SNOWFALL AND SNOWMELT ESTIMATES FOR MODERN AND LAST GLACIAL CONDITIONS WITHIN THE OWENS VALLEY, CALIFORNIA

by

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ABSTRACT

We estimate monthly snowfall, snowmelt, and resulting snowpack for a two-dimensional solution domain which includes California's Owens Valley. A stastically-based local climate model computes monthly temperature and precipitation for a 160 by 290 km domain at a 2 km grid cell spacing for last glacial maximum (LGM) and present boundary conditions. The fraction of monthly precipitation that falls as snow is computed as the double integral of a bivariate normal distribution in temperature and log precipitation. We use the methodology of Martinec et al (1983) to compute monthly snowmelt. The snow model computes monthly snowfall, snowmelt, and resulting snowpack at each grid cell, and it estimates the densification of perennial snowpack and its transformation into glacial ice when appropriate. The model proceeds year-by-year until monthly snow and ice totals equilibrate.

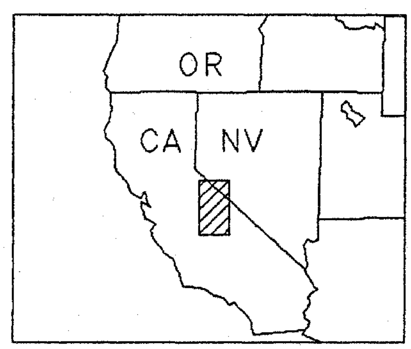
INTRODUCTION

This work is part of an effort to estimate modern and last glacial maximum (LGM) pluvial lake extent within the Great Basin as a function of climate. The overall modeling effort can be subdivided into three distinct hydrologic models (the snow model, the runoff model, and the lake model), all constrained to use only that information available both for modern and LGM conditions. The snow model computes snowfall, snowmelt, and resulting snowpack from climate information. The runoff model computes mean monthly runoff from snowmelt and rainfall matrices, and the lake model computes pluvial lake extent as a function of runoff. Modern and LGM snow model results are presented here.

METHODOLOGY

We apply a statistically-based local climate model (Stamm, 1991) to compute mean maximum monthly temperature (T_{max}) and mean monthly precipitation (both calculated from boundary conditions including terrain elevation, insolation, atmospheric CO_2, monthly windfields, and sea-surface temperatures) for a 160 by 290 km domain, which includes California's Owens Valley, at a 2 km grid cell spacing (Figure 1). Mean monthly temperature, $\dot{T}$ (computed from T_{max} by subtracting a constant), is accepted as the mean of a normal distribution, and the log of the predicted precipitation value is accepted as the mean of a lognormal distribution. The variances of temperature and log precipitation are calculated from calibration station data for each month (for a total of 24 values). Monthly temperature and precipitation correlation coefficients, again

Figure 1. Index map showing location of two-dimensional solution domain.



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calculated from calibration station data, are used to construct a bivariate normal distribution. Snow accumulation is assumed to be the fraction of precipitation that occurs below a critical temperature, T_{crit}. The bivariate normal distribution is integrated from $-\infty$ to $+\infty$ in log precipitation and from $-\infty$ to T_{crit} in temperature to calculate the fraction of precipitation that falls as snow.

The snowfall equations are calibrated with temperature, precipitation, and snowfall records from 52 stations in Nevada and California, representing the central and western Great Basin (U.S. West Optical Publishing, 1988). The difference (δT) between T_{max} and T is varied during the calibration process, and the coefficient of determination (R^2) between modeled and observed snowfall is computed. The optimal δT is that value corresponding to the highest R^2. Varying δT from 8°C to 17°C in steps of 1°C produces an optimal R^2 of 0.800 corresponding to a δT of 12°C. To test whether δT is altered by the occurrence of snow, an additional experiment is performed. In this experiment δT is varied and snowfall is computed and compared to observed snowfall for only those data points in which observed snowfall is equal to or greater than the snowfall threshhold value (Table 1).

Table 1. Results of snowfall code optimization for months when snowfall exceeds a given snowfall threshhold.

