January 1994

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*Submitted for publication in IAHS Proceedings of Symposium on Biogeochemistry of Seasonally Snow Covered Catchments, June 3-14, 1995, Boulder, CO.*

Working Paper WP-94-HWR-DGT/003

November 1994

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# A Spatially Distributed Energy Balance Snowmelt Model

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**Abstract** This paper describes an energy balance snowmelt model developed for the prediction of rapid snowmelt rates responsible for soil erosion and water input to a distributed water balance model. The model uses a lumped representation of the snowpack with two primary state variables, namely, water equivalence and energy content relative to a reference state of water in the ice phase at $0^{o}$C. This energy content is used to determine snowpack average temperature or liquid fraction. This representation of the snowpack is used in a distributed version of the model with each of these state variables modeled at each point on a rectangular grid corresponding to a digital elevation model. Inputs are air temperature, precipitation, wind speed, humidity and radiation at hourly time steps. The model uses physically-based calculations of radiative, sensible, latent and advective heat exchanges. An equilibrium parameterization of snow surface temperature accounts for differences between snow surface temperature and average snowpack temperature without having to introduce additional state variables. Melt outflow is a function of the liquid fraction, using Darcy's law. This allows the model to account for continued outflow even when the energy balance is negative. A detailed description of the model is given together with results of tests against data collected at the Central Sierra Snow Laboratory, California; Reynolds Creek Experimental Watershed, Boise Idaho; and at the Utah State University drainage and evapotranspiration research farm, Logan Utah. The testing includes comparisons against melt outflow collected in melt lysimeters, surface snow temperatures collected using infrared temperature sensors and depth and water equivalence measured using snow core samplers.

## INTRODUCTION

Snowmelt is a significant surface water input of importance to many aspects of hydrology including water supply, erosion and flood control. Snowmelt is driven primarily by energy exchanges at the snow-air interface. The model described here was developed initially to predict the rapid melt rates responsible for erosion. It has also been used to provide the spatially distributed surface water input in a water balance study. In developing a new snowmelt model our goal was to incorporate ideas from the many existing models and parameterize the processes involved in as simple, yet physically correct a manner as possible. We hoped to develop a parsimonious, physically-based model that could be driven by readily available inputs and applied anywhere with no (or minimal) calibration. The striving for simplicity led us to parameterize a snowpack in terms of lumped (depth averaged) state variables so as to avoid having to model the complex processes that occur within a snowpack. We have still, however, attempted to capture important physical differences between bulk (depth averaged) properties and the surface properties that are important for surface energy exchanges. We have relied heavily on an understanding of snowmelt processes gleaned from Gray and Male (1981) and the descriptions of existing models (Anderson, 1973; 1976; Morris, 1982; Leavesley *et al.*, 1983; Kondo and Yamazaki, 1990). We first give a detailed description of the model. We then describe the data sets we used to test the model and show results comparing model calculations to observations.

## MODEL DESCRIPTION

The snowpack is characterized by state variables, water equivalence W [m], energy content U, [kJ/m$^2$] and the age of the snow surface which is only used for albedo calculations. These are, we believe, sufficient to characterize the snowpack for the surface water inputs of interest. The state variable, energy content U, is defined relative to a reference state of water at 0°C in the ice (solid) phase. U greater than zero means the snowpack (if any) is isothermal with some liquid content and U less then zero can be used to calculate the snowpack average temperature T [°C]. Energy content is defined as the energy content of the snowpack plus a top layer of soil with depth $D_e$ [m]. We discuss below the choice of $D_e$ and the role it plays in the model.

The model is designed to be driven by inputs of air temperature $T_a$ [°C], wind speed V [m/s], relative humidity RH, precipitation P [m/hr], incoming solar $Q_{si}$ and longwave $Q_{li}$ radiation [kJ/m$^2$/hr], and ground heat flux $Q_g$ [kJ/m$^2$/hr] (taken as 0 when not known) at each time step. Time steps of 0.5, 1 and 6 hours have been used in data comparisons here. When incoming solar radiation is not available it is estimated as an extra terrestrial radiation (from sun angle and solar constant) times an atmospheric transmission factor Tr, estimated from the daily temperature range using the procedure given by Bristow and Campbell (1984). When incoming longwave radiation is not available it is estimated based on air temperature, the Stefan-Boltzman equation and a parameterization of air emissivity due to Satterlund (1979), adjusted for cloudiness using Tr.

Given the state variables U and W, their evolution in time is determined by solving energy and mass balance equations.

$$\frac{dU}{dt} = Q_{sn} + Q_{li} + Q_p + Q_g - Q_{le} + Q_h + Q_e - Q_m \tag{1}$$

$$\frac{dW}{dt} = P_r + P_s - M_r - E \tag{2}$$

In the energy balance equation terms are (all in $kJ/m^2/hr$): $Q_{sn}$, net shortwave radiation; $Q_{li}$, incoming longwave radiation; $Q_p$, advected heat from precipitation; $Q_g$, ground heat flux; $Q_{le}$, outgoing longwave radiation; $Q_h$, sensible heat flux; $Q_e$, latent heat flux due to sublimation/condensation; and $Q_m$, advected heat removed by meltwater. In the mass balance equation (all in m/hr of water equivalence) terms are: $P_r$, rainfall rate; $P_s$, snowfall rate; $M_r$, meltwater outflow from the snowpack; and E, sublimation from the snowpack. Many of these fluxes depend functionally on the state and input driving variables. We elaborate on the parameterization of these functional dependencies below. Equations (1) and (2) form a coupled set of first order, nonlinear ordinary differential equations. They can be summarized in vector notation as:

$$\frac{d\underline{X}}{dt} = \underline{F}(\underline{X}, \text{driving variables}) \tag{3}$$

where $\underline{X} = (U, W)$ is a state vector describing the snowpack. With $\underline{X}$ specified initially, this is an initial value problem. A large variety of numerical techniques are available for solution of initial value problems of this form. Here we have adopted a Euler predictor-corrector approach (Gerald, 1978).

$$\underline{X}' = \underline{X}_i + \Delta t\, \underline{F}(\underline{X}_i, \text{driving variables}) \tag{4}$$

$$\underline{X}_{i+1} = \underline{X}_i + \Delta t \frac{\underline{F}(\underline{X}_i, \text{driving variables}) + \underline{F}(\underline{X}', \text{driving variables})}{2} \tag{5}$$

where $\Delta t$ is the time step, $\underline{X}_i$ refers to the state at time $t_i$ and $\underline{X}_{i+1}$ refers to the state at time $t_{i+1}=t_i+\Delta t$. This is a second order finite difference approximation, with global error proportional to $\Delta t^2$ (Gerald, 1978, p257). Numerical instabilities sometimes occur under melting conditions when the snowpack is shallow due to the nonlinear nature of the melt outflow parameterization. To deal with this we compare $\underline{X}_{i+1}$ to $\underline{X}'$ and if they differ by more than a specified tolerance (0.025 m for W and 2000 kJ/m$^2$ for U) iterate up to four times setting $\underline{X}'$ to $\underline{X}_{i+1}$ then recalculating $\underline{X}_{i+1}$ at each iteration. If convergence is still not achieved we take the solution that would keep the liquid fraction of the snow constant. Following we describe how each of the processes involved in equations (1) and (2) are parameterized.

