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GENERATION OF DIURNAL SOLAR RADIATION, TEMPERATURE, AND HUMIDITY PATTERNS

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ABSTRACT

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Diurnal patterns of solar radiation, temperature and humidity are needed for energy and water use evaluations of greenhouses and other agricultural structures, as well as for plant growth simulation studies. Moreover, these diurnal weather pattern data are not readily available for most locations around the world. Somewhat more available are mean monthly values of total daily solar radiation, of maximum and minimum air temperature, and of vapor pressure. Some simple models were taken from the literature and modified to generate diurnal patterns from such data. Model values were then compared to actual diurnal data from a total of eleven locations in New Zealand, Denmark and the United States. For solar radiation a simple half-cosine model had an overall root-mean-square (rms) difference from observations of 16.3 $W m^{-2}$, but the daylength was adjusted slightly to attain this low rms value. For air temperature, the model of Parton and Logan was modified to force the nighttime decrease in temperature to the minimum near sunrise. This modified equation had an overall rms difference from the observations of 0.64°C. The model for vapor pressure was a simple constant equal to the daily average, and it had an overall rms difference from observations of 0.042 kPa. Diurnal relative humidity values could then be calculated from the constant vapor pressure and diurnal temperature values. Thus, these simple models should be adequate for generating diurnal solar radiation, temperature, and humidity patterns for many applications.

INTRODUCTION

The diurnal patterns of solar radiation, temperature, and humidity are important data needed for the analysis of energy and water requirements of new designs for greenhouses (Kimball, 1973), swine housing (Goetsch and Muehling, 1984) and other agricultural structures. Such data are also

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needed for plant growth simulation models (Acock et al., 1984) in order to predict photosynthesis rates, canopy temperatures, and evapotranspiration rates. Ultimately, accurate evapotranspiration rate simulations will enable growers to make real time irrigation scheduling decisions (Swaney et al., 1983).

One obvious source for these diurnal weather patterns is to use actual measured observations, but such data are not readily available for most locations in the world. Moreover, this solution involves handling volumes of magnetic tape and long processing times. Another approach is to reconstruct the diurnal patterns from values of daily total radiation, maximum and minimum temperatures, and average daily vapor pressure. Average monthly values of minimum and maximum temperatures, and vapor pressure (or dew point) have been published and therefore are readily available for many locations or representative climates of the world (Landsberg, 1960, for example). For many analyses of long-term energy consumption and thermal performance, average monthly computations are appropriate, although some cases require analyses of all individual days. The purpose of this paper is to present models for generating diurnal patterns from daily total solar radiation, minimum and maximum temperatures, and average daily vapor pressures. The models will be validated using average monthly data for several locations representative of a wide range of climates.

DESCRIPTIONS OF THE MODELS

Solar radiation. A model for generating the diurnal patterns of solar radiation is a simple cosine bell, as first presented by Hirschmann (1974):

$$S = S_{\mathrm{noon}} \cos[\pi(t-12)/D] \tag{1}$$

where S is the instantaneous global solar radiation (W m^{-2}), S_{noon} is the global solar radiation at solar noon, t is solar time (h), and D is the daylength (h). Integrating (1) from time t_b to t_e (h) yields:

$$S_I = S_{\mathrm{noon}}(3600/10^6)(D/\pi)\{\sin[\pi(t_e-12)/D]-\sin[\pi(t_b-12)/D]\} \tag{2}$$

where S_I is the integrated solar radiation (MJ m^{-2}). Denoting S_d as the total daily value of S_I, using $t_b = 12-(D/2)$ (sunrise) and $t_e = 12+(D/2)$ (sunset) in equation (2), and rearranging yields (Hirschmann, 1974):

$$S_{\mathrm{noon}} = \left[\frac{S_d\pi}{2D}\right]\left[\frac{10^6}{3600}\right] \tag{3}$$

The total daily solar radiation, S_d, is available for many locations or representative climates from literature sources, as discussed in the Introduction. The daylength, D, can be computed from the latitude, ℓ (degrees), and day of year, n. First the declination, δ (radians from the equator) can be computed from (Kimball, 1973):

$$\delta = 23.5(\pi/180)\cos[2\pi(n-172)/365] \tag{4}$$

Then, the astronomical daylength, D_a, is calculated from (Sellers, 1965:

$$D_a = 2\{\arccos[-\tan(\ell)\tan(\delta)]\}(180/\pi)/15 \qquad (5)$$

Using data from a few clear days, Hirschmann (1974) and also Jackson et al., (1983) found that a better fit of equation (1) to experimental data could be obtained using:

$$D_h = HD_a \qquad (6)$$

where $H = 0.946$.

