Snow, Snowmelt, Rain, Runoff, and Chemistry in a Sierra Nevada Watershed

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PREFACE

In the Sierra Nevada and much of western North America, snow dominates the hydrologic cycle. The montane snowpack is an integrator of wet and dry atmospheric deposition, which is held in storage until release during a melt period. At higher elevations of the Sierra Nevada, summer rainfall appears to contribute a much smaller quantity of water and solutes in comparison to snowfall. All aspects of the hydrologic cycle were examined in this study, but snow-related processes received most of the attention because of their dominant role.

Snow on the ground is a dynamic material that changes markedly in response to heat and mass transport. Thus, water molecules within the snowpack are transferred by sublimation between the solid and the gaseous phases. Recrystallization of ice generates a physical fractionation of ionic species. Impurities within the ice crystal lattice are segregated on the outer surfaces of snow grains because the impurities are not readily incorporated into the crystalline lattice during recrystalization.

Liquid water moving through the snowpack readily leaches the soluble impurities, but not all chemical species are distributed identically at the percolating melt-front. Montane snowpacks generally exhibit preferential elution of ions, i.e. the removal of some ions from the snowpack more quickly than others. Snow metamorphism, as well as chemical and biological transformations, can alter the chemical concentration and distribution of solutes in the snowpack and lead to release of most of the solutes during the initial phase of melt.

Prediction of melt water chemistry from bulk snowpack concentrations or cumulative concentrations from snow events is not straightforward. The original distribution of ions in the snow grains, the magnitude and type of snowpack metamorphism, the degree of dispersion at the advancing melt-front, and the number and intensity of melt-freeze cycles before runoff all play important roles in determining melt water solute concentrations. However, understanding the dynamics of snowmelt runoff through the montane watersheds is imperative when assessing the sensitivity of alpine environments to polluted atmospheric deposition.

Understanding of snowpack contributions to the chemistry of surface waters in alpine basins is complicated further by the rugged and variable terrain. Large topographic differences over short distances result in spatial and temporal variation in the magnitude of snow accumulation and in the onset of snowmelt within a given watershed. The onset and rate of snowmelt at a particular location in the basin is a function of the complex interactions of aspect, elevation, slope, season, and meteorological parameters. Spatial and temporal differences in the onset of melt within a watershed produce a snowpack with variable chemistry.

Snowmelt is perhaps the dominant event that affects alpine ecosystems on an annual basis. This infusion of concentrated runoff, followed immediately by dilute melt water, may be the controlling abiotic event in alpine aquatic ecosystems. The changes in water chemistry from snowmelt runoff may have dramatic effects on individual organisms, and even on entire communities.

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ABSTRACT

Snow Accumulation and Distribution

Distribution of snow water equivalence (SWE) was measured in the Emerald Lake watershed located in Sequoia National Park, California, by taking hundreds of depth measurements and depth profiles at six locations during the 1986, 1987, and 1988 water years. Elevations range from 2800 to 3416 m, and the total watershed area is about 120 ha. A stratified sampling scheme was evaluated by identifying and mapping zones of similar snow properties based on topographic parameters that account for variations in both accumulation and ablation. Elevation, slope, and radiation values calculated from a digital elevation model were used to determine the zones. The topographic parameters (slope and elevation) do not change between survey dates, but the radiation data vary temporally, providing a physically justified basis for the change in SWE distribution through time. Field measurements of SWE were combined with the physical attributes of the watershed and clustered to identify similar classes of SWE. The entire basin was then partitioned into zones for each survey date. Optimal sampling schemes are calculated based on the observed variance in SWE found in each zone. Although results do not identify which of the classification attempts is superior, net radiation is clearly of primary importance, and slope and elevation appear to be important to a lesser degree. The peak accumulation for the 1986 water year was 2.0 m SWE, about twice the 50-year mean. The peak accumulation for 1987 was 0.67 m SWE, and for 1988 was 0.63 m SWE, both about half the 50-year mean.

Water Balance

A water balance developed for the Emerald Lake basin illustrates the absolute and relative magnitudes of the main water transfers in the catchment over two hydrologic years (1986 and 1987). For the combined water years,

total precipitation (**367 cm**) - total losses to the atmosphere (**80 cm**)
= total streamflow (**283 cm**) + error (**4 cm**).

Snow dominated the water balance, accounting for 95 percent of the precipitation and subsequent streamflow. Snowpack accumulation was the principal hydrologic process from November through March, and snowmelt was the main activity from April through June. Evaporation from snow was the principal water loss to the atmosphere, accounting for about 80 percent of the total evaporation. Groundwater storage and release account for only a small portion of the total quantity of water in the annual water balance of this largely-impermeable basin.

Basin Discharge

We have developed adequate rating curves (stage-discharge relationships) for the outflow and the two major inflows (1 and 2). These three channels were continuously monitored using automatic data-logging devices. Minor inflows were monitored with many manual observations.

The total annual volume of water flowing out of the Emerald Lake basin over the complete period of record (Oct. 1983 - Sept. 1987) ranged from 670,000 m^3 to 2.6 million m^3. The maximum volume during water year 1986 was more than three times the minimum

volume during water year 1985. The total volume of Emerald Lake is about 160,000 m^3. Equivalent depths of water averaged over the basin were 214 cm in water year 1986, 68 cm in water year 1987, and 58 cm through mid-June 1988. Annual streamflow even during the low year was more than twice the national average of 23 cm.

Hydrographs clearly show that the majority of runoff occurred during the months of snowmelt. More than three-quarters of the annual runoff occurred in the months of April through July. Under optimum combinations of conditions favoring high rates of snowmelt runoff, peak discharges approached 1 $m^3 s^{-1}$ during 3 days in 1986. The minimum flow in water year 1986 was about 180 $m^3 day^{-1}$. The minima for the entire period of record were below 20 $m^3 day^{-1}$ and occurred in mid-February to mid-March of 1985 and September and October of 1987.

Climate and Energy Exchange at the Snow Surface

A detailed evaluation of surface climate and energy exchange at the snow surface is presented for the 1986 and 1987 water years. Each form of energy transfer — radiation, sensible and latent heat flux, soil heat flux, and heat flux by mass advection — is evaluated to determine its magnitude and importance in the seasonal energy and mass balance of the snowcover. During snowmelt, radiation accounts for between 75 and 90% of the energy available for melt. Sensible and latent heat transfer during this time are of approximately equal magnitude, but are usually of opposite sign, and therefore cancel. Calculated sublimation during the entire snow season accounted for the loss of about 20% (approximately 50 cm SWE) of the mass of the snowcover in 1986, and about 35% (approximately 23 cm SWE) of the mass of the snowcover in 1987.

Topographic Distribution of Solar Radiation

Among the energy fluxes controlling snow metamorphism and snowmelt in mountainous drainage basins, solar radiation has the largest topographically caused variation. A two-stream atmospheric radiation model calculates solar radiation over alpine terrain in two broad wavelength bands — visible and near-infrared — and a spectral model for the albedo of snow is parameterized to the same wavelength bands to estimate net solar radiation. A least-squares fit to surface measurements finds the necessary atmospheric attenuation parameters, and the topographic variables are calculated from digital elevation data.

The spatial and temporal distribution of solar radiation is characterized by low spatial variance at low magnitudes in the winter, higher spatial variance in the early spring, and low variance at high magnitudes in the late spring and early summer.

Chemistry of Wet Deposition and Snowmelt Runoff

The annual volume-weighted concentration of solutes in wet deposition at the Emerald Lake watershed, for water years 1985 through 1987, was equal to or less than 5 $\mu eq\ L^{-1}$ for each of the major ions. H^+ and NH_4^+ each account for about 18% of the total ionic content, followed closely by NO_3^- (17%), SO_4^{2-} (14%) and Cl^- (11.5%). The remaining portion is divided among Ca^{2+}, Na^+, K^+ and Mg^{2+}. The organic anions CH_3COO^- and $HCOO^-$ comprise 25% of the total anionic content of wet deposition. Dry deposition to the snowpack does not appear to be important during the winter season. Rainfall is acidic, with a H^+ concentration about 6-fold greater than pure water in equilibrium with atmospheric

carbon dioxide. Snowfall supplied 90% of the solute flux to the basin in 1985 and 1986. Rain supplied 66% of the solute flux in 1987.

Interactions among the solutes retained and released from the snowpack, energy flux throughout the basin, and hydrologic pathways are all important to hydrochemistry during snowmelt runoff. Solutes in the initial fraction of snowmelt runoff are 5 to 10-fold more concentrated than the bulk concentration of solutes in the snowpack, an ionic pulse. Spatial and temporal variations in the initiation and intensity of snowmelt prolong the time period of the ionic pulse in the basin. NO_3^- concentrations in streamwater during snowmelt are elevated 100% to 200% above winter concentrations of NO_3^-. The source of the elevated NO_3^- concentrations in streamwater is snowmelt runoff. SO_4^{2-} concentration in streamwater during snowmelt runoff is attenuated with respect to SO_4^{2-} concentrations in meltwater. Hydrogen ion concentration in streamwater during snowmelt runoff indicates strong interactions between runoff and biogeochemical processes: 80% of the H^+ stored in the snowpack in 1986 was removed before reaching Emerald Lake; 90% was removed before reaching the lake in 1987.

I. Summary and Conclusions

A. Snow Accumulation and Distribution

An important measurement in the study of Emerald Lake basin is the amount of snow water equivalence (SWE) stored in the watershed. Water storage and distribution is important in the hydrological and chemical portions of the study. SWE can be easily measured at a point, but it is far more difficult to estimate over an entire watershed. Snow is not deposited uniformly in alpine basins. Wind and avalanches redistribute snow unevenly, and variable energy inputs increase spatial heterogeneity. Intensive snow surveys to measure depth were conducted in the Emerald Lake basin during the 1986, 1987, and 1988 melt seasons, and multiple snow-pits were dug throughout all three seasons to characterize density, which varied little throughout the basin. The combination of these data allowed us to estimate total SWE stored in the basin. Snowfall events were monitored in the basin using snowboards in the 1986 and 1987 water years.

Statewide, the 1986 water year was marked by snowfall that was about 1.5× the 50-year mean. Data from the Tulare River basin showed that by April 1, the cumulative precipitation had reached 140% of the 50-year mean, and in the Kaweah basin it totaled 175%. Precipitation at the Lodgepole Ranger Station in Sequoia National Park reached twice the 50-year mean by the end of February. The duration of these records is much longer than the few seasons at Emerald Lake, and they serve as an index of the magnitude of winter precipitation in the basin. Precipitation at Emerald Lake in 1986 was measured using snow boards placed at several locations in the watershed and sampled shortly after each storm. Two moderate storms in November and one large storm in December were the only significant deposition events prior to a record-breaking storm in February, which deposited over 1 m SWE over much of the basin and brought the basin mean SWE to nearly 2 m. Little precipitation followed this event with the exception of one storm in March that deposited almost 0.5 m SWE.

The 1987 water year was marked by lower than normal precipitation. Statewide precipitation was 65% of the 50-year mean for the 1987 water year, and estimates for the Sierra Nevada were even lower. Data from the Tulare River basin showed that by April 1, the cumulative precipitation had only reached 70% of the 50-year mean, and the Kaweah basin attained only 55%. Statewide snow surveys indicated that the snowpack was just over 50% of normal for April 1 and 20% of normal for May 1. At Emerald Lake, snowboards were sampled at two locations near the inlet after each storm. The first measurable deposition fell after January 1 and only deposited 0.11 m SWE. Four more storms of similar magnitude followed, leaving a cumulative SWE of 0.49 m on March 9. The first week in April was the date of peak accumulation in the Emerald Lake which with about 0.67 m SWE. Little precipitation fell after this date.

In the 1988 water year, early storms provided optimism for a normal or above normal precipitation year, but a dry period that began in mid-January and lasted through most of February showed 1988 to be another critically dry year. At the end of January, the Tulare basin was at 110% of the 50-year precipitation mean, but one month later the record showed only 85%. By the end of March the estimate had dropped to 70%. Records show that by

April 1, the cumulative precipitation had only reached 85% of the 50-year mean for the Kaweah basin. In the Emerald Lake basin the date of maximum accumulation was about the third week in March with a total of 0.63 m SWE. Rapid melting after this date depleted the snowpack. April storms temporarily stopped melt and added a small amount to the snowpack, which brought the regional precipitation up to 83% of the long-term mean for the date. The seasonal snowpack is usually about at 70% of the seasonal maximum on May 1, but this year it was only about 20%.

Distribution of snow water equivalence (SWE) in the Emerald Lake basin was examined during the 1986, 1987, and 1988 water years. The peak accumulation for the 1986 water year was 2 m SWE. The 1987 and 1988 water years were similar in distribution and volume of snow, with peak accumulations of 0.67 m and 0.63 m SWE, respectively.

One objective of this project was to find better methods for measuring and quantifying the distribution of snow water equivalence in an alpine basin. A stratified sampling scheme was evaluated by identifying and mapping zones of similar snow properties based on topographic parameters that account for variations in both accumulation and ablation. Elevation, slope, and radiation values calculated from a digital elevation model were used to determine the zones. Field measurements of SWE were combined with the attributes of the sample locations and clustered to identify similar classes of SWE. The entire basin was then partitioned into zones for each survey date. The topographic parameters of the basin used in the classification (slope and elevation) did not change between survey dates. The radiation data vary temporally, providing a physically justified basis for the change in SWE distribution through time. Although results do not identify which of the classification attempts is superior, net radiation is clearly of primary importance, and slope and elevation appear to be important to a lesser degree. The results show that terrain features and radiation exert some effect on snow distribution and show promise in modeling snow distribution in alpine areas using physically-based parameters. Results show that an optimal sampling scheme can be defined for an alpine basin using a stratified random sample. The optimal survey allows sampling of SWE to a desired level of accuracy and can be based on cost, application needs, or logistical considerations.

B. Water Balance

The water balance for the combination of water years 1986 and 1987, expressed as equivalent water depths averaged over the catchment area was

total precipitation (**367 cm**) - total losses to the atmosphere (**80 cm**)
= total streamflow (**283 cm**) + error (**4 cm**).

A monthly water balance demonstrated the highly seasonal nature of the major hydrologic processes in the Emerald Lake basin. At this shorter time scale, errors did not compensate to the same degree as in the two-year balance and better indicate the uncertainty in some of the components. The peak snowpack water equivalence before the onset of spring melt is the most useful reference for alpine hydrology. In 1986, about 90 percent of the water stored in the snowpack in mid-April plus subsequent precipitation during the remainder of the water year became streamflow. In 1987, about 75 percent of the peak snow storage plus subsequent precipitation became streamflow.

Snow dominated the water balance during the study period, accounting for 95 percent of the precipitation. Snowpack accumulation was the principal hydrologic process from November through March, and snowmelt was the main activity from April through June. The two water years differed greatly in precipitation: about 2.6 m occurred in 1986 and about 1 m occurred in 1987.

Water losses to the atmosphere are the only output from the Emerald Lake basin other than streamflow. Estimated total evaporation from snow, water surfaces, soil, and vegetation at Emerald Lake was 22 percent of the estimated precipitation over both water years. Sublimation was the largest loss, accounting for about 80 percent of the total evaporation. Total sublimation in 1986 was almost twice the amount that occurred in 1987, largely because of the greater duration of snow cover in 1986. Evaporation from non-snow surfaces was limited due to the small proportion of the basin that is covered by water or vegetation.

Groundwater storage and release account for only a small portion of the total quantity of water in the annual water balance of this largely-impermeable basin. However, subsurface water is very important in the seasonal distribution of water. Releases from subsurface storage are the primary water input to Emerald Lake for eight to nine months of the year. The residence time of the groundwater in the basin varies between a few days and a few months. Groundwater discharged during late fall and winter probably has been in subsurface storage for several months or is present due to recharge from autumn rains. The total groundwater storage in the basin was estimated to be equivalent to 10 cm storage averaged over the basin area of 1.2 km^2.

Streamflow occurred primarily during the spring snowmelt period. Flow during the remainder of the year was largely a long recession until the next spring. In 1986, 90 percent of the 2.6 million m^3 of streamflow leaving the basin occurred from April through August. Snowmelt runoff from April through June of 1987 accounted for 86 percent of that year's total runoff of 820,000 m^3.

C. Basin Discharge

Streamflow is one of the most important and informative measurements of the hydrologic mass balance. Streamflow integrates all of the hydrologic processes occurring throughout the basin into a single point measurement. Information about many of the processes occurring in the basin is contained in the streamflow record. Because losses and storage other than those related to the snow cover tend to be small, the basin outflow can provide an indication of the timing and quantity of rainfall and snowmelt. In combination with the knowledge of other components of the water balance, streamflow provides an excellent integration of the hydrology of the basin and its implications for chemical cycling.

We have developed adequate rating curves (stage-discharge relationships) for the outflow and the two major inflows (1 and 2). These three channels were continuously monitored using automatic data-logging devices. Large uncertainties exist in these functions at higher discharges because of scant data collected during short duration events in 1986. Greater confidence will only be attained if pre-calibrated measuring structures (e.g., weirs or flumes) are placed in the channels, and maximum flows are again generated in the basin. Minor inflows were given a lower priority because of their reduced importance in the overall water balance. They were not continuously monitored by data-logging equipment, but as many manual observations as possible were made.

The total annual volume of water flowing out of the Emerald Lake basin over the complete period of record (Oct. 1983 - Sept. 1987) ranged from 670,000 m^3 to 2.6 million m^3. Thus, the total volume during water year 1986 was more than three times the total volume during water year 1985. For comparison, the volume of Emerald Lake is about 160,000 m^3. The equivalent depths of water averaged over the basin area of 1.2 km^2 were 214 cm in water year 1986, 68 cm in water year 1987, and 58 cm through mid-June 1988. Annual streamflow even during the low year was more than twice the national average of 23 cm.

Hydrographs and tabulation of monthly streamflow volumes clearly show that the majority of runoff occurred during the months of spring snowmelt. More than three-quarters of the annual runoff occurred in the months of April through July. May and June were the two months of greatest flow, accounting for at least half of the water year volume in each of the five years. Streamflow declined through summer as snow cover receded and water slowly drained out of soils and other surficial deposits. Some of the snowfall in autumn usually melted within a few days and accounted for increased flow in September, October, and November.

Daily water volume flowing out of the basin illustrates the high variability of streamflow. The highest daily volume of record was about 36,000 m^3 on May 30, 1986. Streamflow exceeded 20,000 $m^3\,day^{-1}$ (1.5 $cm\,day^{-1}$ water depth averaged over the basin area) on 44 days during spring and summer snowmelt in water year 1986. Flows were above 10,000 $m^3\,day^{-1}$ (0.75 $cm\,day^{-1}$) on 71 days in water year 1986 versus 32 days in 1987, and 26 in 1988.

Instantaneous flows rarely exceeded 0.5 $m^3\,s^{-1}$ under snowmelt conditions. Under optimum combinations of conditions favoring high rates of snowmelt runoff, peak discharges approached 1 $m^3\,s^{-1}$ during 3 days in 1986. The greatest instantaneous discharge during the study period occurred on February 15, 1986 when massive avalanches on to the ice cover of Emerald Lake displaced a substantial amount of the water from the lake. Peak flows between 10 and 20 $m^3\,s^{-1}$ were estimated from channel scour. The minimum flow in water year 1986 was about 180 $m^3\,day^{-1}$. The minima for the entire period of record were below 20 $m^3\,day^{-1}$ and occurred in mid-February to mid-March of 1985 and September and October of 1987.

D. Topographic Distribution of Solar Radiation

Among the energy fluxes controlling snow metamorphism and snowmelt in mountainous drainage basins, solar radiation has the largest topographically caused variation and is responsible for the major spatial variations in snowmelt, metamorphism, and ion elution. A two-stream atmospheric radiation model calculates solar radiation over alpine terrain in two broad wavelength bands — visible and near-infrared — and a spectral model for the albedo of snow is parameterized to the same wavelength bands to estimate net solar radiation. A least-squares fit to surface measurements finds the necessary atmospheric attenuation parameters, and the topographic variables are calculated from digital elevation data.

The spatial and temporal distribution of solar radiation is characterized by low spatial variance at low magnitudes in the winter, higher spatial variance in the early spring, and low variance at high magnitudes in the late spring and early summer.

E. Chemistry of Wet Deposition and Snowmelt Runoff

1. Solute Flux from Wet Deposition

The annual volume-weighted mean concentration of solutes in wet deposition at the Emerald Lake watershed, for water years 1985 through 1987, was equal to or less than 5 $\mu eq\ L^{-1}$ for each of the major ions. H^+ and NH_4^+ each account for about 18% of the total ionic strength of precipitation, followed closely by NO_3^- (17%), SO_4^{2-} (14%) and Cl^- (12%). The remaining portion of ionic flux is divided among Ca^{2+}, Na^+, K^+ and Mg^{2+}. The organic anions CH_3COO^- and $HCOO^-$ comprise 25% of the total anionic content of wet deposition. Solute concentrations in rainfall are about 10 times those in snowfall, with the exception of H^+, which is about 2½-fold more concentrated in rain than in snow.

The mean annual flux of solutes for water years 1985 through 1987 was 508 equivalents (eq) per hectare. Snowfall supplied 90% of the solute flux to the basin in 1985 and 1986. Rain supplied 66% of the solute flux in 1987; 87% of the rain in 1987 fell in the month of May. Rainfall and wet snow in the three months of September, October and May, deposited 25% to 65% of the annual solute flux to the basin during these three water years. Event sampling of rainfall and wet snowfall in the spring and autumn is necessary to adequately measure wet deposition to alpine watersheds.

2. Storage and Release of Ions in the Snowpack

Most of the solute flux from snowfall is stored in the seasonal snowpack. Solutes are released from the snowpack in snowmelt runoff, usually from about April 1 through July 15. Thus, 90% of the annual wet deposition to the basin entered the terrestrial and aquatic components of the watershed in a time period of three to four months. Any increase in the acidity of snowfall will be stored during the winter season, to be released to the watershed in a relatively short time span.

The snowpack in alpine areas is a collector of solute flux from wet and dry deposition. Comparison of solutes and water in cumulative snowfall events to solutes and water stored in the snowpack at the same site demonstrated either a slight loss of solutes and water from the snowpack relative to snowfall events, or no difference. If dry deposition is important to the winter snowpack, solute storage in the snowpack should be higher than solute loading by wet deposition, if there is no significant water loss from the snowpack. Our measurements indicate that dry deposition to the snowpack is not important during the winter season.

Solutes in the initial fraction of snowmelt runoff in 1987 where 5 to 10-fold more concentrated than the bulk concentration of solutes in the snowpack, an ionic pulse. The ionic pulse lasted 5-15 days at a particular site, depending on the rate of snowmelt, and took about 4-6 weeks to pass through the entire basin. Any increase in the acidity of snowfall will be magnified 5 to 10 times in the first fraction of snowmelt runoff.

Little spatial variability exists in the chemistry of the basin's snowpack prior to snowmelt. Chemical mass balance calculations can be made with a reasonable degree of confidence at the start and end of the snowmelt season. However, because of the differential release of solutes from the snowpack, spatial variability of solutes in the snowpack increases with time during the snowmelt season. Calculation of changes in chemical mass balance for a shorter time step during the snowmelt season may not be possible with our present sampling protocol, due to the high spatial and temporal variability of snow chemistry during this time period. Combining measurements of snow chemistry with a spatially-distributed snowmelt model may be the only possible method to adequately estimate solute flux to the watershed and Emerald Lake from snowmelt runoff.

3. Solute Concentrations in Streamwater During Snowmelt Runoff

Solute concentrations in stream water during snowmelt runoff are a combination of precipitation input and modification by within-snowpack and watershed processes. Basic cation, silica and HCO_3^- concentrations in stream water are mainly the product of weathering processes in the basin. During the period of snowmelt runoff their concentrations in stream water decrease consistently as a result of groundwater mixing with the more dilute water from snowpack runoff. However, silica and HCO_3^- concentrations decrease at a faster rate than basic cations during the start of snowmelt runoff, because basic cations are eluted from the snowpack at relatively high concentrations.

During the summer and autumn biological uptake results in NO_3^- immobilization. A reduction in biological activity during the winter and spring results in NO_3^- becoming mobile. NO_3^- concentrations in stream water during snowmelt are elevated 100% to 200% above winter concentrations. The source of the elevated NO_3^- concentrations in stream water is snowmelt. Since the acidification potential of NO_3^- is expressed as the number of mobile nitrate ions, any increase in the NO_3^- concentration of snowfall will result in an increase in the acidification of surface waters.

Sulfate concentration in stream water during snowmelt runoff is attenuated with respect to SO_4^{2-} concentrations in melt water. Adsorption-desorption by the clay minerals of the basin is a possible cause of this SO_4^{2-} attenuation. This is unexpected, since soils comprise only 10% of the surface area of the watershed. If a SO_4^{2-} isotherm does exist in the clay minerals of the basin, acidic inputs from increased H_2SO_4 concentration in future deposition will be ameliorated by SO_4^{2-} adsorption to some unknown degree. More knowledge of SO_4^{2-} adsorption-desorption properties by clay minerals in alpine watersheds is essential to determine the sensitivity of these watersheds to potential increases in SO_4^{2-} deposition.

Dilution of groundwater by snowmelt runoff results in a decrease in the sum of basic cations (C_b) in surface waters during spring runoff. While dilution of C_b cannot cause strong acidification (ANC < 0 μeq L^{-1}), it does cause a watershed to be much more sensitive to increases in strong acid anions. Strong acid anions do increase during the initial period of snowmelt runoff. The combination of dilution of the sum of basic cations and increase in the sum of mineral acid anions thus causes surface waters of the Emerald Lake basin to be sensitive to acid deposition during snowmelt runoff.

Hydrogen ion concentration in stream water during snowmelt runoff reflects strong interactions between runoff and biogeochemical processes within the watershed. 80% of the H^+ stored in the snowpack in 1986 was removed before reaching Emerald Lake; 90% was removed before reaching the lake in 1987. The magnitude of H^+ buffering is surprising, given the short residence time of snowmelt runoff in groundwater and soil reservoirs, or in contact with bedrock during overland flow. Titration of HCO_3^- accounts for little of the H^+ buffering. The H^+ consumption in snowmelt runoff appears to be from the complex interaction of accelerated weathering, cation exchange and adsorption.

The consistent and large quantity of H^+ buffering that occurs in snowmelt runoff at present deposition levels has important implications for the susceptibility to acidification of high-elevation watersheds. Surface waters in alpine basins may not be as sensitive to acid deposition as indicated by their characteristically low concentrations of ANC. Alternatively, the biogeochemical processes that at present buffer H^+ inputs from wet deposition may be nearly saturated, and small increases in H^+ flux may cause large increases in acidification. A better understanding of the biogeochemical processes that buffer H^+ flux from wet deposition is imperative in assessing the sensitivity of alpine watersheds to increases in acid deposition.

Interactions among the solutes retained and released from the snowpack, energy flux throughout the basin, and hydrologic pathways are all important to hydrochemistry during snowmelt runoff. Snow metamorphism produces an ionic pulse in the first fraction of melt water to exit the snowpack. The variable topography of the Emerald Lake watershed results in a highly variable energy flux in time and space, which in turn generates spatial and temporal variations in the initiation and rate of snowmelt in the basin. As a consequence the ionic pulse in snowmelt runoff is prolonged within the basin. What route snowmelt runoff takes as it flows towards Emerald Lake partially determines the chemistry of water flowing into the lake during snowmelt runoff. Apparently contact time on the order of hours to days between snowmelt runoff and the terrestrial part of the watershed is enough to consume H^+ and remove or add SO_4^{2-} to snowmelt runoff. Geochemical processes within the watershed are important to surface-water chemistry during snowmelt runoff. These biogeochemical processes cannot be ignored when modeling attempts are made to understand or predict the effects of current or future acidic deposition on alpine watersheds.

4. Rain-on-Snow Event

Surface waters in alpine watersheds are thought to be particularly sensitive to acidic deposition during snowmelt runoff. Rainfall in the spring of 1987 resulted in a natural experiment that permitted us to test this hypothesis. 14 cm of rainfall with a volume-weighted mean pH of 4.9 deposited 2,200 eq of H^+, compared to the 1,600 eq stored in the snowpack, from April 27 to June 8. Prior to the rain events, pH in stream waters was similar to that of water year 1985 (6.1), and much higher than that in water year 1986 (5.7). Stream waters experienced a depression in pH (5.7) from these rain events during snowmelt runoff in 1987 as low as the minimum pH recorded in 1986. Surface waters in alpine basins are therefore more sensitive to acidic rainfall during the period of snowmelt runoff.

II. RECOMMENDATIONS

A. Monitoring

- In future studies of alpine hydrology, streamflow should be measured more precisely by means of a pre-calibrated hydraulic structure.
- Precipitation should continue to be measured by means of an intensive snow survey at peak accumulation, supplemented with recording gages from April through October.
- An aerial survey of snow cover is necessary on at least three occasions during spring snowmelt.
- Continuous monitoring of meteorological variables needed to drive an energy-balance snowmelt model (radiation, windspeed, vapor pressure, and air temperature) is necessary during at least April through October.
- Seasonal and interannual trends in wet deposition to alpine areas need to be quantified.
- Long-term monitoring of wet depositon to alpine areas in the Sierra Nevada is essential to evaluate changes over time in acidic deposition. The Emerald Lake watershed should be one of the monitoring sites.
- Rainfall and autumn snowfall must be monitored on an event basis.
- Monitoring of solutes in winter snowfall needs particular emphasis, as winter snowfall supplies the majority of solute flux to alpine watersheds. Winter snowfall in alpine areas (defined as areas that do not receive rain-on-snow events of sufficient magnitude to cause runoff before spring melt), can be monitored effectively from snowpits at the time of maximum accumulation.
- Organic acids need to be included in the standard analysis of precipitation quality in statewide precipitation networks.
- Frequent sampling of solutes in surface waters during the period of snowmelt runoff is necessary to monitor potential acidification of stream and lake waters from solute pulses during this time period.

B. Research

- Further work is needed on snow accumulation and distribution to adequately estimate water deposition from snowfall to alpine watersheds with a minimum of manpower and cost.
- Variables that control the drift erosion and deposition of snow, such as the rate of change or second derivative of slope, must als be explored to adequately estimate water deposition from snowfall.
- Another major question that needs to be addressed is how to scale up from small headwater basins, such as Emerald Lake, to major river basins in the Sierra Nevada. The only practical solutions to questions on large scale snow distribution depend on our knowledge of the electromagnetic properties of snow and our ability to adequately make

ground observations at this scale.

- Also, by accounting for energy fluxes to the snowpack, one can estimate the temperature profile of the pack and account for loss of snow mass through sublimation and melting. Knowledge of energy fluxes to the snowpack, combined with bulk snow chemistry measurements, may provide a method for estimating the chemistry of runoff to aquatic systems. It is therefore essential to continue investigation into these variables,, to assess the effect of the spatial distribution of snowmelt processes on the release of ions into the soils and streams.
- A spatially-distributed snowmelt model is needed to adequately predict snowpack water release during the spring months.
- A geographic information system should be developed for the Emerald Lake data set.
- More work needs to be done on elucidating how geochemical and biological processes operate in alpine areas.
- A mechanistic and predictive understanding of the processes that produce an ionic pulse in snowpack meltwater is essential to estimate geochemical and biological responses to increases in acidic deposition of snowfall. Any increase in the amount of acidic anions in snowfall to alpine areas can be magnified 5-fold or more in snowpack runoff by snow metamorphism. How and why this occurs needs to be determined. Our scientific understanding of the processes that produce an ionic pulse in snowpack meltwater needs to increase to where predictive assessments can be made of anionic concentrations in snowpack meltwater for a given anionic concentration in snowfall.
- Source-receptor relationships for solutes in wet deposition need to be determined.
- Our knowledge of several geochemical processes needs to improve: SO_4^{2-} sorption; H^+ consumption by accelerated weathering, ion exchange and adsorption; and the role of organic acids in alpine basins. It is imperative that we develop a mechanistic understanding of the above processes if we are to correctly evaluate the sensitivity of alpine watersheds to potential increases in acidic deposition.
- To increase our understanding of geochemical processes that operate during snowmelt runoff, we recommend that a small (100 to 1000 m^2), experimental watershed be selected and instrumented. Various experiments during the time period of snowmelt runoff can then be conducted to determine hydrologic pathways, residence time of snowmelt runoff in vadose and groundwater reservoirs, geochemical responses to different levels of acidification, the importance of ion exchange reactions in soils and unconsolidated deposits, and to determine if accelerated weathering occurs in response to increases in acidic deposition. The fate of organic acids in alpine basins can also be determined using this experimental watershed.