**Depth averaged temperature - T**

The snow and interacting soil layer average temperatures are obtained from the energy content and water equivalence, relative to 0˚C ice phase.

| | | | |
|---|---|---|---|
| If $U < 0$ | $T = U/(\rho_w W C_s + \rho_g D_e C_g)$ | All solid phase | (6) |
| If $0 < U < \rho_w W h_f$ | $T = 0$˚C. | Solid and liquid mixture | (7) |
| If $U > \rho_w W h_f$ | $T = \dfrac{U - \rho_w W h_f}{\rho_g D_e C_g + \rho_w W C_w}$ | All liquid | (8) |

In the above the heat required to melt all the snow water equivalence is $\rho_w W h_f$ [kJ] where $h_f$ is the heat of fusion [333.5 kJ kg$^{-1}$] and U in relation to this determines the solid-liquid phase mixtures. The heat capacity of the snow is $\rho_w W C_s$ [kJ/˚C] where $\rho_w$ is the density of water [1000 kg m$^{-3}$] and $C_s$ the specific heat of ice [2.09 kJ kg$^{-1}$ ˚C$^{-1}$]. The heat capacity of the soil layer is $\rho_g D_e C_g$ [kJ/˚C] where $\rho_g$ is the soil density and $C_g$ the specific heat of soil. These together determine the T when U < 0. The heat capacity of liquid water, $\rho_w W C_w$, where $C_w$ is the specific heat of water [4.18 kJ kg$^{-1}$ ˚C$^{-1}$], is included in (8) for numerical consistency during time steps when the snowpack completely melts.

The parameter $D_e$ is intended to quantify the depth of soil that interacts thermally with the snowpack. Heat flow in snow and soil is governed by Laplace's equation. The depth of penetration of changes in surface temperature can be evaluated from the expression (Rosenberg, 1974):

$$\frac{R_z}{R_s} = \exp\left(- z \left(\frac{\pi}{\alpha P}\right)^{\frac{1}{2}}\right) \tag{9}$$

where $R_s$ is the range of temperature oscillation at the surface, $R_z$ the range of temperature oscillation at depth z, P the period of oscillation, and $\alpha$ the thermal conductivity. For soil $\alpha$ is typically in the range 0.004 to 0.006 $cm^2/s$. Fig. 1 shows $R_z/R_s$ versus z for $\alpha$ = 0.005 $cm^2/s$ for various periods. This figure shows that for oscillations less than one week the effect at 0.4 m is damped to less than 30% and even for monthly oscillations is still damped 50% at 0.4 m depth. This result suggests using $D_e$ = 0.4 m in our model since the time scale of interest is the seasonal accumulation then melting of snow. The state variable U represents energy content above this level. The ground heat flux represents heat transport at this depth and is therefore a long-term average. Oscillating, high frequency, ground heat fluxes above this depth are absorbed into U, the energy stored in the snow and soil above depth $D_e$. This procedure provides a simple approximation of the effects of frozen ground, or snow falling on warm ground.

**Radiation**

Net shortwave radiation is calculated as

$$Q_{sn} = Q_{si}\,(1-A) \tag{10}$$

where albedo A, is calculated based on the age of the snow surface using a parameterization due to Dickinson et al. (1993). The age of the snow surface is retained as a state variable, and is updated with each time step, dependent on snow surface temperature and snowfall. When the snowpack is shallow (depth z < h = 0.1 m) the albedo is taken as r $A_{bg}$ + (1-r) A where
r = $(1-z/h)e^{-z/2h}$. This interpolates between the snow albedo and bare ground albedo with the exponential term approximating the exponential extinction of radiation penetration of snow.

Outgoing longwave radiation is

$$Q_{le} = \varepsilon_s \, \sigma \, T_s^4 \tag{11}$$

where $\varepsilon_s$ is emissivity, $\sigma$ the Stefan Boltzmann constant [2.07 x $10^{-7}$ kJ $m^{-2}$ $hr^{-1}$ $K^{-4}$] and $T_s$ is absolute temperature [K].

**Snow fall accumulation and heat with precipitation**

Measured precipitation rate P, is partitioned into rain $P_r$, and snow $P_s$, (both in terms of water equivalence depth) using the following rule based on air temperature $T_a$, (U.S. Army Corps of Engineers, 1956)

$$\begin{array}{ll} P_r = P & T_a \geq T_r = 3\ ^oC \\ P_r = P(T_a - T_b)/(T_r - T_b) & T_b < T_a < T_r \\ P_r = 0 & T_a \leq T_b = -1\ ^oC \\ P_s = (P - P_r)\,F & \end{array} \tag{12}$$

where $T_r$ is a threshold air temperature above which all precipitation is rain and $T_b$ a threshold air temperature below which all precipitation is snow. The accumulation of snow is sometimes subject to considerable wind redistribution with drifts forming on lee slopes. We account for this in the model through a snow drift factor, F, dependent on location. Ideally F needs to be related to topography. In the application to Reynolds Creek, F was estimated by calibrating the snow water equivalences obtained from the snow model (with F = 1) at each cell, $W_m$, against the observed values, $W_o$. The discrepancy between observations and predictions over an interval between measurements is attributed to drifting and suggests $F = 1 + (W_o - W_m)/P_s$ where $P_s$ is the gage snowfall (calculated from P with F = 1) during the interval. Values of F less than one correspond to locations of depletion or wind scour. This approach models drifting which actually occurs after snowfall as concurrent with snowfall. The calibration of F assumes that the snowmelt model correctly accounts for all other processes (melt, sublimation, condensation, etc.) affecting the accumulation and ablation of snow water equivalence. Further details are given in Jackson (1994).

The temperature of rain is taken as the greater of the air temperature and freezing point and the temperature of snow is the lesser of air temperature and freezing point. The advected heat is the energy required to convert this precipitation to the reference state (0˚C ice phase).

$$Q_p = P_s\, C_s\, \rho_w \min(T_a, 0\ ^\circ C) + P_r \left[ h_f\, \rho_w + C_w\, \rho_w \max(T_a, 0\ ^\circ C) \right] \tag{13}$$

**Turbulent fluxes, $Q_h$, $Q_e$, E**

Sensible and latent heat fluxes between the snow surface and air above are modeled using the concept of flux proportional to temperature and vapor pressure gradients with constants of proportionality, the so called turbulent transfer coefficients or diffusivity a function of windspeed and surface roughness. Considering a unit volume of air, the heat content is $\rho_a\ C_p\ T_a$ and the vapor content $\rho_a$ q, where $\rho_a$ is air density (determined from atmospheric pressure and temperature), $C_p$ air specific heat capacity [1.005 kJ kg$^{-1}$ $^oC^{-1}$], and q specific humidity [kg water vapor per kg air]. Heat transport towards the surface, $Q_h$ [kJ/m$^2$/hr] is given by:

$$Q_h = K_h \rho_a\, C_p \left( T_a - T_s \right) \tag{14}$$

where $K_h$ is heat conductance [m/hr] and $T_s$ is the snow surface temperature. Vapor transport away from the surface (sublimation), $M_e$ [kg/hr] is:

$$M_e = K_e\, \rho_a\, (q_s - q) \tag{15}$$

where $q_s$ is the surface specific humidity and $K_e$ the vapor conductance [m/hr].