As will be demonstrated later, we found that the D_h from equation (6) worked well for short days but progressively worse for longer days. To improve the long-day predictions, we developed an additional daylength adjustment:

$$D = HD_a[1+f(8-D_a)] \qquad (7)$$

where the eight bases the adjustment with respect to an 8-h day. An f of 0.008 produced the best fit to Invercargill, New Zealand, experimental data.

Equation (1) is an half-cosine bell, and therefore it cannot be expected to work well in the summertime at latitudes near the poles where there is little or no night, and the solar altitude more closely follows a full cosine curve. We attempted to remove this seasonal latitude limitation. Similar to equation (1), the instantaneous global solar radiation can be computed from:

$$S = S_N \cos(Z) \qquad (8)$$

where S_N (W m^{-2}) is the solar radiation intensity on a plane perpendicular (normal) to the solar beam, and Z is the angle of incidence (radians) of the beam with a horizontal surface. For this case the angle of incidence is in fact the solar zenith angle. From Sellers (1965) the zenith angle can be computed from:

$$\cos(Z) = K_2 + K_1 \sin[15\pi(t-12)/180] \qquad (9)$$

where $K_1=\cos(\ell)\cos(\delta)$, $K_2=\sin(\ell)\sin(\delta)$, and the expression in the brackets is the hour angle, the angle the earth has turned at time t with respect to solar noon. Integrating (8) from $t=12-(D_a/2)$ to $12+(D_a/2)$ (sunrise to sunset) an expression relating S_N to daily total solar radiation, S_d, is obtained, analygous to equation (3):

$$S_N = \frac{S_d}{K_1 D_a + K_2(2/15)(180/\pi)\sin[(15/2)(\pi/180)D_a]} \qquad (10)$$

Equations (8—10) should be a more general model than equations (1)—(3) for predicting the diurnal variation of solar radiation from daily totals because they are based on astronomical geometry (Sellers, 1965) rather than being purely empirical. However, equations (8)—(10) do not account for variations in attenuation of the solar beam due to differences in atmospheric clarity or path length, nor can simple manipulations of D, such as in equations (6) and (7), be expected to improve the fit to observed data.

Temperature. The following equations adapted from Parton and Logan (1981) offer a means to calculate air temperatures, T (°C), from values of the minimum, T_{min} (°C), and maximum, T_{max} (°C) temperatures.

— daytime:

$$T = T_{min} + (T_{max} - T_{min}) \sin\left[\frac{\pi(t - t_{min})}{D_a + 2a}\right] \tag{11}$$

— nighttime:

$$T = (T_{min} - d) + [T_{set} - (T_{min} - d)] \exp\left[\frac{-b(t - t_{set})}{(24 - D_a + c)}\right] \tag{12}$$

where t_{min} is the time of the minimum temperature and equals $t_r + c$, the sunrise time plus a lag time, and D_a is the astronomical daylength from equation (5). T_{set} is the temperature at sunset computed from equation (11), and t_{set} is sunset time. If t is between 0 (midnight) and t_{min}, then 24 h is added to t for use in equation (12). The d is a displacement that forces T to T_{min} at $t = t_{min}$ and is computed from:

$$d = (T_{set} - T_{min})/[\exp(b) - 1] \tag{13}$$

Parton and Logan (1981) did not have a displacement, d, in their night equation, and therefore, T could never reach T_{min}. Furthermore, because they did not have such a displacement, there was a discontinuity between the end of the night equation and the beginning of the day equation.

Parton and Logan (1981) used a year's worth of hourly temperature data from northeastern Colorado (about 41°N latitude) to find values of a, b, and c that produced a best fit. For air temperature at a height of 1.5 m, these values were $a = 1.86$ h, $b = 2.20$ and $c = -0.17$ h.