C. Policy Recommendations

- Continue to support the development of scientific knowledge on processes relating the sensitivity of alpine ecosystems to changes in precipitation chemistry.

- Develop longer-term records of hydrology and aquatic chemistry at a series of alpine lake basins.
- Establish a detailed monitoring network for precipitation and snowpack chemistry in the alpine zone.
- Continue to support development of hydrochemical models for scenario analysis.
- Begin to consider how different emission standards will affect alpine ecosystems over several decades, in anticipation of eventual standards.

III. SNOW ACCUMULATION AND DISTRIBUTION

A. Introduction

One of the principal properties of concern in snow hydrology is snow water equivalence (SWE), the depth of water at a point that would result if the snow were melted. SWE may be estimated by multiplying the depth by the mean density so

$$SWE = \frac{\int_0^h \rho_s(z)dz}{\rho_w} \tag{1}$$

where:

ρ_s = density of snow layer (kg m^{-3}),
ρ_w = density of water (kg m^{-3}),
z = depth of snow layer (m),
h = depth of snowpack (m),

and mean density is defined as

$$\bar{\rho} = \rho_w \frac{SWE}{z} \tag{2}$$

With the use of both established and recently developed techniques, SWE measurements at a given location are not difficult to obtain. Several accurate methods for measuring density exist, ranging from those involving excavation and sampling pits [Perla and Martinelli, 1978] to the isotope profiling gauge [Kattelmann et al., 1983]. Depth measurement requires only a robust probe and some experience in use.

The persistent question is: how do we accurately interpolate between measurements at points to estimate the total volume of water stored in the snowpack over an entire drainage basin? Snowpack properties may vary greatly over small distances. Numerous studies have been conducted in prairies or regions of mild relief, and snowpack variation in these places is better understood than spatial and temporal variations of snow cover in alpine regions. The factors contributing to variation in SWE (slope, aspect, elevation, vegetation type, surface roughness, energy exchange) are exaggerated in alpine areas, resulting in a heterogeneous snowpack that changes markedly in space and time.

Clearly, we need sampling methods that can capture snowpack variability and characterize it over an area, that have reasonable time and manpower requirements, and yet accurately assess the snowpack. An approach that requires many samples throughout a basin is seldom practical, given logistical constraints of safety and time. In this study we attempted to accurately determine the distribution of SWE over a small alpine basin by identifying and mapping zones of similar snow properties, based on topographic parameters that account for variations in both accumulation and ablation. These zones were calculated from the Emerald Lake digital elevation model (DEM) using image processing techniques. Parameters used were elevation, slope, and daily integrated solar radiation for clear atmospheric conditions. Snow depth and density measurements were obtained in four intensive snow surveys over three melt seasons, providing a large sample of spatial point measurements for model development and testing. The basin was classified into zones of similar physical parameters using different parameters and numbers of zones. Survey data was registered to the classified basin and evaluated to determine if the classifications were meaningful in terms of delineating areas of different snow accumulation.

After discussing snow distribution and its controlling factors, the field methods used in this study are discussed. Results from the field surveys and modeling attempts are presented, followed by a discussion of the results and possibilities for future work.

B. Factors Affecting Snow Distribution

In order to understand the variable distribution of the snow cover, it is necessary to understand the processes controlling distribution. Investigations on snow accumulation and distribution in the last two decades have focused on elevation, vegetation, and topography. Meiman [1968] summarizes many of the earlier studies. Although much of the work has been done in regions of low elevation and minimum relief, many of the results apply to alpine areas. Even in regions with gentle terrain and low altitude, snow accumulation increases with elevation [Steppuhn and Dyck, 1974]. Studies have also examined the relationship between snow accumulation and terrain features and vegetation [Granberg, 1979]. Snow accumulation has been shown to depend on vegetation and topographic roughness through a wide range of scales, from small vegetation and surface roughness to large terrain features such as ridges and valleys. Table 1 lists some selections from recent work on snow distribution and its relationship with topographic and meteorological variables.

Properties of the snowpack (e.g. depth, density, temperature, chemistry) vary in space and time. Snow depth and density are controlled by both accumulation and ablation. On a large scale, these processes are controlled by meteorological patterns and major terrain features, and on a small scale by redistribution, new snow properties and micro-meteorology. Accumulation consists of two processes: snowfall itself and redistribution of the original snowfall by wind transport or by sloughing and avalanching. Ablation occurs by melting, sublimation, and deflation.

1. Accumulation

a. *Snowfall*

Precipitation, including snowfall, is a highly stochastic process and its variability must be considered on a wide range of scales. Regional climate and latitude affect snowfall, but neither of these vary significantly within most alpine basins. Elevation is considered the single most important factor in snow cover distribution by most of the studies cited in Table 1, but the relationship is not independent of climate or slope. Orographic effects depend more on slope and wind speed than on elevation [Gray, 1979]. Elevation within alpine basins, however, may have a range of more than 1000 meters, and the resulting differences in air temperature and vapor pressure affect snow crystal morphology [LaChapelle, 1969; Perla and Martinelli, 1978] and, therefore, affect density of new snow. Rhea and Grant

[1974] found a positive relationship between the topographic slope of the 20 km upwind fetch and long-term average precipitation. Snowfall also depends on the number of upstream barriers able to deplete the moisture supply of the air mass. Wind affects the amount of fragmentation that crystals undergo during and after deposition, and heavy fragmentation leads to higher density.

b. Redistribution

Much of the spatial heterogeneity of SWE in alpine regions is the result of redistribution. Even if snowfall were uniform over an area, the final deposition pattern would be highly irregular, because snow is typically moved by wind and redeposited during the precipitation event. The low density of the deposited snow, and the large surface area of many flakes compared to their mass, allows transport over irregular terrain and large areas. Variation in storm patterns and wind direction further complicates the problem.

Recently, much work has been done on blowing snow, because of its economic effects [reviewed by Schmidt 1982a]. Snow may be transported by wind-induced creep, saltation or entrainment into the air mass [Mellor, 1965; Radok, 1977; Schmidt, 1980]. Blowing snow is a two-phase process where ice crystals represent the solid phase and air represents the fluid. It becomes a three-phase process when solids sublimate to vapor during transport. Saturation is reached when the air cannot carry more solid load. Estimates for the fetch necessary to reach saturation are between 200 and 500 m, so it seldom occurs in rugged topography, because barriers are spaced too closely and they effectively trap the solid load.

In order for snow already on the ground to become entrained, electrostatic forces, surface tension and ice bonds must be overcome. Only exceptional winds are capable of entrainment in maritime climates where well-bonded surfaces develop rapidly. Saltation may occur where the surface is not too well-bonded, and there is a source of impact crystals from snow collected in trees, surface hoar or newly precipitated snow. The impact of crystals hitting the surface and dislodging other crystals will redistribute snow [Martinelli and Ozment, 1985].

Dyunin and Kotlyakov [1980] differentiate between storm types and their depositional characteristics. Upper snow storms are those in which snow falls without further transport. Deflation snowstorms do not contribute new precipitation but move previously deposited snow by saltation. Suspension snow storms are similar to deflation storms, but the previously deposited snow is entrained into the airmass by turbulent diffusion. Deflation storms are capable of drift and cornice formation on lee slopes and ridges, while upper storms have the ability to transport large volumes of snow into belts on lee slopes.

Like other sediments [cf. Bagnold, 1966] snow tends to accumulate in areas where air decelerates or flow is divergent, and it tends to erode in areas of acceleration or convergent flow. Föhn [1980], Schmidt [1984] and Schmidt et al. [1984], found maximum drift flux on an alpine ridge to be on the upwind side within a few meters of the crest, with scoured areas on windward slopes and deposition on lee slopes. Small disturbances in airflow lead to drift formation. Deflation hollows form adjacent to objects such as trees or boulders while immense drifts lie nearby. Largely due to the work of Mellor [1965], Schmidt [1980, 1982b, 1984], Schmidt et al. [1984], Tabler [1985], and Anno [1985, 1986], snow drift over simple uniform barriers is well understood. Where terrain irregularities and wind patterns are consistent in time, drifts and scoured areas tend to repeat in form and location, year after year. However, the problem is considerably more difficult and remains largely unresolved for complicated three-dimensional terrain found in alpine areas. Drifts may shift between storms as the storm track changes. Over a season, consistent patterns still often emerge.

Considerable volumes of snow may be moved by avalanches in a watershed. Regions in upper parts of basins accumulate snow in avalanche starting zones. When released, the snow is transported downslope to a resting point. Additional snow in the track or runout zone may be entrained and redeposited by the moving mass. Snow may repeatedly slough from slopes that are sufficiently steep, especially at low temperatures, for which metamorphic bonding processes are slow. Avalanching and sloughing are important in the nourishment and mass balance of glaciers [Tushinsky, 1975]. Alford [1973] identified cirque glaciers in Colorado nourished almost entirely by avalanches, although most were nourished by a combination of avalanching and drifting snow.

Avalanching does not change the total mass of snow in a drainage basin, but correct estimates of the volume in avalanche deposits are hydrologically important because these deposits may contain large amounts of water. Zalikhanov [1975] found that 30 to 64% of the alpine snow cover in the Caucasus may be transported to valley bottoms by avalanches. The retarded melt rate in the very deep deposits is usually not offset by the increase in energy available to melt the snow at lower elevations. Weir [1979] observed a single event at Mount Hutt, New Zealand that moved approximately half of the basin's SWE to an elevation far below the snowline. Melting snow produced a thin mantle as entrained debris collected on the surface. Melt rates may have then been retarded and runoff increased by reduced evaporation and infiltration. Iveronova [1966] and Sosedov and Seversky [1966], working in the Zailiysky Alatau of Russia, showed that displacement of snow to valley bottoms retarded melt and attenuated peak runoff. Martinec and de Quervain [1975] found that accelerated melt and increased runoff in the early season from avalanche deposits attenuated the peak seasonal discharges in the Dischmatal, Switzerland.

2. Ablation

A common method to evaluate ablation and snowmelt is through evaluation of the surface energy exchange. Snowpack ablation is controlled by energy exchanges at the air/snow and snow/ground interfaces. Energy inputs may come from solar and emitted atmospheric radiation, sensible heat exchange, latent heat exchange, heat flux from the underlying substrate, and advective heat transfer.

Of the available energy sources, it is well documented that in most cases solar and longwave radiation dominate [Zuzel and Cox, 1975]. Turbulent transfer processes -sensible and latent heat exchange- are important in snowmelt in some conditions, but are usually of opposite sign. Values for these fluxes are difficult to derive and the processes driving them, such as wind, are highly stochastic. Even with sophisticated instrumentation it is difficult to accurately estimate their contribution to melt. Advective heat transfer from rainfall is small, especially at the high elevations of the southern Sierra Nevada, where weather during snowmelt is usually clear.

Energy exchange through the ground/snow interface is important in some cases, such as areas with higher than normal geothermal exchange, but this energy is usually negligible in comparison to surface exchanges [Davis, 1980]. Heat flux from the ground may control accumulation at the onset of winter by melting snow as it falls on the surface. This depends on the thermal and optical properties of the substrate. Areas capable of absorbing and storing significant amounts of energy, typically areas with a low albedo and high heat capacity, will melt new snowfall. The effect may carry into the winter on features too steep to accumulate snow. The energy absorbed may be emitted as longwave radiation and melt out depressions around the features. Olyphant [1986b] found that radiation emitted from exposed rock faces reduced net longwave losses by 37% to 63% in Colorado.

More importantly, radiation affects net accumulation through ablation at the surface. If the melt only percolates into the snowpack and refreezes, then depth and density have changed but SWE has not. Once melt water reaches an ice lens or the ground, however, it may move laterally and the SWE at that point will change. Radiation thus influences the spatial element of accumulation as it may effectively move SWE from discrete parts of the basin where the energy balance is sufficient or remove SWE when runoff leaves the basin.

Melt water production can only take place once the pack is locally isothermal at 0^{o}C. It is important to note that it is not necessary for the entire pack to be ripe in order to produce melt because of the thermal diffusivity and conductivity of snow. Further, the production of surface melt does not necessarily lead to runoff. Surface melt may percolate down into an unripe portion and refreeze where the energy necessary to maintain the liquid phase is no longer sufficient. At this point the latent heat of fusion is released, an ice lens may form, and the pack temperature is raised a corresponding amount. Ice lenses may also be formed when liquid water encounters a layer of reduced permeability -a buried surface lens, wind slab, etc.- and spreads out horizontally on the incongruity. Refreezing may then take place, leaving an ice lens. This is a common scenario in the Sierra and other maritime snow environments and is a principal method of ripening or removing the cold content. In maritime environments ice lenses occur throughout the season, while in continental snowpacks they are primarily found in spring.

In predicting areas of melt for a given set of conditions, it is necessary to examine a number of factors. Besides the basic energy exchange components, it is necessary to look at the different physiographic characteristics of the point in question. Factors such as slope, aspect, latitude and horizon must be taken into account, especially in rugged terrain. In locations where radiation inputs are relatively low (high latitude), melt and rainfall tend to have a uniform effect on the snow cover [Adams, 1976]. In areas where radiation is both important and variable (lower latitudes and high elevations with rough topography) variability in snowpack parameters is increased. Some parts of alpine basins may go one or two months in the winter without receiving direct solar radiation, while adjacent areas may receive large amounts of direct radiation and experience occasional melt throughout the winter season. Breaking a subarctic basin into different topographic units has resulted in successful modeling of snowmelt from the units using an energy balance approach [Price et al., 1976]. Obled and Harder [1979] showed that topography controlled snow distribution during the accumulation season and accounted for the observed spatial diversity in snowmelt during the ablation season. Rugged, alpine terrain has a pronounced effect on the total energy balance, both by controlling incoming radiation and by variable emission of longwave radiation from terrain features [Olyphant, 1984, 1986a].

C. Field Methods

An exhaustive field measurement program was undertaken to measure SWE in the Emerald Lake basin. The program resulted in hundreds of depth measurements and excavation of numerous pits over the basin, which could be used to validate the results of the snow distribution model. Snowboards were used in several locations to monitor storm deposition. Snow stakes were used to measure ablation.

1. Snowfall

One objective of the field program was to sample each precipitation event for SWE and chemistry, with periodic sampling of pits to examine these same parameters in the accumulated snowpack. Event sampling was carried out at three locations in 1986, the inlet, pond, and ridge, and at the inlet in 1987. Snowboards were used for event sampling following established protocols [Perla and Martinelli, 1978]. A 1-m square plywood board placed on the old snow surface insures that the observer will not confuse the old snow with the new snow. The boards used were painted to minimize chemical contamination and a PVC tube was attached orthogonally to the board so it could be located after a heavy snowfall.

Two snowboards were placed several meters apart at each site on flat surfaces. As soon after each event as possible, each board was sampled. Melting and wind may remove snow from the boards; this leads to erroneous estimates of precipitation. Several samples were collected from each board using clear, graduated, PVC tubes. These tubes were inserted vertically until they reached a clean spatula inserted at the board surface. The sample was transferred to a clean plastic bag and weighed using a spring scale. Stepped samples were taken when the new snow depth exceeded the length of the tube (50 cm). Depth was read from the graduations on the tube after insertion. Sample volume was calculated by multiplying tube cross-sectional area by the depth. The weight of the sample divided by the density of water, multiplied by the volume gave sample density. Multiple samples were taken from both boards to get a mean value for the site.

2. Snow Water Equivalence

Variability in both depth and density must be considered in evaluations of snow distribution. Density measurements involve excavating snowpits and sampling the pit wall, which is labor intensive and time consuming. Conversely, depth measurements simply involve probing, and a large number of samples may be taken in the time required for a single pit. Depth varies more than density in alpine areas, so the major source of variation in SWE is variation in depth, especially during the melt season [Logan, 1973]. Fortunately, this makes field sampling feasible since many easily obtainable depth measurements can be combined with a smaller number of density profiles. A sampling scheme requires that the number of depth and density measurements be proportional to the parameter variances [Goodison et al., 1981]. Prior to the onset of melt, density exhibits considerable spatial variability, and measurements throughout the entire drainage area are needed to characterize this variability. SWE estimates are then computed from depth measurements, using an interpolated value for density. Ripening of the snowpack before runoff leads to less variation in grain size or density because large grains grow at the expense of smaller ones [Colbeck, 1982, 1983; Marsh, 1987]. Once the snowpack is ripe, fewer samples are necessary to characterize the density variation [Hasholt, 1973; Logan, 1973; Adams, 1976].

a. Snow Depth

Sample survey points were selected by several techniques described in detail below. Ordinarily, a stratified random sample is preferred for statistical reasons [Cochran, 1977]. In this study, however, the survey data were used to test our classification, and stratifying the basin before the surveys were completed would have biased the results, implying *a priori* knowledge of the distribution. Locations of the points were transferred from the DEM to orthographically corrected aerial photographs used by the field teams, and depth measurements were taken at each accessible point. Some of the selected survey points fell on locations that were too difficult for the survey teams to reach. These points were discarded if they appeared to have any snow on them, since accurate estimate from afar was not possible. However, a point was retained and a depth of zero recorded if it could be positively determined that it was located on a rock outcrop.

The field teams used the orthographic photographs, topographic maps, close-up photos and compasses to precisely locate the points in the field. At each location the survey team recorded aspect, slope angle, and snow depth at the point, as well as depths 4 m away in the four cardinal directions. The five depths were then averaged to minimize local variation of depth, caused by underlying boulders. Slope angle and aspect were obtained using compasses, and depths were found using interlocking aluminum probes usable up to 10 m depth. At depths greater than 10 m, friction and icing on the probes made it impossible to obtain a sample. Slope angle and aspect observations were used only as a check on location. For data analysis, the values for these variables were calculated from the DEM. Four surveys were completed in 1986, starting at the date of peak accumulation in the basin and following at approximately three-week intervals thereafter. Four surveys were carried out during the 1987 melt season at approximately two-week intervals. One survey was carried out as close to the date of peak accumulation as possible during the 1988 melt season.

b. Snow Density

Snow pits were dug at selected sites throughout the watershed to obtain density profiles. Pit locations were retained for each survey to minimize labor, because several of the pits in 1986 exceeded 6 m depth. The pit wall was excavated inward to a point that had not been subjected to the environmental changes caused by exposure of the previous sampling margins. Locations of the pit and snowboard sites are shown in Figure 1. Locations were chosen to give a range of exposures and elevations characteristic of the basin.

During this study we found that there was, and still is, no acceptable method available for simultaneously obtaining snow density and chemistry samples. PVC tubes work well for chemistry but poorly for density. Initially we used the PVC tubes for both measurements. When we determined that the PVC tubes were not producing reliable density measurements, we looked for a proven method for sampling snow density. No acceptable, commercially-available method existed at the beginning of this project. The conventional method involves inserting a small, 500 mL metal tube (also called a CRREL or SIPRE tube) into the pit wall, isolating the sample and weighing it [Goodison et al., 1981; Avalanche Research Centre, 1981]. This technique is labor intensive and slow [Dexter, 1986] and is not feasible for continuous sampling of deep pits. Wedge-shaped cutters have the advantage of extracting a sample without requiring additional excavation, which makes them a great deal faster. The only commercially-available, wedge-shaped cutter at the time worked well in low-density, continental snowpacks, but its large surface-area-to-volume ratio of 1.44 leads to edge effect errors, and the thin fabrication material deforms or is destroyed in dense snowpacks. Edge effects are important because they may be the largest source of error in density calculations. When a cutter is inserted, the edge isolates the sample by breaking the bonds or the grains separating it from the rest of the snowpack. The cutter edge may force the broken grains into the sampler or exclude them from the sample. This may result in under- or over-sampling. This error source is unquantifiable, but as the volume of the sample increases, this error diminishes. The surface-area-to-volume ratio is important because as the ratio decreases, the error because of edge effect decreases. Another widely used method is the Mount Rose (or Federal) sampler. It extracts a core of snow equal to the entire depth of the snowpack, thus only a mean density of the profile is obtained. The accuracy of this instrument has also been questioned although it is sufficient for its purpose as an index tool [Work et al., 1965].

The cutter instrument and measurement technique we adopted was developed by R. Perla of Environment Canada. The sampler is a wedge-shaped cutter 20-cm long, 10-cm wide, and 10-cm high giving a 1000-mL volume. To minimize deformation, 14 gage stainless steel was used. Occasionally it was necessary to pound the cutter into the pit wall with a rubber mallet and a less robust design would not withstand this treatment. The large volume gives a reasonable surface-area-to-volume ratio of 0.72, that minimizes under- or over-sampling. Edges are beveled to further reduce the edge effect. All cutters were carefully calibrated and found to have less than 1% volume error. In spite of the small error, the calculated correction was applied to all data. Continuous density profiles were taken in 10 cm increments in each pit, and dual profiles were taken when time allowed. The 1000-mL cutter was used in conjunction with a top-loading digital scale with an accuracy of ±1 g. The sample was weighed in the cutter because the tare weight of the cutter was removed before sampling, making the process fast, precise, and accurate. A 6-m pit could be sampled for density in approximately 1 hour with one person sampling and the other recording the data (not including excavation time). All density measurements taken after 3 May, 1986 used this technique. We continued to use the PVC tubes for chemistry samples because they proved to be the best sampler.

3. Ablation

Ablation measurements were made at intervals of four to seven days. Sampling strategies varied between the three years. In the 1986 season, 50 stakes were subjectively located for measurement convenience throughout the 120 hectare basin. The bias created by subjectively choosing points was removed in the 1987 season when locations were chosen randomly from a 25 m square grid map. This sampling design provided sufficient ablation data for all elevations, slopes, and aspects in the basin. A dense network is necessary in rough alpine terrain where factors affecting snowmelt vary greatly over short distances. On alpine glaciers, a density of 10 stakes per km^2 was found to be adequate to measure net ablation with an accuracy of better than ±5 percent [Hoinkes and Rudolph, 1962]. The density of our network was about five times greater than in the study mentioned above. In Wyoming, researchers had to increase sample size as variability in ablation increased with decreasing snow cover [Bartos and Rechard, 1974].

Development of the water balance for Emerald Lake required estimates of snowmelt throughout the basin. Amounts of melt were expected to vary to a large degree around the catchment due to variability in exposure and shading. Estimation of the spatial and temporal variability of snowmelt requires snowpack data representative of each type of terrain within a drainage basin obtained at frequent intervals. However, detailed snowpack measurements can only be made at a limited number of sites. Ablation stakes provided an alternative means of measuring snowmelt quickly at a large number of points. Ablation stakes are simply a narrow rod or tube inserted vertically into the snowpack to provide a reference for measurements of surface lowering. Snow depth stakes extending through the snowpack and touching or attached to the ground surface offer a better reference, but they cannot be used in many snow climates. Ablation stakes have logistical advantages where seasonal snow depth exceeds 2 m and where creep and glide would distort or damage snow depth stakes. Ablation stakes have been used extensively in studies of glacial mass balance [e.g., LaChapelle, 1959] where the ground surface may be far below the ice surface. A study in Wyoming found that the local ablation determined from stake measurements was within 2 percent of that determined with a more precise surveying technique [Bartos and Rechard, 1974].

Ablation stakes can be made of almost any material that does not result in excessive reradiation melting of the surrounding snow. We found white, thin-wall (Sch. 125) PVC pipe to be well suited for this use. After snowpack settling has decreased to a low rate in spring, ablation stakes were installed to a depth of at least one meter in the vertical direction (not perpendicular to the slope). The height that the stake extends above the snow surface was measured after installation and then at intervals of a few days. Because the snow surface is irregular and a small depression will melt out next to the stake, the height to be measured must incorporate some average of the surface immediately around the stake. Our measurement protocol involved laying a meter stick across the slope just upslope of the stake as a means of objectively and consistently averaging the local snow surface. In theory, the rate of lowering of the peaks of the microtopography (points in contact with the straight edge) may differ from that of the overall surface [Muller and Keeler, 1969]. However, over repeated intervals of a few days each, this difference should be negligible.

The change in snow surface height over the measurement interval was corrected for settling and multiplied by the density of the ablated snow. The amount of settling can be determined by comparison of several stakes of varying length at one site [LaChapelle, 1959]. In this method, differences in height of each stake over the interval are plotted against depth of each stake. If settlement is uniform with depth, greater settlement should be associated with longer stakes. The net change in snow surface height can then be determined by extending the plotted line back to a stake depth of zero [LaChapelle, 1959]. We did not find any consistent differences in ablation measured with ablation stakes of 1.5 m length and ablation measured with 3 m stakes. Alternatively, the product of the proportional change in density and the change in height should be subtracted from the change in height. The corrected change in height can then be multiplied by the mean density of the layer for that interval to obtain the change in snow water equivalent. The density value used for calculations is perhaps best approximated by measuring the density of the upper 0.5 m of snow on the days of stake measurements.

4. Snow Covered Area

Snow covered area was estimated from many oblique photographs obtained during the surveys and throughout the melt season. The nature of the basin topography allowed us to get adequate views of nearly the entire watershed from opposing ridges. Aerial overflights were also effective, but we experienced considerable cloud cover during most of the surveys in 1987.

D. Modeling Methods

The basin was classified into areas of similar snow characteristics using a 5 m grid from the DEM. The large number of grid points (48048) in the basin made it necessary to employ image processing techniques for analysis and classification. The basin was divided into regions in a two step process. First, a random sample of 1000 points was drawn from the 5m DEM. The corresponding values of radiation, slope, and elevation were clustered to identify the structure of similar groups within the basin. The entire basin was then classified using a Bayesian classifier based on the statistics generated from the clustered subimages [Richards, 1986]. This technique was repeated for several variations of the parameters and for two different numbers of classes (8 and 12). These numbers were arrived at as a compromise between being operationally and computationally small enough and still providing adequate resolution and information. The actual number of classes varied since the classifier omitted some classes identified by the clustering algorithm. The combinations are listed in Table 2, and for simplicity, acronyms are used for the stratifications hereafter. The acronyms include initials for each parameter used and a number indicating the number of classes (e.g. RSE12 represents radiation, slope and elevation with 12 classes).

Slope and radiation images were generated using Image Processing Workbench software [Frew and Dozier, 1986]. The methods used to calculate net radiation are described by Dozier [1980]. These three parameters were chosen because they represent physically based parameters that affect accumulation and ablation of snow. Slope and elevation are fixed in time for the purposes of this study, but radiation varies markedly through the seasons and provides the time- dependent element needed to model the change in the distribution of SWE over the basin. This is intuitively clear if one thinks of lingering frost, snow or ice patches that melt as the sun reaches them, either on a diurnal or seasonal scale. The net radiation images used in the classification are indices, where the daily net radiation was calculated for a clear sky (a condition that persists in the Sierra) for the 15th of each month from December through June. These were then summed for all months prior to the survey date for which they were used. Two radiation images, early and late season, are shown in Figure 2; examination shows the marked increase in radiation through time, particularly on the west wall of the basin. All slopes greater than a critical amount (60^o for 1986, and 55^o for 1987) were assigned a zero value for SWE to take into account the persistently bare cliffs in the basin. The areal extent of these zones matched snow-free areas in early season aerial photographs. A mask of snow covered area was made from aerial and oblique photographs for the 1988 survey results, and this included the persistently bare steep slopes.

E. Results and Discussion

1. Accumulation

The following is a summary of the winter precipitation and the snowpack that developed in the 1986, 1987, and 1988 water years. Depth, density, and SWE data collected from storms and snowpits in the Emerald Lake basin are presented. Data from Lodgepole, Kaweah basin, and the Tulare region are used as an index of the seasonal snowfall and are also discussed. Examination of these supplementary data serves two purposes: (1) it indicates whether the Emerald Lake data are representative of a larger area than the small basin itself, and (2) the regional data indicate that systematic errors are unlikely in the Emerald Lake data since the trends there corroborate the regional condition. It is important to determine the representativeness of the Emerald Lake data. These indices are also valuable in looking at long-term trends because both the Kaweah and Lodgepole records exceed 50 years. "Normal" in the following discussion refers to the mean value from these 50 year records.

a. Emerald Lake Basin and Regional Snowfall Event Summary

Precipitation data from the California Cooperative Snow Survey (CCSS) has served as an index of snowpack conditions in California for many years. CCSS data do not exist for Emerald Lake, but precipitation data for the Tulare Region and Kaweah Basin subregion, of which Emerald Lake is a part, can be used as an index for regional precipitation. The Lodgepole ranger station also has a long record and is worth examining as the geographically closest index to Emerald Lake basin precipitation trends. The Lodgepole gage is located about 5 km ESE and 470 m below Emerald Lake. The precipitation data are collected from precipitation gages as it accumulates. Cumulative precipitation is the sum of the individual precipitation events.

1986 Water Year. The 1986 water year was characterized by heavy precipitation throughout the Sierra Nevada. Estimates of the accumulation were about 150% of the 50 year mean [California Cooperative Snow Survey, 1986]. The Tulare region and Kaweah basin were both substantially above the 50 year mean on the date of peak accumulation, showing about

150% and 180% normal, respectively (Table 3). Precipitation at the Lodgepole ranger station (Table 4) also indicated a much greater than normal precipitation year with 180% of the normal cumulative precipitation on April 1.

Accumulation was monitored throughout the season at Emerald Lake using snowboards, which were sampled as close to the end of each storm as possible. Total accumulation at the inlet site at Emerald Lake was approximately 2.65 m SWE. Results are summarized in Table 5. October storms produced little notable accumulation. The first major event occurred in the second week of November, followed a week later by another storm with similar deposition. December storms produced less than 0.5 m SWE, and January had only one storm, which brought the cumulative precipitation close to 1 m SWE. In mid February, a large storm deposited over 1 m SWE over the entire basin. This storm buried all the snowboards and measurements were taken only at the inlet for the remainder of the season. Following one more notable precipitation event in March, which deposited nearly 0.5 m SWE, no significant deposition occurred through the rest of the season. The overburden from deep accumulation, coupled with the warm temperatures, produced a deep, high density, snowpack.

Table 5 shows early season variability in accumulation between the inlet, pond, and ridge sites. This difference may be genuine, because of variable precipitation in the basin, or it may be due, in part, to sampling. A snowboard in one part of the basin may be scoured by wind before it is sampled while another location is left untouched. As the season progresses, between-site variability diminishes. By the second week in January, cumulative differences in all three sites have smoothed out, varying by a maximum of 4% on January 8 and by less than 5% by February 3.