By comparison with the usual expressions for turbulent transfer in a logarithmic boundary layer profile (Male and Gray, 1981; Anderson, 1976; Brutsaert, 1982) for neutral condition, one obtains the following expression:

$$K_h = K_e = \frac{k^2\, V}{\left[ 1n\, (z/z_o) \right]^2} = K \tag{16}$$

where V is wind speed [m/hr] at height z [m]; $z_o$ is roughness height at which the logarithmic boundary layer profile predicts zero velocity [m]; and k is von Karman's constant [0.4]. Recognizing that the latent heat flux towards the snow is:

$$Q_e = -h_v M_e \tag{17}$$

and using the relationship between specific humidity and vapor pressure and the ideal gas law, one obtains:

$$Q_e = K_e \frac{h_v\, 0.622}{R_d\, T_a^{abs}} \left(e_a - e_s(T_s)\right) \tag{18}$$

where $e_s$ is the vapor pressure at the snow surface snow, assumed saturated at $T_s$, and calculated using a polynomial approximation (Lowe, 1977); $e_a$ is air vapor pressure, $R_d$ is the dry gas constant [287 J $kg^{-1}$ $K^{-1}$] and $h_v$ the latent heat of sublimation [2834 kJ/kg]. The water equivalence depth of sublimation is:

$$E = -\frac{Q_e}{\rho_w\, h_v} \tag{19}$$

When there is a temperature gradient near the surface, buoyancy effects may enhance or dampen the turbulent transfers. This effect can be quantified in terms of the Richardson number or Monin-Obukhov length. Adjustments to the neutral transfer coefficients to account for these effects exist and were tried based on the temperature difference between the air and snow surface. However we found that it was quite common that large temperature differences and low wind speeds resulted in unreasonable correction factors, beyond the range for which they had been developed, so for the purposes of the results presented here we have used neutral transfer coefficients.

**Snow Surface Temperature, $T_s$**

Since snow is a relatively good insulator, $T_s$ is in general different from T. This difference is accounted for using an equilibrium approach that balances energy fluxes at the snow surface. Heat conduction into the snow is calculated using the temperature gradient and thermal diffusivity of snow, approximated by:

$$Q = \kappa \rho_s C_s (T_s - T)/Z_e = K_s \rho_s C_s (T_s - T) \tag{20}$$

where $\kappa$ is snow thermal diffusivity [$m^2$ $hr^{-1}$] and $Z_e$ [m] an effective depth over which this thermal gradient acts. The ratio $\kappa/Z_e$ is denoted by $K_s$ and termed snow surface conductance, analogous to the heat and vapor conductances. A value of $K_s$ is obtained by assuming a depth $Z_e$ equal to the depth of penetration of a diurnal temperature fluctuation calculated from equation (9) (Rosenberg, 1974). $Z_e$ should be chosen so that $R_z/R_s$ is small. Here $K_s$ is used as a tuning parameter, with this calculation used to define a reasonable range. Then assuming equilibrium at the surface, the surface energy balance gives.

$$Q = Q_{sn} + Q_{li} + Q_h(T_s) + Q_e(T_s) + Q_p - Q_{le}(T_s) \tag{21}$$

where the dependence of $Q_h$, $Q_e$, and $Q_{le}$ on $T_s$ is through equations (14), (18) and (11).

Analogous to the derivation of the Penman equation for evaporation the functions of $T_s$ in this energy balance equation are linearized about a reference temperature T*, and the equation is solved for $T_s$:

$$T_s = \frac{Q_{sn} + Q_{li} + Q_p + KT_a\rho_a C_p - 0.622 K h_v \rho_a \left( e_s(T^*) - e_a - T^*\Delta \right)/P_a + 3\varepsilon_s \sigma T^{*4} + \rho_s C_s T K_s}{\rho_s C_s K_s + K \rho_a C_p + 0.622 \Delta K h_v \rho_a / P_a + 4 \varepsilon_s \sigma T^{*3}} \tag{22}$$

where $\Delta = de_s/dT$ and all temperatures are absolute [K]. This equation is used in an iterative procedure with an initial estimate $T^* = T_a$, in each iteration replacing T* by the latest $T_s$. The procedure converges to a final $T_s$ which if less than freezing is used to calculate surface energy fluxes. If the final $T_s$ is greater than freezing it means that the energy input to the snow surface cannot be balanced by thermal conduction into the snow. Surface melt will occur and the infiltration of meltwater will account for the energy difference and $T_s$ is then set to 0˚C.

**Meltwater Outflux, $M_r$ and $Q_m$**

The energy content state variable U determines the liquid content of the snowpack. This result, together with Darcy's law for flow through porous media, is used to determine the outflow rate.

$$M_r = K_{sat}\, S^{*3} \tag{23}$$

where $K_{sat}$ is the snow saturated hydraulic conductivity and S* is the relative saturation in excess of water retained by capillary forces. This expression is based on Male and Gray (1981, p. 400, eqn 9.45). S* is given by:

$$S^* = \frac{\text{liquid water volume - capillary retention}}{\text{pore volume - capillary retention}} = \left(\frac{L_f}{1 - L_f} - L_c\right) / \left(\frac{\rho_w}{\rho_s} - \frac{\rho_w}{\rho_i} - L_c\right) \tag{24}$$

where $L_f=U/(\rho_w h_f W)$ denotes the mass fraction of total snowpack (liquid and ice) that is liquid, $L_c$ [0.05] the capillary retention as a fraction of the solid matrix water equivalence, and $\rho_i$ the density of ice [917 kg $m^{-3}$]. This melt outflow is assumed to be at 0˚C so the heat advected with it, relative to the solid reference state is:

$$Q_m = \rho_w\, h_f\, M_r \tag{25}$$

**Forest Cover**

The presence of vegetation, especially forests, significantly influences energy exchanges at the snow surface. A forest canopy reduces windspeed, thus reducing sensible and latent heat transfers. It also affects the radiation exchanges. The penetration of radiation through vegetation has been widely studied (Sellers *et al.*, 1986; Verstraete, 1987a; 1987b; Verstraete *et al.*, 1990; Dickinson et al., 1993), and models developed that discretize the canopy into layers treating the energy balance of each layer separately (Bonan, 1991). Here we avoid these complexities and adopt a pragmatic parameterization modeled after the representation of snowmelt used by the WEPP winter routines (Young *et al.*, 1989; Hendrick *et al.*, 1971). Forest cover is parameterized by the canopy density parameter Fc, representing the canopy closure fraction (between 0 and 1). Windspeed, and therefore the corresponding heat and vapor fluxes, are reduced by a factor (1-0.8Fc). Radiative fluxes $Q_{sn}$, $Q_{li}$ and $Q_{le}$ in equation (1) are reduced by a factor (1-Fc). Adjustments are also made to the radiation terms in the calculation of snow surface temperature (equation 22).