We attempted to make the temperature equations more general and independent of empirical constants from a particular location. Like Parton and Logan, we chose a cosine bell for daytime, but rather than fix the time of the temperature maximum at 1.86 h after solar noon, we varied it with daylength. Relying on experience in Phoenix, the temperature maximum was placed midway between noon and sunset at $t_x = 12 + (D_a/4)$ h. The minimum temperature was assumed to occur at a time $t_{min} = t_r + \alpha D_a$ after sunrise where $\alpha = 0.06$:

— daytime:

$$T = T_{min} + (T_{max} - T_{min}) \cos\left[\frac{\pi(t - t_x)}{2(t_x - t_{min}}\right] \tag{14}$$

At night the air temperature was assumed to decay toward the average temperature of the night sky starting at αD_a h before sunset:

— nighttime:

$$T = T_{sky} + (T_{set} - T_{sky}) \exp\{-k[t - (12 + (D_a/2) - \alpha D_a)]\} \tag{15}$$

where again if t is between 0 (midnight) and t_{min} then 24 h are added to t. The maximum temperature at the start of the decay is computed from (14) with $t = 12 + (D_a/2) - \alpha D_a$:

$$T_{set} = T_{min} + (T_{max} - T_{min}) \cos\left[\frac{\pi(1-4\alpha)}{2(3-4\alpha)}\right] \tag{16}$$

The decay constant is chosen to force $T = T_{min}$ at $t = t_{min}$:

$$k = \frac{\ln\left[\dfrac{T_{set} - T_{sky}}{T_{min} - T_{sky}}\right]}{24 - D_a + 2\alpha D_a} \tag{17}$$

Idso and Jackson (1969) developed the following equation for computing sky emittance:

$$\epsilon = 1 - 0.261 \exp(-7.77 \times 10^{-4}\, T^2) \tag{18}$$

where T is in °C. Using an average night temperature of $(T_{set} + T_{min})/2$ in (18), the sky temperature can then be computed from:

$$T_{sky} = [(T_{set}+273 + T_{min}+273)/2]\, \epsilon^{1/4} - 273 \tag{19}$$

Vapor pressure and humidity. Plots of vapor pressure against time of day generally do not reveal any consistent patterns, and the fluctuations are relatively small. As discussed by Geiger (1959) and Campbell (1977), the vapor pressure close to a wet surface decreases with dew fall at night and increases during the daytime, but at screen level (1.5 m above ground) the diurnal fluctuations are relatively small. Over a dry surface the fluctuations are even smaller with little or no pattern. Therefore, the basic model for humidity is simply that vapor pressure is constant and equal to the average for the day. Of course, it follows that dew point, water vapor density, and humidity ratio are also constant. Relative humidity, wet bulb temperature, and enthalpy do change greatly during the course of a day because of changes in dry bulb temperature. These parameters can be computed using the constant average vapor pressure and the dry bulb temperature from equations (11) and (12) — or (14) and (15) — using the usual psychrometric relationships (ASHRAE, 1972; Weiss, 1977; Kimball, 1981).

DATA FOR MODEL VALIDATION

Hourly average observations of temperature, relative humidity, and solar radiation were available for four New Zealand locations (Leslie and Trethowen, 1977). As listed in Table 1, these were Auckland, Wellington, and Invercargill with data from November 1969, through October 1974, and Christchurch with data from November 1967, through October 1974. For each hour of the day for each month for each of these locations, all of these data were averaged together to obtain 'average days' for each month of the year. For example, all of the 01:00 a.m. readings taken in January for each of the 5 years of data for Auckland were averaged together (31 days in January × 5 years = 155 observations) to become the 01:00 a.m. value for the average January day for Auckland. Since these were hourly average data, equations (1), (8), (11), (12), (14) and (15) were integrated so that integrated hourly average model values could be compared to the hourly average data. The hourly data were collected at integer local standard time, so the solar times corresponding to these local times also had to be calculated for use in computing the predicted values.

