DEVELOPING TIME-SERIES CLIMATE SURFACES TO DRIVE TOPOGRAPHCIALLY DISTRIBUTED ENERGY- AND WATER-BALANCE MODELS

David Susong[1], Danny Marks[2], Tim Link[3], David Garen[4]

ABSTRACT

Topographically distributed energy- and water-balance models can accurately simulate both the development and melting of a seasonal snowcover in mountain basins. The models require time-series climate surfaces of air temperature, humidity, wind, precipitation, and solar and thermal radiation. If data are available, these parameters can be adequately estimated at time steps of 1 to 3 hours. Unfortunately, climate monitoring in mountain basins is very limited, and the full range of elevations and exposures that affect climate conditions, snow deposition, and snowmelt is seldom sampled. However, detailed time-series climate surfaces have been successfully developed using limited data and relatively simple methods. A synopsis of the tools and methods used to combine limited data with simple is presented.

STUDY AREA AND TIME-SERIES DATA

A topographically distributed energy-balance snowmelt model (Garen and Marks, 1996; Marks and others, 1997; Marks and others, 1998) was used to simulate snowmelt over the Central Wasatch Range near Park City, Utah (fig. 1). Elevations in the model area range from 1,560 to 3,370 meters. There are three SNOTEL stations, two National Weather Service (NWS) climate stations and one USGS climate station. In addition to these climate data sites, two data sites in the Salt Lake Valley to the west of the model area were used to develop time-series climate surfaces to drive the model. The stations are:

Station Name	Elevation (meters)	Parameters
Brighton, SNOTEL	2,667	T_a,Pcp,SWE
Mill D,SNOTEL	2,731	T_a,Pcp,SWE
Thaynes Canyon, SNOTEL	2,843	T_a,Pcp,SWE
Park City Fire Station, NWS	2,106	T_a,Pcp
Snyderville, NWS	2,770	T_a,Pcp
Salt Lake City Airport	1,236	T_a,ea,u,Pcp
Cottonwood Air Monitoring	1,335	T_a,ea,u,Pcp
Park City Reference Site, USGS	2,636	T_a,ea,u,S_n,I

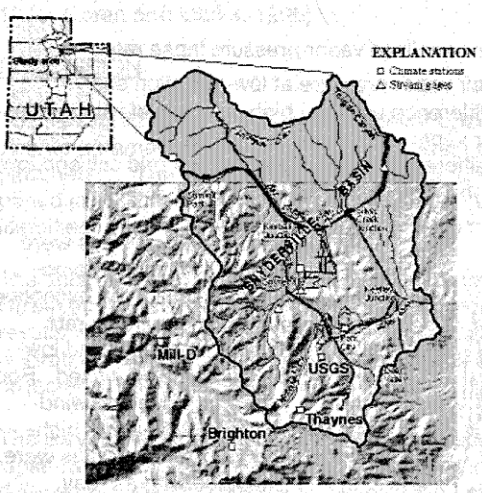

Figure 1. Location of model area and central Wasatch Range, Utah.

The time series varied at the different stations from 15-minute averages to daily average, maximum, and minimum values. The model was run at a 3-hour time step over a 75-meter digital elevation model. From the station data, 3-hour time-series climate surfaces were generated to drive the model for the following input parameters:

Air Temperature	T_a
Net Solar Irradiance	S_n
Thermal Irradiance	I
Precipitation Mass	Pcp
Precipitation Temperature	Pct
Vapor Pressure	ea
Wind Speed	u

[1] U. S. Geological Survey, Salt Lake City, Utah, ddsusong@usgs.gov
[2] USDA, Agricultural Research Service, Boise, Idaho, danny@quercus.ars.pn.ubr.gov
[3] Oregon State University, EPA NHEERL, Corvallis, Oregon, tim@heart.cor.epa.gov
[4] USDA, Natural Resources Conservation Service, Portland, Oregon, dgaren@gw.wcc.nrcs.usda.gov

METHODS TO GENERATE CLIMATE SURFACES

Three methods were used to generate climate surfaces. Solar and thermal irradiance surfaces were simulated. Elevational gradients were used to calculate vapor pressure and wind-speed surfaces, and a detrended kriging algorithm was used to generate air temperature, precipitation, and snow surfaces.