1987 Water Year. This year was marked by lower than normal precipitation. Statewide precipitation was 65% of the 50 year mean for the 1987 water year, and estimates for the Sierra Nevada were even lower [California Cooperative Snow Survey, 1987]. Data from the Tulare River basin (Table 3) showed that by April 1, the cumulative precipitation had only reached 70% of the 50-year mean, and the Kaweah basin attained only 55%. Statewide snow surveys indicated that the snowpack was just over 50% of normal for April 1 and 20% of normal for May 1. This illustrates the relationship between low precipitation and rapid depletion of the thin snowpack found in a dry year. Lodgepole data showed a similar trend with only about 50% of the 50 year mean at maximum accumulation (Table 6).

Snowboards were sampled at two locations near the inlet after each storm and results are listed in Table 7. The values are means of the measurements from the two sites. The first measurable deposition fell after January 1 and only deposited 0.11 m SWE. Four more storms of similar magnitude followed, leaving a cumulative SWE of 0.49 m on March 9.

1988 Water Year. The 1988 water year proved to be very similar to 1987. Early storms appeared to be following a normal trend, but a dry period that began in mid January and lasted through most of February put the cumulative precipitation well below normal. At the end of January, the Tulare basin was at 110% of the 50 year precipitation mean (Table 3), but one month later the record showed only 85%. By the end of March the estimate had dropped to 70%, with forecasters calling it another critically dry year. Records show that by

April 1, the cumulative precipitation had only reached 85% of the 50-year mean, for the Kaweah basin as well [California Cooperative Snow Survey, 1988]. April storms helped the problem to some degree by temporarily stopping melt and adding a small amount to the snowpack, which brought the regional precipitation up to 83% of the long-term mean for the date. The seasonal snowpack is usually about at 70% of the seasonal maximum on May 1, but this year it was only about 20% [California Cooperative Snow Survey, 1988]. The observed distribution and volume of snow in the Emerald Lake basin was very similar in the 1987 and 1988 water years. The April storms caused differences in the snowmelt and runoff scenario, but changed the time table only, delaying the melt in 1988. Snowboards were not monitored in the basin in 1988. The precipitation data from Lodgepole was not available from CCSS at the time of publication.

2. Redistribution

There is visible evidence for snow redistribution in the Emerald Lake watershed. Large cornices form on the uppermost ridges and generally face into the basin. These may be formed during southeasterly storms (the basin faces north) or may be the result of considerable erosion and scouring of the hillslopes outside the basin. Other large storms come from the northwest and travel up the basin, leaving large upslope drifts on the pronounced benches. These drifts account for a significant amount of deposition and are present in this particular watershed, both in years of high and low precipitation. Similar deposits have been observed in an alpine basin by Weir [1979] in New Zealand.

Storms in the Sierra Nevada are usually associated with air temperatures near the melting point. At these high temperatures equilibrium metamorphic processes are rapid and result in a strong well-bonded surface. Snow deposited in the few trees in the basin quickly melts, and surface hoar is an anomaly here. Thus the only likely source for impact crystals is newly precipitated snow. During and immediately following a storm, loose snow may be easily moved and even disaggregate the old snow surface, incorporating dislodged crystals into the redistribution. Once the surface develops, little snow movement takes place even in high winds, and the majority of snow loss under these conditions is from sublimation. For these reasons, most redistribution occurs during or immediately following the precipitation event.

Many of the snow patches that persist for the longest period into the melt season in the Emerald Lake watershed are avalanche deposits or snowbanks found at the foot of steep cliffs fed by sloughing from above. The February 1986 storm produced large accumulation over a short period of time and led to an avalanche cycle, which moved a large portion of snow from mid-elevation in the basin to the lake surface. Depths of drifts and avalanche deposits during the 1986 season sometimes exceeded 10 m, and sloughing from steep rock faces produced many depths exceeding 8 m. With the exception of cornices and drifts, snow deposition smooths the Emerald Lake basin features. This smoothing happens on a small scale, including talus and small boulders, and on a larger scale, including gullies, depressions and large boulders. All but the largest boulders were obscured completely during the 1986 winter. Effects of redistribution, in the form of drifts and deflation areas, were found in the same locations all three years in spite of radical differences in precipitation.

3. Snow Density from Field Measurements

Three different approaches were used in the 1986, 1987, and 1988 melt seasons to calculate mean snow densities for the basin. Changes were necessary because of the large difference in accumulation between 1986 and the two following seasons and because of the two different measurement techniques used during the field study.

1986 Water Year. The 1986 water year produced a deep snowpack; the heavy overburden and warm temperatures caused a relatively high, uniform density over the basin. Density did not vary appreciably through the entire melt season with a mean of $520\,kg\,m^{-3}$ and a standard deviation of $44.0\,kg\,m^{-3}$, less than 10% of the mean. The small deviation allowed us to apply the mean density value to all depths to obtain SWE estimates. No attempt was made to fit a time-related function to the density, because the deep snow constrained us to a small number of pits and the data were not sufficient for anything other than a seasonal mean. Data from all the 1986 pits are summarized in Table 8.

1987 Water Year. Lower snow accumulation during the 1987 water year allowed us to dig over fifty pits, giving us excellent density data with high spatial and temporal resolution. In the absence of strong temperature and vapor pressure gradients in the snowpack, snow density increases throughout the season from overburden pressure and mixed metamorphic processes. Mean density showed an increase from February through June. Early season densities were low, corresponding to the low temperatures and thin snowpack. As temperatures increased and accumulation proceeded, mean density increased asymptotically to about $470\,kg\,m^{-3}$. This value was considerably lower than the previous year's $520\,kg\,m^{-3}$, because of the decreased overburden. The increase was rapid during the early part of the melt season and slowed as the snowpack ripened. The data were nonstationary with a changing mean through the season, large variance in early season measurements, and relatively small variance after melt began.

Data from all the pits dug in 1987 are listed in Table 9, where values represent means of several profiles in some cases, and are summarized in Table 10, where values represent all density measurements taken over the entire basin within 1 day of the given date. From the date of the first survey forward, the standard deviation of the density throughout the basin was less than 10% of the mean in all cases except one, when it reached 11%. A linear model was fitted by simple correlation of the data after April 1, where mean density ($\bar{\rho}$) in $kg\,m^{-3}$ was a function of day of year (D) in the following form:

$$\bar{\rho} = 1.028\,D + 315.9 \qquad r = 0.406 \tag{3}$$

The correlation was tested using a t test (Zar, 1984, p. 309) and was shown to have a correlation coefficient different than zero at the 99.9% confidence level. Values of predicted density from Eq. 3 are also listed in Table 10. The data show a consistent increase in time, with two exceptions late in the melt season. These decreases probably result from relocation of the pits or experimental error, rather than a basin-wide decrease in density. The predicted values are close to the observed values and are within one standard deviation of the observed means except for the week of 27 May. The snowpit in the cirque was moved during this week and anomalously low densities were recorded at the new location. The predicted mean density for each survey was calculated using Eq. 3 and the mean date of the survey. The results (Table 11) vary less than 5% from the observed values for the closest date in Table 10. These values were then used to calculate SWE from the survey depth data.

1988 Water Year. The same pit locations were used during the 1988 field season to measure density. We were again able to dig many pits (58), giving us data with high spatial and temporal resolution. Data from all the pits dug in 1988 are listed in Table 12. The values represent means from several profiles in most cases.

Early season densities were low, corresponding to the low temperatures and thin snowpack. As temperatures increased and accumulation proceeded, mean density increased asymptotically to about $510\,kg\,m^{-3}$. The increase was rapid during the early part of the melt season and slowed as the snowpack ripened. In an effort to improve our basin-wide SWE estimates from the surveys, we examined the local variation and temporal change in density for each of the pit sites. Enough data existed at each site, except the pond, to fit a linear model to the data and have predicted density as a function of date and location. This is important because the basin does not behave similarly on a temporal or spatial scale. In some instances, the east side of the basin is producing runoff while the west side is still subfreezing in some portions. The variable energy inputs caused by the rough terrain lead to this heterogeneity and one manifestation is the spatially variable density patterns for any given date. The commonly observed temporal change is for the density at a point to remain relatively low (300–350 $kg\,m^{-3}$) until the energy inputs become large enough to cause a rapid warming of the snowpack and a resultant rapid increase in density. Once the densification reaches a critical value (500–550 $kg\,m^{-3}$), the process slows considerably and approaches a maximum asymptotically. The period of accelerated densification may last several weeks to several months depending on the local topography and meteorology. Clearly inclement weather associated with low temperatures retards the process, while clear weather or inclement, warm weather (e. g., rain-on-snow events) will accelerate the densification process.

The spatial and temporal variability of densification makes it desirable to evaluate the process for all areas in the basin. The best approach to solving this problem is to model each density data site separately as a function of date and interpolate the results over the basin. This approach was taken for the 1988 water year and the peak SWE estimates reported herein are a result of this treatment. Five linear equations were derived by linear regression to model density as a function of date for the inlet, bench, ramp, hole, and cirque. An indicator function set density to the maximum observed value for the location after the date where observations showed it to be approaching the asymptote. Scant data forced us to use the bench relationship on the pond site. Such an assumption is not bad one because the two sites do not differ radically in elevation or aspect and available data shows similar densification trends. Predicted density from these equations was interpolated over the entire basin for the peak accumulation survey date. The depth survey points were then registered to the interpolated density image and SWE calculated for each point.

4. Estimations of Snow Water Equivalence from Survey Data

It is possible to estimate the basin SWE ($\overline{SWE}$) simply from the mean of the point SWE values obtained from the snow surveys when the sample size is large enough. The snow depth sample size during the 1986 water year exceeded 125 in all but the first survey, where it was 86, ranged between 256 and 328 during the 1987 water year, and was 354 in 1988. Each sample point represents the mean of five measurements. A statistical summary of the depth measurements is found in Table 13. $\overline{SWE}$ was calculated for each survey using the mean depth and density calculated as discussed above. Snow-covered area was implicitly accounted for in the calculations because the survey points without snow were averaged into the mean snow depth. With a large, randomly located sample, this procedure should be sufficient. The values of $\overline{SWE}$ calculated from this method were used to evaluate the results of $\overline{SWE}$ based on the classification of the basin by terrain features discussed below and will be referred to as the "expected" values.

$\overline{SWE}$ was also calculated using Thiessen polygons following the algorithm presented by Renka [1984]. This technique produced similar results to the simple arithmetic mean and all estimates except one were within 7% of the expected value (Table 14). Thiessen polygons and other areal interpolation techniques fail to account for the abrupt changes in SWE dictated by abrupt changes in the terrain and produce a smooth snow distribution over the basin. It was hoped that classification of terrain into similar zones of SWE on a 5 m scale would improve this problem.

1986 Water Year. $\overline{SWE}$ was obtained by multiplying the mean snow depth for each survey date by the seasonal mean density ($520\,kg\,m^{-3}$). Total volume of water stored in the basin was calculated by multiplying the total basin area by $\overline{SWE}$ and results from all four surveys are listed in Table 14.

The first survey covered only the northeast wall of the basin. The following three surveys encompassed the entire basin, and the distribution of points from the second survey is typical of the 1986 surveys. Some justification is necessary for using snow depths from the subsample of the basin obtained in the first survey to represent the entire basin. The effect of radiation on the spatial distribution of snow in the basin appears to be negligible early in the season. At this time the energy flux is small and its difference between points within the basin is minimal. The spatial variation of SWE is also at a minimum. As the season progresses and the sun angle changes, some portions of the basin receive a great deal more energy than others. Significant ablation may take place in some areas of the basin before melt is initiated in others, leading to increased variation in SWE. To characterize this difference through time, an analysis of variance (ANOVA) was used to test the means of depth from the northeast wall against the means from the entire basin for the second, third, and fourth surveys. The null hypothesis (H_o) is: there is no difference between the mean depth for the northeast wall and the mean depth of the entire basin. An F test showed, at the 95% confidence level, that there was no difference in $\overline{SWE}$ for the locations in either the second or third survey, and that the means for the subarea fell well within the 95% confidence intervals for the entire basin. The null hypothesis was rejected for the fourth survey, indicating that the northeast wall was no longer representative of the entire basin.

The mean and standard deviation of the three surveys can be found in Table 15 and the results from the F tests are listed in Table 16. The mean from this survey was applied to the entire basin to calculate the values for the first survey in Table 14. This result is important because this survey was completed close to the date of maximum accumulation and preceded any significant ablation, and it was possible to obtain a reliable estimate of basin $\overline{SWE}$ for the first survey date using the data available from the northeast wall.

1987 Water Year. Four surveys were completed during the 1987 water year, each of these encompassing the entire basin. Sample point locations were randomly located on a 25 m grid registered to the DEM. The first survey was completed shortly after the date of peak accumulation in the basin. Discharge measurements showed that some melt water generation and runoff occurred prior to the first survey and this difference had to be accounted for to get an accurate estimate of peak accumulation for the water year. The volume of runoff before the survey was equal to 5.2 cm SWE and sublimation from the snowpack was calculated to be 2.4 cm SWE. Thus, an accurate estimate of peak SWE can be made by adding 7.6 cm SWE to the first survey estimate over the entire basin, which gives a total of 67.4 cm SWE at peak accumulation. Once melt began, the relatively thin snowpack diminished quickly, making it necessary to complete surveys on approximately two week intervals. $\overline{SWE}$ was obtained by multiplying density by the mean snow depth from each survey.

1988 Water Year. A survey was completed during the 1988 water year, as close to the date of maximum snow accumulation as possible. Sample locations were chosen using a randomly located grid scheme. First, a 25 m square grid was placed on the DEM using a randomly selected starting point. Points that fell on a 100 m square grid coincident with the initial point were then selected. Approximately 115 points were obtained in this manner. Another 247 points were randomly selected from a 25 m square grid also coincident with the original point. The total number of points chosen was 362. The total number of points was based on an estimate of what the field team could reasonably accomplish. As described above, those points in the field that could not be reached and could not be verified to be snow free were discarded. The field effort was successful and obtained 354 measurements from the 362 possible points. Figure 3 shows the distribution and number of points sampled for snow depth. There was little runoff from the basin prior to the survey date and most of the change between the date of maximum accumulation and the survey date were in densification of the snowpack on the eastern side of the basin; thus estimates of SWE at this time do not differ appreciably from the peak value for this season. SWE was calculated using the density, modeled for six pit locations and interpolated over the basin, multiplied by the depth from the survey points.

Due to the relatively large sample sizes and the field techniques employed, we have confidence in the values presented in Table 14, but these large surveys are seldom practical and new techniques must be developed to obtain a similar degree of accuracy with a reasonable time, manpower, and economic investment.

5. Snowmelt From Ablation Stakes

In 1986, ablation measurements began May 13 and continued through August. The above normal snow accumulation of water year 1986 allowed the long observation period. Data were analyzed through the end of July because there were few remaining points in the month of August. By the time measurement was initiated, the density was similar throughout the basin and settlement resulting from further densification was minimal or nonexistent. All measurements of change in depth at the ablation stakes were multiplied by a mean basin density of 520 $kg\ m^{-3}$ to obtain the snow water equivalence that had melted or sublimated.

During the 1987 water year, measurements began on April 19 and continued through June 8. The shorter observation period resulted from the much reduced snowpack, which was essentially gone by the end of June. Snow density was still increasing in the basin over the observation time period. Temporal variability made it necessary to calculate change in SWE using a simple linear model for density mentioned above. Over the measurement intervals, ablation varied from less than 1 cm of water equivalence per day in April to about 3 cm per day in July. Measured ablation varied with weather conditions and exposure to insolation as expected. The highest ablation rates were noted at stakes surrounded by exposed boulders.

Ablation images were created by interpolating the point measurements of change in SWE over the entire basin area using a Thiessen polygon algorithm [Renka, 1984]. The mean change in SWE over the entire basin for each observation period was then calculated by summing the the change in each 25 m^2 pixel and dividing by the basin area. This result represents the expected change in SWE if the entire basin were snow covered, so each value was multiplied by the percent snow covered area in the basin for the date to obtain actual change in SWE. The change in SWE for each time period could be used as a check on our snowmelt and runoff estimates for the same time periods. The ablation stake data appears to provide a better relative estimate of snowmelt than an absolute one. However, there was poor correspondence between snowmelt estimated from the ablation data and snowmelt estimated from the snow survey data (Table 44). The reasons for these discrepancies are presently unknown.

At Emerald Lake, ablation stakes were an easy means of acquiring spatially distributed data throughout three snowmelt seasons. These results showed which portions of the basin were generating snowmelt at various rates throughout the season. The ablation data can be combined with physical parameters calculated from a digital elevation model such as slope, exposure, elevation, and net radiation. This combined information should help quantify some of the effects of terrain-related factors on snowmelt.

6. Results of Modeling Attempts

a. Validation of Results

Clustering and classification are not rigorous statistical techniques, and formal statistical approaches for validating results do not yet exist. The results in this study have been evaluated qualitatively by our intimate knowledge of the basin and the observed snow distribution, and quantitatively by two methods. First, a single classification ANOVA was used where the null hypotheses for the ANOVA was stated as: there is no difference between the means of the groups identified in the classification. If the null hypothesis is accepted, similar information can be found in more than one class and a poor classification has resulted. Rejection of the null hypothesis shows that the classes contain different information or represent different populations, which is the desired result. Second, standard errors (SE) from the classifications were compared to the basin-wide data. In any classification attempt the SE should be reduced for the classified groups when compared to the whole data set, but a significant reduction in SE suggests a successful classification.

All data were checked for the assumption of a normal distribution. The data for radiation, slope, and elevation were close to normal with no hope for improvement through transformation. SWE data were normally distributed except for the many zeros. This was partly taken care of by masking the steep snow-free areas in the basin and removing them from statistical analysis.

b. Analysis of Parameter vs. SWE Classifications

1986 Water Year. The results from 1986 were unexpectedly poor. Scatter plots of SWE against radiation, slope, and elevation showed no discernible relationship. All attempts to cluster and classify the basin into zones of similar SWE distribution produced unacceptable results. Two different methods were used. First, the radiation, slope, and elevation images were subsampled and then clustered; the results were then used in classification attempts described above. In addition, the points for which there existed SWE data were grouped into classes of similar SWE and the basin was classified based on the terrain and radiation parameters for these points. Combinations of the parameters and group sizes were explored for both techniques without producing any viable results.

There are several explanations for the disappointing results. The orthophotograph described earlier was not yet available during the field season, and lower quality oblique photographs were used by the field teams to locate the sample points in the field. The points were subsequently transferred to the orthophotograph and the corresponding UTM coordinates and DEM values were used in the analysis. There is considerable room for error in both the use of the oblique photographs and the post-survey transfer to the orthophotograph. The large volume of snow encountered on that year may account for some problems. A deep snowpack dictated decreased sample sizes and in some cases field crews could not reach the base of the snowpack due to icing of the probe or depths exceeding 10 m. When this occurred, the maximum depth of penetration was recorded by the field crew resulting in substantial undersampling in some cases. It may also be that the deep snow reduces the effects by terrain features on the distribution patterns of snow as previously discussed. What we do not know is at what level this becomes important. The large snowfall also made location within the basin more difficult because many of the distinguishing features in the basin were completely obscured. Our best estimates of basin SWE for this year remain the estimates described earlier where the mean depth was multiplied by mean density and basin area. Fortunately the 1987 and 1988 water year data proved to be much better.

1987 Water Year. Scatter plots of radiation, slope, and elevation against SWE for the surveys showed that the relationship between SWE and radiation is the strongest, but they were all rather weak. Stepwise linear regression supported the weak relationship. Radiation and slope together accounted for 40% of the observed variation and inclusion of elevation made negligible improvement of less than 1%. This is contrary to the SE results, which follow and show that elevation may be important. ANOVA results in Table 17 are highly significant for all classifications. Examination of the various classifications shows some inadequacies. Only the best classification image has been displayed for each of the four surveys in 1987 and the one survey in 1988. The term "best" is somewhat subjective since it is based on qualitative comparisons to field observations, as well as the ANOVA and standard error tests described above. In some cases it was not clear which image classification was superior. Groups with similar SWE were combined into a distinguishable number of classes to ease visual interpretation.

RSE12 placed the maximum accumulation below steep slopes on the west side, upper benches, and cirque. RSE8 has the desirable attributes of RSE12, but shows greater definition and more realistic distribution on the east wall of the basin (Figure 4). RS12 places a uniform snowpack over the entire east wall, which has not been observed, and places relatively deep deposits low in the basin. RS8 showed an increased SE over the random survey SE, which indicates that a poor classification resulted and the image supported this conclusion, being too chaotic to evaluate. RE12 appears to be reasonable, but may be too simplistic with large homogeneous regions. RE8 is worse with no definition in several areas. Results of the predicted basin SWE from each classification are listed in Table 18. These values were calculated by multiplying the zone mean SWE by the zone area and summing all the zone values for the basin. All values were lower than the expected basin mean described earlier, but by less than 5%.

All ANOVA results for the second survey (Table 19) were significant at greater than the 99% level except for RE8 (96%). RSE12 and RSE8 appear to incorrectly locate the maximum SWE at low elevation, which observations indicate should be located beneath the steep cliffs in the upper reaches of the basin where sloughs accumulate and radiation is low. RS12 and RS8 are good approximations of observed SWE distribution. RS8 does the best job of locating the maximum SWE deposits and differentiates between the east and west wall in a realistic manner. RE12 and RE8 are again, too coarse to be useful. Only classifications including a slope parameter improved the SE. Stepwise regression again shows radiation and slope to be the most important, however, the R^2 improves substantially (from 0.36 to 0.46) when elevation is included. Table 20 indicates that all classifications overpredicted SWE except RE8. RSE12 and RE12 overpredicted SWE by more than 5%.

Results from the third survey ANOVAs showed several of the classification attempts to be poor (Table 21), and only RSE8, RE12, and RE8 were significant. RSE8 locates snow deposits fairly well in the higher elevations of the basin, but does not show the observed difference between the east and west walls and does not point to the large difference seen in the west joint, which stores little snow by this date. RE12 and RE8 show little spatial variation and are clearly too elevation dependent. RSE8, RE12, and RE8 improved the SE. All SWE estimates were within 5% of the basin mean except RS12 (Table 22). Stepwise regression showed the importance of elevation as the R^2 improved from 0.23 with radiation and slope, to 0.35 with elevation.

ANOVA results for the fourth survey were highly significant only for RS12 and RE8 (Table 23). None of the classifications, except RS12, differentiate between the west and east aspects of the basin, which is a result of the radiation balance becoming more uniform over the basin by this late date. RS12 does separate the subtle differences in aspects. Again, stepwise regression showed the importance in all three parameters. Only RS12 over-predicts basin SWE by more than 5% (Table 24).

1988 Water Year. Results from the peak accumulation survey in 1988 are similar to those of 1987, probably partially because of the similarity in the snowpacks. The correlation coefficients between SWE and radiation, slope, and elevation are not as good as the 1987 results, but the classifications appear to be better. Stepwise regression showed radiation to be the most import variable relating to SWE, explaining 16% of the observed variance. Elevation was again important as the R^2 improved from 0.22 with radiation and slope, to 0.27 with elevation.

All the classification attempts produced reasonable results. RSE12 located the maximum SWE deposits correctly under the steep north-facing walls on the upper benches and in the cirque, but was inadequate in differentiating between the east and west walls lower in the basin. RSE8 was similar to RSE12, but homogenized even more of the basin. RS12 appeared to be the best classification and is shown in Figure 5. Maximum SWE is correctly located on the upper benches and in the cirque. Significant deposits also lie on the east wall as drifts on the benches, and at lower elevations under the cliffs on the west wall. RS8 places the maximum deposits well, but simplifies the rest of the basin into a homogeneous snow cover. Both RE12 and RE8 over-simplify the entire basin and place maximum SWE deposits away from the cliffs and at low altitudes.

The single factor ANOVA tests showed that all classification attempts were significant at the 95% level, with RSE12, RE12, and RE8 significant at the 99% confidence level. The standard error of the mean from the classified groups improved the random standard error considerably, but consistently, such that no single attempt was proven superior by this test. Results from both tests are summarized in Table 25. Total basin water volume differed from the expected volume by very little in all cases (Table 26).

7. Design of Optimal Surveys

One objective of this study was to find an optimal method for surveying snow water equivalence. It is seldom practical to conduct high resolution field surveys such as those used in this study because of time, money, and personnel constraints. The large data set collected in this study provides good estimates of the mean snow water equivalence found in each zone from the classification attempts. Using these data and standard statistical procedures we can estimate the number of samples required to be certain, at a specified confidence level, that our estimate of the sample mean represents the true population mean. An estimate of sample size required to be within a given measurement error of zone $\overline{SWE}$ was made for each survey and all classification attempts using the formulation presented by Bhattacharyya and Johnson [1977, p. 273], where the required sample size is determined by the sample standard deviation, a desired confidence level, and a desired absolute error in the measurement. An error of ±5 cm was used with required sample size calculated for 95% and 99% confidence levels.

In most cases an unreasonable number of measurements would be necessary to obtain the desired level of confidence in the estimate of $\overline{SWE}$ if all the zones were to be sampled to the degree recommended for a minimum sample. The small zones, in terms of areal extent, account for the majority of required samples because they had few field measurements in them and have the greatest error in estimation of the zone $\overline{SWE}$. However, these areas account for a small percentage of total basin area and total stored water, which means that errors in the basin $\overline{SWE}$ are relatively insensitive to errors in these group means. Eliminating such areas from a sampling scheme allows us to complete a survey in a reasonable amount of time and makes very little difference in the accuracy of the estimate for basin $\overline{SWE}$. A further benefit in eliminating the smallest zones comes from the fact that these are the most difficult zones to locate in the basin when the field data are being collected. The larger zones can be sampled with a great amount of confidence that the sample point is actually within the zone boundaries.

To test the effect of selectively eliminating small zones, the zones that held less than 5 and 10 field measurements were eliminated from the calculation of basin SWE in all classification attempts from all surveys in 1987 and 1988, giving 30 test cases. The area they covered was given the mean snow water equivalence of all the remaining field measurements and then added to the basin SWE. This value was compared to the value that was estimated using all field measurements and all zones. Elimination of these small zones made little difference in most cases. Results are listed in Table 27. The percent difference in basin SWE when zones having less than five field measurements were removed was less than ±1% in 80% of the cases, and never exceeded ±5%. The maximum area covered by the discarded areas was 3.8% of the basin. When zones having less than 10 field measurements were removed, 47% of the volume estimates were ±1% or less, five were greater than ± 5%, and 2 were greater than ±10% of the total basin SWE. Over 73% of these cases involved discarding total zone areas of less than 5% of the basin area. The few large volume differences corresponded to zones that covered larger areas of the basin, up to 12.5% in one extreme case.

Removing these small zones from a field data collection effort dramatically decreases the number of required samples to characterize basin SWE in most cases, with very little compromise in the estimate of basin SWE. Table 28 shows the required sample size for all 30 cases. Table 29 shows details of the determination of the required sample sizes for one example and similar results were found for most of the surveys and classification attempts. The dependence of required sample size on both standard error and original sample size is apparent. The required sample size to characterize the entire classified basin with a stratified random sample and an error of ±5 cm at the 95% and 99% confidence levels is 85 and 144, respectively. In a rugged alpine basin these numbers represent a minimum of two to three days work for an field team of two persons to measure snow depth alone. However, if the small zones requiring the greatest amount of samples are removed from the sampling scheme, the required sample size drops significantly to 22 and 36 for the 95% and 99%

confidence levels, respectively. These sample sizes could be collected by an experienced team in a single day. Note that spatial autocorrelation of *SWE* was not accounted for, therefore, the estimates of required sample size are probably high and may be thought of as an upper bound. According to Table 29, zone 1 only requires a sample size of 1. This is partly a function of the large number of field samples that fell in this zone when the classification was carried out. In reality one would want to sample this zone more intensively. By eliminating the small zones the effort can be transferred to the larger zones to insure a good estimate. From the example in Table 29 it may be desirable to eliminate zone 3, which covers less than 1% of the basin, and transfer the sampling effort to zone 1.

The results show that terrain features and radiation exert some effect on snow distribution. We have shown that slope and elevation may be used as static terrain features to model SWE in this basin, and net radiation provides a physically based, temporally dynamic variable necessary to explain the changing distribution through the melt season. These three variables do not tell the whole story as evidenced by the low correlations they produce with SWE and by the ambiguities existing in the choice of classification parameters. The large proportion of the variation not explained by the regression equations also indicates that other factors control snow distribution or that the variables used interact in a complex, non-linear fashion. At this point it is not clear which scheme or parameters produce the best results. It is only clear which combinations produce poor results and further, it is not clear why. Of the parameters used, radiation appears to be the most important. It consistently shows the highest correlation with SWE and produces the best results if only one parameter is used. Slope is important because some slopes are too steep to retain snow and the slopes lying below accumulate the sloughing snow from above. More important than slope itself may be information about the neighboring slopes. The elevation parameter used in conjunction with radiation and slope appears to have improved some of the classifications. Elevation itself is not an adequate parameter for partitioning alpine basins into zones of similar stored SWE. This result is important because many sampling efforts have been based on elevation as the controlling parameter. Elevation has minor importance in this basin because other factors overshadow its effects, however, it does seem to become more important as the season progresses. Early in the season elevational effects are balanced by the bare steep slopes found in the upper reaches of the basin. Later in the season melt has been most vigorous at the lower elevations, and the large deposits remaining at the bases of the cliffs in the upper basin produce a stronger positive relationship between elevation and SWE.

Although it is not clear which classification scheme provides the best result, it is clear that a stratified random sample based on terrain and radiation parameters provides the the basis for an optimal sampling scheme. A combination of radiation, slope, and elevation leads to a division of the basin into classes of similar snow water equivalence that provides a superior stratified random sampling scheme in contrast to a simple basin-wide random sample. Specific parameter choices must be based on the users tools and needs. If a high resolution DEM and an adequate computing environment are available, a similar method as described in this study may be employed. If topographic maps are all that is available, a basin may be divided up based on slope, aspect and elevation classes derived from the map itself, however this method would require experience to obtain an objective, meaningful result. Familiarity with a specific basin would be a great help, as in applying the isohyetal method of mapping precipitation. A DEM is an ideal tool for this problem because it provides the basis for objective division of the basin into zones of similar snow accumulation properties, which can then be sampled according to the within-zone variability of SWE. Intelligent decisions can then be made as to which zones should be sampled and which should be discarded from the sampling scheme. This decision can be based on several constraints to provide an optimal survey, based on the users specific criteria. For example, only zones of a specified minimum area may be sampled. Zones expected to contain less than a specified percentage of the total basin SWE may be eliminated. Homogeneity may be the constraint, where only zone of sufficient size and continuity are sampled to insure the sample points fall within the zone boundaries. If field expenses are critical and only a rough estimate of basin SWE is needed, then perhaps only the largest zone containing the majority of stored water should be sampled. However, the poor results from the 1986 season followed by markedly improved results in 1987 and 1988 suggest the importance of a good DEM and careful field techniques if a high level of accuracy is desired. It is felt that the 1986 results were largely affected by the difficulties in locating the sampling points in the field and then assigning the points to the correct DEM grid location for data analysis. The extraordinarily deep snowpack may have effectively modified the terrain in such a way as to significantly change the outcome of the classification attempts. The DEM has proved to be a valuable tool in both the field and data analysis portions of this study. The DEM was used extensively in sample design before going into the field and the high resolution contour map derived from the DEM was invaluable in the field for locating the sampling points. The DEM also provided the basis for extending the point measurements of density over the basin and registering density with the many depth measurements. The classification parameters of slope and net radiation were both calculated using the DEM.