## DATA

In this paper data collected at the Central Sierra Snow Laboratory (CSSL); Utah State University drainage and evapotranspiration research farm and Reynolds Creek Experimental Watershed are used to calibrate and test the model.

### Central Sierra Snow Laboratory

The CSSL located 1 km east of Soda Springs, California, measures and archives comprehensive data relevant to snow. It is at latitude 39˚19'N and at elevation 2100m. We obtained the meteorological and snow observation data for the winter of 1985 - 1986. The meteorological data is reported each hour and consists of temperature, radiation, humidity, precipitation, and wind measurements at two levels in a 40 x 50 m clearing and in a mixed conifer fir forest with 95% forest cover. Only data from the clearing are used here. Snow depths and water equivalence are measured daily (except on weekends) and eight lysimeters record melt outflow each hour. We used the temperature, precipitation, radiation (incoming solar and net), humidity and wind measurements to drive our model and compared model output to measurements of snow water equivalence, melt outflow and snow surface temperature (infrared sensor).

### USU drainage and evapotranspiration research farm

An experiment to measure snow energy balance and sublimation from snow the winter of 1992 - 1993 is described more fully by Tarboton (1994). Data from this work included measurements of snow water equivalence, snow surface temperature and the meteorological variables necessary to drive our model. The USU drainage and irrigation experimental farm is located in Cache Valley near Logan, Utah, USA (41.6˚ N, 111.6˚ W, 1350m elevation). The weather station and instrumentation are in a small fenced enclosure at the center of a large open field. There are no obstructions to wind in any direction for at least 500m. Cache valley is a flat bottomed enclosed valley surrounded by mountains that reach elevations of 3000m. During the period of this experiment the ground was snow covered from November 20, 1992 to March 22, 1993. Air temperatures ranged from -23 ˚C to 16 ˚C and there was 190 mm of precipitation (mostly snow, but some rain). The snow accumulated to a maximum depth of 0.5 m with maximum water equivalent of 0.14 m.

### Reynolds Creek Experimental Watershed

Upper Sheep Creek is a 26 ha catchment within the semi-arid Reynolds Creek experimental watershed. Snowmelt is the main hydrologic input and its areal distribution is heavily influenced by wind induced drifting. Detailed descriptions of the various features of the area are given in Flerchinger et al. (1992) and references therein. Snow water equivalence measurements are made biweekly (as weather permits) on a 30.48 m (100 ft) grid over the watershed. A digital elevation model (DEM) was constructed from a 1:1200 map with 0.61 m (2 ft) contour interval developed from low-level aerial photography. The DEM grid was constructed to coincide with the grid used for field measurements and provided slope and aspect inputs to the model radiation calculations. Fig. 2 shows the topography and grid over Upper Sheep Creek together with locations of the instrumentation. Data from the winters of 1985 - 1986 and 1992 - 1993 were used in this study to test the model running in a distributed mode at each grid cell. Snow melt outputs were used as hydrologic inputs for a water balance study (Jackson, 1994; Tarboton *et al*., 1995).

## RESULTS

The model was calibrated against the CSSL data for the winter 1985 - 1986. The energy balance and overall accumulation and ablation of the snowpack is governed primarily by surface energy exchange processes. The adjustable parameters involved in these are $z_o$ and $K_s$, which were adjusted to obtain a match between modeled and observed water equivalence (shown in Fig. 3), and modeled and observed snow surface temperatures (Fig. 4), with the model driven by the measured net radiation input. We then used measured incoming solar radiation to drive the model and found that the melt is delayed (Fig. 3). Discrepancies were analyzed and attributed to differences in daytime net radiation, primarily affected by albedo. The albedo parameterization (Dickinson et al., 1993) has parameters $A_{vo} = 0.95$ and $A_{nir} = 0.65$ which represent the albedo of new snow in the visible and infrared ranges. $A_{vo}$ was reduced to 0.85 to match the daytime net radiation when compared to measured CSSL 1985 - 1986 data (Fig. 5). The resulting snow water equivalence comparison (Fig. 3) indicates that some early season melt is not modeled resulting in slight over accumulation, but the main melt is well modeled. In all results except the line indicated on Fig. 3, $A_{vo}$ = 0.85 was used. Melt outflow rate was compared to the average from the eight melt lysimeters, with $K_{sat}$ adjusted to get a good fit. Results are shown in Fig. 6.

Table 1 lists the adjustable parameters that were calibrated against the CSSL data. Table 2 lists the remaining model parameters which were held fixed at their nominal values. The model was tested against the data from Reynolds Creek and USU drainage and evapotranspiration research farm without further adjustment of parameters. The Reynolds Creek study applied the model to each 30.48 x 30.48 m grid cell over Upper Sheep Creek (Fig. 2). The drift factor to adjust snow input was estimated from the observed grided snow data for 1985-1986 (Jackson, 1994). Fig. 7 shows the drift factors and Fig. 8 compares measured and modeled spatial distribution of snow about halfway through the snowmelt phase in 1992-1993. Due to space limitations not all of the comparisons are shown. These results indicate that the model correctly represents the spatial accumulation and melt patterns. Fig. 9 compares measured and modeled snow water equivalence at the USU drainage and evapotranspiration research farm.

## CONCLUSIONS

The tests described have shown that this simple, depth averaged, mass and energy balance snowmelt model is able to capture the essential physics of the snow accumulation and melt processes and provide distributed hydrologic inputs. Using parameter values calibrated against CSSL data the model performed well when tested at other locations. This comparison suggests that the model is transportable and parameter values listed may be acceptable for wider application. However, further testing against additional data is necessary. In particular we need to test the parameterization of forest cover and further evaluate the parameterization of albedo and the effect of atmospheric stability on turbulent fluxes.

The model is available electronically on the internet from David Tarboton (dtarb@cc.usu.edu).

**Acknowledgements** Thank you Bruce McGurk for access to the CSSL data, and Keith Cooley and the USDA ARS Northwest Watershed Research Center staff for access to and collaboration in Reynolds Creek. This work was funded in part by the US Department of the Interior, Geological Survey, under USGS Grant No. 14-08-0001-G2110, and the US Department of Agriculture, Forest Service joint venture agreement INT-92660-RJVA. The views and conclusions are those of the authors and should not be interpreted as necessarily representing the official policies, either expressed or implied, of the U.S. Government.