Clear-sky solar-irradiance surfaces were simulated using a clear-sky spectral solar-radiation model (Dozier, 1980; Marks and others, 1991). Clear-sky irradiance was then modified for cloud-cover effects based on the daily diurnal temperature range (Bristow and Campbell, 1984). Clear-sky thermal-irradiance surfaces were simulated with a clear-sky long-wave radiation model (Marks and Dozier, 1979) and then modified for cloud-cover effects.

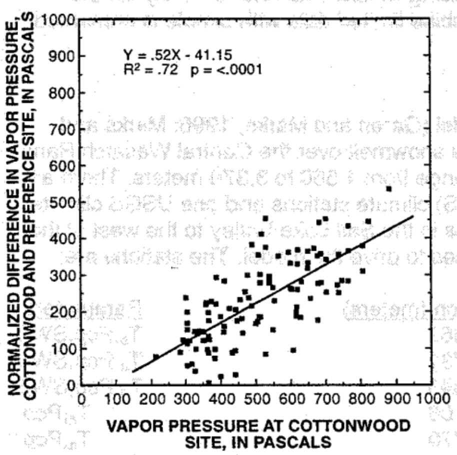

Figure 2. Linear regression of normalized differences in vapor pressure between Cottonwood and USGS reference sites and vapor pressure at the Cottonwood site.

Elevational gradients were calculated between the USGS Park City Reference site and lower-elevation stations to the west in the Salt Lake Valley. This was necessary because of the lack of complete time-series of vapor-pressure data at high-elevation stations in the model area. The procedure to calculate a lapse rate was to divide the high- by the low-elevation data and then normalize this difference to the low-elevation site. This vector was fit with linear regression to the low-elevation data and divided by the elevation difference between the sites, resulting in a lapse rate as a function of the low-elevation data (fig. 2). The lapse rate was multiplied by the difference between the pixel elevation and low-elevation site and subtracted from the low-elevation data to generate the value at each grid cell (fig 3). For example, vapor pressure at a grid cell was calculated by:

$$ea = (ea_L - (ea_{dz}) \times pdz)$$
$$ea_{dz} = (0.52\,x - ea_L\,41.5)/rdz$$

where
ea_{dz} = normalized vapor-pressure lapse rate
ea_L = air vapor pressure at low-elevation site
rdz = difference between high- and low-elevation sites
ea = air vapor pressure at a grid cell
pdz = difference in elevation between grid cell and low-elevation site

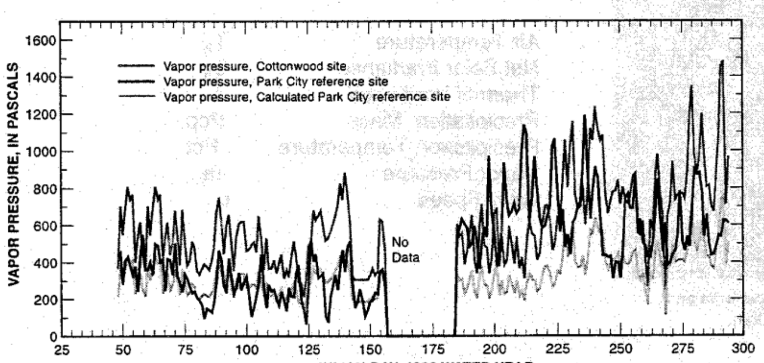

Figure 3. Measured vapor pressure at Cottonwood and Park City Reference site and calculated vapor pressure Park City reference site, Park City, Utah.

Wind-speed surfaces were calculated with the same procedure, with the exception that the mean lapse rate between the high- and low-elevation site was used. Both vapor-pressure and wind-speed vectors at the high- and low-elevation sites were smoothed with a 7-day moving average to reduce the local site effects but preserve the overall air-mass change.