Temporal aspects of snow distribution should be addressed when sampling is being planned. Our field work shows us that near peak accumulation when the snow covered area is highest, the variance in SWE is also highest. A full range of values of SWE are represented in the basin as all roughness scales from small vegetation to boulders and gullies collect snow. As the melt season proceeds the thin areas melt most rapidly, which reduces the basin-wide variance. In late melt season only isolated snow patches exist and variance tends toward a minimum, both because the snowpack is homogeneous and it is asymptotically approaching the zero value of the rest of the basin. Table 14 shows the decrease in the range of the 90% confidence interval through time in both the 1986 and 1987 seasons. It is interesting to note that although the confidence interval decreases through time, the basin SWE is decreasing at a faster rate. The 90% confidence interval becomes a more significant portion of the basin SWE as time progresses. However, these values are based on basin-wide estimates and include all measurements. The required sample size to be within a given error at a specified confidence level in the stratified random sampling schemes shows a steady decrease as time proceeds. The stratified random sampling scheme allows sample location in the homogeneous zones where the standard error is low.

From this study, it appears that stratified random sampling methods based on objective classification is desirable, and that conventional methods of arbitrarily or subjectively locating snow courses will not give accurate estimates of basin SWE in alpine watersheds. Although the snow courses have provided a useful index of basin water storage for many years, an accurate estimate of the actual basin SWE must come from a sampling scheme that represents the areal distribution of the variability in SWE. The trade-offs in selecting a sampling method between the extremes of a simple snow course and an extensive, high-resolution survey must be evaluated by the user and his or her needs and constraints. The technique presented in this study may be used to design an optimal sampling scheme where the zones of similar SWE distribution are determined by areas of similar terrain features with no *a priori* information about the snowpack. Areas containing the significant accumulation can be concentrated on without wasting time or energy on the relatively unimportant portions of the basin. The number of samples required to describe the zone SWE to a desired level of accuracy can be determined by completing a quick on-site pilot survey. This method has obvious benefits where cost and manpower are prohibitive.

Perhaps the best evaluation of the method presented here is the accuracy of the volume measurements since it is basin storage that we are after. The agreement between the basin SWE volumes produced by most of the classifications and the simple statistical means is encouraging. Although this agreement does not indicate which attempts are superior, poor results would raise serious questions about the classifications. It does appear that the techniques and parameters used here will generate good estimates of volume.

F. Future Work

Currently it appears that we are doing an adequate job of modeling the change in distribution of SWE through use of the radiation index, and we have begun to explain the accumulation through the slope and elevation variables. However, we have yet failed to effectively model the component of accumulation due to redistribution of the snow. We have touched on this through slope, which accounts for sloughing and persistently bare areas, but other factors in the terrain controlling redistribution must be identified. Better results may be obtained if the parameters are weighted according to their importance, rather than being evenly weighted or excluded altogether as has been done in this study. Variables that control the drift erosion and deposition, such as the rate of change or second derivative of slope, must be explored. Clearly, we must also use snow-covered area in future attempts if this work is to provide accurate spatial information about SWE, necessary as input to spatial snowmelt models.

The other major question we have not addressed is scale. Are these results and this technique singular both to this basin and a 5 m DEM? Would we get better or worse results if we were to degrade the resolution of the DEM? As the snowpack depth increases does resolution become less important? Perhaps in deep snowpacks a 25 m DEM would be adequate or more appropriate than a 5 m DEM. If we can model snow distribution on the scale of this small basin, can it be extended to large basins, or perhaps to regions such as the entire Sierra Nevada? In the future we need to couple remote sensing techniques into such investigations. The only practical solutions to questions on large scale snow distribution depend on synergism of our knowledge of the electromagnetic properties of snow and our ability to accurately characterize ground observations. More work is needed in both areas.

TABLE 1. Summary of Previous Work on Snow Distribution

Authors	Location	Elevation (m)	Dependent Variable(s)	Independent V
Meiman [1968]	Reviews previous studies		SWE	elevation, aspe forest canopy
Leaf and Kovner [1972]	Fraser Experimental Forest and Fool Creek, Colorado	not specified	SWE	elevation
Alford [1973]	Front Range, Colorado	3440-4040	SWE	elevation, cirque orientat
Engelen [1973]	Snow courses in Colorado and New Mexico	not specified	depth, SWE	elevation, topo vegetation, dat latitude, longit
Logan [1973]	Wilmot Creek Basin, Ontario	76-373	depth, density, SWE	elevation, seas air temperature barometric pre liquid precipit
Grant and Rhea [1974]	Snow courses in Colorado	2370-3440	density	geographic loc
Rhea and Grant [1974]	Colorado and Utah	2700-3400	SWE	elevation, topo
Steppuhn and Dyck [1974]	Beir Basin, Yukon; Bad Lake Basin, Saskatchewan; Battle Basin, Alberta	not specified	SWE	terrain, vegetat land use
Storr and Golding [1974]	Marmot Creek Experimental Watershed, Alberta	1585-2805	SWE	elevation
Young [1974, 1975]	Peyto Glacier, Alberta	2100-3200	SWE	elevation, slop
Caine [1975]	San Juan Mountains, Colorado	2650-3500	SWE	elevation
Adams [1976]	Peterborough, Ontario	~220	depth, density, SWE	vegetation
Dickison and Daugharty [1979]	Nashwaak Experimental Watershed, New Brunswick	195-480	depth, SWE	elevation, slop aspect, vegetat vegetation bas
Dingman et al. [1979]	Snow courses in New Hampshire and Vermont	90-760	depth, density, SWE	elevation, date
Granberg [1979]	Timmins 4 Permafrost Experimental Site, Quebec	755-795	SWE	topographic an vegetative rou
Weir [1979] and Weir and Owens [1981]	Mount Hutt, New Zealand	1345-2077	depth, density, SWE, stratigraphy	elevation, aspe slope angle, to
Woo and Heron [1979]	Resolute, Northwest Territories	85-200	SWE	terrain
Rawls et al. [1980]	Reynolds Creek Experimental Watershed, Idaho	1400-2195	depth, density, SWE	slope angle, as vegetation, dri non-drift area
Dexter [1986]	Front Range, Colorado	2500-4000	depth, density, SWE, stratigraphy	elevation, slop
Haston [1986]	Emerald Lake Watershed, California	2800-3416	SWE	elevation, slop

TABLE 2. Abbreviations for Parameters Used in Basin Classification

abbreviation	classification parameters	number of classes
RSE12	radiation, slope, elevation	12
RSE8	radiation, slope, elevation	8
RS12	radiation, slope	12
RS8	radiation, slope	8
RE12	radiation, elevation	12
RE8	radiation, elevation	8

TABLE 3. Regional Precipitation - 1986, 1987, and 1988
Percent of 50-Year Mean

	Region	February 1	March 1	April 1	May 1
1986	Tulare region	108%	140%	150%	140%
	Kaweah	125%	175%	180%	165%
1987	Tulare region	57%	70%	75%	71%
	Kaweah	40%	55%	65%	65%
1988	Tulare region	110%	85%	70%	83%
	Kaweah	110%	85%	75%	80%

data source: California Cooperative Snow Survey

TABLE 4. 1986 Water Year Precipitation (cm) - Lodgepole Ranger Station (1943 m a.s.l.)

	Oct	Nov	Dec	Jan	Feb	Mar	April	May	June	July	Aug	Sept
1986 water year	4.4	30.7	15.5	23.6	80.4	27.7	6.9	1.5	0.0	1.5	0.3	7.4
50 year mean	4.0	10.2	20.1	22.1	20.7	17.0	11.0	4.1	1.1	0.4	0.5	1.9
% difference	109%	302%	77%	106%	389%	162%	63%	37%	0%	407%	55%	397%
1986 cumulative	4.4	35.2	50.7	74.2	154.6	182.2	189.2	190.7	198.7	192.1	192.4	199.8
50 year cumulative	4.4	14.2	34.3	56.4	77.1	94.1	105.1	109.2	110.3	110.7	111.2	113.1
% difference	109%	248%	148%	131%	200%	193%	180%	174%	173%	173%	173%	177%

data source: California Cooperative Snow Survey, 1986

TABLE 5. Snow Board Data Summary - 1986 Water Year

Site	Date		Depth (m)	Density ($kg\,m^{-3}$)	SWE (m)	Cum. SWE (m)
inlet	6	Oct	0.06	267	0.02	0.02
	8	Oct	0.04	275	0.01	0.03
	21	Oct	0.30	126	0.04	0.07
	11	Nov	0.79	150	0.12	0.19
	20	Nov	0.46	250	0.12	0.31
	3	Dec	1.42	270	0.38	0.69
	11	Dec	0.12	300	0.04	0.73
	8	Jan	0.38	410	0.15	0.88
	3	Feb	0.78	230	0.18	1.06
	6	Feb	0.09	55	0.01	1.07
	18	Feb	2.02	410	0.83	1.90
	19	Feb	0.60	400	0.24	2.14
	19	Mar	1.47	290	0.43	2.57
	10	Apr	0.11	328	0.04	2.61
	16	Apr	0.09	151	0.01	2.62
	4	May	0.11	215	0.02	2.64
	7	May	0.04	241	0.01	2.65
pond	20	Nov	0.43	271	0.13	0.13
	18	Dec	1.37	380	0.52	0.65
	8	Jan	0.45	410	0.18	0.83
	3	Feb	0.72	240	0.17	1.00
ridge	20	Nov	1.10	235	0.26	0.26
	11	Dec	1.38	290	0.40	0.66
	8	Jan	0.47	380	0.18	0.84
	3	Feb	0.89	250	0.22	1.06

TABLE 6. 1987 Water Year Precipitation (cm) - Lodgepole Ranger Station (1943 m a.s.l.)

	Oct	Nov	Dec	Jan	Feb	Mar	April	May	June	July	Aug	Sept
1987 water year	1.4	2.3	2.2	9.9	20.6	15.4	4.6	5.4	1.8	0.0	0.1	1.1
50 year mean	4.0	10.2	20.1	22.1	20.7	17.0	11.0	4.1	1.1	0.4	0.5	1.9
% difference	34%	23%	11%	45%	100%	90%	41%	130%	169%	7%	25%	58%
1987 cumulative	1.4	3.7	5.9	15.8	36.4	51.8	56.4	61.8	63.6	63.6	63.7	64.8
50 year cumulative	4.4	14.2	34.3	56.4	77.1	94.1	105.1	109.2	110.3	110.7	111.2	113.1
% difference	32%	26%	17%	28%	47%	55%	54%	57%	58%	58%	57%	57%

data source: California Cooperative Snow Survey, 1987

TABLE 7. Snow Board Data Summary - 1987 Water Year

Site	Date	Depth (m)	Density ($kg\,m^{-3}$)	SWE (m)	Cum. SWE (m)
inlet	15 Jan	0.47	230	0.11	0.11
	5 Feb	0.43	250	0.11	0.22
	20 Feb	0.58	225	0.13	0.35
	3 Mar	0.21	260*	0.05	0.40
	9 Mar	0.37	230	0.09	0.49

* estimated - ice on board

TABLE 8. Snowpit Data Summary - 1986 Water Year

Site	Date	Depth (m)	Density (kg m^{-3})	SWE (m)
tower	18 Feb	2.48	422	1.05
	3 Mar	3.70	485	1.79
	4 Mar	3.50	429	1.50
inlet	18 Jan	1.65	461	0.76
	6 Feb	2.30	365	0.84
	5 Mar	3.20	461	1.48
	2 May	4.05	593	2.40
	21 May	3.57	554	1.98
	27 June	2.03	590	1.20
pond	5 Feb	2.25	392	0.88
	12 Apr	3.17	524	1.66
	6 May	2.90	475	1.38
	24 May	2.10	520	1.09
	26 June	2.20	588	1.29
ridge	17 Jan	1.98	411	0.81
	4 Feb	2.45	365	0.90
	13 Apr	6.00	548	3.29
	6 May	5.90	520	3.17
	23 May	4.65	572	2.66
	27 June	2.50	578	1.44
hole	3 May	3.50	497	1.74
	7 May	4.80	485	2.33
	24 May	3.90	513	2.00
	26 June	2.00	557	1.11
cirque	11 Jan	3.05	400	1.22

TABLE 9. Snowpit Data Summary - 1987 Water Year

Site	Date	Depth (m)	Density ($kg\,m^{-3}$)	SWE (m)
inlet	4 Mar	1.22	310	0.38
	16 Mar	1.80	280	0.50
	2 Apr	1.50	425	0.64
	10 Apr	1.40	450	0.63
	17 Apr	1.02	435	0.44
	23 Apr	0.92	480	0.44
	29 Apr	0.63	480	0.30
	7 May	0.35	415	0.15
bench	4 Mar	1.20	350	0.42
	18 Mar	1.50	330	0.50
	2 Apr	1.33	420	0.56
	10 Apr	1.16	435	0.50
	17 Apr	0.86	435	0.37
	22 Apr	0.61	440	0.27
	29 Apr	0.32	400	0.13
pond	4 Mar	1.43	330	0.47
	17 Mar	2.05	303	0.62
	2 Apr	1.90	380	0.72
	9 Apr	1.60	450	0.72
	17 Apr	1.40	420	0.59
	22 Apr	1.20	470	0.56
	29 Apr	0.98	450	0.44
	7 May	0.71	460	0.33
	13 May	0.50	460	0.23
hole	31 Mar	1.90	310	0.59
	9 Apr	1.73	380	0.66
	17 Apr	1.48	430	0.63
	22 Apr	1.30	400	0.52
	29 Apr	1.04	450	0.46
	7 May	0.78	465	0.36
	13 May	0.60	480	0.29
	22 May	0.84	440	0.37
	27 May	0.80	475	0.38
ramp	18 Apr	1.34	405	0.54
	22 Apr	1.25	480	0.60
	29 Apr	1.07	480	0.51
	7 May	1.00	480	0.48
	13 May	0.68	470	0.32
	22 May	0.25	415	0.10
cirque	31 Mar	2.35	320	0.75
	9 Apr	2.36	355	0.84
	17 Apr	2.10	405	0.85
	22 Apr	1.94	410	0.79
	29 Apr	1.67	460	0.77
	7 May	1.52	465	0.71
	13 May	1.25	470	0.59
	22 May	1.31	440	0.58
	27 May	1.37	440	0.61
	27 May	2.40	400	0.97

TABLE 10. Mean Basin Snow Density from Snowpits - 1987 Water Year
Observed and Predicted Weekly Means

date	year date	observed density $kg\,m^{-3}$	standard deviation $kg\,m^{-3}$	sample size n	predicted density $kg\,m^{-3}$
19 Feb	50.5	290.4	50.86	28	*
3 Mar	62.5	324.1	40.82	79	*
17 Mar	76	332.7	43.96	72	*
1 April	91	357.3	64.40	109	*
9 April	99.5	405.8	56.36	77	418.2
17 April	107.5	418.0	37.93	75	426.4
22 April	112.5	440.2	47.35	67	431.6
29 April	119	464.9	32.97	50	438.9
7 May	127	468.3	31.87	37	446.5
13 May	133	470.9	22.03	29	452.6
22 May	142	450.5	29.37	18	461.9
27 May	147	424.8	40.12	40	467.0
11 June	162	491.8	43.13	32	482.4

* predicted density relationship used only after 1 April.

TABLE 11. Predicted Mean Basin Snow Density
During Survey Periods - 1987 Water Year

date	year date	predicted density ($kg\,m^{-3}$)
17-19 April	108	426.9
8-10 May	129	448.5
21-23 May	142	461.9
5 June	156	476.3

TABLE 12. Snowpit Data Summary - 1988 Water Year

Site	Date	Depth (m)	Density ($kg\,m^{-3}$)	SWE (m)
inlet	9 Jan	1.67	285	0.48
	22 Jan	1.84	305	0.56
	3 Feb	1.81	354	0.64
	18 Feb	1.68	367	0.62
	9 Mar	1.78	374	0.67
	21 Mar	1.67	433	0.72
	3 Apr	1.38	458	0.63
	11 Apr	1.09	520	0.57
bench	10 Mar	1.66	407	0.68
	21 Mar	1.49	482	0.72
	30 Mar	1.18	446	0.53
	3 Apr	1.04	493	0.51
	11 Apr	0.77	523	0.40
	13 May	0.50	503	0.25
pond	23 Mar	1.25	435	0.54
	5 Apr	0.87	505	0.44
	13 May	0.90	480	0.43
hole	24 Mar	1.50	350	0.53
	4 Apr	1.80	432	0.78
	13 May	1.58	475	0.75
ramp	24 Mar	1.80	363	0.65
	4 Apr	1.57	381	0.60
	12 Apr	1.56	411	0.64
	13 May	1.80	455	0.82
	21 May	1.30	480	0.62
cirque	14 Jan	2.19	300	0.66
	17 Feb	2.23	359	0.80
	8 Mar	2.80	400	1.12
	21 Mar	2.10	385	0.81
	23 Mar	2.57	413	1.06
	5 Apr	2.30	429	0.99
	12 Apr	1.94	479	0.93
	21 May	1.55	478	0.74

TABLE 13. Summary of Depth Surveys - 1986, 1987, and 1988 Water Years

survey date		n	mean depth (cm)	std. err. of mean	90% conf. interval
1986	15-17 April	86	384	13.69	22.77
	2-5 May	127	378	19.43	32.20
	23-26 May	157	292	15.70	25.97
	24-27 June	166	107	12.64	20.91
1987	17-19 April	256	140	5.45	9.00
	8-10 May	295	79	4.73	7.80
	21-23 May	328	47	3.61	5.95
	5 June	279	28	3.07	5.07
1988	20-23 March	354	153	5.11	8.43

TABLE 14. Summary of Snow Water Equivalence - 1986, 1987, and 1988 Water Years

survey date	mean depth (cm)	mean density $kg\,m^{-3}$	SWE (cm)	expected SWE volume m^3	90% conf. int.	interpolated SWE volume m^3
1986 15-17 April	384	520	199.7	2,398,560	142,230	*
2-5 May	378	520	196.6	2,361,080	201,130	2,534,260
23-26 May	292	520	151.8	1,823,900	162,210	1,789,680
24-27 June	107	520	55.6	668,350	130,610	901,010
1987 17-19 April	140	427	59.8	718,320	46,160	667,050
8-10 May	79	449	35.5	426,430	42,070	416,160
21-23 May	47	462	21.7	260,660	33,020	280,380
5 June	28	476	13.3	159,760	28,990	159,010
1988 20-23 March	153	411	63.0	750,700	41,470	721,080

* not calculated

TABLE 15. Depth Statistics - Entire Basin vs. Northeast Wall (1986)

	entire basin			northeast wall		
survey date	n	mean depth (cm)	variance	n	mean depth (cm)	variance
May 2-5	127	378	47,939	50	357	21,870
May 23-26	157	292	38,655	34	269	13,005
June 24-27	166	107	26,528	42	49	7,727

TABLE 16. F Test Results - Entire Basin vs. Northeast Wall (1986)

survey date	May 2-5	May 23-26	June 24-27
F_{est}	0.399	0.453	12.19
$F_{(0.05)(1)(1)(v)}$	3.89	3.89	3.90
accept/reject H_0	accept	accept	reject
% conf. level	0.001	0.001	0.002
basin mean depth	378.10	292.46	107.43
NE wall mean depth	356.84	268.97	49.29
95% conf. interval for basin mean depth	±38.45	±31.01	±24.95

TABLE 17. ANOVA and Standard Error Evaluations, 17-19 April, 1987

stratification	*F* ratio	conf. level	total df	no. of classes	SE cm	% of random SE
RSE12	5.126	0.003	233	11	2.13	91
RSE8	6.688	0.011	233	7	2.16	93
RS12	11.117	0.0002	233	10	2.02	87
RS8	6.242	0.013	233	7	2.47	106
RE12	11.786	0.0004	233	9	2.18	94
RE8	10.989	0.006	233	6	2.10	90

SE for random sample = 2.33 cm, n = 256

TABLE 18. Basin SWE Volume Estimates from Classifications, 17-19 April, 1987

stratification	estimated volume (m^3)	% difference from expected
RSE12	714,520	1%
RSE8	699,480	3%
RS12	694,130	3%
RS8	686,570	4%
RE12	693,580	3%
RE8	714,120	1%

expected volume = 718,320 m^3

TABLE 19. ANOVA and Standard Error Evaluations, 8-10 May, 1987

stratification	*F* ratio	conf. level	total df	no. of classes	SE cm	% of random SE
RSE12	10.538	0.003	268	9	1.99	94
RSE8	6.467	0.007	268	8	2.09	98
RS12	9.110	0.0003	268	11	1.98	93
RS8	15.514	0.003	268	6	1.99	94
RE12	13.289	0.0001	268	10	2.74	129
RE8	4.039	0.040	268	7	2.16	102
SE for random sample = 2.13 cm, n = 295						

TABLE 20. Basin SWE Volume Estimates from Classifications, 8-10 May, 1987

stratification	estimated volume (m^3)	% difference from expected
RSE12	462,990	9%
RSE8	439,370	3%
RS12	428,670	0%
RS8	435,400	2%
RE12	453,470	6%
RE8	423,190	2%
expected volume = 426,430 m^3		

TABLE 21. ANOVA and Standard Error Evaluations, 21-23 May, 1987

stratification	*F* ratio	conf. level	total df	no. of classes	SE cm	% of random SE
RSE12	1.172	0.456	296	8	1.75	105
RSE8	18.825	0.005	296	5	1.58	95
RS12	1.039	0.559	296	6	1.75	105
RS8	3.998	0.139	296	4	1.73	104
RE12	16.166	0.0001	296	9	1.48	89
RE8	11.569	0.002	296	7	1.59	95

SE for random sample = 1.67 cm, n = 328

TABLE 22. Basin SWE Volume Estimates from Classifications, 21-23 May, 1987

stratification	estimated volume (m^3)	% difference from expected
RSE12	254,410	2%
RSE8	268,730	3%
RS12	239,790	8%
RS8	249,720	4%
RE12	265,320	2%
RE8	258,850	1%

expected volume = 260,660 m^3

TABLE 23. ANOVA and Standard Error Evaluations, 5 June, 1987

stratification	*F* ratio	conf. level	total df	no. of classes	SE cm	% of random SE
RSE12	3.824	0.023	246	9	1.54	105
RSE8	3.279	0.066	246	7	1.57	107
RS12	4.790	0.017	246	8	1.52	104
RS8	3.636	0.157	246	4	1.58	108
RE12	11.934	0.002	246	7	1.43	98
RE8	4.484	0.075	246	5	1.57	108
SE for random sample = 1.46 cm, n = 279						

TABLE 24. Basin SWE Volume Estimates from Classifications, 5 June, 1987

stratification	estimated volume (m^3)	% difference from expected
RSE12	163,830	3%
RSE8	164,930	3%
RS12	174,830	9%
RS8	165,800	4%
RE12	167,190	4%
RE8	164,000	3%
expected volume = 159,760 m^3		

TABLE 25. ANOVA and Standard Error Evaluations, 20-23 March, 1988

stratification	*F* ratio	conf. level	total df	no. of classes	SE cm	% of random SE
RSE12	5.029	0.009	284	9	1.47	72
RSE8	4.272	0.053	284	6	1.52	74
RS12	4.483	0.040	284	6	1.51	73
RS8	5.630	0.029	284	6	1.49	72
RE12	6.589	0.006	284	8	1.45	71
RE8	7.458	0.008	284	7	1.45	71
SE for random sample = 2.06 cm, n = 354						

TABLE 26. Basin SWE Volume Estimates from Classifications, 20-23 March, 1988

stratification	estimated volume (m^3)	% difference from expected
RSE12	730,280	3%
RSE8	734,760	2%
RS12	726,530	3%
RS8	729,830	3%
RE12	730,300	3%
RE8	736,400	2%
expected volume = 750,700 m^3		

TABLE 27. Error In Volume Estimates with Elimination of Small Zones

			≥5 field measurements		≥10 field measurements	
			% change	% basin	% change	% basin
survey date		class	SWE volume	area	SWE volume	area
1987	17-19 April	RSE12	+0.04	2.4	+0.04	2.4
		RSE8	+0.03	1.4	-0.9	4.4
		RS12	-0.9	1.6	-8.5	12.5
		RS8	-0.4	0.4	-5.3	1.6
		RE12	-0.4	0.3	-0.4	0.3
		RE8	0.0	0.0	+0.2	4.5
	8-10 May	RSE12	-4.4	3.7	-20.5	8.7
		RSE8	-0.4	0.2	-2.5	6.3
		RS12	-3.6	3.3	-6.9	9.3
		RS8	-2.1	3.8	-2.1	3.8
		RE12	+0.02	0.3	-1.3	3.2
		RE8	+0.3	0.9	+2.3	4.5
	21-23 May	RSE12	-1.9	0.6	-1.2	1.4
		RSE8	+0.6	0.6	+0.6	0.6
		RS12	0.0	0.0	+3.0	6.4
		RS8	0.0	0.0	0.0	0.0
		RE12	+0.3	0.5	-16.3	3.7
		RE8	+0.3	0.5	+1.0	3.4
	5 June	RSE12	0.0	0.0	+1.1	1.1
		RSE8	0.0	0.0	0.0	0.0
		RS12	+0.4	0.7	+2.6	2.7
		RS8	0.0	0.0	+4.0	3.5
		RE12	+2.8	2.6	+2.8	2.6
		RE8	+0.8	0.8	+0.8	0.8
1988	20-23 March	RSE12	-0.3	2.4	+0.5	5.1
		RSE8	+0.03	2.2	-0.3	3.5
		RS12	+0.03	0.9	-0.3	1.9
		RS8	-0.2	1.7	-0.2	3.2
		RE12	0.0	0.0	-0.8	6.4
		RE8	+0.5	0.4	-0.6	3.1

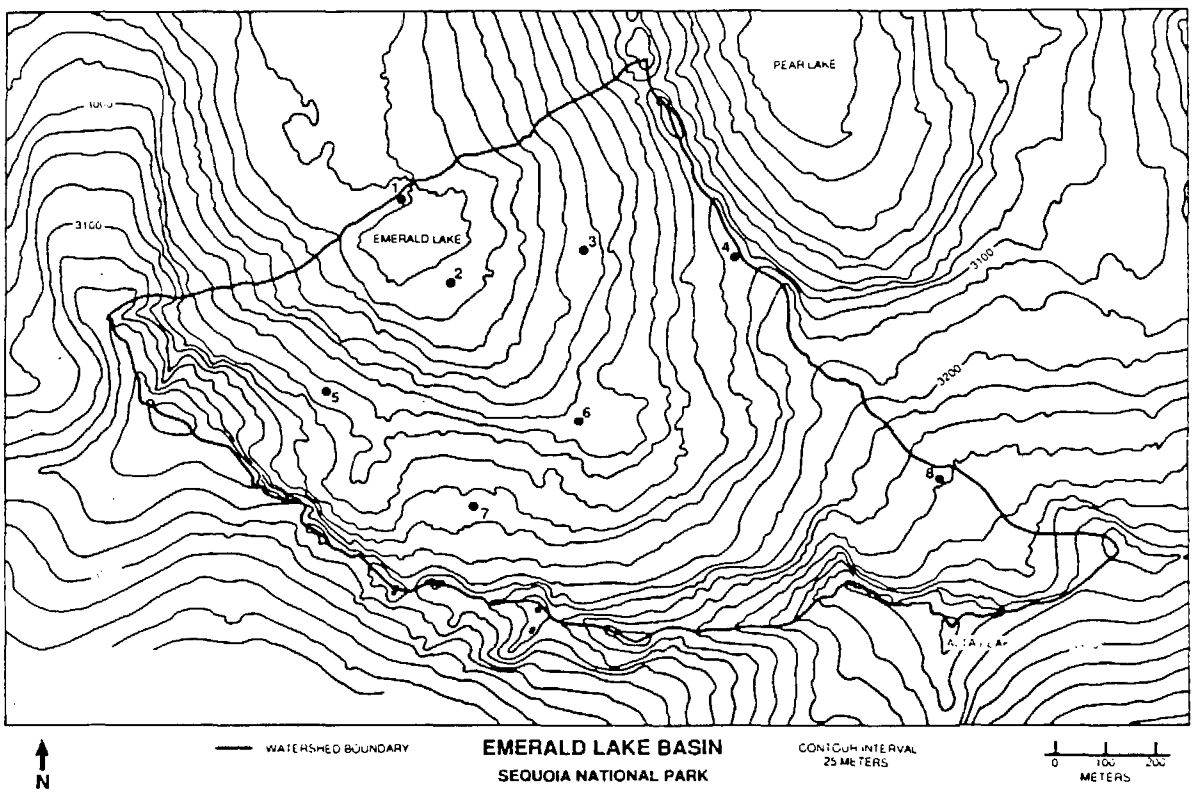
TABLE 28. Required Sample Sizes for Specified Confidence Level and 5 cm Error

survey date		class	95% confidence level all zones	95% confidence level ≥5 field measurements	95% confidence level ≥10 field measurements	99% confidence level all zones	99% confidence level ≥5 field measurements	99% confidence level ≥10 field measurements
1987	17-19 April	RSE12	120	71	71	206	122	122
		RSE8	50	33	20	84	55	33
		RS12	112	83	25	206	139	40
		RS8	61	41	20	102	68	33
		RE12	57	57	57	95	95	95
		RE8	33	33	27	55	55	45
	8-10 May	RSE12	112	95	16	190	161	24
		RSE8	158	157	40	272	270	67
		RS12	142	138	38	245	238	64
		RS8	98	98	57	168	168	97
		RE12	74	74	24	126	126	41
		RE8	112	56	42	174	78	54
	21-23 May	RSE12	48	47	45	78	76	74
		RSE8	18	18	18	29	29	29
		RS12	31	27	16	48	42	24
		RS8	11	11	11	18	18	18
		RE12	52	38	36	88	63	60
		RE8	39	25	13	65	40	20
	5 June	RSE12	35	35	34	59	59	58
		RSE8	25	25	25	43	43	43
		RS12	8	8	8	12	12	12
		RS8	3	3	3	5	5	5
		RE12	17	17	17	26	26	26
		RE8	26	26	26	42	42	42
1988	20-23 March	RSE12	151	77	13	259	132	23
		RSE8	85	61	22	144	103	36
		RS12	69	47	11	118	81	19
		RS8	51	21	11	88	36	18
		RE12	50	38	10	84	64	17
		RE8	23	23	17	40	40	30

TABLE 29. Sample Size Requirements, 20-23 March, 1988, RSE8

zone	$\hat{\sigma}$ (m)	# points sampled in field	# points 95% conf. ±0.05 m	# points 99% conf. ±0.05 m	% of basin area	% of snow covered area
1	0.019	191	1	1	50.7	64.9
2	0.061	18	6	10	4.0	5.1
3	0.089	10	13	22	2.4	3.1
4	0.124	3	24	41	2.2	2.8
5	0.158	7	39	67	1.3	1.7
6	0.029	56	2	3	17.5	22.4
# total points		285	85	144	78.1	100.0
# points ≥ 5			61	103	75.9	97.2
# points ≥ 10			22	36	74.6	95.5

Figure 1. Emerald Lake Basin - Snowpit Locations



Locations of snowpits are as follows: 1—Tower, 2—Inlet, 3—Bench, 4—Ridge, 5—Ramp, 6—Pond, 7—Hole, 8—Cirque.

Figure 2. Cumulative Net Radiation in March and in June

Cumulative net radiation, December through March (left) and December through June (right). In the early season (left) mean value is $78\,W\,m^{-2}$ and standard deviation is $33.7\,W\,m^{-2}$. By late season (right) there is less difference between the northeast and southwest portions of the basin, because of a higher solar zenith angle and greater azimuthal range. Steep, north-facing slopes at the bottom of the image still receive much less net radiation than the remaining basin. Mean value is $309\,W\,m^{-2}$ and standard deviation is $77.9\,W\,m^{-2}$.