SNOWFALL THRESHHOLD	OPTIMAL R^2	δT
0 mm	0.800	12°C
5	0.743	12
10	0.719	12
15	0.708	12
20	0.700	12

The optimal coefficient of determination consistently corresponds to a δT of 12°C for small snowfall threshhold values, so the optimal δT is accepted as 12°C for this study. The optimal δT exceeds the mean observed δT (approximately equal to 10.5°C) from the snowfall stations. This could reflect a difference in distributional parameters of temperature and precipitation on days with precipitation versus days without precipitation (which make up the majority of the days of the month in this study area). The local climate model distributional parameters are computed with data for all days of the month, thus biasing them toward non-precipitation conditions.

To calculate snowmelt we use the methodology of Martinec et al (1983) (wherein air temperature is the index of energy exchange across the upper snow surface) because it seems most appropriate for this simulation based on the information available from the local climate model. The scale of the basins in the Owens River system is well within the scale of the basins to which this methodology has been successfully applied. Snowmelt depth, M, is proportional to degree-days, T_d,

$$M = a \cdot T_d \cdot S \qquad (1)$$

where a is a degree-day factor, and S is the fraction of the catchment area covered by snow. The degree-day factor is in turn related to average snow specific gravity, τ,

$$a = 1.1 \cdot \tau \qquad (2)$$

During the addition of new snow S is very close to 1.0, but approaches zero with increasing ablation. In this study, each grid cell behaves as an individual catchment; no elevation zoning or temperature extrapolation is required since each cell is assumed to be at a single elevation with a specified mean monthly air temperature. No snow cover information is available, so S is assumed to be unity. It is expected that the fine resolution in temperature and elevation will at least partially offset the lack of snowcover information. The number of degree days is the mean monthly temperature (°C) multiplied by the number of days per month, N, and monthly snowmelt depth becomes,

$$M = a \cdot T_d = a \cdot N \cdot T \qquad (3)$$

In modeling snowpack, the possibility exists that annual snow accumulation will exceed annual snow ablation, resulting in a perennial snowpack. Over time, a perennial snowpack may deepen enough that it produces glacial ice. This scenario must be accounted for in the snow model, as areas within the Owens River solution domain, particularly the Sierra Nevada, were glaciated during the last glacial maximum. Even today, 497 small glaciers reside within a 25,000 km^2 area of the Sierra Nevada (Brown, 1989).

Snow density increases with depth, as snow is compressed by the column of snow above it. In this study variations in snowpack density will be assumed to be linear, with a mean firn-ice transition depth of about 30 meters (Paterson, 1983). Assuming a new snow density of 0.1 g/cm^3, an ice density of 0.9 g/cm^3, and a linear change with depth, then the average density of the snowpack above the firn-ice transition is 0.5 g/cm^3, and the water equivalent of the 30 m snow column is 1500 cm. If the snow model, then, predicts a perennial snowpack for a particular grid cell, the water equivalent of the snow column is compared to the firn-ice transition water equivalent. If the snow column water equivalent exceeds the transition water equivalent, then the excess is assumed to be ice at a constant density of 0.9 g/cm^3.

Ice ablation is computed using a degree-day methodology similar to that used in computing snow ablation. Johannesson et al (1993) demonstrate that the degree-day factor for ice is very stable. Published values (in cm/°C/day) range from about 0.55 to 0.78, and mean values are 0.63 for studies in the Canadian Arctic (Braithwaite, 1981), 0.73 and 0.71 in Greenland (Braithwaite and Olesen, 1984; Braithwaite, 1984), and 0.77 and 0.64 in Iceland and Norway (Johannesson et al, 1993). Monthly icemelt (I) is computed from degree days and an icemelt degree-day factor (α).

$$I = \alpha \cdot T_d = \alpha \cdot N \cdot T \qquad (4)$$

Ice melts only when it is not covered by snow (Gottlieb, 1980), thus ice forming under a perennial snowpack must be routed elsewhere to ablate. If one assumes that glacial ice is a perfectly plastic medium, a simple relationship exists between limiting ice thickness and terrain slope (Paterson, 1983). A perfectly plastic material exhibits zero strain rate below a yield stress, then an infinite strain rate when the applied stress exceeds the yield stress. Maximum ice thickness (h) is related to yield stress (σ), density (r), acceleration due to gravity (g), and terrain slope (Θ) as follows.

$$h = \sigma/(r \cdot g \cdot \sin\Theta) \qquad (5)$$

Yield stress has been measured for many glaciers; a mean value of 100 kPa will be used here (Paterson, 1983). Maximum ice thickness may then be determined based on terrain slope calculated from the digital elevation matrix.

