brown Soils and Talus
MM	Moist Meadow Soils
MMC	Moist Meadow Soil Complex
RG	Rock covered Glaciers
RO	Rock Outcrop
ROB	Rock Outcrop and Alpine Brown Soils
RR	Rocky Riparian Soil Complex
RRT	Rocky Riparian Soil and Talus Complex
RVB	Rock Outcrop and Volcanic Brown Soils
T	Talus
TB	Talus and Alpine Brown Soils
TR	Talus and Rubble
TRB	Talus, Rubble and Alpine Brown Soils
TS	Talus and Scree
VB	Volcanic Brown Soils
VBR	Volcanic Brown Soils and Rock
VBS	Volcanic Brown Soil and Scree
VBT	Volcanic Brown Soil and Talus
VMM	Volcanic Moist Meadow Soils
VST	Volcanic Scree and Talus
VWM	Volcanic Wet Meadow Soils
WM	Wet Meadow Soils
WMC	Wet Meadow Soil Complex

Alphabetical by Name in Group

Map symbol	**Group** Mapping Unit Name
	-----------**Predominantly Soil Units**----------------
Br	Alpine Brown Soils
BC	Alpine Brown Soil-Meadow Complex
MM	Moist Meadow Soils
VB	Volcanic Brown Soils
VMM	Volcanic Moist Meadow Soils
VWM	Volcanic Wet Meadow Soils
WM	Wet Meadow Soils
	---------**Soil Dominated Mixed Units**--------------
BrR	Alpine Brown Soils and Rock
BrS	Alpine Brown Soils and Scree
BT	Alpine Brown Soils and Talus
MMC	Moist Meadow Soil Complex
RR	Rocky Riparian Soil Complex
RRT	Rocky Riparian Soil and Talus Complex
VBR	Volcanic Brown Soils and Rock
VBS	Volcanic Brown Soil and Scree
VBT	Volcanic Brown Soil and Talus
WMC	Wet Meadow Soil Complex
	------**Non-Soil Dominated Mixed Units**-----------
BLB	Boulders and Alpine Brown Soils
ROB	Rock Outcrop and Alpine Brown Soils
RVB	Rock Outcrop and Volcanic Brown Soils
TB	Talus and Alpine Brown Soils
TRB	Talus, Rubble and Alpine Brown Soils
	---------**Predominantly Non-Soil Units**------------
BL	Boulders
RG	Rock covered Glaciers
RO	Rock Outcrop
T	Talus
TR	Talus and Rubble
TS	Talus and Scree
VST	Volcanic Scree and Talus

Table 3. Tabulation of the numbers of delineations in each watershed, and totals

Map symbol	Group / Mapping Unit Name	—SW Sierra—			–E Central Sierra–			-N-	
		Pear	Topaz	Mini	Ruby	Spull.	Cryst.	Lost	Totals
	—Predominantly Soil Units—								
Br	Alpine Brown Soils		5		3	2			10
BC	Alpine Brown Soil-Meadow Complex	9	18		7	9		14	57
MM	Moist Meadow Soils	1	1		1	2			5
VB	Volcanic Brown Soils						1		1
VMM	Volcanic Moist Meadow Soils						5		5
VWM	Volcanic Wet Meadow Soils						1		1
WM	Wet Meadow Soils		2					1	3
	—Soil Dominated Mixed Units—								
BrR	Alpine Brown Soils and Rock	6	13	3	22	10		9	63
BrS	Alpine Brown Soils and Scree				5	1			6
BT	Alpine Brown Soils and Talus	6			2	3			11
MMC	Moist Meadow Soil Complex	6	5		4	5		4	24
RR	Rocky Riparian Soil Complex	10	6		14	2	3	6	41
RRT	Rocky Riparian Soil and Talus Comple	6			4	27		5	42
VBR	Volcanic Brown Soils and Rock						14		14
VBS	Volcanic Brown Soil and Scree						9		9
VBT	Volcanic Brown Soil and Talus						5		5
WMC	Wet Meadow Soil Complex	1	4						5
	—Non-Soil Dominated Mixed Units—								
BLB	Boulders and Alpine Brown Soils				2	5			7
ROB	Rock Outcrop and Alpine Brown Soils	23	46	3	35	22		8	137
RVB	Rock Outcrop and Volcanic Brown Soils						26		26
TB	Talus and Alpine Brown Soils	10			16	6		1	33
TRB	Talus, Rubble and Alpine Brown Soils							7	7
	—Predominantly Non-Soil Units—								
BL	Boulders	1			2	4			7
RG	Rock covered Glaciers				2				2
RO	Rock Outcrop	41	7	4	18	17	17	15	119
T	Talus	9			12	4	8	1	34
TR	Talus and Rubble							12	12
TS	Talus and Scree	5			5				10
VST	Volcanic Scree and Talus						3		3
Totals		134	107	10	154	119	92	83	699

Table 4. Summary of watershed characteristics. Basin area is the amount of drainage area above the outflow gauging station.