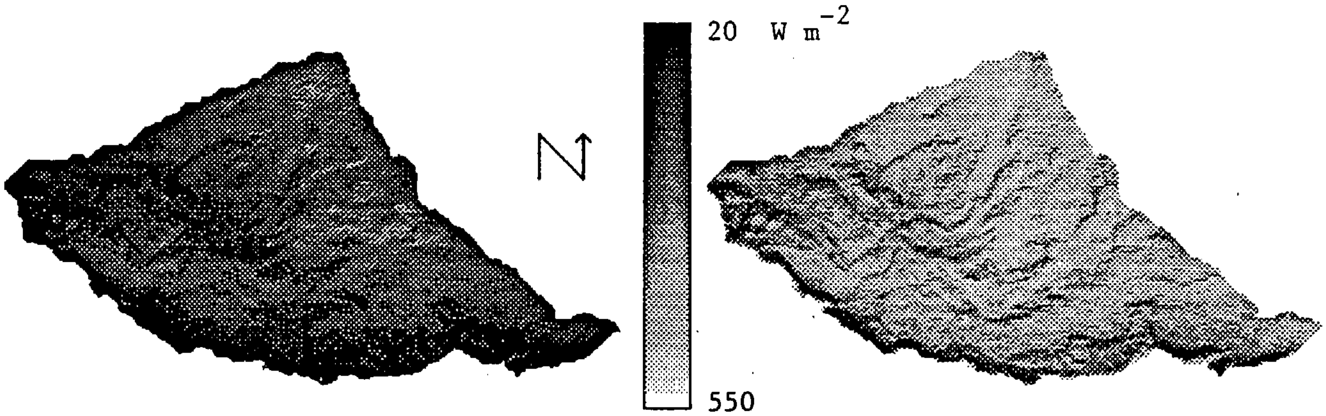


Figure 3. Distribution of Survey Points, 20-23 March, 1988

Distribution of survey points, 20-23 March, 1988. Points were chosen by two methods. (1) 115 points were located from a randomly located 100m square grid. (2) 247 points were randomly located from a 25m square grid coincident with the 100m square grid.

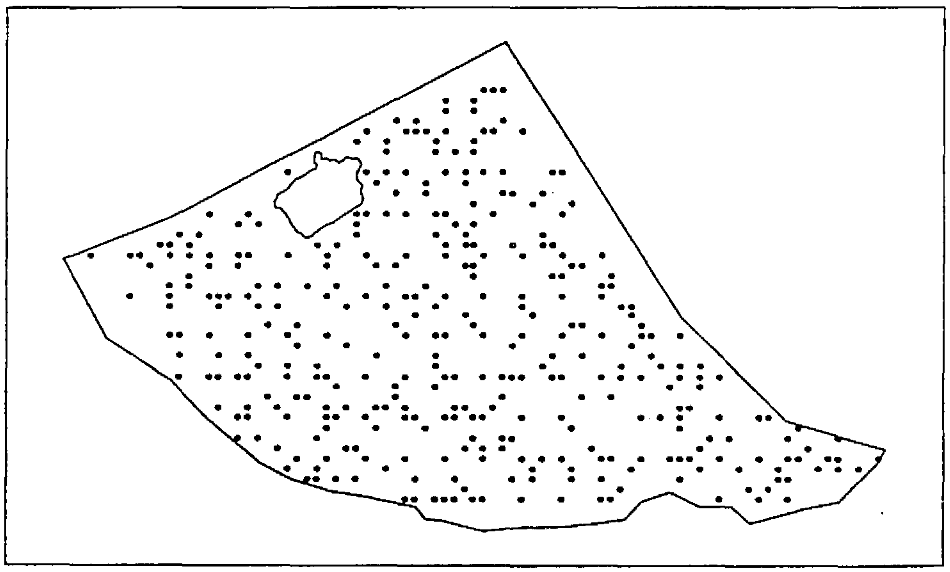


Figure 4. Survey Classification Results from RSE8, 17-19 April, 1987

Survey classification results from RSE8, 17-19 April, 1987. Greatest SWE values are found on north-facing slopes below steep faces. The southwest-facing wall has the least amount of water stored on it. Black values represent areas of no snow, extent was determined by identifying all areas of the basin with slopes of 55^o or greater.

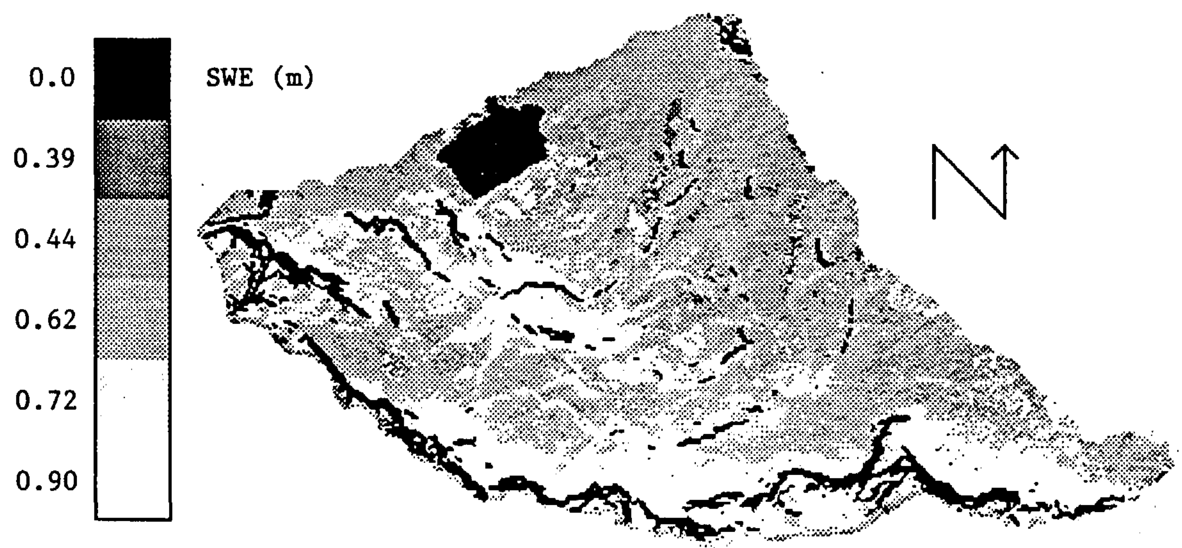
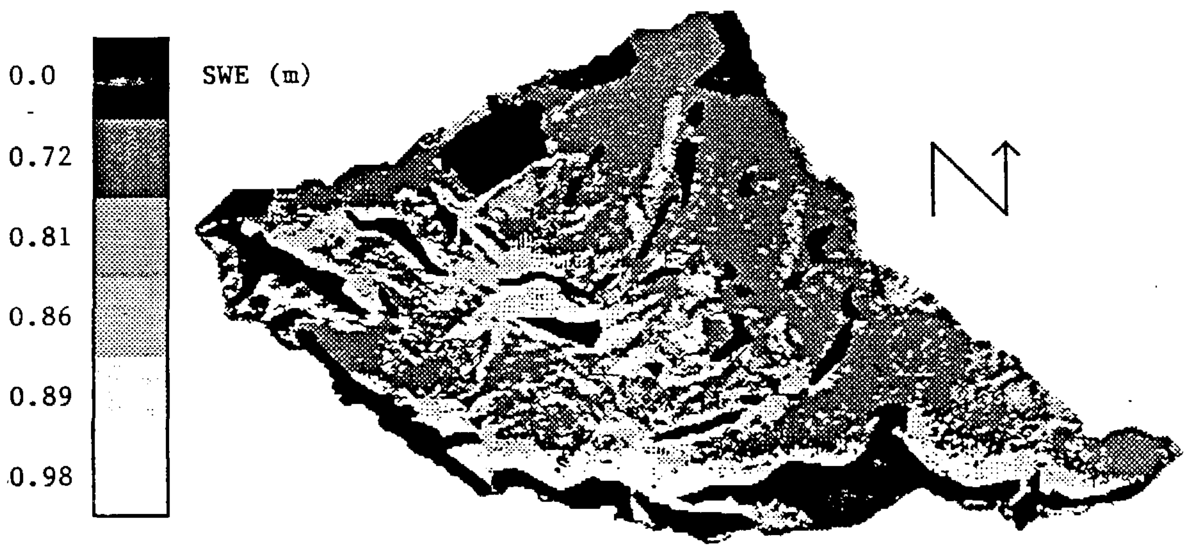


Figure 5. Survey Classification Results from RS12, 20-23 March, 1988

Survey classification results from RS12, 20-23 March, 1988. Greatest deposits of SWE are found on the flat areas below the cliffs and at the upper elevations. The southwest-facing wall shows the least amount of SWE, but still identifies the significant drift deposits found on the benches in this area. Black values represent areas of no snow, extent was determined from field notes and oblique photographs.



G. References

California. Dept. of Water Resources, *California Cooperative Snow Surveys Bulletin 120-86, Fall Report*, Department of Water Resources, Sacramento, CA, 1986.

California. Dept. of Water Resources, Water conditions in California: special dry-year report, *California Cooperative Snow Surveys Bulletin 120-87*, Department of Water Resources, Sacramento, CA, 1987.

Adams, W. P., Areal differentiation of snow cover in east central Ontario, *Water Resour. Res.*, *12*, 1226-1234, 1976.

Alford, D. L., Cirque glaciers of the Colorado Front Range: mesoscale aspects of a glacier environment, Ph.D. Thesis, 198 pp., University of Colorado, Boulder, Boulder, CO, 1973.

Alford, D. L., Cirque glaciers of the Colorado Front Range: mesoscale aspects of a glacier environment, Ph.D. Thesis, 198 pp., University of Colorado, Boulder, 1973.

Anno, Y., Development of a snowdrift wind tunnel, *Cold Regions Sci. Technol.*, *10*, 153-161, 1985a.

Anno, Y., Froude number paradoxes in the modeling of a snowdrift, *Cold Regions Sci. Technol.*, *10*, 191-192, 1985b.

Avalanche Research Centre, Guidelines for weather, snowpack, and avalanche observations, *Tech. Mem. 132*, 60 pp., National Research Council Canada, Associate Committee on Geotechnical Research, Ottawa ontario, 1981.

Bagnold, R. A., An approach to the sediment transport problem from general physics, *Prof. Pap. 422-L*, U.S. Geol. Survey, Washington, D. C., 1966.

Bartos, L. R. and P. A. Rechard, Ablation characteristics of an alpine snowfield in summer, in *Advanced Concepts in the Study of Snow and Ice Resources*, edited by H. S. Santeford and J. L. Smith, pp. 90-98, National Academy of Sciences, Washington, D.C., 1974.

Bhattacharyya, G. K. and R. A. Johnson, *Statistical Concepts and Methods*, Wiley, New York, 1977.

Caine, N., An elevational control of peak snowpack variability, *Water Resources Bulletin*, *11*, 613-621, 1975.

California Cooperative Snow Surveys, *Bulletin 120-88, Reports 1-4*, Calif. Dept. Water. Resour., Sacramento, CA, 1988.

Cochran, W. G., *Sampling Techniques, Third Edition*, Wiley, New York, 1977.

Colbeck, S. C., An overview of seasonal snow metamorphism, *Rev. Geophys. Space Phys.*, *20*, 45-61, 1982.

Colbeck, S. C., Snow particle morphology in the seasonal snow cover, *Bull. Amer. Meteorol. Soc.*, *64*, 602-609, 1983.

Davis, R. E., Simulation of an alpine soil thermal regime beneath a snowpack, M.A. Thesis, 60 pp., Dept. Geogr., Univ. Calif., Santa Barbara, 1980.

Dexter, L. R., Aspect and elevation effects on the structure of the seasonal snowcover in Colorado, Ph. D. Thesis, University of Colorado, Boulder, Colorado, 1986.

Dickison, R. and D. Daugharty, A square grid system for modeling snow cover in small watersheds, in *Proceedings of a Meeting on Modeling Snow Cover Runoff*, edited by S. C. Colbeck and M. Ray, pp. 71-76, U.S. Army Cold Regions Res. Engrg. Lab., Hanover, NH, 1979.

Dingman, S. L., E. Henry, and R. Hendrick, Variation of snow properties with elevation in New Hampshire and Vermont, in *Proceedings of a Meeting on Modeling Snow Cover Runoff*, edited by S. C. Colbeck and M. Ray, pp. 93-100, U.S. Army Cold Regions Res. Engrg. Lab., Hanover, NH, 1979.

Dozier, J., A clear-sky spectral solar radiation model for snow-covered mountainous terrain, *Water Resour. Res.*, *16*, 709-718, 1980.

Dyunin, A. K. and V. M. Kotlyakov, Redistribution of snow in the mountains under the effect of heavy snow-storms, *Cold Regions Sci. Technol.*, *3*, 287-294, 1980.

Engelen, G. B., A graphical and statistical approach to the regional study of snowpack in mountain areas, with a special reference to Colorado and New Mexico, in *The Role of Snow and Ice in Hydrology*, edited by J. C. Rodda, IAHS-AIHS Publ. No. 107, pp. 885-889, Intl. Assoc. Hydrol. Sci., Wallingford, UK, 1973.

Föhn, P. M., Snow transport over mountain crests, *J. Glaciol.*, *26*, 469-480, 1980.

Frew, J. and J. Dozier, The Image Processing Workbench—portable software for remote sensing instruction and research, *Proceedings IGARSS '86*, 271-276, European Space Agency, Paris, 1986.

Goodison, B. E., H. L. Ferguson, and G. A. McKay, Measurement and data analysis, in *Handbook of Snow*, edited by D. M. Gray and D. H. Male, pp. 191-274, Pergamon Press, New York, 1981.

Granberg, H., Snow accumulation and roughness changes through winter at a forest-tundra site near Schefferville, Quebec, in *Proceedings of a Meeting on Modeling Snow Cover Runoff*, edited by S. C. Colbeck and M. Ray, pp. 83-92, U.S. Army Cold Regions Res. Engrg. Lab., Hanover, NH, 1979.

Grant, L. O. and J. O. Rhea, Elevation and meteorological controls on the density of new snow, in *Advanced Concepts and Techniques in the Study of Snow and Ice Resources*, pp. 169-181, Natl. Acad. Press, Washington, DC, 1974.

Gray, D. M., Snow accumulation and distribution, in *Proc. Modeling Snow Cover Runoff*, edited by S. C. Colbeck and M. Ray, pp. 3-33, U.S. Army CRREL, Hanover, New Hampshire, 1979.

Hasholt, B., Random sampling technique in measuring snow-water equivalent in a drainage basin, in *Role of Snow and Ice in Hydrology*, IAHS-AIHS Publ. 107, pp. 680-687, Intl. Assoc. Hydrol. Sci., Wallingford, UK, 1973.

Haston, L., Mapping zones of similar terrain for improved estimation of snow volume in an alpine watershed, M.A. Thesis, 57 pp., Dept. Geogr., Univ. Calif., Santa Barbara, 1986.

Hoinkes, H. and R. Rudolph, Mass balance studies on the Hintereisferner, Otztal Alps, 1952-1961, *J. Glaciol.*, *4*, 266-280, 1962.

Iveronova, M. I., Le rôle hydrologique des avalanches, in *Scientific Aspects of Snow and Ice Avalanches*, IAHS-AIHS Publ. No. 69, pp. 73-77, Intl. Assoc. Hydrol. Sci., Gentbrugge, Belgium, 1966.

Kattelmann, R. C., B. J. McGurk, N. H. Berg, J. A. Bergman, J. A. Baldwin, and M. A. Hannaford, The isotope profiling snow gage: twenty years of experience, *Proc. Western Snow Conf.*, *51*, 1-8, 1983.

LaChapelle, E. R., Errors in ablation measurements from settlement and subsurface melting, *J. Glaciol.*, *3*, 458-467, 1959.

LaChapelle, E. R., *Field Guide to Snow Crystals*, University of Washington Press, Seattle, 1969.

Leaf, C. F. and J. Kovner, Sampling requirements for areal water equivalent estimates in forested subalpine watersheds, *Water Resour. Res.*, *8*, 713-716, 1972.

Logan, L., Basin-wide water equivalent estimation from snowpack depth measurements, in *Role of Snow and Ice in Hydrology*, edited by J. C. Rodda, IAHS-AIHS Publ. 107, pp. 864-884, Intl. Assoc. Hydrol. Sci., Wallingford, UK, 1973.

Marsh, P., Grain growth in a wet arctic snow cover, *Cold Regions Sci. Technol.*, *14*, 23-31, 1987.

Martinec, J. and M. R. de Quervain, The effect of snow displacement by avalanches on snowmelt and runoff, in *Snow and Ice*, IAHS-AIHS Publ. No. 104, pp. 364-377, Intl. Assoc. Hydrol. Sci., Wallingford, UK, 1975.

Martinelli, Jr., M. and A. Ozment, Some strength features of natural snow surfaces that affect snow drifting, *Cold Regions Sci. Technol.*, *11*, 267-283, 1985.

Meiman, J. R., Snow accumulation related to elevation, aspect and forest canopy, in *Snow Hydrology; Proceedings of a Workshop Seminar*, pp. 35-47, Canadian National Committee for the International Hydrological Decade, Ottawa, Ontario, 1968.

Mellor, M., Blowing Snow, *CRREL Monograph III-A3c*, Cold Regions Research and Engineering Laboratory, Hanover, NH, 1965.

Muller, F. and C. M. Keeler, Errors in short-term ablation measurements on melting ice surfaces, *J. Glaciol.*, *8*, 91-105, 1969.

Obled, C. and H. Harder, A review of snowmelt in the mountain environment, in *Proc. Modeling Snow Cover Runoff*, edited by S. C. Colbeck and M. Ray, pp. 179-204, U.S. Army CRREL, Hanover, New Hampshire, 1979.

Olyphant, G. A., Insolation topoclimates and potential ablation in alpine snow accumulation basins: Front Range, Colorado, *Water Resour. Res.*, *20*, 491-498, 1984.

Olyphant, G. A., The components of incoming radiation within a mid-latitude alpine watershed during the snowmelt season, *Arctic Alpine Res.*, *18*, 163-169, 1986b.

Olyphant, G. A., Longwave radiation in mountainous areas and its influence on the energy balance of alpine snowfields, *Water Resour. Res.*, *22*, 62-66, 1986a.

Perla, R. and M. Martinelli, *Avalanche Handbook*, Agric. Hbk. 489, rev. ed., USDA Forest Service, Washington, DC, 1978.

Price, A. G., T. Dunne, and S. C. Colbeck, Energy balance and runoff from a subarctic snowpack, *Rep. 76-27*, Cold Regions Research and Engineering Laboratory, Hanover, N.H., 1976.

Radok, U., Snow drift, *J. Glaciol.*, *19*, 123-139, 1977.

Rawls, W. J., T. J. Jackson, and J. F. Zuzel, Comparison of areal snow storage sampling procedures for rangeland watersheds, *Nordic Hydrol.*, *11*, 71-82, 1980.

Renka, R. J., Algorithm 624, Triangulation and interpolation at arbitrarily distributed points in the plane, *ACM Trans. Math. Software*, *10*, 440-442, 1984.

Rhea, J. O. and L. O. Grant, Topographic influences on snowfall patterns in mountainous terrain, in *Advanced Concepts and Techniques in the Study of Snow and Ice Resources*, pp. 182-192, Natl. Acad. Press, Washington, DC, 1974.

Richards, J. A., *Remote Sensing Digital Image Analysis*, Springer-Verlag, Berlin, 1986.

Schmidt, R. A., Threshold wind-speeds and elastic impact in snow transport, *J. Glaciol.*, *26*, 453-467, 1980.

Schmidt, R. A., Properties of blowing snow, *Rev. Geophys. Space Phys.*, *20*, 39-44, 1982a.

Schmidt, R. A., Vertical profiles of wind speed, snow concentration, and humidity in blowing snow, *Boundary-Layer Meteorol.*, *23*, 223-246, 1982b.

Schmidt, R. A., Measuring particle size and snowfall intensity in drifting snow, *Cold Regions Sci. Technol.*, *9*, 121-129, 1984.

Schmidt, R. A., R. Meister, and H. Gubler, Comparison of snow drifting measurements at an alpine ridge crest, *Cold Regions Sci. Technol.*, *9*, 131-141, 1984.

Sosedov, I. S. and I. V. Seversky, On hydrological role of snow avalanches in the northern slope of the Zailiysky Alatau, in *Scientific Aspects of Snow and Ice Avalanches*, IAHS-AIHS Publ. No. 69, pp. 78-85, Intl. Assoc. Hydrol. Sci., Gentbrugge, Belgium, 1966.

Steppuhn, H. W. and G. D. Dyck, Estimating true basin snow cover, in *Advanced Concepts and Techniques in the Study of Snow and Ice Resources*, pp. 314-328, Natl. Acad. Press, Washington, DC, 1974.

Storr, D and D. Golding, A preliminary water balance evaluation of an intensive snow survey in a mountainous watershed, in *Advanced Concepts and Techniques in Study of Snow and Ice Resources*, pp. 294-303, National Academy of Sciences, Washington, DC, 1974.

Tabler, R. D., Ablation rates of snowfence drifts at 2300 meters elevation in Wyoming, *Proc. Western Snow Conf.*, *53*, 1-12, 1985.

Tushinsky, G. K., The part avalanches play in the formation and dynamics of glaciers and snow patches in the territory of the USSR, in *Snow and Ice Symposium*, edited by J. C. Rodda, IAHS-AIHS Publ. No. 104, pp. 381-389, Intl. Assoc. Hydrol. Sci., Wallingford, UK, 1975.

Weir, P. L., Topographic influences on snow accumulation at Mount Hutt, M.A. Thesis, Univ. Canterbury, New Zealand, 1979.

Weir, P. L. and I. F. Owens, Snowpack structure at Mount Hutt, Canterbury, in 1979, *New Zealand Journal of Science*, *24*, 95-102, 1981.

Woo, M. and R. Heron, Modeling basin snow storage in a high arctic environment, in *Canadian Hydrology Symposium, 79: Cold Climate Hydrology, Proceedings*, pp. 206-216, Natl. Res. Council Canada, Ottawa, Ontario, 1979.

Work, R. A., H. J. Stockwell, T. G. Freeman, and R. T. Beaumont, Accuracy of field snow surveys, *CRREL Tech. Rep. 163*, U.S. Army Cold Regions Res. Engrg. Lab., Hanover, NH, 1965.

Young, G. J., A stratified sampling design for snow surveys based on terrain shape, *Proc. Western Snow Conf.*, *42*, 14-22, 1974.

Young, G. J., Accumulation and ablation patterns as functions of the surface geometry of a glacier, in *Snow and Ice Symposium*, edited by J. C. Rodda, IAHS-AIHS Publ. No. 104, pp. 134-138, Intl. Assoc. Hydrol. Sci., Wallingford, UK, 1975.

Zalikhanov, M. Ch., Hydrological role of avalanches in the Caucasus, in *Snow and Ice*, IAHS-AIHS Publ. No. 104, pp. 390-394, Intl. Assoc. Hydrol. Sci., Wallingford, UK, 1975.

Zar, J. H., *Biostatistical Analysis, Second Edition*, Prentice-Hall, Englewood Cliffs, N.J., 1984.

Zuzel, J. F. and L. M. Cox, Relative importance of meteorological variables in snowmelt, *Water Resour. Res.*, *11*, 174-176, 1975.

IV. WATER BALANCE OF THE EMERALD LAKE BASIN

A. Introduction

A central part of the Integrated Watershed Study was to identify and quantify the hydrologic fluxes in the Emerald Lake basin. These included rain, snowfall, streamflow entering and leaving Emerald Lake, groundwater storage, and evaporation from open water, snow, soils and vegetation. The Emerald Lake basin was instrumented to directly measure or allow indirect estimation of the most important processes. Measurement and estimation of the various water fluxes through the basin were useful to studies of the aquatic chemistry and biology, soils, and terrestrial vegetation of the basin as well as in modeling efforts and assessment of the susceptibility of an alpine ecosystem to changes in precipitation chemistry.

This study was the most detailed hydrologic investigation of a non-glacierized, alpine basin conducted to date. Despite the importance of alpine areas in generating streamflow for lowland uses [e.g., Martinelli, 1975; Leaf, 1975; Kattelmann and Berg, 1987], alpine hydrology has received little scientific attention. A variety of studies of limited scope have been carried out in mountain areas [e.g., Glen, 1982; Young, 1985], but there have been few integrated hydrologic investigations of alpine basins. Reviews of literature relating to alpine hydrology have been written by Caine [1974], Slaymaker [1974], Dozier [1987], and Clark [1988].

Only three other water balance studies in alpine basins of North America not presently glaciated are known to have been completed. The most thorough of these investigations examined the water balance of a 2 km^2 basin in the Colorado Rocky Mountains for six months [Carroll, 1974 and 1976]. Additional work continues in this basin which is designated as a long term ecological research site [e.g., Brendecke et al., 1984].

Detailed studies of a 0.04 km^2 alpine basin in the Coast range of British Columbia revealed that virtually all of the snowmelt left the catchment as surface runoff which lasted for less than two months [Jordan, 1978]. In an alpine watershed in southeastern Alaska, runoff accounted for 85 percent of measured precipitation [Stednick, 1981]. Although not in a truly alpine setting, another water balance study was conducted in a snow-dominated basin with little vegetation in northern Hokkaido [Motoyama et al., 1986]. The water balance for this 1 km^2 basin was determined for 4 two-month long snowmelt seasons. Most other mass balance studies in alpine areas have involved basins with glaciers [Slaymaker, 1974; Young, 1985] and are not particularly applicable here. Several recent or current studies of chemical cycling in alpine basins involve some hydrologic data: Gem Lake, California [Stoddard, 1987]; Eastern Brook Lake, California [Nodvin, 1987]; Mexican Cut, Colorado [Harte et al., 1985]; Loch Vale, Colorado [Baron et al., 1986]; and West Glacier Lake, Wyoming [Clow et al., 1988].

The primary objective of the investigation described here was to determine the absolute and relative magnitudes of the main water transfers within an alpine lake basin over two hydrologic years. Examination of the water balance should identify the relative importance of various hydrologic processes in this environment and provide an improved basis for assessing the impacts of acidic precipitation on high elevation ecosystems. Although much can be inferred about the hydrology of alpine basins from general hydrologic knowledge, studies at lower elevation, and various measurements obtained for other purposes, a detailed field study was necessary to adequately describe the processes occurring in a typical alpine headwater basin. Prior to this work, a comprehensive basin study had never been conducted in the alpine zone of the Sierra Nevada.

B. Measurement Program and Estimation Methods

All work reported here was contained within the 1.2 km^2 Emerald Lake basin described by Dozier et al., [1987, chapter 2] and Dracup et al. [1988, chapter 1]. Initial measurements of streamflow leaving the basin were begun in October 1983 by the U.S. Geological Survey in cooperation with the National Park Service. Studies sponsored by the California Air Resources Board began in autumn of 1984. Most hydrologic and micro-meteorologic instrumentation was installed during the summer of 1985 and was removed by June 1988. Within this period, hydrologic processes were monitored for two complete water years (October 1985 to September 1987). Data collected outside of this period were helpful in assessing natural variability and confirming results. Although simple in concept, the water balance can be difficult to measure adequately. Many components involve high spatial variability and large measurement uncertainty. Difficulties increase in a remote location under hazardous conditions combined with administrative constraints. The heavy use of the study area for backcountry recreation required that our instrumentation and field activities maintain low visibility. Measurements, instruments and analytic procedures for this project are described in detail by Marks et al., [1986]; Dozier et al., [1987]; and Dracup et al., [1988].

Ideally, the water balance of a river basin involves measuring or estimating all hydrologic inputs, losses, and changes in storage, and then comparing these to outflow from the catchment:

$$\text{Input} - \text{Losses} - \Delta\text{Storage} = \text{Outflow} + \text{Residual}$$

where:

Input is the total water input to the basin,
Losses are the water lost to evaporation and seepage from the basin,
ΔStorage is the change in water storage in the basin,
Outflow is stream discharge from the watershed, and
Residual is the quantity not accounted for,
with all terms expressed as volumes over the same time period.

If all terms in this equation are precisely measured or known, it will balance with an residual term equal to zero. However, such knowledge is generally not possible for any real hydrologic system larger than a few m^2. In the absence of complete information, the water balance can be calculated from modeled or estimated input parameters and their distribution in time and over the watershed. The error illustrates the uncertainty in the estimates and the importance of processes not accounted for. When the components balance with little residual error, either the processes are well accounted for or the errors happened to compensate for one another. Therefore, the uncertainty in individual components needs to be well defined. Uncertainty can be large due to the estimation of spatially distributed quantities from very few point measurements. Of all the water balance components at Emerald Lake, only streamflow leaving the basin is confined to a single location. All other processes have high spatial variability due to differences in water and energy availability and the physical and biological characteristics of the basin.

Water balance studies are a common and standard method of hydrological investigation [Solokov and Chapman, 1974; Van der Beken and Herrmann, 1985]. They have been used at a range of scales from plots of less than a square meter to the entire earth [e.g., UNESCO, 1971; Dooge, 1984]. Although a water balance can be simplified to just estimating combined evaporative losses and changes in storage as a simple difference between precipitation and streamflow, the procedure is more useful if each component can be identified and estimated independently. As more processes are included, a better understanding of the pathways that water takes through the basin becomes possible. When the spatial distribution of the active processes is considered over short time periods, the water balance becomes very complex. Therefore, detailed spatial and temporal resolution must be limited to maintain a manageable field program.

1. Precipitation

a. Snow Winter snowfall at Emerald Lake was measured as part of the Snow Hydrology project beginning in water year 1985 [Dozier et al., 1987]. Snowfall water equivalence was determined by measuring the density of snowfall samples of known depth from snowboards placed on the old snow surface. Event measurements are described in chapter III.

For the water balance calculations, winter precipitation was estimated from intensive snow surveys in April and May near the time of peak accumulation. An average basin-wide snowpack water equivalence was estimated from surveys of 85 to 325 points distributed throughout the basin. This estimation procedure is thoroughly described in chapter 8 of the Snow Hydrology project final report [Dozier et al., 1987] and in chapter III of this report. In an environment such as Emerald Lake, snow accumulates throughout the winter and does not begin significant melting until spring. The snow cover in early to mid-April, prior to the onset of spring melt, represents the cumulative precipitation for the winter (after adjusting for sublimation and minor midwinter melt). Therefore, an intensive snow survey was thought to be a much better approximation of accumulated precipitation than event measurements at two sites. The 90 percent error bounds calculated for the snow surveys at peak accumulation were approximately ± 10 percent in each year. Sublimation estimated with an energy balance method was 22 cm and 19 cm in the two winters. Midwinter melt was estimated from the volume of non-displacement basin outflow after accounting for rainfall runoff and groundwater discharge. Midwinter melt was estimated to be less than 10 cm in each of the years. Because most of the annual precipitation is present in the snowpack at peak accumulation, the total error in estimating precipitation from snow surveys is likely to be much less than from conventional techniques [Woo et al., 1983].

b. Rainfall Rainfall measurements at Emerald Lake were started by the Hydrology project in July 1985. Three recording rain gages (Belfort weighing bucket with a 20 cm orifice) were installed at elevations of 2800 m, 3000 m, and 3200 m. Eight pairs of non-recording rain gages (plastic funnel type with a 10 cm orifice) were distributed throughout the basin over the full range of elevations and exposures and observed after each storm. All gages were removed during the winter. Because wind reduces the collection of rainfall and snowfall by precipitation gages, actual rainfall was estimated to exceed measured rainfall by at least 10 percent. Actual snowfall during the gaged period was believed to exceed snowfall measured by the precipitation gages by at least 20 percent. These assumed correction factors were based on a review of relevant literature [e.g., WMO, 1970; Larson and Peck, 1974; Bergman, 1982; Sevruk, 1986].