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Table 1. Adjustable parameter recommended values.

| Parameter | Notation | Calibrated Value |
|---|---|---|
| Surface aerodynamic roughness | $z_o$ | 0.005 m |
| Surface conductance | $K_s$ | 0.02 m/hr |
| Saturated hydraulic conductivity | $K_{sat}$ | 20 m/hr |
| New snow visible albedo | $A_{vo}$ | 0.85 |

Table 2. Snowmelt model fixed parameters.

| Parameter | Notation | Reference Value |
|---|---|---|
| Ground heat capacity | $C_g$ | 2.09 kJ kg$^{-1}$ ˚C$^{-1}$ |
| Density of soil layer | $\rho_g$ | 1700 kg m$^{-3}$ |
| Snow density | $\rho_s$ | 450 kg m$^{-3}$ |
| Capillary retention fraction | Lc | 0.05 |
| Emissivity of snow | $\varepsilon_s$ | 0.99 |
| Temperature above which precipitation is rain | $T_r$ | 3˚C |
| Temperature below which precipitation is snow | $T_s$ | -1˚C |
| Wind/air temperature measurement height | z | 2 m |
| Soil effective depth | $D_e$ | 0.4 m |
| Bare ground albedo | $A_{bg}$ | 0.25 |
| Albedo extinction depth | h | 0.1 m |

## Figure Captions

Figure 1. Depth of penetration of temperature fluctuations into soil with thermal conductivity $\alpha$ = 0.005 $cm^2/s$.

Figure 2. Upper Sheep Creek topography and instrumentation.

Figure 3. Comparison between observed and modeled snow water equivalence, CSSL.

Figure 4. Comparison between observed and modeled snow surface temperatures, CSSL. Net indicates model driven by measured net radiation. Solar indicates model driven by measured solar radiation.

Figure 5. Comparison between observed and modeled net radiation, CSSL. Measured solar radiation is input.

Figure 6. Comparison between observed and modeled melt outflow rate, CSSL. Measured solar radiation is input.

Figure 7. Drift factor from Jackson (1994). Contours at 0.5, 0.9, 1.5, 2.5, 4 and 6.

Figure 8. Observed and modeled spatial distribution of snow at Upper Sheep Creek, April 8, 1993.

Figure 9. Observed and modeled snow water equivalence, USU research farm.

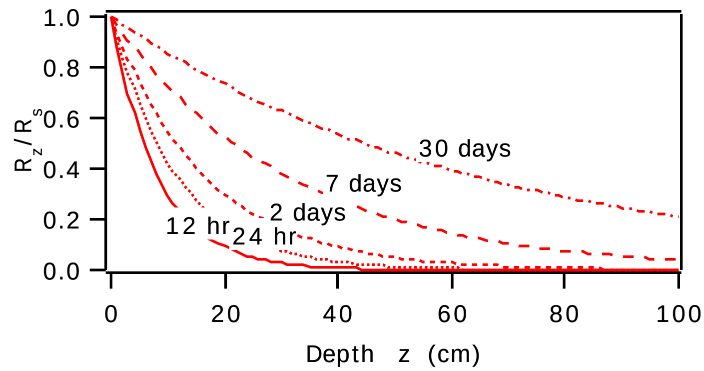



Figure 1.