Lake/Basin	Basin Area (ha)	Basin Relief (m)	Lake Area (ha)	Outlet Elev. (m)
Crystal	135	293	5.0	2,951
Emerald	120	616	2.7	2,800
Lost	25	160	0.7	2,475
Pear	136	471	8.0	2,904
Ruby	441	812	12.6	3,390
Spuller	97	537	2.2	3,131
Topaz	178	275	5.2	3,218
Mini-watersheds	7	19	0	2,950

Table 5. Potential phases of genetic soils, horizonation and possible taxonomic subgroups

Genetic Soil	Major Variants	Potential Phases	Horizonation	Major Taxa (Subgroups)
Alpine Brown Soils	Deep, Stable	Modal Phase	A-Bw-C; A-Bw-C-R	Typic Cryumbrepts, Dystric Cryochrepts
		Highly Leached Phase	O-A-Bw-C; O-EA-Bs-C	Dystric Cryochrepts, Typic Cryumbrepts
		High Elev. Minimally Dev. Ph.	A-(Bw)-C; A-C	Entic Cryumbrepts, Typic Cryumbrepts
		Highly rocky Phase	A-Bw-C; A-Bw-C-R	Typic Cryumbrepts, Dystric Cryochrepts
		Iron oxide rich Phase	A-Bw-C; A-Bw-C-R	Typic Cryumbrepts, Dystric Cryochrepts
	Deep , Stratified	Modal Stratified Ph., (w/ Bw)	A-Bw-2A-2C; A-AB-2A-2C	Typic Cryumbrepts, Entic Cryumbrepts
		Min. Dev. Ph. (wo/ clear Bw)	A-C-2A-2C; A-(Bw)-C-2A...	Entic Cryumbrepts
		Min. Dev., high elev. (wo/ Bw)	A-C-2A-2C	Entic Cryumbrepts, Typic Cryorthents
		Stratified highly rocky Phase	A-Bw-2A-2C; A-AB-2A-2C	Typic Cryumbrepts, Entic Cryumbrepts
		V. deep, Multiple buried soil Ph.	A-Bw-2A-2C-3A-3C....	Typic Cryumbrepts, Entic Cryumbrepts
	Shallow, Intermediate or Variable Depths	Very Shallow (<25 cm)	A-R; O-A-R	Lithic Cryumbrepts, Lithic Cryorthents
		Shallow (~25 to 50 cm)	A-R, A-Bw-R	Lithic Cryumbrepts, Lithic Cryorthents
		Shallow highly rocky phase	A-R, A-Bw-R	Lithic Cryumbrepts, Lithic Cryorthents
		Mod. deep Ph. (50 to 100 cm)	A-Bw-R; A-R; O-A-Bw-R	Typic Cryumbrepts, Dystric Cryochrepts
		Highly variable depth Phase	A-R; A-Bw-R; O-A-R	Ruptic-Lithic Cryumbrepts, Lithic-Ruptic-Entic Cryumbrepts
		Shallow, high elevation Phase	A-R	Lithic Cryorthents, Lithic Cryumbrepts
		Shallow highly leached phase	O-EA-R	Lithic Cryochrepts, Lithic Cryorthents
		Shallow iron oxide rich phase	A-R, A-Bw-R	Lithic Cryumbrepts, Lithic Cryorthents
Moist Meadow Soils	Deep, Stratified	Modal Phase	A-Bw-Cg; A-C-2A-2Cg	Aquic Cryumbrepts, Aquic Cryofluvents, Aquic Cryochrepts
		Rocky Phase	A-Bw-Cg; A-C-2A-2Cg	& Oxyaquic Cryumbrepts (apply to modal & Rocky Phases)
		Min. dev. young Phase	A-C-Cg; A-C-2C	Aquic Cryofluvents, Oxyaquic Cryofluvents<Aquic Cryumbrepts
	Shallow, Intermediate or Variable Depths	Shallow to v. Shallow Phase	A-R; A-Bw-R; A-Cg-R	Lithic Cryumbrepts, Lithic Cryorthents, Oxyaquic Cryofluvents
		Shallow rocky phase	A-R; A-Bw-R; A-Cg-R	Lithic Cryumbrepts, Lithic Cryorthents, Oxyaquic Cryofluvents
		Highly variable depth Phase	A-R; A-Bw-R; A-Cg-R	Lithic Cryorthents, Oxyaquic Cryorthents
Wet Meadow Soils	Deep, Stratified	Modal Phase	A-Cg-2A-2Cg; O-A-Cg	Typic Cryaquents, Humic Cryaquepts, Histic Cryaquepts
		Rocky Phase	A-Cg-2A-2Cg; O-A-Cg	Typic Cryaquents, Humic Cryaquepts, Histic Cryaquepts
		Minimally dev., young Ph.	A-C; O-C; C-2C	Typic Cryaquents, Lithic Cryaquepts
		Highly organic Phase	O-A-Cg; O-Cg; O-O...	Histic Cryaquepts, Terric Cryofibrists, Fluvaquentic Cryofibrists
	Shallow, Intermediate or Variable Depths	Shallow to v. Shallow Ph.	A-Cg-R; A-R; O-A-R	Lithic Cryaquepts, Typic Cryaquents, Lithic Cryorthents
		Shallow rocky Phase	A-Cg-R; A-R; O-A-R	Lithic Cryaquepts, Typic Cryaquents, Lithic Cryorthents
		Highly variable depth Ph.	A-Cg-R; A-R; O-A-R	Lithic Cryaquepts, Histic Lithic Cryaquepts, Typic Cryaquents