Precipitation and air temperature time-series surfaces and snowcover-initial condition surfaces were developed using a detrended kriging algorithm (Garen and others, 1994, Garen, 1995). Kriging is a spatial interpolation procedure that estimates an unmeasured site as a weighted sum of nearby measurements. Detrending is required to account for nonstationarity of the field due to topography. This is done by calculating linear parameter-elevation relations from the measured data and performing spatial interpolation on the residuals. The idea of the procedure is to separate the sources of variability into elevation and distance components.

Daily precipitation surfaces were calculated using the detrended kriging algorithm and data from the three SNOTEL stations and two NWS cooperative stations. Three-hour precipitation surfaces were derived by a simple fractioning approach. The fraction of the daily precipitation at sites where there were hourly precipitation data was summed into 3-hour values. These values were then subjectively lumped and smoothed to produce a daily set of 3-hour fractions of daily precipitation. The 3-hour fractions were multiplied by the daily surfaces to produce 3-hour surfaces. The precipitation density for each 3-hour surface was calculated as a function of the dew point temperature for that period. Precipitation density and the amount of precipitation that is snow is set by:

Temperature (°C)	Percent snow	Snow density (kg/m^3)
$T < -5$	100	75
$-5 \le T < -3$	100	100
$-3 \le T < -1.5$	100	150
$-1.5 \le T < -0.5$	100	175
$-0.5 \le T < 0$	75	200
$0 \le T < 0.5$	25	250
$0.5 \le T$	0	0

Air-temperature surfaces were calculated with the same detrending algorithm. Daily maximum and minimum surfaces were created. Three-hour surfaces were obtained by passing an average diurnal cycle through the maximum and minimum temperature surfaces. The diurnal cycle was calculated using a procedure similar to that used in the National Weather Service HYDRO-17 snow model (Anderson, 1976; Garen and Marks, 1996).

DISCUSSION

The sensitivity of the simulated snowcover energy and mass balance to the simplified estimates of climate parameters was examined to evaluate our methods for distributing and estimating climate parameters. The model adequately simulated snow accumulation and melt during the March – June period in 1994 and 1995. However, as part of the sensitivity evaluation a number of problems were identified with the generated climate surfaces.

The solar and thermal models are well documented and simulate clear-sky solar and thermal radiation. However, the modification of the climate surfaces for cloud-cover effects is problematic. Although good results were achieved with the snowmelt model, actual solar- and thermal-radiation data are much more desirable. Future work will compare the simulated solar- and thermal-radiation surfaces with data collected in the Boise River Basin.

The elevational lapse-rate-derived vapor-pressure surfaces worked well during March and April but the lapse-rate relation became more complicated in May and June. At this time land surfaces are very different between the high- and low-elevation site. All the snow is melted, stream discharge is high and irrigation has begun at low elevations. Evapotranspiration becomes significant at the low-elevation site, while the high-elevation site remains snow covered. The low elevation climate site (Cottonwood) is in the Salt Lake Valley at a significantly lower elevation than the Park City study area. During winter, conditions at this site are similar to higher elevation sites, but during spring, this is no longer the case. To derive reasonable high-elevation vapor pressure surfaces, a lapse rate using only winter data had to be developed..

There is a pronounced rain shadow from west to east across the model area. The detrended kriging algorithm was unable to simulate a rain shadow with the limited number of climate stations on the east side of the model area. To correct for this, an elevational, exponential decay function was applied to the precipitation surfaces adjusting the function until simulated snow accumulation and melt matched field observations. This solution worked because the western half of the model area is all at high elevations and thus the decay function did not affect this area. Other solutions that would be more widely applicable include using synthetic data stations in the rain shadow to control the kriging or using other spatial-estimation techniques.

Simple methods like simulations, elevational lapse rates, and detrended kriging for generating distributed climate surfaces are adequate for distributed snowmelt modeling at watershed and regional scales. However, methods need to be more generalized and formalized. As part of data collection efforts, high-elevation data on humidity, wind, temperature, and radiation are critical for improving estimates of sublimation and runoff especially during the spring months. Solar and thermal radiation data are rarely measured, but if available, simplify the correction of clear-sky solar and thermal radiation for cloud-cover effects.

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