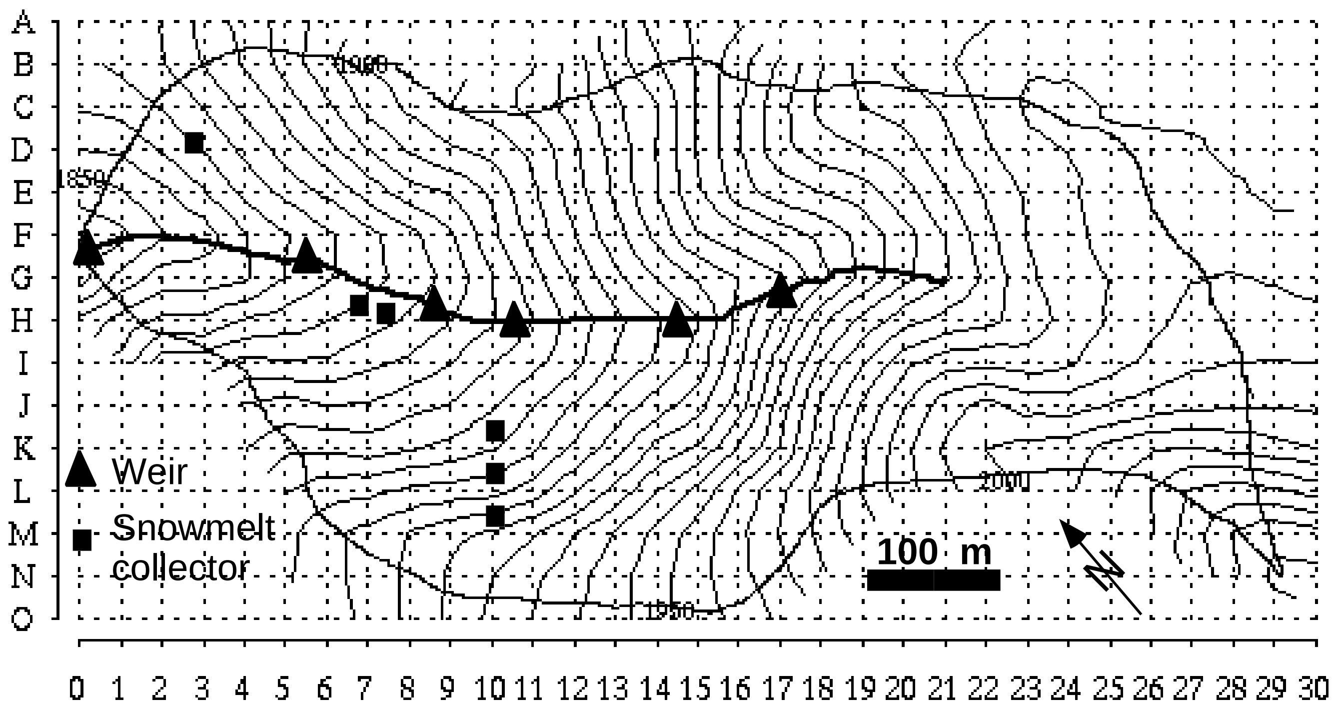
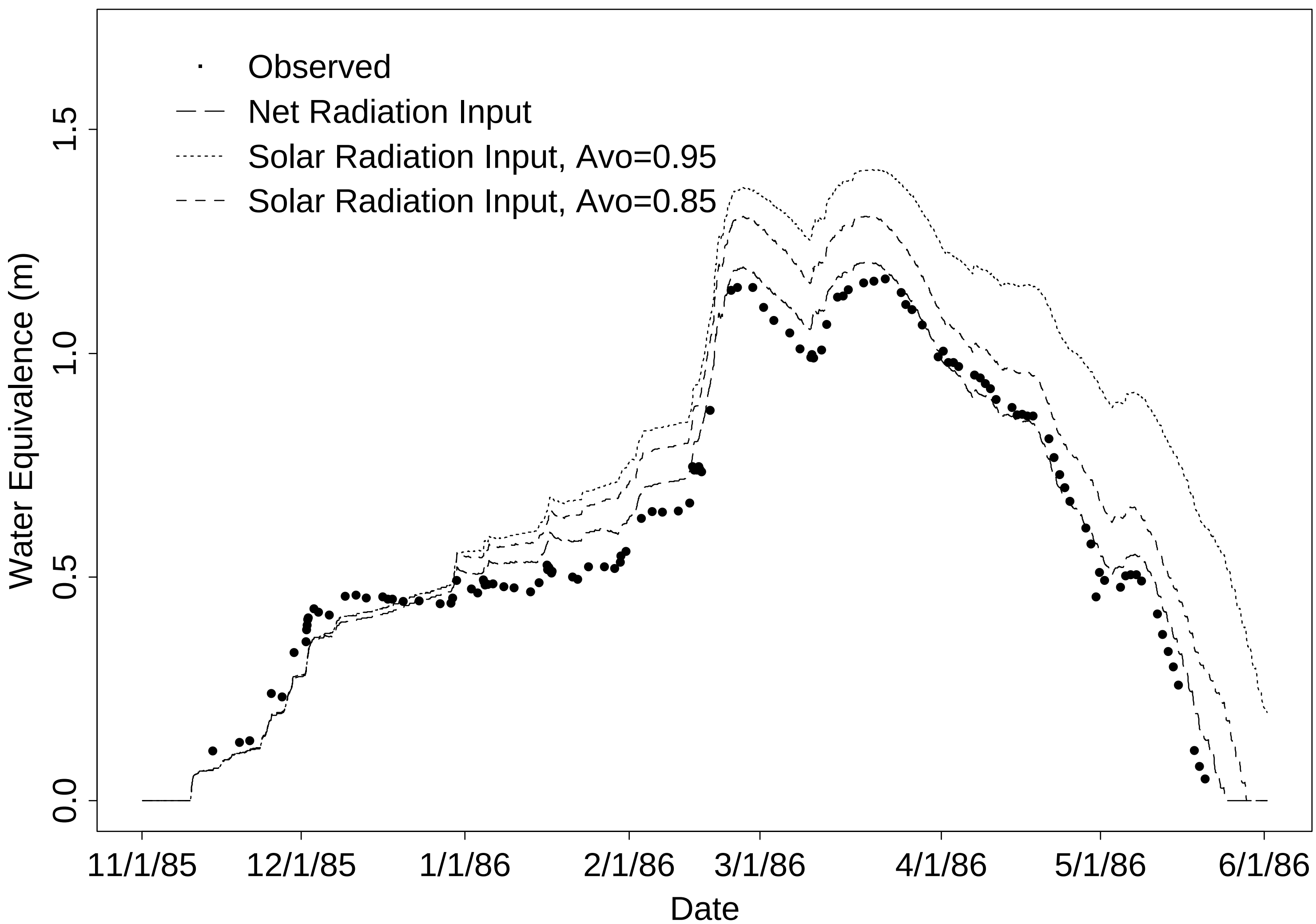
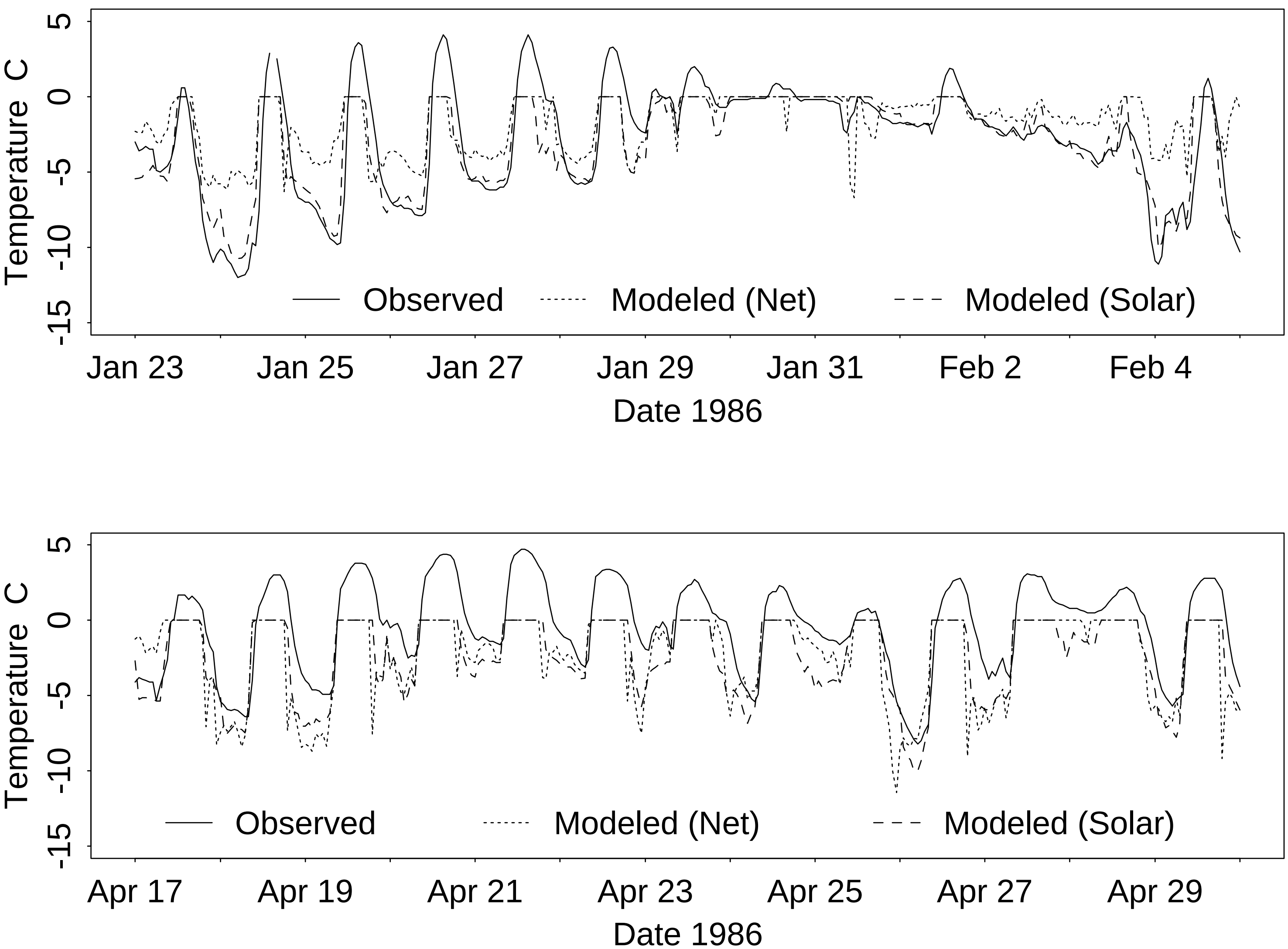



Figure 2.



Observed
Net Radiation Input
Solar Radiation Input, Avo=0.95
Solar Radiation Input, Avo=0.85
Water Equivalence (m)
0.0
0.5
1.0
1.5
11/1/85
12/1/85
1/1/86
2/1/86
3/1/86
4/1/86
5/1/86
6/1/86
Date


Figure 3.