Table 5. (cont.) Potential phases of genetic soils, horizonation and possible taxonomic subgroups

Genetic Soil	Major Variants	Potential Phases	Horizonation	Major Taxa (Subgroups)
Riparian Soils	Bedrock	Modal Phase	A-R; A-C-R	Lithic Cryumbrepts, Lithic Cryorthents
		Variable Depth Phase	A-C-R; A-R; A-C-A-R	Ruptic-Lithic Cryumbrepts
		High elevation Phase	A-R; A-C-R	Lithic Cryorthents, Lithic Cryumbrepts
		Low rock content phase	A-R; A-C-R	Lithic Cryumbrepts, Lithic Cryorthents
	Talus	Modal Talus Phase	A-C	Oxyaquic Cryumbrepts
		Stratified, variable depth Ph.	A-C-2A-2C	Oxyaquic Cryumbrepts
	Volcanic	Modal Volcanic Phase	A-R; A-C-R	Lithic Cryumbrepts, Lithic Cryorthents
		Stratified, variable depth Ph.	A-R; A-C-R; A-C-2A-R	Ruptic-Lithic Cryumbrepts, Vitrandic Cryorthents
Volcanic Brown Soils	Deep, Stable	Modal Phase	O-A-Bw-C; A-Bw-C	* *The volcanic soils may classify into about 15 to 20 subgroups, some of which (especially very shallow and wet ones) are listed above. In addition Vitrandic and Aquandic subgroups of the common Great Groups listed above appear very likely. Andisols may be present and if so are Cryands and Cryaquands. About 5 to 10 specific tests on about 50 samples are needed to initially estimate Soil Taxonomy with any degree of reliability.
		Highly Leached Phase	O-EA-Bw-C; O-E-Bs-C	*
		Highly rocky Phase	O-A-Bw-C; A-Bw-C	*
	Deep, Stratified	Modal stratified phase	O-A-C-2A-2Bw-2C	*
		Min. dev. & unstable Phase	A-C-2C; O-A-(Bw)-C-2A-2C	*
	Shallow, Intermediate or Variable Depths	Very Shallow (<25 cm)	A-R; O-R; O-A-R	*
		Shallow (~25 to 50 cm)	O-A-R; A-R; O-A-Bw-R; A-Bw-	*
		Mod. deep Ph.(50 to 100cm)	O-A-Bw-R; A-Bw-R	*
		Highly variable depth Phase	A-R; O-A-R; O-A-Bw-R	*
		Shallow highly rocky phase	O-A-R; A-R; O-A-Bw-R; A-Bw-	*
		Shallow highly leached phase	O-EA-R; O-E-R; O-E-Bs-R	*
Volcanic Moist Meadow Soils	Deep, Stratified	Modal Phase	A-Bw-Cg; A-C-A-Cg; A-Cg	*
		Rocky Phase	A-Bw-Cg; A-C-A-Cg; A-Cg	*
		Minimally dev., young Phase	A-C; A-C-Cg; C-C	*
	Shallow, Intermediate or Variable Depths	Shallow to very Shallow Ph.	A-R;O-A-R; A-Cg-R	*
		Shallow rocky phase	A-R;O-A-R; A-Cg-R	*
		Highly variable depth Phase	A-R;O-A-R; A-Cg-R, A-Bw-R	*
Volcanic Wet Meadow Soils	Deep, Stratified	Modal Phase	O-A-Cg; A-Cg-2A-2Cg	*
		Highly Organic Phase	O-Cg-2O-2Cg; O-A-Cg-2O-	*
		Young Minimally dev. Phase	C-Cg; A-C-Cg; A-Cg; O-C	*