Additional temperature data were also available for six New Zealand locations: Auckland airport, Ohakea, Wellington-Kelburn, Christchurch, Dunedin, and Invercargill. Listed in Table 1 with an '(MS)' after each site name, these data were hourly averages for an average day for each month

TABLE 1

Locations for which weather data were analyzed and the types and length of record of the data

City	State/Country	Latitude	Longitude	Data Type[a]			
		(degrees)		Solar radiation	Temperature	Vapor pressure	Record length year
Auckland[b]	New Zealand	36.9S	174.8E	A	A	A	5
Wellington[b]	New Zealand	41.28S	174.77E	A	A	A	5
Christchurch[b]	New Zealand	43.53S	172.62E	A	A	A	7
Invercargill[b]	New Zealand	46.2S	168.4E	A	A	A	5
Auckland (MS)[c]	New Zealand	36.9S	174.8E	—	A	—	16
Ohakea (MS)[c]	New Zealand	40.16S	175.55E	—	A	—	21
Wellington (MS)[c]	New Zealand	41.28S	174.77E	—	A	—	19
Christchurch (MS)[c]	New Zealand	43.53S	172.62E	—	A	—	21
Dunedin (MS)[c]	New Zealand	45.85S	170.52E	—	A	—	19
Invercargill (MS)[c]	New Zealand	46.2S	168.4E	—	A	—	21
Brownsville[d]	Texas, U.S.A.	25.9N	97.4W	A	I	I	1
Phoenix[d]	Arizona, U.S.A.	33.43N	112.02W	A	I	I	1
New York[d]	New York, U.S.A.	40.8N	73.0W	A	I	I	1
Bismarck[d]	N. Dakota, U.S.A.	46.77N	100.75W	A	I	I	1
Hojbakkegaard[e]	Denmark	55.68N	12.56E	A	A	—	10

[a]A, hourly averages; I, instantaneous.
[b]From Leslie and Trethowen (1977).
[c]From New Zealand Meteorological Service (1984). The '(MS)' has been added after these city names to distinguish these data from other New Zealand data[b].
[d]From typical Meteorological Year (TMY) data, National Climatic Center (1978).
[e]From H.C. Aslying and K.J. Kristensen, unpublished data (1977). Coordinates are for Copenhagen.

of the year from about 20 years' worth of observations. They were averaged by the New Zealand Meteorological Service (1984).

We also had a set of the Typical Meteorological Year (TMY) Tapes (Sandia Laboratories, 1978; National Climatic Center, 1978). As listed in Table 1, four sites were selected which represent a wide range of climates: Brownsville, TX; Phoenix, AZ; New York, NY; Bismarck, ND. The data for each month of the year in a TMY data set are actual data for a month from some particular year that was judged to be most typical of that month. Thus these data were already preselected to represent typical or near average, rather than extreme conditions. The data for each hour were averaged for each of these months to form the desired average days for each month of the year for each location. The solar energy data were hourly averages, whereas temperature and dew point data were instantaneous values measured at 1 or 3-h intervals. Vapor pressures were calculated from the dew points, and comparisons were made with predicted instantaneous model calculations for temperature and vapor pressure at 3-h intervals and with hourly averages for solar radiation, again all corrected from solar to local time.

An additional set of solar radiation and temperature data was furnished by Aslying and Kristensen (unpublished data) for Hojbakkegaard, Denmark, near Copenhagen. From hourly average observations, they compiled 'average days' for each month using 10 years' worth of data collected from 1966 through 1975.

RESULTS AND DISCUSSION

Solar radiation

The results of the solar radiation computations and comparisons are presented in Table 2 and at the top of Fig. 1. Referring to Table 2, the half-cosine model using the full daylength (H = 1) had a root-mean-square (rms) difference averaged over all locations for whole days of 29.9 W m^{-2}. This is a low value considering that peak solar radiation can exceed 1000 W m^{-2}, and therefore the simple cosine model would be adequate for many applications. However, the more sophisticated astronomical equation model was consistently better for all locations whether considering just the rms differences for whole days or just at midday (Table 2).