Temperature C
5
0
-5
-10
-15
Observed
Modeled (Net)
Modeled (Solar)
Jan 23
Jan 25
Jan 27
Jan 29
Jan 31
Feb 2
Feb 4
Date 1986
Temperature C
5
0
-5
-10
-15
Observed
Modeled (Net)
Modeled (Solar)
Apr 17
Apr 19
Apr 21
Apr 23
Apr 25
Apr 27
Apr 29
Date 1986

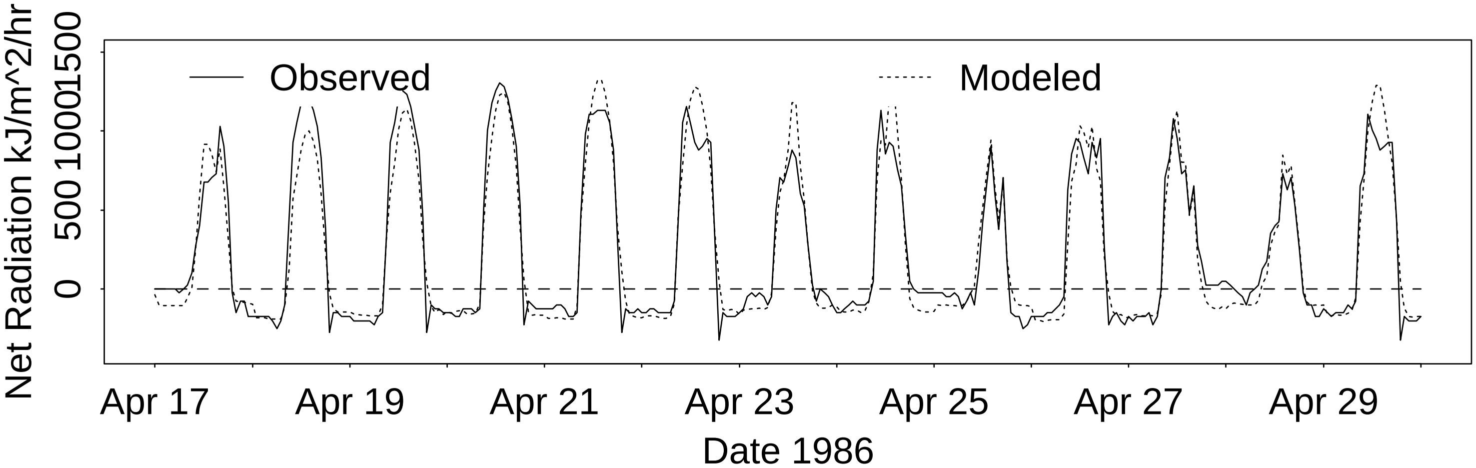



Figure 5.