The estimation of areal precipitation from point gage measurements is a persistent difficulty in hydrology. Several interpolation methods have been developed at varying levels of complexity [e.g., Rainbird, 1976]. Measured precipitation at Emerald Lake varied between gages by a different amount for each storm. However, since this difference was only a few millimeters, simple averaging was believed to be adequate to estimate average basin-wide precipitation. The non-recording gages were believed to be less reliable than the recording gages. They also tended to measure less precipitation than the recording Belfort gages. Measurements from the non-recording gages were first averaged as a group. This average was then included in an average with the three recording gages. The average of all gages was then multiplied by 1.1 in the case of rain or by 1.2 in the case of snow to estimate basin-wide precipitation. In a few cases, the basin-wide values were subjectively modified to account for other observations, such as measurements of snowfall depth and density or known gage problems. Although the final results include some subjective judgement, they represent our best estimates of basin-wide precipitation given all available information. Assessment of errors in gaged precipitation is difficult due to the absence of knowledge of "true" precipitation, but errors are often assumed to range from 10 to 60 percent, with mountainous regions being a worst case [Winter, 1981].

2. Evaporation

Evaporative losses in the Emerald Lake basin depend largely on water availability and may be distinguished on that basis as evaporation from snow, vegetation and soils, and open water. Methods for estimating evaporation are still far from certain [Brutsaert, 1989] and involve spatial extrapolation leading to considerable uncertainty.

a. Losses from Open Water and Snow Evaporation from water surfaces and snow were treated similarly by adapting a series of equations for estimating turbulent transfer of energy and mass described by Brutsaert [1982]. The computations are described in detail by Marks [1988] and Dozier et al. [1987, chapter 9.2]. The procedure required continuous measurements of air temperature, water surface temperature, relative humidity, and windspeed. Measurements obtained 5 minutes apart were averaged and recorded at 15 minute intervals at two sites (south side of the lake at 2820 m and ridge crest east of the lake at 3085 m). Data from the site adjacent to Emerald Lake were used to estimate lake evaporation. The evaporative flux from the lake surface was multiplied by the lake surface area of 28,500 m^2 to obtain the volume of evaporation from the lake during the ice-free part of the year. Data from both sites were used to calculate sublimation at those points. Basin-wide average sublimation was estimated as the mean of values from these two points. The average flux was multiplied by the snow covered area to obtain estimates of the volumetric losses per day. The term sublimation as used here includes all evaporative losses from the snow surface.

Although this procedure is physically reasonable, it could not be tested directly in this study. Extensive work at other locations during development and verification of the energy balance method for estimating lake evaporation has confirmed its excellent reliability when all of the inputs and coefficients can be accurately determined [Brutsaert, 1982]. In general, energy balance methods such as this one are considered to provide the most accurate estimates of evaporation with uncertainties usually less than 10 percent [Winter, 1981]. The lack of a physically-defensible extrapolation procedure introduces additional uncertainty and limits the need for "perfect" point estimates. Uncertainty in snow covered area adds to the error of the sublimation estimates.

b. Evaporation from Vegetation and Soils The limited availability of water during the summer months in a basin such as Emerald Lake requires careful distinction between potential and actual evapotranspiration. Potential evapotranspiration was estimated with the Penman [1948] technique. This method combines a simplified energy balance with a mass-transfer term that accounts for wind movement and vapor pressure gradient. It is explained in detail as a standard method in most hydrology texts [e.g., Dunne and Leopold, 1978]. Although it is more than forty years old, this technique is still considered the state of the art as a simplified approach to estimating potential evapotranspiration [Brutsaert, 1982; Calder et al., 1983] and has been used in other high elevation areas [Henning and Henning, 1981; Najjar et al., 1981]. In addition to measurements of temperature, vapor pressure, and wind speed, as with estimating sublimation and lake evaporation, radiation data were also needed. Net all-wave radiation was estimated from measurements of incoming solar radiation, an assumed average albedo of 0.3 [Morgan and Slusser, 1978], and longwave radiation estimated with the Brunt equation [Brunt, 1932; Anderson, 1954].

Estimates of potential evapotranspiration obtained with the Penman equation were compared with pan evaporation. Although evaporation pans do not provide a reliable measure of lake evaporation [e.g., Miller, 1977], they "probably provide the best method of obtaining an index of potential evapotranspiration" [Dunne and Leopold, 1978, p. 128]. Despite their inherent shortcomings, evaporation pans do provide an easily observed measure of evaporation at the location of interest [Thom et al., 1981]. Two evaporation pans were maintained near the lake throughout the snow-free part of the year. The pans were observed and refilled at irregular intervals of 1 to 12 days. Consequently, the water level in the pans varied beyond the preferred range (which keeps the volume relatively constant). The pan data were used for the potential evapotranspiration estimates in October 1985 when there was inadequate data to use the Penman method. A simple method of estimating potential evapotranspiration requiring only mean daily air temperature [Hamon, 1963] was also tried, but later abandoned because of poor correspondence to the pan data.

Estimation of actual evapotranspiration employed an accounting procedure that was simple in concept but required substantial subjective judgement and was based on field observations and several unverifiable assumptions. Although each part of the method and the assumptions can be debated, the overall procedure is physically reasonable and represents one estimate of water loss from soils. The accounting procedure is a simple stepped function of total storage similar to the root constant model of Calder et al. [1983]. As each idealized soil reservoir dries out, water becomes less accessible to plants, and the ratio of actual evapotranspiration to potential evapotranspiration declines in a series of steps. This means of determination of actual evapotranspiration was based on observations that indicated relatively little of the basin area was capable of contributing to evapotranspiration except immediately after rainfall when the entire basin was wet. Therefore, the soil moisture accounting procedure had to accommodate different vegetative cover densities, rooting depths, recharge characteristics, and change through time.

For estimation of actual evapotranspiration, the basin area of 120 ha was divided into 6 zones of water availability and use:

1. Lake, pond, and streams —3 ha— (dealt with separately)
2. Bare rock and colluvium —85 ha— 65% of this area was assumed to be non-transpiring and to lose water only after rainfall; 30% consisted of unvegetated soils that were assumed to evaporate at up to the rate of 0.5 mm/day for 5 days after rainfall; and 5% of this area was assumed to be covered by scattered small plants with access to 5 cm of soil moisture storage.
3. Dry grasses and forbs —14 ha— 30% of this area was vegetated and assumed to have access to 5 cm of soil moisture storage.
4. Shrubs —10 ha— 60% of this area was vegetated; 4 ha were assumed to have 5 cm of available water, 1 ha was assumed to have 10 cm of available water, and 1 ha of phreatophytes had unlimited water.
5. Wet meadows and wet rocks —5 ha— all of this area was assumed to have unlimited water for 30 days following snowmelt or rainfall. Water for these areas was supplied from upslope with little on-site storage capability.
6. Trees —3 ha— the stand of western white pine (*Pinus monticola*) in the east joint was assumed to have access to 30 cm of available water following snowmelt.

Vegetation coverage and densities were obtained from a map of Emerald Lake vegetation [Rundel et al., 1985; Rundel, 1988].

A variety of additional assumptions determined water availability in each zone. While the pine stand was still snow covered (mid-April to late May 1986 and mid-April 1987), actual evapotranspiration losses from this zone were assumed to average 4 mm per day or about 100 m^3 per day. Following snow cover disappearance in the pine stand, the trees were assumed to transpire at the potential rate until 15 cm of water storage was depleted. The remaining 15 cm of water was assumed transpired at one-half of the potential rate for 5 cm, then at 1 mm per day (or at the potential rate if potential evapotranspiration was less than 1 mm per day) for the next 5 cm, and then at 0.5 mm per day for the remaining 5 cm. Rainfall replenished part of the storage.

Maps and photographs of snow covered area allowed estimation of the dates of exposure of the different zones. Of the combined area with 5 cm of available water, 1 ha was thought to be exposed by June 15, 1986; 5 ha by July 1, 1986; and all 12 ha by July 8, 1986. Depletion of water from areas first exposed was assumed to be balanced by recharge in areas last exposed until all areas were snow-free. Depletion of the 5 cm of storage started from these areas on July 10, 1986 at half of the potential rate.

The phreatophytes were not thought to become exposed until July 1, 1986. After some of the ephemeral channels dried up, half (1 ha) of the phreatophytes were assumed to begin depleting 10 cm of stored water. Of the wet meadows in 1986, 1 ha was considered exposed by June 15 and all 5 ha were snow-free by July 1. The meadows were assumed to begin drying out after September 1. Two hectares of rocks were assumed to be exposed and wet by June 1 as snow retreated. Evaporation was assumed to occur on 5 ha of wet rocks as snow cover became patchy from June 5 to 25. The area of wet rocks then declined to 2 ha by July 15, to 1 ha by August 1, and to none by August 15. This zone of overland flow for several meters below retreating snow patches has been reported in other studies [Slaymaker, 1974; Clark, 1988].

Storms of July 21-24, August 20, September 18, September 23-24, October 1, and October 18, 1986 were assumed to provide 1 mm [Davis and Dewiest, 1966] of evaporation from surface storage throughout the basin. This evaporation was assumed to occur on the day following rainfall or was spread out over a few days when snowfall occurred. These storms replenished some of the available water in different zones and allowed the resumption of evapotranspiration at the potential rate. Actual evapotranspiration slowed greatly during October and November as energy input decreased, and it stopped completely on November 10, 1986 when the winter snow cover began.

In 1987, snow cover depletion occurred earlier and more rapidly than in 1986. The trees were assumed to transpire at the potential rate in May with soil moisture depletion beginning May 25. Fifteen centimeters of soil moisture and rainfall recharge were calculated to have been used by the trees by July 7. Transpiration by the trees continued at half the potential rate until July 20, and then at 1 mm or less. One hectare of phreatophytes was exposed by May 15 and the second became snow free by June 1. The one hectare of phreatophytes along ephemeral channels was estimated to have depleted its 10 cm of storage by June 28. Evapotranspiration from this area was assumed to occur at 1 mm per day or less until the next rain.

Wet meadows were assumed to be exposed on 2 ha by May 1, on 3 ha by May 12, and on all 5 ha by June 1. During June and July, the wet meadows dried out with an increasing proportion first transpiring at half potential and later at 1 mm per day. The combined zones with 5 cm of storage increased in area from 3 ha on May 1 to all 12 ha by June 1. The 5 cm of storage was depleted throughout the zone by June 18.

Area of wet rocks expanded to 3 ha by May 10 and contracted to zero by June 15. Small rain storms on nine days of summer in 1987 temporarily increased actual evapotranspiration when all of the basin was wet following the rain, as well as briefly replenishing some of the soil moisture. All of the assumptions and guesswork involved in the estimation of actual evapotranspiration is in error to one degree or another, particularly on a daily basis. However, the methods described above illustrate the general trend of water availability at

Emerald Lake during the spring and summer.

3. Groundwater Storage

Total groundwater storage for the Emerald Lake basin was estimated independently by different methods [Kattelmann, 1989]. Because these methods inferred storage from several other estimated quantities, uncertainties in the estimates may exceed ± 50 percent. The independently derived estimates of total storage were combined to produce a "best-estimate" of basin-wide storage. The unconsolidated materials and the fracture system were considered separately because of their different properties. Storage in the unconsolidated materials was calculated from estimates of areas, depths, and specific yields of each major water-bearing unit and independently for each category of deposit. These estimates were then combined. Areas for the individual deposits and types of deposits were obtained from an orthophoto of the basin and reports of the soils projects that were part of the Integrated Watershed Study [Huntington and Akeson, 1986; Lund et al., 1987]. Depths of non-soil deposits were estimated from visual surveys of the confining topography of deposits and guessing at an average depth that could be saturated at peak snowmelt over the area of the deposits. Soil depths obtained from drilling were found in the soils mapping report [Huntington and Akeson, 1986]. Specific yields (ratio of volume of water that can be drained by gravity to the volume of initially saturated material) were obtained from values reported in the literature for materials similar to those at Emerald Lake.

Water storage in the bedrock was estimated independently in three ways. In the first method, average specific yields of 0.01 and 0.1 percent were assumed for the upper 10 m of rock over the basin area. For the second method, the fracture volume was estimated from surface observations of fracture density. The average fracture volume was estimated to be in the range of 0.1 to 2 m^3 per 100 m^2 of surface area. A third estimate was based on the average depth of water that might be contained in the bedrock on a unit area basis. Values between 0.5 and 2 cm were estimated to be reasonable. A range of results based on extremes of reasonable values was calculated for these three methods.

Estimates of storage in unconsolidated materials and bedrock were combined into an estimate of total subsurface storage in the basin. Estimates of the total basin storage remaining after most of the snow had melted were derived from analyses of the streamflow recessions in late summer of 1985, 1986, and 1987. Because the groundwater storage estimates were not crucial to the water balance presented here, details of the procedures are not repeated but may be found in Dracup et al. [1988].

4. Snowmelt

Snowmelt was estimated with three different techniques. One method used a network of about 50 ablation stakes as described in chapter III of this report. An interpolation procedure [Renka, 1984] provided estimates of total snowpack loss over irregular periods of 4 to 8 days. The loss values were adjusted for sublimation and snow covered area to obtain a volume of snowmelt over the period. Snowmelt in August 1986 was estimated by assuming an average rate of snowmelt of 2 cm per day (from ablation stake data) over a declining snow covered proportion of 0.2 on August 1 to 0.1 on August 31. Another method determined monthly snowmelt as the difference in snowpack water equivalence extrapolated from the intensive snow surveys described in chapter III. Although the survey results are regarded as highly accurate, extrapolation to other dates introduced considerable uncertainty because these estimates were subjectively adjusted for precipitation and patterns of generated runoff. Independent estimates were made by two people and then averaged. These differences in water equivalence were then adjusted for sublimation over the month to estimate snowmelt. Estimates of snowmelt for the spring and summer of 1986 were also obtained from an energy-balance snowmelt model [Marks, 1988]. Although estimates from this model are probably a good physical representation of snowmelt at the two points providing input data, extrapolation to the basin scale is problematic. The monthly estimates of snowmelt used here are simply an average of calculations from the two measurement sites. Results from the three independent methods were averaged to obtain an estimate of monthly snowmelt for use in the water balance. Unfortunately, there were considerable differences between these estimates, and the uncertainty in the average must be regarded as up to ± 50 percent.

Crude estimates of snowmelt during months not covered by the methods described above were based on measured precipitation, observed patterns of snow cover disappearance, and snowmelt rates observed from the ablation stake data. Of the 7 cm of snowfall that occurred in October 1985, 7 cm was assumed to melt on 25 percent of the basin (21000 m^3), 3 cm was assumed to melt on another 25 percent of the basin (9000 m^3), 2 cm was assumed to melt on another 25 percent of the basin (6000 m^3), and no melt was assumed to occur on the remainder. In November 1985, 2 cm of new snowfall was assumed to melt on 25 percent of the basin (6000 m^3), and 1 cm was assumed to melt on 50 percent of the basin (6000 m^3). Snowmelt in March 1986 was estimated from streamflow adjusted for snow loading displacement and a guess of subsurface recharge (4 cm over 20 percent of the basin = 10000 m^3). About 25000 m^3 of snow water equivalence was estimated to remain on 10 percent of the basin area on September 1, 1986. Snowmelt was estimated to occur at an average rate of 1 cm per day for the first 20 days of the month as the proportion of snow covered area declined from 0.1 to 0.05. Total melt for this period was estimated as 18000 m^3. Storms on September 18 and 23-24 deposited about 13 cm of snow water equivalence in the basin. By the end of the month, all of this snow was assumed to melt in the quarter of the basin with the highest energy input (39000 m^3), 10 cm was assumed to melt on another 25 percent of the basin (30000 m^3), 7 cm was assumed to melt on another 25 percent of the basin (21000 m^3), and 3 cm was assumed to melt in the most shaded quarter of the basin (9000 m^3). In October 1986, storms deposited 3 cm of snow water equivalence. All of this new snow was assumed to melt on 75 percent of the basin and none on the remaining quarter (27000 m^3). In addition, 3 cm of the snow remaining from the September storms was assumed to melt from 25 percent of the basin (9000 m^3). In November 1986, the 1 cm of new snow was assumed to melt on 75 percent of the basin (9000 m^3), and 1 cm of residual snow was assumed to melt on one quarter of the basin (3000 m^3). Snowmelt in March 1987 was estimated from streamflow and an assumed contribution of 10000 m^3 to subsurface storage as in March 1986. Snowmelt in July, August, and September of 1987 was estimated by partitioning the snow storage remaining at the end of June among the three months on the basis of declining snow covered area.

5. Snow Covered Area

Many of the hydrologic processes in the Emerald Lake basin are dependent on snow covered area. The spatial extent of snow determines or influences the volumes of snowmelt, sublimation, and evapotranspiration. Snow covered area was estimated from a combination of aerial and ground-based photographs obtained at intervals of 7 to 14 days. Snow cover was mapped from the photographs, and the combined area of snow fields and snow patches was estimated from the maps. Snow cover depletion was interpolated between the observations. The accuracy of snow covered area determinations has been estimated as ± 5 percent for very large basins [Wiesnet, 1974]. Errors in the snow covered areas estimates for Emerald Lake could easily be of this size but seem unlikely to be greater than ± 10 percent. Gross errors are limited by the progressive nature of snow cover depletion and independent determination of snow cover for each of the dates.

6. Streamflow

Streamflow out of the Emerald Lake basin was perhaps the most important quantity monitored during this project. Streamflow integrates all of the other hydrologic processes occurring throughout the basin and is available for measurement at a single point. Despite its importance, our measurements of streamflow were less than ideal and involve considerable uncertainty. Because of the high recreational value of the site, we were not permitted to install a flow measuring structure in the channel. Therefore, we relied on an empirical rating of the stream.

Water level in the stream a few meters below the outlet was recorded by this project since August 1985. Stream discharge (volume per unit time) was calculated from the stage records with an empirical stage-discharge relationship (rating curve). Instantaneous discharge was measured with a dilution technique. The error in the discharge measurements was estimated to be about ± 10 percent. The coefficients of determination for the stage-discharge equations were above 0.95. Methods of determining streamflow are described in chapter V of this report.

7. Compilation of the Water Balance

The basic water balance equation as used in this study is of the form:

$$\text{Precipitation} + \text{Evaporation} - \Delta\text{Snow Storage} = \text{Streamflow out of the basin} + \Delta\text{Subsurface Storage and Error}$$

where precipitation is snowfall and rainfall;
evaporation is sublimation, evapotranspiration, and lake evaporation where these losses are negative in sign;
ΔSnow Storage is change in the snowpack (gross accumulation [positive] + snowmelt [negative] + sublimation [negative]);
Streamflow is measured discharge at the basin outlet; and
ΔSubsurface Storage and Error is the residual quantity.

This version of a water balance for the Emerald Lake basin was compiled from the various components described above. Starting with data recorded at 15-minute intervals, estimates of the important fluxes were extrapolated over the basin area and integrated into volumes at a monthly time step. These monthly values were combined into a water balance for each water year as well as the complete two-year period. A daily water balance was calculated at one point, but it was essentially useless due to our inability to estimate changes in storage over short periods.

Snowfall and avalanches onto the lake surface displaced about 108,000 m^3 of lake water into the outlet stream in 1986 and about 10,000 m^3 in 1987. Most of this displacement occurred during the massive storm of February 1986 when about 80,000 m^3 of water left the lake. Although this water was not generated by the usual processes of rainfall, snowmelt or groundwater drainage, it had to be considered in the total balance. However, this displaced water was deducted from streamflow values used in the water balance shown here to reduce the number of terms and simplify the presentation. Snowfall responsible for the displacement (assumed equal to the volume of water displaced) was similarly deducted from monthly snowfall in the tabulation.

C. Results

1. Precipitation

a. Precipitation Type During the period July 1985 through June 1988, 55 precipitation events (defined here as distinct storms or shower periods separated by several hours without precipitation) were measured in at least one rain gage (Table 30). In addition, about 40 snowfall events occurred during the three winters. Out of the 95 events, the precipitation type was rain in 26 cases. Twelve events consisted of mixed rain and snow. The precipitation type of the remaining 57 events was snow. Virtually all winter precipitation observed by the Snow Hydrology project fell as snow [Dozier et al., 1987]. Rain occurred briefly during two storms in winter of 1986 and once in winter of 1987. In terms of total precipitation, snowfall contributed the overwhelming majority of water to the basin. During the three years of the project, 95 percent of the precipitation fell in the form of snow (assuming the mixed events were half snow and half rain).

b. Precipitation Quantities Tabulation of monthly precipitation (Table 31) shows major differences in precipitation regime between seasons. Precipitation during the months of May through October contributed less than 15 percent of the total. Less than 2 percent of the total precipitation fell in June, July, and August. This observation was in agreement with an earlier study indicating that this same proportion of annual precipitation could be expected on average in summer throughout the high Sierra Nevada [Hannaford and Williams, 1967]. Although most of the annual precipitation in the Sierra Nevada typically falls during the months of December, January, and February [Smith, 1982], temporal distribution of precipitation was greatly different between the three winters. Each winter was notable for at least one very dry month. Monthly precipitation throughout most of California was among the lowest amounts on record in each of January 1986, December 1986, and February 1988. At the other extreme, the massive storms of February 1986 deposited almost twice the precipitation of each of the other winters.

A total of approximately 2.6 meters of precipitation fell in the Emerald Lake basin during water year 1986. About one meter of precipitation occurred in the other two years of the study. Because almost of all of these large amounts of water input fell as snow, the hydrology of the Emerald Lake basin is dominated by accumulation of a deep snowpack in winter and release of the stored precipitation in spring.

The snow surveys in April and May, 1986 indicated a total basin-wide snow water equivalence of about 200 cm at the onset of melt. Early season snow surveys in the other years indicated that average snowpack accumulation was about 60 cm plus the amount lost to evaporation in each year. These values were used as indices for precipitation which accumulated throughout the winter. Uncertainty in the snow surveys at peak accumulation was estimated to be ± 10 percent in each of the years.

c. Precipitation Intensity In general, average precipitation intensities tend to be relatively low at Emerald Lake, particularly during winter storms resulting in snowfall in the basin. During most storms, the intensity of precipitation rarely exceeds 1 or 2 mm per hour. However, during some summer thundershowers, intensities greater than 5 mm per hour may occur. Convective activity in mountain areas can result in short-term bursts of rainfall of up to 1 mm per minute.

Recording weighing-bucket rain gages at Emerald Lake permitted examination of rainfall rates (Table 32). During the 1986 water year, storms consisting solely of rain occurred on only five days. Adequate records for intensity analysis were obtained from nine events in WY 1987 and from one event in WY 1988. This limited record of rainfall provides only an indication of the range of rainfall intensities at the study area. The most intense rainfall recorded occurred on June 8, 1987 and exceeded 12 mm/hour. That storm was the only event during the three years of record that included a short-term intensity greater than 10 mm/hour. Otherwise, hourly rates were much lower. Intense rainfall could be particularly effective in mobilizing contaminants resulting from dry deposition. A longer period of record would be necessary to adequately assess rainfall intensity at Emerald Lake or other locations in the alpine Sierra Nevada.

d. Interception Interception of precipitation by vegetation plays only a minor role in the disposition of water input to the Emerald Lake basin. Less than 20 percent of the basin area is covered by vegetation, and only 3 percent of the basin area is covered by trees (primarily western white pine, *Pinus monticola*). With the exception of the trees, the basin's vegetation is snow-covered early in winter and does not intercept any precipitation except during the relatively dry, snow-free part of the year.

Most of the snow intercepted by trees later falls to the ground as wet clumps or drip [Miller, 1964]. Conifers can store from 0.3 to 9 mm of water in the form of snow [US Army, 1956]. We assumed 5 mm storage and 20 percent evaporative loss from storage [Satterlund and Haupt, 1970] to estimate a 1 mm loss from each snowstorm over the tree-covered area. The amount of water lost was negligible in the overall basin water balance.

Interception loss from rainfall was estimated as 2 mm from trees and 1 mm from shrubs, forbs, and grass. These estimates were based on a compilation of interception storage data [Zinke, 1967]. Again, the magnitude of the loss is a small fraction of a millimeter over the basin area and is much less than the uncertainty associated with the rainfall estimates. Therefore, interception loss was not measured or included in calculation of the water balance.

2. Evaporation

Water losses to the atmosphere are the only output from the Emerald Lake basin other than streamflow. These evaporative losses are small compared to the discharge out of the basin. Although evaporation dominates the water balance in most of the United States, consuming about two-thirds of the average precipitation [Dunne and Leopold 1978], the proportion is much less in mountain areas [e.g., Carroll, 1976; Johnson and Brown, 1979]. In the European Alps, evaporation is regarded as less than 10 percent of precipitation [Lang, 1981]. In water years 1986 and 1987, estimated total evaporation of 963,000 m^3 from snow, water surfaces, soil, and vegetation at Emerald Lake was 22 percent of the estimated precipitation of 4.4 million m^3. In 1986 alone, the percentage was 19. In 1987, it was 32.

Evaporation from snow was the principal water loss to the atmosphere, accounting for about 80 percent of the total evaporation (84% in 1986 and 73% in 1987). Evaporation from non-snow surfaces is limited due to the small proportion of the basin that is covered by water or vegetation. Calculations of potential evapotranspiration indicate that 4 or 5 mm/day of water could evaporate under typical conditions of spring and summer. Up to 7 mm/day could evaporate under ideal conditions (high vapor pressure gradient and wind). However, relatively few areas of the basin provide such an opportunity by having water available at or near the surface. The greatest opportunity for evaporation is during snowmelt runoff when large areas below snow patches are wet. Transpiration by vegetation is also at a maximum when soil moisture is high from snowmelt recharge. Evaporation from the lake and streams occurs at a relatively high rate for an extended period of time, but the area of open water is less than 3 ha.

a. Evaporation from Open Water Open water surfaces in the Emerald Lake basin provide the most obvious opportunity for evaporative losses. However, because such surfaces occupy less than 3 percent of the basin area, the total volume of evaporation is small. Evaporation from lakes has been studied extensively around the world, and a wide variety of methods have been developed to estimate lake evaporation [e.g., Harbeck and Meyers, 1970; Morton, 1983; Veihmeyer, 1964]. However, relatively little work on evaporation has been done at high altitudes. A review of water balance techniques concluded, "For mountain regions, there are no reliable methods of measurement of evaporation..." [Solokov and Chapman, 1974, p. 47]. Nevertheless, a variety of measurements and estimates of evaporation in the mountain environment have been made.

The decline in atmospheric pressure with increased elevation had long been assumed to allow increased evaporation. However, the lower air and water temperatures and reduced vapor pressure gradients should theoretically result in a decrease in evaporation with increasing elevation [Horton, 1934; Price, 1981; Veihmeyer, 1964]. Field measurements of evaporation from small pans along elevation profiles have demonstrated a substantial decline with elevation [Blaney, 1958; Fortier, 1907; Longacre and Blaney, 1962; Peck and Pfankuch, 1963]. However, confounding influences prevent development of a simple lapse rate of evaporation with altitude.

Data from studies in mountain areas provide an indication of the magnitude of evaporation to be expected from Emerald Lake. Water loss was measured from evaporation pans maintained for 13 years at Kaiser Pass, a site that coincidentally is at the same elevation as Emerald Lake (2800 m) and only 85 km to the north. Estimated monthly lake evaporation at that site was 12, 15, 14, 12, 8, and 5 cm for the months of June through November, respectively [Longacre and Blaney, 1962]. Mean daily pan evaporation at three sites in the Wasatch mountains of Utah ranged from 6 to 8 mm [Peck and Pfankuch, 1963]. Average evaporation from a small pond in the White Mountains (just east of the Sierra Nevada) at 3500 m in mid-July was calculated from energy balance measurements to be about 5 mm/day [Terjung et al., 1969]. The average value for the Sierra Nevada from a map of annual lake evaporation for the United States is less than 90 cm [Kohler et al., 1959]. If this amount of evaporation is assumed to occur during an ice-free period of about 150 days, then the average rate would be about 6 mm/day. Evaporation in a mountain stream has been measured in Utah as about 1-2 mm/day [Croft, 1948].

Lake evaporation was calculated with the energy balance method for the ice-free portion of the year (Tables 33 and 34). In water year 1986, the lake was open for only 100 days. In water year 1987, Emerald Lake was free of ice for almost six months. Total evaporation from the lake during the two years only accounted for about one-half of one percent of the total precipitation. However, the daily flux averaged 3 to 5 mm/day in summer. Larger lakes ice-free for longer periods could lose substantial amounts of water to the atmosphere under these conditions. The monthly totals of 9 cm in June ('87), 13 cm in July ('87), 11 and 15 cm in August (both years), 10 and 12 cm in September, and 6 and 7 cm in October were quite close to the values reported at Kaiser Pass [Longacre and Blaney, 1962]. The monthly totals for the mid-summer months also corresponded closely (less than 15 percent difference) with evaporation pan data. Lake surface temperatures were 2° to 4°C warmer in August and September of 1987 than in the same months of 1986 and should explain the greater calculated evaporation in 1987.

b. Evaporation from Vegetation and Soils Potential evapotranspiration was calculated using the Penman [1948] method from June through November of 1986 and May through September of 1987 (Tables 35 and 36). Average daily values increased from about 2 mm in May ('87) to 4 - 5 mm in June, July, and August and then declined to 2 - 3 mm in September and about 1 mm in October and November. Potential evapotranspiration was estimated from evaporation pan data for some of these months (Table 37). Remarkably good correspondence was found between the estimates obtained obtained with both methods for the periods late August-September 1986 (110 vs 102 mm), July 1987 (145 vs. 146 mm), August 1987 (163 vs. 154 mm), and early September 1987 (37 vs. 39 mm). However, this comparison may indicate that the Penman estimates are high for the basin as a whole because the lake area where the pans are located should receive more energy than the basin average.

The monthly amounts of potential evapotranspiration fall within the ranges calculated for various high-altitude sites by Henning and Henning [1981], although none of their sites have a combination of latitude, elevation, and precipitation similar to that of Emerald Lake. An average potential evapotranspiration rate of 6 mm/day from alpine tundra vegetation was calculated at higher elevation (3580 m) in the more arid White Mountains in mid-July [Terjung et al., 1969]. Transpiration from grassland at 2560 m in the Austrian Alps was estimated from an energy balance to average 2.5 mm/day in July and August [Staudinger and Rott, 1981]. A water balance study using the Hamon [1963] method in Colorado estimated potential evapotranspiration in summer to average about 1.8 mm/day [Carroll, 1976]. Transpiration from alpine plants in the Sierra Nevada has been observed to range from 1.5 to 3 mm/day at a moist site and to be less than 1 mm/day at a dry site [Mooney et al., 1965]. In Colorado, actual evapotranspiration from alpine tundra has been thought to range from 0.3 to 0.8 mm/day [Webber, 1974; cited by Carroll, 1976].