Both the half-cosine model with full daylength (H = 1) and the astronomical model tended to underestimate the peak radiation at midday and to overestimate the radiation near sunrise and sunset. A similar observation led Hirschmann (1974) to propose multiplying the daylength used in the half-cosine model by 0.946, and the results from this modification are also presented in Table 2 and Fig. 1. The rms values were lowered to a whole day average of 18.7 W m^{-2} for all locations. After studying plots of all the data, however, there appeared to be excellent agreement for winter days, but on summer days the model still underpredicted midday values and overpredicted dawn and dusk values, as illustrated for a winter

TABLE 2

Root-mean-square differences (W m^{-2}) between solar radiation observations and values computed from an astronomical model and a half-cosine model using various values of the H and f factors to adjust daylength (equation 7)

Site	Model type			
	Astro-nomical	Half-cosine		
		$H = 1.0$, $f = 0.0$	$H = 0.946$, $f = 0.0$	$H = 0.946$, $f = 0.008$
Equations:	(8)–(10)	(1), (3), (5)	(1), (3), (5), (6)	(1), (3), (5), (7)
Whole-day differences:				
Auckland	21.8	24.7	11.8	12.3
Wellington	23.9	27.8	14.7	9.4
Christchurch	26.0	29.9	17.3	9.1
Invercargill	24.2	28.3	17.2	11.9
Brownsville	31.7	32.9	22.3	23.8
Phoenix	32.3	34.7	20.1	22.3
New York	35.6	37.6	30.5	28.4
Bismarck	27.6	32.6	22.3	18.7
Hojbakkegaard	14.0	21.0	12.3	10.6
Average	26.3	29.9	18.7	16.3
Midday differences:				
Auckland	41.4	47.6	19.8	13.8
Wellington	41.7	50.0	25.1	12.1
Christchurch	53.5	60.9	35.9	13.4
Invercargill	44.4	52.6	28.9	8.1
Brownsville	50.6	53.2	26.3	28.3
Phoenix	44.8	48.8	12.0	29.0
New York	48.4	54.0	30.4	17.1
Bismarck	43.6	54.3	32.5	22.3
Hojbakkegaard	23.4	37.6	19.1	10.8
Average	43.5	51.0	25.6	17.2

and a summer month for two locations in Fig. 1. To improve the fit, we added the daylength adjustment of equation (7). Using Hirshmann's value of 0.946 for H, we adjusted f to produce the minimum year-long rms difference for the Invercargill data. This optimum value for f was 0.008, which was then used to compute modified half-cosine model rms differences in the right-hand column of Table 2 ($H = 0.946, f = 0.008$).

The results of using the half cosine model with the adjusted daylength ($H = 0.946, f = 0.008$) was another reduction of the rms differences averaged