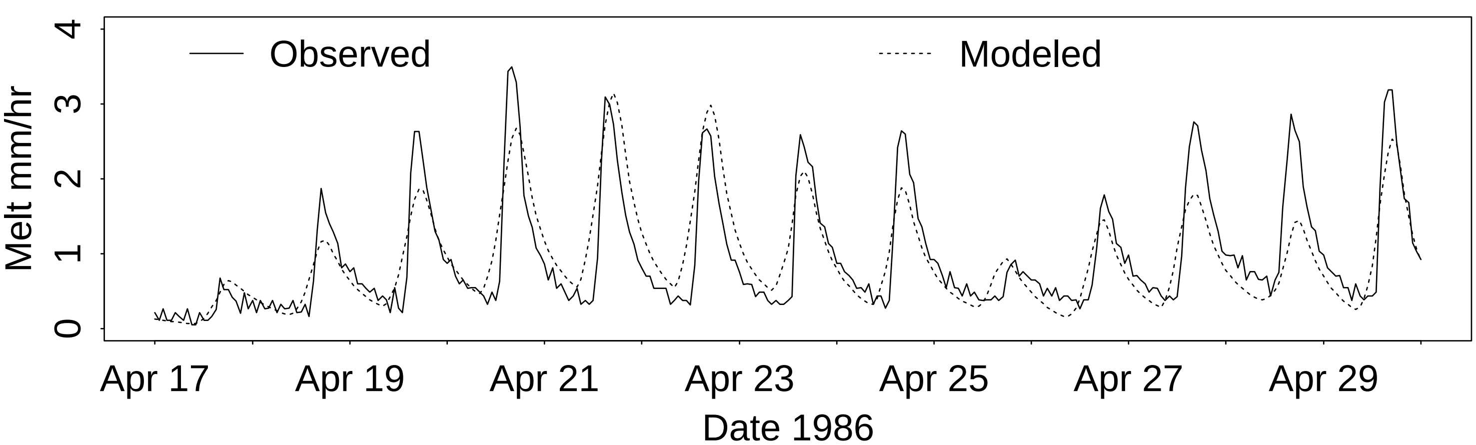
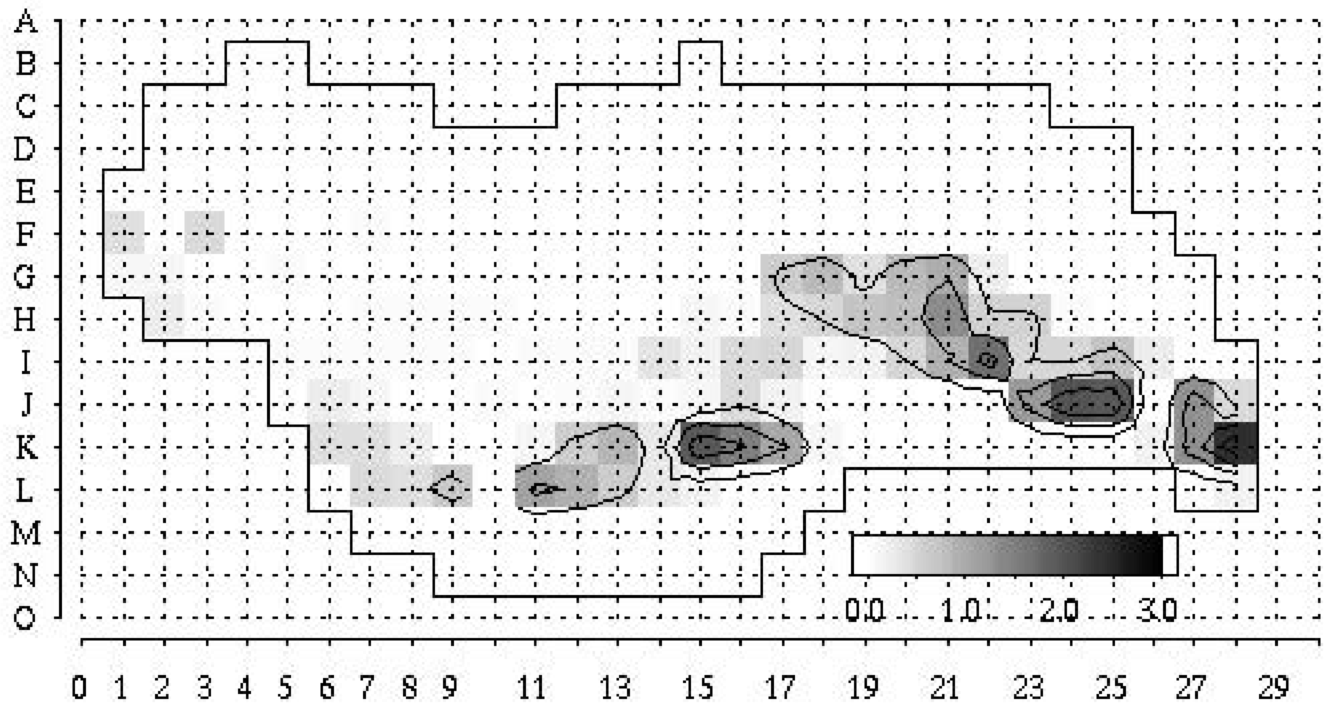
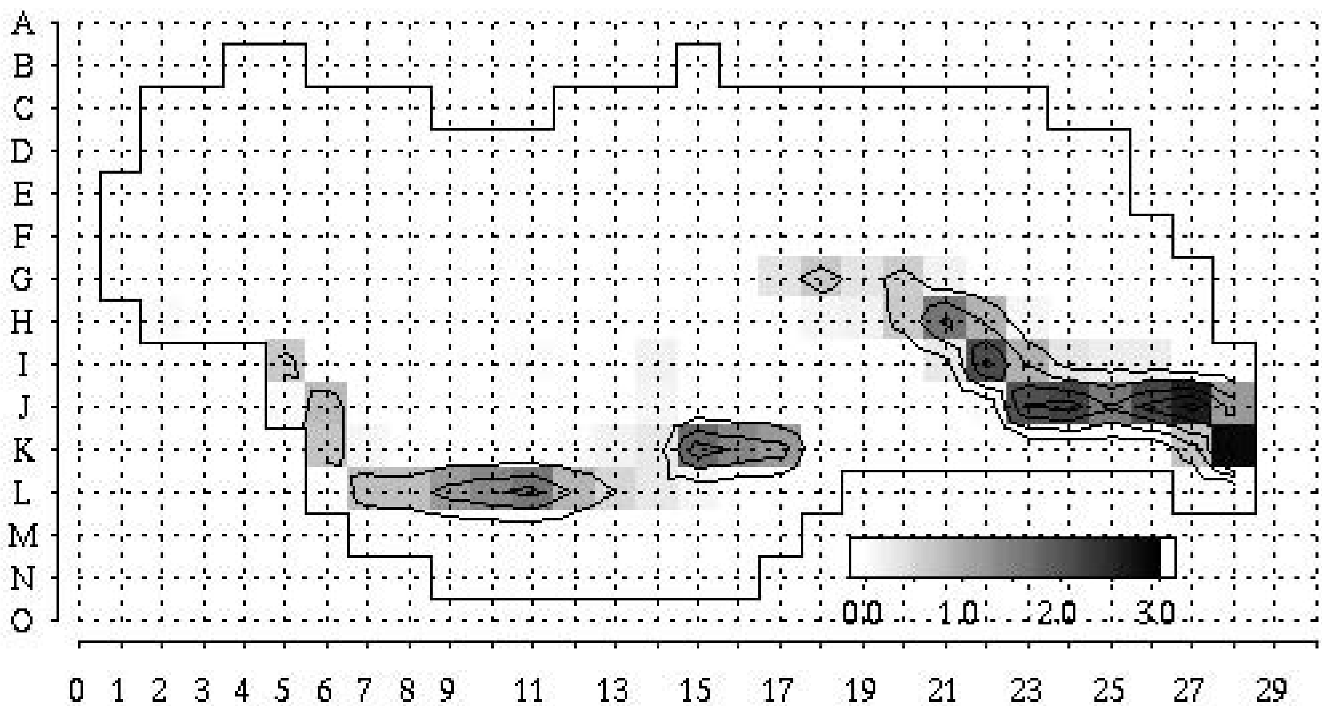



Figure 6.


Figure 7.

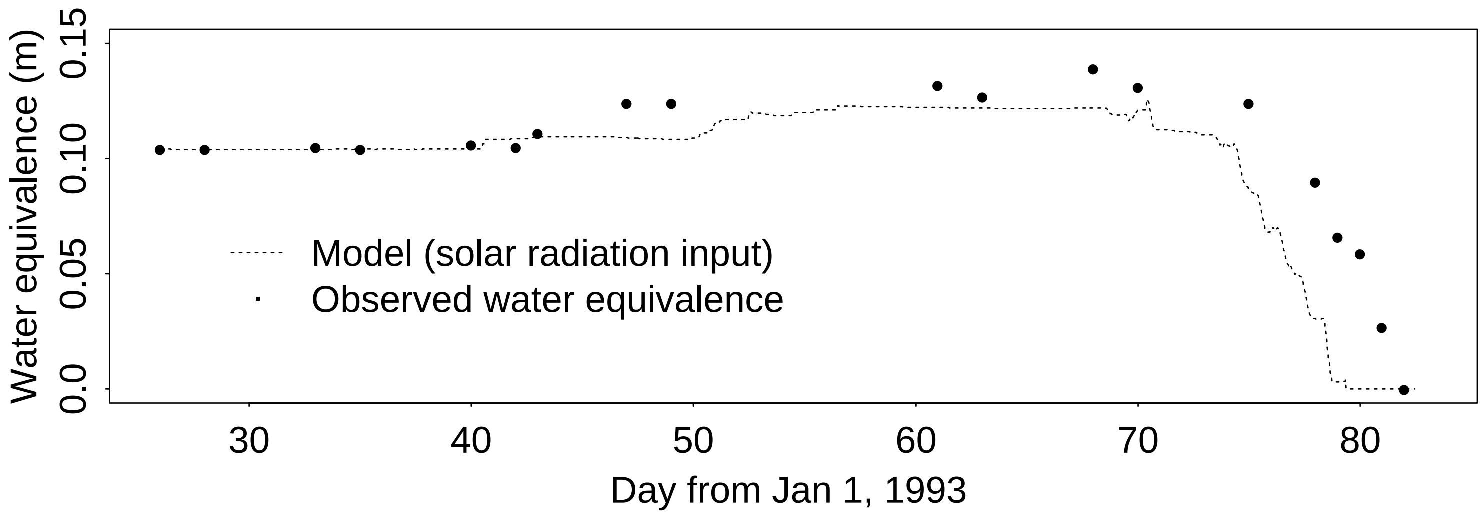



a) Observed, April 8, 1993




b) Modeled, April 8, 1993

Figure 8.


Water equivalence (m)
0.0
0.05
0.10
0.15
30
40
50
60
70
80
Day from Jan 1, 1993
Model (solar radiation input)
Observed water equivalence