Actual evapotranspiration reflects the availability of water and the opportunity for evaporation. Actual evapotranspiration estimated over 15 non-winter months totaled 166,000 m^3 of water or less than 4 percent of the precipitation during the two water years (Tables 38 and 39). Total evapotranspiration was about the same in the two years: 84,800 m^3 in 1986 and 81,800 m^3 in 1987. Therefore, actual evapotranspiration was a smaller proportion of precipitation in 1986 (.03) than in 1987 (.07). Persistent snow cover in 1986 delayed evapotranspiration losses compared to 1987. Peak monthly losses of more than 25,000 m^3 occurred in July 1986 and June 1987. Most of the high daily losses (exceeding 1000 m^3) occurred following precipitation when the entire basin was briefly wetted. Because of the layers of assumptions and inability to verify the estimates, daily values of evapotranspiration could easily be in error by large amounts (75 percent less to several hundred percent greater than the estimates). However, the compensating nature of some of the assumptions should limit potential error in the seasonal totals to between half and three times the calculated values in the extreme case. These error estimates, like the evapotranspiration estimates themselves, are limited to guesswork based on field observations and some physical limits.

c. *Evaporative Losses from Snow* Sublimation was calculated as an average flux (Tables 40 and 41) and as a total volume after adjusting for snow covered area (Tables 42 and 43). Calculated sublimation was typically between 1 and 2 mm/day with extremes related to high vapor pressure gradients and high winds. A literature review of dozens of studies around the world [Slaughter, 1970] and an analysis of atmospheric conditions favorable to evaporation from snow [Stewart, 1982] suggest that these values are reasonable. Unfortunately, there is no direct means of verifying these estimates. Comparison of snowfall measurements and snowpack water equivalence measured in nearby snowpits indicates the calculated sublimation values are too high. However, the uncertainty due to spatial variability in this comparison prohibits any definite conclusions. Nevertheless, this comparison suggests that the evaporation calculations are not likely to be underestimates. The procedure used has a strong physical basis, but has not been verified for snow to the same extent as energy-balance estimates of evaporation from free water surfaces. Mid-winter estimates (December through March) are not directly included in the water balance since they were used in estimating precipitation from the snow survey data and effectively cancel themselves. Monthly total fluxes varied inversely with storm duration. Low fluxes calculated in May and June 1987 resulted from a weather pattern that favored cloudiness and high humidity.

Evaporative losses from snow were the largest component of total evaporation, accounting for 80 percent of the total loss to the atmosphere. Sublimation over both water years was 18 percent of total precipitation. In water year 1986, sublimation was 16 percent of total precipitation, and in 1987, it was 23 of the annual precipitation. Total sublimation in 1986 was almost twice that in 1987, largely because of the greater duration of snow cover.

3. Groundwater

Field observations and streamflow characteristics allowed some generalizations to be made about the nature of subsurface water in the Emerald Lake basin and its role in the water balance. Groundwater storage and release account for only a small portion of the total quantity of water in the annual water balance of the Emerald Lake basin. However, subsurface water is very important in temporal distribution of water. Releases from subsurface storage are the primary water input to Emerald Lake for eight to nine months of the year. Although the quantity of this water is small compared to snowmelt runoff, groundwater discharged from springs and seeps has the potential to control the lake chemistry for more than two-thirds of the year. Streamflow from April through August of 1986 was 90 percent of the annual total. In water year 1987, streamflow from April through June was 85 percent of the annual amount. Most of the other 10 to 15 percent of streamflow in these years can be assumed to have had longer contact with soils and rocks than the water generated during the spring and summer snowmelt seasons. Groundwater discharge provides Emerald Lake with up to 2000 m^3 per day inflow in summer and a few hundred m^3 per day of water during winter.

The principal groundwater reservoirs of the basin are glacial till of Alta Cirque, unconsolidated deposits in both parts of the master joint, soils on Aaron's Bench, and talus of the upper bench and area west of Danny's Hole. Additional storage is present in talus and colluvial deposits of smaller areal extent, scattered areas of shallow soil, and the fracture system of the bedrock.

In general, groundwater storage in the Emerald Lake basin appears to change rapidly in response to input and drainage. During active snowmelt, melt water takes a variety to paths to the streams. Much of the water flows over impermeable or saturated surfaces and has minimal contact with geologic materials. Some of the water infiltrates and moves quickly through a soil or colluvial deposit, possibly entering a stream channel within a few hours of infiltration. Other portions of the melt water remain below the ground surface for longer periods, slowly moving downslope over days, weeks, and months. Storage should be at a maximum during snowmelt just before the snow covered area begins to decline. Until that time, recharge from snowmelt should maintain high water levels and may completely fill the subsurface void space over much of the basin. When water input from snowmelt or drainage from upslope ceases at a given place, water drains from the large pores within a few hours or days. Drainage from the small pores then continues at a slow and declining rate.

As snow cover disappears, water drains from the exposed areas scattered throughout the basin. During late spring and early summer, streamflow is generated from a mosaic of melting snowpatches and areas of slowly-draining bare ground. When snow covers less than about 5 percent of the basin or when snow is not melting, groundwater release should account for most of the streamflow. Because of steep slopes and the high proportion of total volume assumed to exist in large pores, most of the groundwater drainage is assumed to be rapid and to occur while the basin is still largely snow covered. Groundwater storage is partially recharged by precipitation in summer and fall. Small quantities of streamflow (<500 m^3 per day) are generated during winter from a combination of groundwater discharge and snowmelt caused by ground heat.

a. Storage in Surficial Deposits and Bedrock Combination of two independently-derived estimates indicates that total storage in the unconsolidated materials is about 100,000 m^3 [Dracup et al., 1988]. Considering alternate values of depths and specific yields up to extremes that appear to be physically possible, total storage in the unconsolidated materials would be unlikely to be below 50,000 m^3 or above 200,000 m^3.

The granitic bedrock of the Emerald Lake basin is essentially impervious [Moore and Wahrhaftig, 1984]. Water flow and storage can take place in the fracture systems of the crystalline rock. However, fracturing in the absence of weathering does not increase overall porosities by more than 2 to 5 percent [Davis and DeWiest, 1966]. Furthermore, cracks in granitic rock are usually closed and rarely more than 2 mm wide [Davis and DeWiest, 1966]. Bore holes in the Wolverton area (4 km west of Emerald Lake) revealed no fractures below 10 m [Akers, 1984; cited by Moore and Wahrhaftig, 1984]. Estimates of annually-exchanged storage in the fractures obtained with the three methods described by Dracup et al., [1988] ranged from 2,000 m^3 to 20,000 m^3. Additional water may be stored in deep fractures that are not flushed on an annual basis.

The volume of available storage in the fractures is a small proportion of that in the unconsolidated materials. Because the fracture storage is well within the bounds of uncertainty surrounding the estimate of storage in the unconsolidated materials, the total groundwater storage in the Emerald Lake basin may be estimated as 120,000 $m^3 \pm 60{,}000\ m^3$. This volume is equivalent to 10 cm storage averaged over the basin area of 1.2 km^2.

Considering the nature of the bedrock in the basin, there is no reason to believe that any appreciable amount of water crosses the surface drainage divides through subsurface routes. Loss of water from the lake was of particular concern. However, field surveys downslope of Emerald Lake found no springs or other evidence of subsurface drainage from the lake. During August 1985 and September 1987, periods of minimum streamflow, all of the water entering the lake could be accounted for by the lake outflow and estimated evaporation from the lake surface.

Recession coefficients calculated for the periods July 31 to August 28, 1985, August 30 to September 30, 1986, and July 1 to August 30, 1987 were 0.941, 0.930, and 0.939, respectively. These periods were chosen to minimize snowmelt and precipitation input. Storage volumes calculated from streamflow recession were 11000 m^3 at the end of July 1985, 34000 m^3 at the end of August 1986, and 35000 m^3 at the beginning of July 1987. Storage in the eastern part of the master joint was calculated in a similar manner and was found to be about 2300 m^3 on July 15, 1986 and about 3000 m^3 on May 27, 1987 or about

20-25 percent of the estimated total storage. Considerable amounts of groundwater were likely to have drained out before these dates while snowmelt still dominated streamflow. Therefore, these estimates of storage from recession analysis are quite conservative and represent only the residual filled-storage that will supply streamflow in late summer and early fall.

b. Groundwater Residence Time The residence time of the groundwater in the basin varies between a few days and a few months. Groundwater discharged in late summer probably has been in contact with the rocks for 1 to 4 months. Groundwater discharged during late fall and winter probably has been in subsurface storage for several months or is present due to recharge from autumn rains. The total quantity of water with long subsurface residence time appears to be small (<10000 m^3). However, it may be the main water input to the lake during several months of the year.

4. Snowmelt and Snow Covered Area

Estimates of monthly snowmelt were obtained with several different methods over the study period (Table 44). When available data allowed more than one method to be used for the same month, estimates obtained with the different methods were considerably different in some of these months. In June, July and August of 1986 and May of 1987, the largest of the estimates was about 50 percent greater than the smallest. Fortunately, snowmelt was used only to alter the timing of the monthly water balance. Better estimates of snowmelt will have to await the completion of a spatially-distributed snowmelt model.

Snow covered area was crucial to a variety of other estimates but was not used directly in the water balance. Depletion of snow cover occurred much more slowly in 1986 than in 1987 (Figure 6) due to the thicker snowpack in 1986. Snow coverage declined from 90 percent to 10 percent in 130 days in 1986, but the same amount of depletion took only half as long in 1987.

5. Streamflow

Streamflow characteristics are described in chapter V of this report. The main points relevant here are the obvious concentration of annual flow during the spring snowmelt period, the long recession through summer, autumn, and winter, and the threefold difference between the two water years. In 1986, 90 percent of the 2.6 million m^3 of streamflow leaving the basin occurred from April through August. Snowmelt runoff in 1987 from April through June accounted for 86 percent of the annual amount of 820,000 m^3. Most of the flow during the winter months was a combination of slow drainage out of subsurface storage, displacement of lake water by snowfall and avalanches, and a minor amount of snowmelt from stored ground heat. Consequently, streamflow in the winter months was a small fraction of that occurring in the spring. The uncertainties in daily discharge reported in chapter V would suggest that uncertainties in the monthly totals were 10 to 15 percent in October through January, March, and September of WY1986 as well as in October through March and July through September of WY1987. Uncertainty in the monthly values for the other months with relatively high flow was estimated to be about 15 to 20 percent.

6. Water Balance

The water balance for Emerald Lake during the study period can be examined at different time scales. At the coarsest level, a simplified balance for the entire period of two years was calculated. Roughly similar conditions of a dry summer and September storms occurred before both the beginning and end of the period, so snowpack and subsurface storage were considered approximately equal at the start and end. Total precipitation (4,401,000 m^3) - total losses to the atmosphere (960,000 m^3) = total streamflow (3,396,000 m^3) - residual (45,000 m^3). Expressed as water depths averaged over the catchment area, total precipitation (367 cm) - total losses to the atmosphere (80 cm) = total streamflow (283 cm) + residual (4 cm). The residual is about 1 percent of total precipitation. This remarkably good fit seems to good to be true. And, of course, it is. Unfortunately, the good closure of the water balance is due to the fortuitous combination of compensating errors.

The actual degree of error begins to appear on an annual basis. In 1986, precipitation (3,164,000 m^3) - losses to the atmosphere (589,000 m^3) = streamflow (2,575,000 m^3). Despite the apparent perfect fit, storage conditions of snow cover and groundwater at the beginning and end of the year differed by about 150,000 m^3. Nevertheless, this error is still less than 5 percent of the precipitation. In water year 1987, precipitation (1,237,000 m^3) - losses (370,000 m^3) = streamflow (821,000 m^3) + residual (46,000 m^3). The storage error for 1987 is of the same order of magnitude as in 1986 but in the opposite direction. Because storage was higher at the beginning of the water year than at the end, this storage error adds to the simple-balance residual for a total residual of about 200,000 m^3. This residual is about 16 percent of precipitation.

Error in the water balance was also estimated by combining errors in individual components. The procedure used here was similar to that used by Winter [1981] and LaBaugh [1985]. Estimated proportional error bounds were multiplied by the total volume of water in each component for each year to produce a set of minimum and maximum volumetric errors (Table 45). The total error was calculated as the square root of the sum of squared errors. In water year 1986, this uncertainty was estimated to be between 430,000 and 620,000 m^3, or 14 to 20 percent of the annual precipitation. In water year 1987, the same procedure yielded an estimated uncertainty in the water balance of 150,000 to 220,000 m^3, or 12 to 18 percent of the year's precipitation. Examination of Table 46 indicates that streamflow accounts for the largest part of the total error. Errors in evaporation from the lake, rainfall, and evapotranspiration contribute relatively little to the total error. These estimates of error are one reasonably objective assessment of the overall reliability of the water balance for Emerald Lake.

The monthly water balance for the Emerald Lake basin (Table 46) demonstrates the highly seasonal nature of the major hydrologic processes in this mountain catchment. The change in snowpack storage summarizes most of the activity in the basin. This term is positive through March as the snowpack accumulates and is negative in April through August as the snowpack melts. The magnitude of the streamflow reflects these changes in the snowpack. The individual components have been examined in the sections above.

The *subsurface storage and residual* term is the quantity remaining after subtracting the other terms from precipitation. The *net storage and residual* term accumulates the monthly residuals after starting with an estimated groundwater storage of 48,000 m^3. This term enables evaluation of the monthly water balance. The values of the cumulative residual should not exceed the estimate of total subsurface storage capacity and, therefore, indicate reasonable results through April of 1986. Our best estimate of available groundwater storage was about 120,000 m^3 or within the range of 60,000 to 180,000 m^3, given the high degree of uncertainty in this estimate. The cumulative residual values in excess of 200,000 for May and June suggest serious errors in estimated melt and/or streamflow in May when the monthly residual is 263,000 m^3. Similarly, the residual in July of -233,400 m^3, which led to a negative value in net storage the following month, suggests an opposite imbalance between snowmelt and streamflow. Although groundwater recharge can be expected in May and groundwater drainage can be expected in July, the quantities are simply too large to be accomodated by the limited subsurface storage capacity of this basin.

In water year 1987, the accumulated residual does not indicate any problems except a minor degree of excess depletion in August. This negative value could easily be a result of lingering error from 1986. All of the monthly residuals in 1987 appear reasonable in magnitude and sign considering the time of year. However, the sum of the Δ Snow Storage column suggests a serious imbalance between estimated snowfall and snowmelt in 1987. A storm near the end of September 1986 resulted in an initial snow storage value of 50,000 to 100,000 m^3 at the beginning of water year 1987. At the end of the year, snow storage in the basin was known to be less than 10,000 m^3. Therefore, the calculated excess of 96,000 m^3 in snow storage indicates that estimated snowfall exceeded estimated snowmelt by 150,000 to 200,000 m^3. Snowmelt estimates have a much weaker basis than the snowfall estimates, but large errors in snowmelt suggest that streamflow in spring of 1987 must also be in serious error because these two terms are in reasonable agreement for that period. Despite the errors, the monthly water balance illustrates the changing relative importance of different components throughout the year. The hydrologic behavoir of the Emerald Lake basin can be separated into three distinct seasons dominated by different processes: snow accumulation from November through March, snowmelt from April through June or July, and a drainage and drying period with some snowmelt and precipitation from July or August through October.

D. Summary and Conclusions

Water balances for water years 1986 and 1987 indicate that snow accounted for 95 percent of the precipitation and eventual streamflow. Total precipitation values in the two years were about 2.6 m and 1.0 m. The snow surface provided 80 percent of the total evaporative losses. Groundwater storage and release affected only a small quantity of the water moving through this largely impermeable basin, but sustained flow into the lake during about eight months.

The peak snowpack water equivalence before the onset of spring melt is a reference for alpine hydrology. In 1986, about 90 percent of the water stored in the snowpack in mid-April plus subsequent precipitation became streamflow. In 1987, about 75 percent of the peak snow storage plus subsequent precipitation became streamflow. Evaporation in spring and summer was roughly the same for the two years.

The quantitative estimates from this water balance must be tempered by uncertainties in the measurements. Errors in streamflow were of greatest significance because stream discharge was a key component of the water balance. If streamflow had been known to be reliable within ± 5 percent, it could have provided a solid basis for evaluating errors in the other components. Hydraulic flow-measuring structures (weirs or flumes) would improve the accuracy of discharge measurements. Particular effort should be directed toward streamflow measurement because it can be measured at a single location, whereas all of the other components involve problems of spatial variability. The use of an intensive snow survey at peak snowpack is the best means of evaluating winter precipitation. A spatially-distributed snowmelt model and careful measurement of a dense network of ablation stakes provide reliable estimates of snowmelt.

TABLE 30. Gaged Precipitation in Emerald Lake Basin (mm)

Date	Type	Lake	Mid-level	Cirque	Average of Non-recording	Basin-wide Estimate
1985						
July 18-19	Rain	NR	8	9	7	9
July 24-25	Rain	3	5	5	3	4
Sept. 3-4	Snow	15	17	18	16	20
Sept. 10-11	Snow	29	30	24	NR	33
Sept. 18	Snow	13	11	10	10	13
Sept. 27	Rain	3	3	2	2	3
WY 1986						
Oct. 6	Rain/Snow	16	17	17	14	18
Oct. 8	Snow	11	13	11	9	13
Oct. 21	Snow	38	39	35	NR	45
July 21-24	Rain	22	NR	NR	21	24
Aug. 20	Rain	4	3	4	3	4
Sept. 18	Rain/Snow	10	11	11	NR	12
Sept. 23-24	Snow	76	89	91	NR	125
WY 1987						
Oct. 1	Snow	19	21	24	NR	26
Oct. 18	Snow	2	NR	NR	NR	2
May 8	Rain	NR	NR	NR	10	11
May 9	Rain	NR	NR	NR	8	9
May 12	Rain	4	8	NR	6	7
May 13	Rain	tr	0.5	NR	1	1
May 14	Rain	1	1	NR	NR	1
May 15	Rain	15	16	NR	11	15
May 21-24	Snow/Rain	24	29	NR	20	29
June 6	Rain	12	9	8	13	12
June 8	Hail/Rain	14	14	10	12	14
July 13	Rain	0.5	1	1	NR	1
Aug. 3	Rain	0.5	tr	NR	NR	0.5
Aug. 24	Rain	2	4	4	3	4
Aug. 28						tr
Aug. 30						tr
Aug. 31	Rain	2	1	2	2	2
Sep. 1	Rain	4	4	4	3	4
Sep. 2	Rain	7	8	8	7	8
Sep. 12	Rain	4	3	3	4	4

Gaged Precipitation in Emerald Lake Basin (mm) (continued)

Date	Type	Lake	Average of Non-recording	Basin-wide Estimate
Water Year 1988				
Oct. 12	Rain	4	NR	4
Oct. 21	Rain	2	2	2
Oct. 22	Rain/Snow	11	11	13
Oct. 23	Rain/Snow	13	13	15
Oct. 24	Rain/Snow	10	10	12
Oct. 27	Rain	8	8	9
Oct. 28	Rain/Snow	11	11	13
Oct. 31	Rain/Snow	3	3	3
Nov. 1	Rain/Snow	11	10	12
Nov. 2	Rain/Snow	3	NR	3
Nov. 3	Rain/Snow	14	NR	16
Nov. 4	Rain/Snow	8	NR	9
Nov. 5	Rain/Snow	5	7	7
Nov. 6	Snow	14	11	15
Nov. 13	Snow	5	6	7
Nov. 17	Rain/Snow	10	9	11
Nov. 20	Snow	12	12	14
Apr. 15	Snow	11	NR	13
Apr. 16	Snow	6	NR	7
Apr. 20	Snow	3	NR	4
Apr. 21	Snow	10	NR	12
Apr. 22	Snow	2	NR	2
Apr. 24	Snow	43	NR	52
Apr. 25-May 2	Snow	8	NR	10
May 6	Snow	10	NR	12
May 8	tr			
May 28-29	Snow	NR	NR	~30
June 18-20	Rain	5	NR	6
July 22-23	Rain	15	NR	18
July 29-30	Rain	NR	NR	~2
Aug 25-Sep 2	Rain	70	NR	~80
Sep. 21	Rain	NR	NR	~10

TABLE 31. Estimated Basin-wide Precipitation by Month (cm)

Month	WY 1986	WY 1987	WY 1988
October	8	3	7
November	32	1	9
December	15	2	36
January	10	23	33
February	125	26	2
March	40	26	6
April	11	10	12
May	5	7	5
June	0	3	1
July	2	0.1	2
August	0.4	1	7
September	14	2	2
Total	262	104	122

TABLE 32. Hourly Rainfall Depths from Belfort Weighing Bucket Rain Gages (mm)

(notes: gage catch as recorded (unadjusted for wind effects) time is PST; only those few storms in which all of the precipitation fell as rain are included; mid-level and cirque gages were still buried under snow through July '86)

WY 1986	Lake		Mid-level		Cirque	
July 21	1715-1815	7.6				
	2000-2100	0.3		NR		NR
	2300-2400	2.8				
July 22	0800-0900	0.5				
	0900-1000	0.8				
	1000-1100	0		NR		NR
	1100-1200	2.5				
	1200-1300	2.6				
	1300-1400	0.5				
July 23	1200-1230	1.8				
	1400-1500	0.5		NR		NR
	1500-1600	0.7				
July 24	1400-1500	0.8		NR		NR
Aug. 20	0900-1200	2.5	0800-0900	1.0	0800-0900	0.5
			0900-1000	1.3	0900-1000	1.3
					1000-1100	1.5
	1400-1500	1.3	1400-1500	0.8	1400-1500	0.5

Hourly Rainfall Depths (continued)

WY 1987	Lake		Mid-level		Cirque	
May 12	1500-1600	0.8				
	1700-1800	0.3		NR		NR
	1800-1900	3.0				
May 15	1200-1300	5.1				
	1300-1400	6.4		NR		NR
	1400-1500	3.8				
	1500-1600	0.3				
June 6	0400-0500	6.4	0400-0500	5.1	0400-0500	3.0
	0500-0600	0.8	0500-0600	0.8	0500-0600	1.3
	0600-0700	5.3	0600-0700	3.0	0600-0700	1.5
					0700-0800	2.5
June 8	1400-1500	1.3	1400-1500	1.3		NR
	1600-1730	13.0	1600-1730	12.7		
Aug. 24	1830-1845	1.8		NR		NR
Aug. 31	1900-1915	1.8	1900-1915	1.3	1900-1915	1.8
Sept. 1	1500-1600	1.5	1500-1600	1.3	1500-1600	0.8
	1600-1700	2.0	1600-1700	2.5	1600-1700	3.0
Sept. 2	1500-1600	2.0	1500-1600	2.5	1500-1600	0
	1600-1700	2.5	1600-1700	4.1	1600-1700	5.8
	1700-1800	1.3	1700-1800	0.3	1700-1800	1.8
	1800-1900	0.8	1800-1900	0.3	1800-1900	0
	1900-2000	0.5	1900-2000	0.5	1900-2000	0.8
Sept. 12	1700-1800	2.0	1700-1800	0.8		snow
	1800-1900	1.8	1800-1900	1.3		
	1900-2000	0	1900-2000	0.8		
	2000-2100	0	2000-2100	0.3		
WY 1988						
Oct. 27	1700-1800	3.3				
	1800-1900	1.3		NR		NR
	1900-2000	1.3				

TABLE 33. Estimated Evaporation from Lake Surface — WY 1986 (m^3)

Water Year 1986	1985 October	1985 November	1986 August	1986 September
1	69	76	**thaw**	120
2	58	54	86	120
3	86	98	100	137
4	99	92	111	160
5	122	92	114	154
6	99	92	91	140
7	89	119	103	111
8	82	91	97	71
9	86	81	108	91
10	87	**freeze**	100	94
11	59		103	97
12	67		91	80
13	69		77	94
14	99		94	88
15	115		103	77
16	86		94	91
17	42		111	74
18	27		88	71
19	33		83	66
20	25		68	66
21	29		108	88
22	20		108	77
23	19		103	46
24	19		108	134
25	26		111	63
26	22		114	68
27	27		103	37
28	28		97	68
29	19		111	74
30	15		108	74
31	25		114	
Monthly Total	1750	800	3090	2730

TABLE 34. Estimated Evaporation from Lake Surface — WY 1987 (m^3)

Water Year 1987	1986 October	1986 November	1987 May	1987 June	1987 July	1987 August	1987 September
1	75	142		90	67	152	127
2	60	98		81	111	165	141
3	85	148		104	79	145	107
4	100	107		82	80	124	98
5	123	86		75	101	149	86
6	101	60		31	107	148	116
7	93	45		41	105	113	116
8	86	41		58	79	131	147
9	81	56		40	77	168	170
10	71	60		61	64	120	127
11	50	74		93	86	111	118
12	56	73		94	80	109	54
13	60	60		72	111	72	74
14	89	46		80	122	80	118
15	100	39		62	153	157	109
16	75	37		74	118	147	105
17	51	33		69	203	128	105
18	87	**freeze**		77	98	117	117
19	60		**thaw**	76	105	117	118
20	45		29	48	141	152	138
21	48		20	72	97	186	153
22	46		23	93	106	217	135
23	46		32	94	96	156	94
24	51		26	95	92	152	64
25	66		38	119	140	160	100
26	52		21	165	224	174	93
27	62		23	153	217	179	105
28	59		33	120	127	145	111
29	43		14	135	111	139	121
30	35		35	116	144	147	137
31	54		43		144	134	
Monthly Total	2110	1210	410	2570	3590	4390	3400

TABLE 35. Potential Evapotranspiration — WY 1986 (mm) (Based on Method of Penman, 1948)

Water Year 1986	**June** 1986	**July** 1986	**August** 1986	**September** 1986
1	6	5	5	4
2	5	6	5	5
3	5	7	5	5
4	4	5	5	5
5	5	4	5	5
6	5	4	5	4
7	4	3	5	3
8	5	4	5	2
9	6	4	5	4
10	6	5	5	3
11	5	5	5	4
12	5	4	4	3
13	6	5	4	3
14	5	4	5	3
15	4	4	3	2
16	5	5	4	2
17	5	5	5	1
18	5	4	1	2
19	6	5	4	1
20	6	4	1	1
21	6	3	5	2
22	7	1	4	2
23	7	1	4	0
24	6	2	5	0
25	6	5	4	0
26	5	2	4	1
27	6	3	3	0
28	4	4	4	1
29	5	5	4	2
30	5	5	3	2
31		5	4	
Total	160	128	130	72

TABLE 36. Potential Evapotranspiration — WY 1987 (mm) (Based on Method of Penman, 1948)

Water Year 1987	**October** 1986	**November** 1986	**May** 1987	**June** 1987	**July** 1987	**August** 1987	**September** 1987
1	0	3	2	5	3	7	3
2	0	3	3	6	5	6	3
3	1	5	4	6	4	4	3
4	2	4	4	5	4	4	3
5	4	2	5	5	5	6	3
6	2	2	5	2	5	5	3
7	2	1	2	2	5	5	4
8	2	1	2	2	5	6	4
9	2	1	0	2	4	6	5
10	2	1	0	4	4	5	4
11	1	2	1	6	5	3	4
12	1	1	1	7	5	4	1
13	2	2	2	5	3	3	2
14	2	1	2	4	5	2	4
15	2	1	0	3	4	5	3
16	2	1	0	4	3	5	2
17	1	1	1	4	4	5	3
18	0		3	5	3	5	4
19	1		4	5	3	5	4
20	1		0	4	4	5	4
21	1		0	4	4	6	5
22	1		1	5	5	7	5
23	1		2	5	5	6	1
24	1		2	6	5	5	1
25	2		1	6	6	6	2
26	1		1	5	8	6	2
27	2		2	7	7	6	3
28	1		1	6	6	4	3
29	1		1	6	5	4	4
30	0		3	5	6	4	4
31	1		3		6	4	
Total	40	30	56	141	146	154	96

TABLE 37. Potential Evapotranspiration Estimated from Evaporation Pan Data (mm)

Date	October 1985	August 1986	Sept 1986	July 1987	August 1987	Sept 1987
1	4	nr	3	4	7	0
2	4	nr	4	5	6	0
3	4	nr	4	5	2	2
4	4	nr	5	5	7	4
5	3	3	5	6	7	4
6	3	2	5	6	6	5
7	3	2	5	6	6	5
8	0	2	5	6	7	5
9	2	2	4	5	6	4
10	3	2	4	5	6	4
11	4	2	4	5	6	4
12	3	2	4	5	5	0
13	3	2	3	1	5	nr
14	3	2	3	4	5	
15	3	2	2	4	5	
16	3	2	2	4	5	
17	3	2	1	4	5	
18	3	2	nr	3	6	
19	3	3		3	7	
20	2	1		4	7	
21	0	3		4	7	
22	2	4		4	7	
23	3	5		4	6	
24	3	5		4	2	
25	3	5		4	3	
26	3	5		5	5	
27	3	5		6	5	
28	2	4		6	5	
29	3	4		6	3	
30	2	4		6	3	
31	2	3		6	1	
Total	86	80	61	145	163	37

TABLE 38. Estimated Actual Evapotranspiration — WY 1986 (m^3)

Water Year 1986	1985 October	1985 November	1986 May	1986 June	1986 July	1986 August	1986 September
1	500	200	100	300	900	800	300
2	500	200	100	300	1100	800	300
3	500	200	100	300	1400	800	300
4	500	200	100	200	1100	800	300
5	400	200	100	300	900	800	300
6	0	200	100	300	900	800	200
7	200	100	100	200	700	800	200
8	0	100	100	300	1000	800	100
9	0	100	100	400	1000	800	200
10	900	100	100	400	1000	800	200
11	600	100	100	400	1000	800	200
12	500		100	400	800	600	200
13	500		100	400	1000	600	200
14	400		100	400	700	600	200
15	400		100	400	700	300	100
16	400		100	500	800	400	100
17	400		100	500	800	400	100
18	400		100	500	700	100	200
19	400		100	600	800	300	1200
20	300		100	700	700	100	100
21	0		100	700	500	1400	400
22	300		100	800	200	500	400
23	900		100	1000	200	500	0
24	500		100	800	300	600	0
25	500		100	900	1700	400	0
26	500		100	800	500	400	1100
27	400		100	900	600	200	0
28	300		100	600	800	300	300
29	400		100	800	900	300	400
30	300		100	800	900	200	400
31	300		100		800	300	
Monthly Total	12200	1700	3100	15900	25400	17300	8000

TABLE 39. Estimated Actual Evapotranspiration — WY 1987 (m^3)

Water Year	1986		1987					
1987	October	November	April	May	June	July	August	September
1	0	200		200	1100	400	300	100
2	0	200		300	1400	500	300	100
3	100	400		400	1400	400	200	1200
4	800	300		400	1200	400	200	400
5	900	200		500	1200	500	300	400
6	500	200		600	100	500	200	400
7	400	100		200	100	400	200	400
8	400	100		100	1500	400	300	400
9	400	100		0	600	300	300	400
10	400	100		0	1100	300	200	300
11	100	200		400	1500	300	200	300
12	100	100		100	1700	300	200	100
13	300	200		100	1300	200	200	1200
14	300	100		100	900	1400	100	400
15	300	100		100	600	400	200	400
16	300	100	100	100	900	300	200	300
17	100	100	100	700	900	300	200	400
18	0		100	600	1100	200	200	400
19	600		100	800	800	200	200	300
20	600		100	0	800	300	200	300
21	200		200	0	800	200	300	300
22	200		200	100	900	300	300	300
23	200		200	100	900	300	300	100
24	200		200	100	1000	300	100	100
25	300		200	200	900	300	1400	100
26	100		200	700	700	400	500	100
27	300		0	500	900	300	500	200
28	100		0	300	800	300	400	200
29	100		0	300	700	200	400	200
30	0		0	700	500	300	400	200
31	200			600		300	300	
Monthly Total	8500	2800	1700	9300	28300	11200	9300	10000

TABLE 40. Estimated Sublimation — WY 1986 (mm)