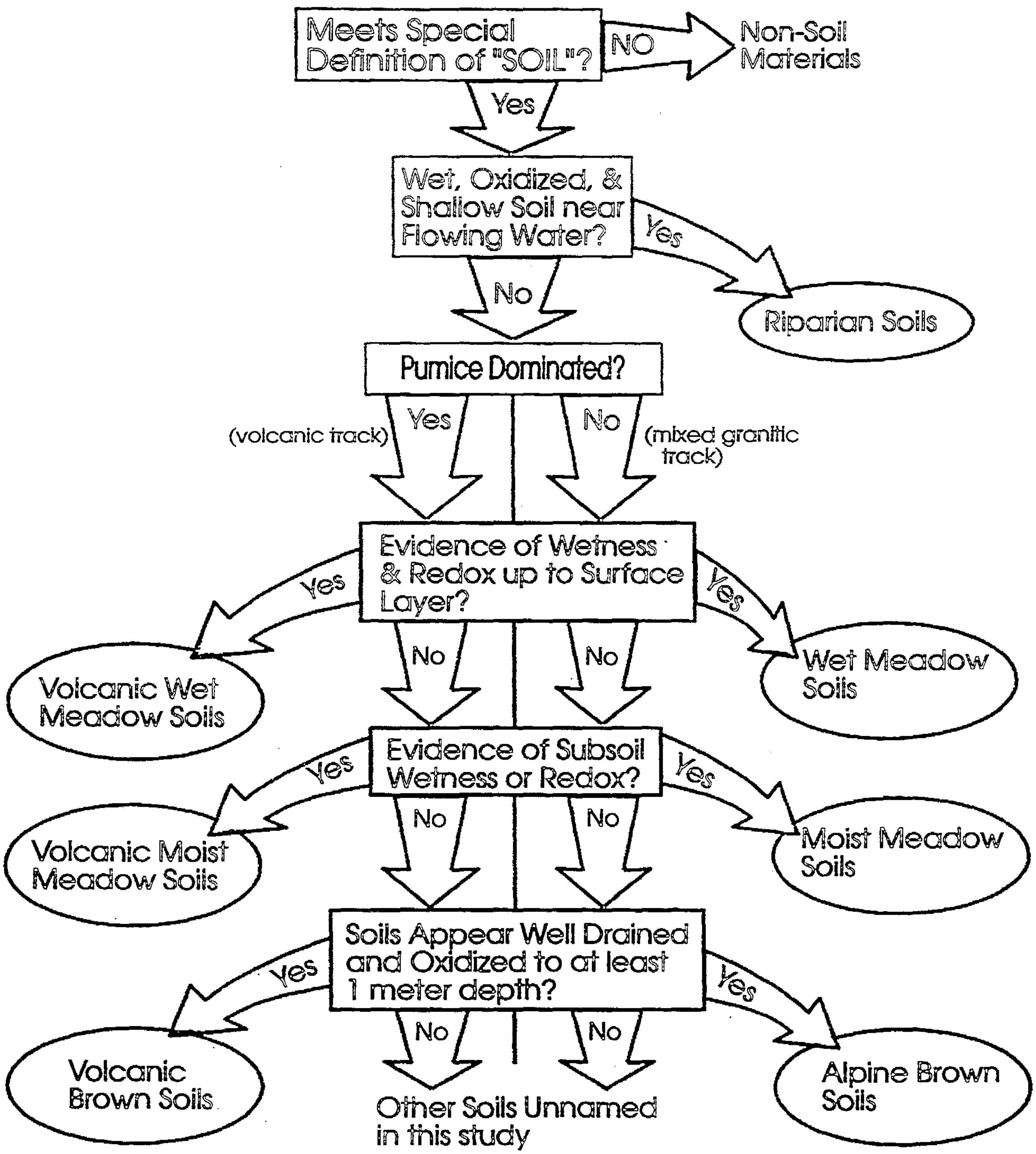

Figure 1. Schematic of relationships between named genetic soils