If a column of ice exceeds h, the excess is transferred to the lowest adjacent grid cell under the assumption that the ice will flow down the steepest gradient path. Snow and ice computations are made until the entire ice mass is at equilibrium, that is, supply via snow densification and ice transfer from higher elevation grid cells is equal to the sum of ablation and transfer to lower elevation grid cells at each grid cell within the solution domain.

RESULTS

The snow model begins with no initial snowpack and runs year-by-year until equilibration of monthly snow and ice volumes (measured in water equivalents). A modern snow model run equilibrates in 39 years with a perennially snow-covered area of 136 km^2, while an LGM run equilibrates in 165 years with a perenially snow-covered area of 3376 km^2 and a glaciated area of 44 km^2. Snowpack for the entire solution domain varies from a maximum of 38 cm snow water equivalent (s.w.e.) in March to a minimum of 0.4 cm s.w.e. in October for modern climate. LGM climate produces a maximum snowpack of 83 cm s.w.e. in April and a minimum of 8 cm s.w.e. in September and October. Maximum modern snowmelt occurs in May and June with an average value of 11 cm, while LGM snowmelt peaks at 23 cm in June. The fraction of the solution domain that is covered by snow varies from 0.76 in January to 0.003 in October for modern conditions, and from 0.90 in January to 0.077 in September for LGM conditions. Snow model solutions indicate that LGM snowpack exceeds modern snowpack both in extent and volume, with the maximum difference of 50 cm s.w.e. occurring in April. LGM snowmelt also greatly exceeds modern, with a maximum departure of 12.5 cm in June. Figure 2 shows modeled modern and LGM snowcover for May superimposed on terrain. Light colored regions represent grid cells covered by snow, while the darker shades of gray correspond to terrain elevation. 32% of the solution domain is predicted to sustain a mean monthly snowpack for modern climate in May, compared to 53% of the solution domain for LGM climate. The Sierra Nevada (trending northwest to southeast) clearly represents the largest snow-covered area for both modern and LGM conditions.

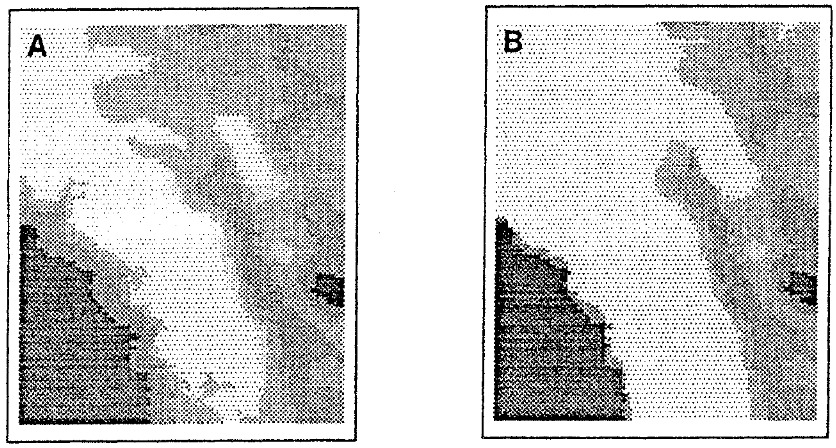

Figure 2. Modern (A) and LGM (B) May snowcover superimposed on terrain. The lightest area represents snowcover, while the darker shades of gray correspond to terrain elevation.

CONCLUSIONS

This model has been developed for the study of long-term changes in the climatic/hydrologic system. Representation of snowfall and snowmelt processes is critical to the evaluation of Owens Valley hydrology because annual runoff is sensitive to, and dominated by, snowmelt from the Sierra Nevada. This was even more important at the time of the last glaciation. We infer that the snow model will be a useful tool in the reconstruction of LGM pluvial lakes within the Great Basin.

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