Water Year 1986	1985 Nov	1985 Dec	1986 Jan	1986 Feb	1986 Mar	1986 Apr	1986 May	1986 Jun	1986 Jul	1986 Aug
1	1.3	0.8	1.7	2.3	2.0	2.9	2.9	2.6	1.9	1.7
2	1.7	0.8	1.5	1.8	2.2	2.3	3.7	2.1	2.3	1.6
3	1.3	0.9	1.3	1.3	2.0	2.1	4.1	2.0	2.9	1.9
4	1.3	1.6	0.7	1.5	1.9	1.8	1.9	2.1	1.5	2.2
5	1.7	1.0	0.7	2.2	2.2	2.5	2.2	2.0	1.6	2.0
6	1.5	1.8	2.2	1.9	2.5	1.5	1.5	1.9	1.6	2.0
7	1.7	1.8	2.5	1.4	2.3	0.8	1.0	2.1	1.2	2.2
8	2.3	1.9	2.3	1.1	1.7	1.3	1.0	2.6	1.6	1.9
9	2.2	1.5	2.3	1.6	0.7	2.3	1.3	2.9	1.6	2.2
10	1.3	1.3	1.6	2.1	0.4	2.9	2.0	2.4	2.3	2.3
11	0.4	1.6	2.0	1.5	0.8	1.3	1.2	2.2	2.0	2.1
12	0.1	0.7	2.8	0.9	0.6	2.7	1.6	1.8	1.6	1.9
13	-0.2	1.1	2.7	1.3	0.6	1.8	1.4	2.4	1.9	1.6
14	1.7	1.2	1.6	2.3	0.6	2.8	1.3	2.3	1.6	1.9
15	2.2	1.0	1.6	1.2	1.5	3.3	2.0	1.5	1.8	1.3
16	1.3	1.6	2.0	0.5	0.4	1.6	1.6	1.9	2.0	2.0
17	0.7	1.6	2.2	0.6	0.6	1.1	1.6	2.3	1.9	2.0
18	1.4	1.8	2.2	1.2	1.6	1.3	1.7	2.1	1.7	1.3
19	1.5	1.9	2.4	1.4	1.0	1.5	1.2	3.1	2.2	1.9
20	1.0	2.3	2.7	0.8	0.7	1.3	1.5	2.6	2.0	1.2
21	1.7	2.4	2.0	1.4	1.4	1.8	1.4	2.5	1.7	2.9
22	1.9	2.8	2.3	1.0	1.4	1.8	1.4	2.8	0.5	2.1
23	1.2	2.7	2.8	1.2	3.0	2.6	1.9	3.0	0.9	2.1
24	1.5	2.3	2.4	1.4	2.9	2.2	2.0	2.5	1.5	2.4
25	1.4	1.8	2.3	1.4	1.4	3.7	2.0	2.3	2.0	1.8
26	2.7	1.9	2.1	2.0	2.2	2.0	2.2	2.0	1.3	1.9
27	1.4	1.9	2.0	1.2	2.5	1.6	1.9	2.6	1.5	1.7
28	1.0	1.4	2.5	1.2	2.6	2.7	2.4	2.7	1.9	2.7
29	1.5	1.0	2.3		2.9	2.4	2.7	1.9	1.6	2.7
30	1.3	1.6	1.9		2.8	2.6	2.6	1.7	1.7	2.1
31		1.2	2.3		3.0		2.6		1.8	2.3
Monthly Total	42	49	64	40	52	63	60	69	54	62

+ indicates condensation

TABLE 41. Estimated Sublimation — WY 1987 (mm)

1987	January	February	March	April	May	June
1	1.3	3.5	1.1	1.4	0.9	1.1
2	3.3	3.8	1.7	1.9	0.8	0.9
3	2.3	1.7	3	0.7	0.8	1.4
4	0.8	3.1	4.5	0.3	0.9	0.5
5	0.5	4.8	0.7	0.7	1.9	0.4
6	0.8	5.7	0.1	0.6	2.3	+0.9
7	0.3	4.3	0.5	1.1	0.8	+0.5
8	1.7	3.1	1.1	1.8	+0.3	+0.1
9	2.7	0.5	2	0.9	+0.4	+0.8
10	3.5	0.4	2.5	0.3	+0.2	+0.6
11	4.7	0.5	1.4	0.8	+0.3	1.8
12	4	0.9	1.6	2.8	+0.5	1.3
13	3.6	0.5	1.4	1.9	+0.6	+0.3
14	2.5	2	1.1	4.6	0	+0.2
15	1.6	1	0.4	3.3	+0.6	0.6
16	1.4	1.6	1.3	3.8	+0.9	0.9
17	4	1.3	1.9	1.1	+0.7	0.6
18	2.3	2.3	1.6	4.3	0.8	0.3
19	3	2.5	0.7	4.2	1.6	+0.1
20	4.6	2.6	0.8	2.2	0.3	+1.3
21	3.4	1.8	0.4	2.6	+0.1	+0.5
22	4	2	0.2	2.4	0.3	1.4
23	1.6	1.1	0.3	1.4	0.8	0.4
24	0.6	0.3	1.1	1.6	0.3	0.5
25	3	0.7	1.4	0.9	0.4	0.6
26	4	1	1.9	0.3	+0.1	1.3
27	2.3	1.1	2.1	0.6	+0.1	0.8
28	1.9	0.8	2.6	+0.1	0.1	0.5
29	2.3		2.1	0.1	+0.3	0.2
30	1.3		1.4	0.3	0.5	+0.5
31	2.5		2.1		+0.6	
Monthly Total	76	55	45	49	8	9

+ indicates condensation

TABLE 42. Estimated Sublimation — WY 1986 (m^3)

Water Year 1986	1985 Nov	Dec	1986 Jan	Feb	Mar	Apr	May	Jun	Jul	Aug
1	800	820	1840	2460	2150	3180	2940	2280	1140	360
2	1000	820	1590	1990	2420	2460	3750	1760	1320	340
3	800	960	1370	1430	2110	2250	4220	1670	1640	390
4	800	1680	800	1660	2100	1930	2340	1770	860	450
5	1000	1100	780	2370	2370	2750	2610	1690	880	400
6	900	1940	2410	2050	2660	1650	1850	1540	890	400
7	1000	1970	2730	1510	2510	860	1150	1700	650	430
8	1400	2050	2440	1240	1830	1390	1250	2080	800	370
9	1300	1610	2450	1740	710	2510	1520	2290	790	410
10	800	1380	1770	2280	420	3080	2290	1880	1130	440
11	200	1770	2180	1600	870	1380	1370	1710	960	380
12	50	800	3050	960	690	2910	1780	1390	740	330
13	+100	1190	2960	1360	600	1940	1490	1810	820	290
14	1000	1300	1740	2470	620	2980	1370	1730	680	350
15	2340	1100	1780	1290	1610	3520	2100	1100	760	220
16	1420	1710	2180	540	420	1730	1680	1410	840	330
17	760	1770	2330	630	650	1140	1600	1680	740	340
18	1480	1980	2420	1290	1760	1360	1750	1540	650	220
19	1650	2080	2630	1510	1100	1590	1180	2200	830	290
20	1110	2520	2950	920	780	1360	1480	1830	720	190
21	1790	2610	2190	1530	1560	1910	1390	1750	600	460
22	2110	3070	2460	1070	1470	1880	1280	1950	180	330
23	1260	2960	2990	1330	3200	2750	1770	2010	280	300
24	1580	2490	2560	1460	3160	2320	1900	1640	470	340
25	1560	1980	2450	1490	1560	3920	1890	1520	620	260
26	2950	2050	2260	2110	2380	2090	2020	1270	380	270
27	1510	2040	2200	1290	2670	1650	1770	1640	390	230
28	1120	1470	2680	1280	2820	2860	2180	1660	480	350
29	1670	1060	2430		3080	2500	2410	1160	390	350
30	1400	1740	2050		2970	2680	2290	1000	390	280
31		1320	2430		3250		2240		400	270
Monthly Total	36660	53340	69100	42860	56500	67250	60860	50660	22420	10370

+ indicates condensation

TABLE 43. Estimated Sublimation — WY 1987 (m^3)

1987	January	February	March	April	May	June
1	1400	3800	1200	1500	800	300
2	3600	4100	1800	2100	600	300
3	2500	1800	3200	800	600	400
4	900	3300	4900	300	600	100
5	500	5200	800	800	1300	100
6	900	6200	100	600	1600	+200
7	300	4600	500	1200	500	+100
8	1800	3300	1200	1900	+200	+20
9	2900	500	2200	1000	+200	100
10	3800	400	2700	300	+100	100
11	5100	500	1500	900	+200	300
12	4300	1000	1700	3000	+300	200
13	3900	500	1500	2100	+300	+40
14	2700	2200	1200	4900	0	+30
15	1700	1100	400	3500	+300	80
16	1500	1700	1400	4000	+500	100
17	4300	1400	2100	1100	+400	70
18	2500	2500	1700	4300	400	40
19	3200	2700	800	4200	700	+10
20	5000	2800	900	2200	100	+100
21	3700	1900	400	2500	+40	+50
22	4300	2200	200	2200	100	200
23	1700	1200	300	1300	300	40
24	600	300	1200	1400	100	40
25	3200	800	1500	800	200	50
26	4300	1100	2100	300	+40	90
27	2500	1200	2300	600	+40	60
28	2100	900	2800	+100	40	40
29	2500		2300	100	+100	10
30	1400		1500	300	200	30
31	2700		2300		+200	
Monthly Total	81800	59200	48700	50100	5220	2200

+ indicates condensation

TABLE 44. Estimated Basin-Wide Snowmelt by Month (m^3)

Month	Δ S.W.E.	Ablation Stakes	Melt Model	Other Estimates	Average
Water Year 1986					
October				36000	
November				12000	
March				36000	
April			160000		
May	730000		873000		800000
June	1003000	681000	791000		825000
July	303000	386000	453000		380000
August	163000	134000		106000	134000
September				117000	
Water Year 1987					
October				36000	
November				12000	
March				18000	
April	194000	178000			186000
May	359000	247000			303000
June	176000	142000			159000
July				10000	
August				4000	
September				2000	

TABLE 45. Estimated Error in Water Balance, Water Year 1986

Component	Amount ($1000\ m^3$)	Lower fraction	Lower amount	Upper fraction	Upper amount
Snowfall	3101	.05	155.1	.10	310.1
Rainfall	63	.10	6.3	.20	12.6
Evaporation from snow	496	.20	99.2	.30	148.8
Evaporation from lake	8	.10	0.8	.20	1.6
Evapotranspiration	85	.20	17	.50	42.5
Streamflow	2575	.15	386.3	.20	515.0
Total Error			428		621

Estimated Error in Water Balance, Water Year 1987

Component	Amount ($1000\ m^3$)	Lower fraction	Lower amount	Upper fraction	Upper amount
Snowfall	1,107	.05	55.4	.10	110.7
Rainfall	129	.10	12.9	.20	25.8
Evaporation from snow	272	.20	54.4	.30	81.6
Evaporation from lake	18	.10	1.8	.20	3.6
Evapotranspiration	81	.20	16.2	.50	40.5
Streamflow	821	.15	123.2	.20	164.2
Total Error			147		220

TABLE 46. Water Balance Data Summary, Monthly Totals (m^3)
Emerald Lake Watershed, wy86 and wy87

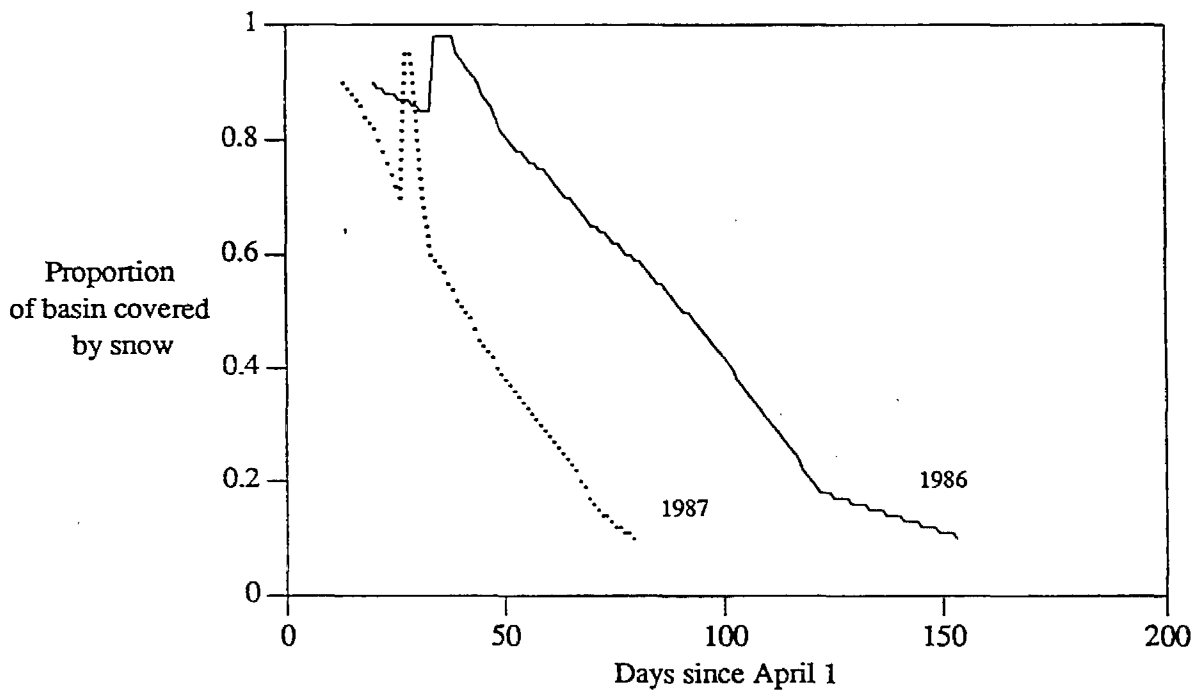
Date	Precipitation		Evaporation			ΔSnow	Outflow	ΔStorage	Net
	Snow	Rain	Snow	ET	Lake	Storage	Discharge	+Residual	SR+ E
									48000
WY86									
Oct	85000	11000	16000	12200	1750	33000	20800	12250	60250
Nov	374000	0	36700	1700	800	325300	12600	-3100	57150
Dec	174000	0	53300	0	0	120700	12100	-12100	45050
Jan	112000	10000	69100	0	0	42900	26300	-16300	28750
Feb	1450000	0	42900	0	0	1407100	7000	-7000	21750
Mar	475000	5000	56500	0	0	382500	36300	4700	26450
Apr	132000	0	67300	1200	0	-95300	135000	23800	50250
May	60000	0	60900	3100	0	-800900	534000	262900	313150
Jun	0	0	50700	15900	0	-875700	825000	-15900	297250
Jul	0	24000	22400	25400	0	-402400	612000	-233400	63850
Aug	0	5000	10400	17300	3090	-144400	204000	-85390	-21540
Sep	160000	8000	10000	8000	2730	33000	36400	77870	56330
Total	3022000	63000	496200	84800	8370	25800	2461500		
WY87									
Oct	33600	0	6000	8500	2110	-8400	44300	-18910	37420
Nov	12000	0	6000	2800	1210	-6000	6800	1190	38610
Dec	24000	0	12000	0	0	12000	2800	-2800	35810
Jan	273000	0	81800	0	0	191200	2300	-2300	33510
Feb	307000	0	59200	0	0	247800	3400	-3400	30110
Mar	310000	0	48700	0	0	243300	12600	5400	35510
Apr	120000	0	50100	1700	0	-116100	176000	8300	43810
May	18000	69600	5220	9300	410	-290220	303000	59890	103700
Jun	0	31200	2200	28300	2570	-161200	229000	-69670	34030
Jul	0	1200	200	11200	3590	-10200	24500	-28090	5940
Aug	0	7800	100	9300	4390	-4100	4800	-6690	-750
Sep	0	19200	100	10000	3400	-2100	1600	6200	5450
Total	1097600	129000	271600	81100	17680	95980	811100		
Total:	4119600	192000	768000	166000	26100	122000	3273000		

Precipitation - Evaporation - Δ Snow Storage - Outflow Discharge = (Δ [Subsurface] Storage + Residual)

Δ Snow Storage = Snow[fall] - Melt - Sublimation

Net S + R [Net Subsurface Storage + Residual] = Cumulative total of Δ [Subsurface] Storage + Residual

Figure 6. Depletion of Snow Covered Area



E. References

World Meteorological Organization, Guide to Hydrometeorological Practices, *WMO Publication 168*, Secretariat of the World Meteorological Organization, Geneva, 1970.

UNESCO, *Scientific Framework of World Water Balance*, Tech. Papers in Hydrology 7, UNESCO Press, Paris, 1971.

Anderson, E. R., Energy budget studies, in *Water-Loss Investigations, Lake Hefner Studies : Technical Report*, U.S Geological Survey Professional Paper 269, pp. 71-119, U.S. Government Printing Office, Washington, D.C., 1954.

Baron, J., S. Norton, D. Beeson, and R. Herrmann, Sediment, diatom, and metal stratigraphy from Rocky Mountain lakes with special reference to acid deposition, *Canadian Journal of Fisheries and Aquatic Science*, *43*, 1350-1362, 1986.

Bergman, J. A., Two standard precipitation gages and the gamma transmission snow gage : a statistical comparison, *Research Note PSW-358*, USDA Forest Service, Pacific Southwest Station, Berkeley, 1982.

Blaney, H. F., Evaporation from free water surfaces at high altitudes, *Transactions of the American Society of Civil Engineers*, *123*, 385-404, 1958.

Brendecke, C. M., D. R. Laiho, D. C. Holden, T. J. Charles, and E. James, Hydrologic water balance investigation of high altitude alpine watersheds of Colorado, (Final Report to Colorado Commission of Higher Education), Dept. of Civil, Environmental, and Architectural Engineering, University of Colorado, Boulder, 1984.

Brunt, D., Notes on radiation in the atmosphere, *Journal of the Royal Meteorological Society*, *58*, 389-420, 1932.

Brutsaert, W., *Evaporation into the Atmosphere*, D. Reidel, Dordrecht, 1982.

Brutsaert, W., The formulation of evaporation from land surfaces, in *Advances in Modelling of Hydrologic Systems*, edited by D. Bowles and P. E. O'Connell, NATO Ser., D. Reidel, Dordrecht, Holland, 1989 - in press.

Caine, N., The geomorphic processes of the alpine environment, in *Arctic and Alpine Environments*, edited by J. D. Ives and R. G. Barry, pp. 133-158, Meuthen, London, 1974.

Calder, I. R., R. J. Harding, and P. T. W. Rosier, An objective assessment of soil moisture deficit models, *J. Hydrol.*, *60*, 329-355, 1983.

Carroll, T., The water budget of an alpine catchment in central Colorado, *M.S. thesis*, University of Colorado, Boulder, 1974.

Carroll, T., An estimate of water efficiency for a Colorado alpine basin, *Proc. Western Snow Conf.*, *44*, 69-77, 1976.

Clark, M. J., Periglacial hydrology, in *Advances in Periglacial Geomorphology*, edited by M. J. Clark, pp. 315-473, Wiley, Chichester, 1988.

Clow, D. C., N. Swoboda-Colberg, J. I. Drever, and F. S. Sanders, Chemistry of snowmelt, soil water and stream water at the West Glacier Lake watershed, Wyoming, *Eos, Trans. Amer. Geophys. Union*, *69*, 1202, 1988.

Croft, A. R., Water loss by stream surface evaporation and transpiration by riparian vegetation., *Transactions, American Geophysical Union*, *29*, 235-239, 1948.

Davis, S. N. and R. J. M. Dewiest, *Hydrogeology*, Wiley, New York, 1966.

Dooge, J. C. I., The waters of the Earth, *Hydrological Sciences Journal*, *29*, 149-176, 1984.

Dozier, J., Recent research in snow hydrology, *Rev. Geophys.*, *25*, 153-161, 1987.

Dozier, J., J. M. Melack, D. Marks, K. Elder, R. Kattelmann, and M. Williams, Snow deposition, melt, runoff, and chemistry in a small alpine watershed, Emerald Lake Basin, Sequoia National Park, Final Rep., Contr. A3-106-32, Calif. Air Resour. Board, Sacramento, CA, 1987.

Dracup, J. A., J. M. Melack, D. Marks, R. Kattelmann, and J. O. Sickman, The hydrologic mass balance of the Emerald Lake Basin Integrated Watershed Study, Sequoia National Park, California, *Final Rep., Contr. A4-123-32*, 78 pp., Calif. Air Resour. Board, Sacramento, 1988.

Dunne, T. and L. B. Leopold, *Water in Environmental Planning*, W. H. Freeman and Co., New York, 1978.

Fortier, S., Evaporation losses in irrigation, *Engineering News*, *58*, 304-307, 1907.

Glen, J. W., editor, *Hydrological Aspects of Alpine and High-Mountain Areas*, IAHS Publ. No. 138, Intl. Assoc. Hydrol. Sci., Wallingford, UK, 1982.

Hannaford, J. F. and M. C. Williams, Summer hydrology of the high Sierra, *Proc. Western Snow Conf.*, *35*, 73-84, 1967.

Harbeck, G. E. and J. S. Meyers, Present day evaporation measurement techniques, *Journal of the Hydraulics Division, American Society of Civil Engineers*, *96*, 1381-1390, 1970.

Harte, J., G. Lockett, R. Schneider, H. Michaels, and C. Blanchard, Acid precipitation and surface water vulnerability of the western slope of the high Colorado Rockies, *Water, Air Soil Poll.*, *25*, 313-320, 1985.

Henning, I. and D. Henning, Potential evapotranspiration in mountain geoecosystems of different altitudes and latitudes, *Mtn. Res. Devel.*, *1*, 267-274, 1981.

Horton, Robert E., Water-losses in high latitudes and at high elevations, *Transactions of the American Geophysical Union*, *15*, 351-379, 1934.

Huntington, G. L. and M. Akeson, Pedologic investigations in support of acid rain studies, Sequoia National Park, CA, (Tech. Rept.), Department of Land, Air and Water Resources, Univ. Calif., Davis, 1986.

Johnson, R. S. and R. W. Brown, Hydrologic aspects related to the management of alpine areas, in *Special Management Needs of Alpine Ecosystems*, edited by D. A. Johnson, pp. 65-75, Society for Range Management, Denver, 1979.

Jordan, P., The snowmelt hydrology of a small alpine watershed, *M.S. thesis*, University of British Columbia, Vancouver, 1978.

Kattelmann, R., Groundwater contributions in an alpine basin in the Sierra Nevada, *Proceedings of the Headwaters Hydrology Symposium*, 361-369, Amer. Water Resour. Assoc., Bethesda, MD, 1989.

Kattelmann, R. and N. Berg, Water yields from high elevation basins in California, in *Proceedings, California Watershed Management Conference*, edited by R. Z. Callaham and J. J. DeVries, Report No. 11, pp. 79-85, Univ. Calif. Wildland Resources Center, Berkeley, 1987.

Kattelmann, R. C., Seasonal hydrologic processes in an alpine catchment, in *Proceedings American Geophysical Union Hydrology Days*, vol. 9, pp. 107-118, Colorado State University, Fort Collins, 1989.

Kohler, M. A., T. J. Nordenson, and D. R. Baker, Evaporation maps for the United States, *U.S. Weather Bureau Tech. Pap. 37*, Washington, D.C., 1959.

LaBaugh, James W., Uncertainty in phosphorus retention, Williams Fork Reservoir, Colorado, *Water Resour. Res.*, *21*, 1684-1692, 1985.

Lang, H., Is evaporation an important component in high alpine hydrology?, *Nordic Hydrol.*, *12*, 217-224, 1981.

Larson, L. W. and E. L. Peck, Accuracy of precipitation measurements for hydrologic modeling, *Water Resour. Res.*, *10*, 857-863, 1974.

Leaf, C. F., Watershed management in the central and Southern Rocky Mountains : a summary of the status of our knowledge by vegetation types, *Research Pap. RM-142*, USDA Forest Service, Rocky Mountain Forest and Range Experiment Station, Fort Collins, Colo., 1975.

Longacre, L. L. and H. F. Blaney, Evaporation at high elevations in California, *Journal of the Irrigation and Drainage Division, American Society of Civil Engineers*, *88*, 33-54, 1962.

Lund, L. J., A. D. Brown, M. A. Lueking, S. C. Nodvin, A. L. Page, and G. Sposito, Soil processes at Emerald Lake Watershed, *Final Rep., Contr. A3-105-32*, Calif. Air Resour. Board, Sacramento, 1987.

Marks, D., Climate and energy exchange at the snow surface in Emerald Lake Basin, Sierra Nevada, Ph.D. Thesis, 159 pp., Dept. Geogr., Univ. Calif., Santa Barbara, 1988.

Marks, D., R. Kattelmann, J. Dozier, and R. E. Davis, Monitoring snowcover properties and processes in a small alpine watershed, in *Proceedings of Sixth International Northern Research Basins Symposium/Workshop*, edited by H. S. Santeford, pp. 259-275, Mich. Technol. Univ., Houghton, 1986.

Martinelli, M., Watershed yield improvement from alpine areas : the status of our knowledge, *Research Pap. RM-138*, USDA Forest Service, Rocky Mountain Forest and Range Experiment Station, 1975.

Miller, D. H., Interception processes during snowstorms, *Research Pap. PSW-18*, 24 pp., USDA Forest Service, Pacific Southwest Forest and Range Experiment Station, Berkeley, 1964.

Miller, D. H., *Water at the Surface of the Earth*, Intl. Geophysics Ser., vol. 27, Academic Press, New York, 1977.

Mooney, H. A., R. D. Hillier, and W. D. Billings, Transpiration rates of alpine plants in the Sierra Nevada of California, *American Midland Naturalist*, *74*, 374-386, 1965.

Moore, J. G. and C. Wahrhaftig, Geology of Emerald Lake basin in relation to acid precipitation, in *Geomorphology and Glacial Geology, Wolverton and Crescent Meadow Areas and Vicinity, Sequoia National Park, California*, Open File Report 84-400, pp. 36-39, U.S. Geol. Survey, Reston, Va., 1984.

Morgan, D. L. and W. F. Slusser, A pilot study of upslope winds in the Sierra Nevada, in *Proceedings : Conference on Sierra Nevada Meteorology*, pp. 55-60, American Meteorological Society, Boston, 1978.

Morton, F. I., Operational estimates of lake evaporation, *J. Hydrol.*, *66*, 77-100, 1983.

Motoyama, H., D. Kobayashi, and K. Kojima, Water balance and runoff analysis at a small watershed during the snow-melting season, in *Proceedings of the Cold Regions Hydrology Symposium*, pp. 297-304, Amer. Water Resour. Assoc., Bestheda, MD, 1986.

Najjar, G., B. Ambroise, and J. L. Mercier, Measurement and mapping of potential evapotranspiration in a small mountainous watershed, *Nordic Hydrol.*, *12*, 195-205, 1981.

Nodvin, S. C., Processes affecting the chemistry of water flowing through a high elevation Sierra Nevada watershed, in *Proceedings of the UNESCO/IHP International Symposium on Acidification and Water Pathways*, vol. 1, pp. 101-106, Norwegian National Committee for Hydrology, Oslo, 1987.

Peck, E. L. and D. J. Pfankuch, Evaporation rates in mountainous terrain, *International Association of Hydrological Sciences Publication*, *63*, 267-278, 1963.

Penman, H. L., Natural evaporation from open water, bare soil and grass, *Proc. Royal Soc. London*, *A193*, 120-145, 1948.

Price, L. W., *Mountains and Man*, Univ. Calif. Press, Berkeley, 1981.

Rainbird, A. F., Methods of estimating areal average precipitation, *IHD Rep. 3*, World Meterological Organization, Geneva, 1976.

Rundel, P. W., Laboratory of Biomedical and Environmental Sciences, University of California at Los Angeles, personal communication.

Rundel, P. W., T. W. St. John, and W. Westman, Vegetation Process Studies, Final Rep., Contr. A3-097-32, Calif. Air Resour. Board, Sacramento, CA, 1987.

Satterlund, D. R. and H. F. Haupt, Disposition of snow caught by conifer crowns, *Water Resour. Res.*, *6*, 649-652, 1970.

Sevruk, B., Correction of precipitation measurements: summary report, in *Correction of Precipitation Measurements*, edited by B. Sevruk, Züricher Geographische Schriften No. 23, pp. 13-26, Geographisches Institut, Eidgenössische Technische Hochschule, Zürich, 1986.

Slaughter, C. W., Evaporation from snow and evaporation retardation by monomolecular films -- a review of literature, *Special Rep. 130,* U.S. Army Cold Regions Res. Engrg. Lab., Hanover, N.H., 1970.

Slaymaker, H. O., Alpine hydrology, in *Arctic and Alpine Environments*, edited by J. D. Ives and R. G. Barry, pp. 133-158, Meuthen, London, 1974.

Smith, J. L., The historic climatic regime and the projected impact of weather modification upon precipitation and temperature at the Central Sierra Snow Laboratory, in *The Sierra Ecology Project*, edited by J. L. Smith and N. H. Berg, vol. 3, pp. 1.vii-1.43, USDI-Bureau of Reclamation, Engineering and Research Center, Denver, 1982.

Solokov, A. A. and T. G. Chapman, *Methods for Water Balance Computations*, UNESCO Press, Paris, 1974.

Staudinger, M. and H. Rott, Evapotranspiration at two mountain sites during the vegetation period, *Nordic Hydrol., 12*, 207-216, 1981.

Stednick, J. D., Hydrochemical balance of an alpine watershed in Southeast Alaska, *Arctic Alpine Res., 13*, 431-438, 1981.

Stewart, B. J., Sensitivity and significance of turbulent energy exchange over an alpine snow surface, M.A. Thesis, Dept. Geogr., Univ. Calif., Santa Barbara, 1982.

Stoddard, J. L., Alkalinity dynamics in an unacidified alpine lake, Sierra Nevada, California, *Limnol. Oceanogr., 32*, 825-839, 1987.

Terjung, W. H., R. N. Kickert, G. L. Potter, and S. W. Swarts, Energy and moisture balances of an alpine tundra in mid July, *Arctic Alpine Res., 1*, 247-266, 1969.

Thom, A. S., J. L. Thony, and M. Vauclin, On the proper employment of evaporation pans and in estimating potential transpiration, *Quart. J. Royal Meteorol. Soc., 103*, 345-375, 1981.

U.S. Army Corps of Engineers, *Snow Hydrology*, 456 pp., North Pacific Division, Portland, 1956.

Van der Beken, A. and A. Herrmann, editors, *New Approaches in Water Balance Computations*, IAHS Publ. 148, Intl. Assoc. Hydrol. Sci., Wallingford, UK, 1985.

Veihmeyer, F. J., Evapotranspiration, in *Handbook of Applied Hydrology*, edited by V. T. Chow, pp. 11.1-11.38, McGraw-Hill, New York, 1964.

Wiesnet, D. R., *The role of satellites in snow and ice measurements*, NOAA Tech. Mem. NESS58, NOAA, U.S. Dept. of Commerce, Washington, D.C., 1974.

Winter, T. C., Uncertainties in estimating the water balance of lakes, *Water Resources Bulletin, 17*, 82-115, 1981.

Woo, Ming-ko, R. Heron, P. Marsh, and P. Steer, Comparison of weather station snowfall with winter snow accumulation in high Arctic basins, *Atmosphere-Ocean, 21*, 312-325, 1983.

Young, G. J. ed., *Techniques for prediction of runoff from glacierized areas*, IAHS-AIHS Publ. 149, 149 pp., Intl. Assoc. Hydrol. Sci., Wallingford, UK, 1985.

Zinke, P. J., Forest interception studies in the United States, in *Forest Hydrology*, edited by W. E. Sopper and H. W. Lull, pp. 137-160, Pergamon Press, Oxford, 1967.