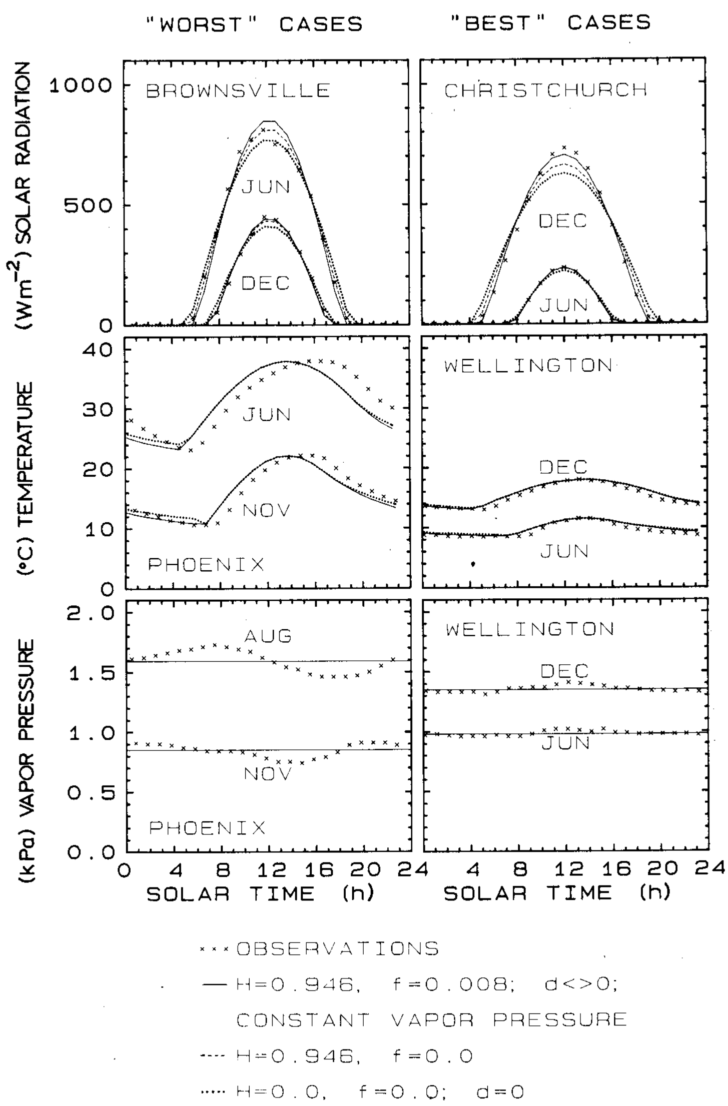

Fig. 1. Diurnal patterns of predicted and observed solar radiation, air temperature, and vapor pressure for selected locations and months. The cases on the right were selected because they had the best or near best fit of predicted to observed data, and similarly on the left for worst cases.

over all locations. However, for two locations, Brownsville and Phoenix, using the adjustable daylength produced larger rms differences. Considering that Hirshmann developed his 0.946 factor from clear day data and that Phoenix probably has the greatest number of clear days of the sites considered here, it is not too surprising that using H = 0.946 with f = 0 produced the better results for Phoenix. For Brownsville in summer, there appears to be a definite asymmetric decrease in solar radiation at midday (Fig. 1), probably caused by increasing clouds, and this asymmetry prevented obtaining lower summertime rms values there. Overall, the half-cosine model using the adjusted daylength ($H = 0.946$, $f = 0.008$) produced excellent results, the mean whole day rms value averaged for all locations was only 16.3 W m^{-2}.

Thus, the simple half-cosine model with a daylength adjustment produced the best fit to the comparison data sets, which do represent a wide range of climatic conditions. However, the fit is likely to become worse closer to the poles for summertime data, and for these extreme high and low latitudes, possibly the astronomical model (equations 8—10) would be better.

Temperature. The results of the air temperature computations and comparisons are presented in Table 3 and the middle of Fig. 1. Referring to Table 3, the Parton-Logan equations (11)—(12) as presented by them with no displacement ($d = 0$) did a good job of simulating the diurnal variation of air temperature, the rms difference averaged over all locations being 0.71°C.

However, the Parton-Logan equations are discontinuous at dawn, as already mentioned, and as can be seen in the November curve for Phoenix. Furthermore, their night equation does not go through the minimum temperature. Adding the displacement (d, equation 13) makes the night equation reach the minimum, and it improved the fit, reducing the overall rms difference to 0.64°C even with the same empirical constants found by Parton and Logan without a displacement.

Scanning the Parton-Logan rms differences in Table 3, it is obvious that the fit is much worse for Phoenix than for other locations. Being from Phoenix, we were familiar with the fact that the temperature maximums here occur close to 16 hours, and thus we were rather skeptical of the Parton-Logan equations which used a constant of 1.86 to fix the temperature maximum at 13.86 h. We had already constructed the sky temperature model (equations 14—19) which had the temperature maximum halfway between noon and sunset and which decayed at night toward sky temperature. Our model had a lower rms difference for Phoenix of 0.92°C compared to 1.66°C for Parton and Logan's. Because our model could accomodate changes in daylength, we expected it to be quite general, but on the contrary it had a higher rms difference than Parton and Logan's for every other location besides Phoenix considered in this study. The overall average rms difference for the 'sky temp' model was 1.10°C compared to 0.71°C for Parton and Logan's.

Thus, the Parton-Logan model, with or without added displacement, produced good fits to the comparison data sets. This result is somewhat surprising, considering that the coefficients came from one location, Colorado. Like the simple half-cosine model for solar radiation, however, we

TABLE 3

Root-mean-square differences between temperature observations and values computed from a model that featured night decay toward sky temperature and from the Parton-Logan model without and with a displacement; also the root-mean-square differences between vapor pressure observations and a model that simply makes vapor pressure a constant each day

Site	Temperature (°C)			Constant vapor pressure (kPa)
	Sky temperature	Parton-Logan		
		Without displacement	With displacement	
Equations:	(14)–(19)	(11)–(12)	(11)–(13)	
Auckland	0.72	0.44	0.45	0.043
Wellington	0.98	0.49	0.43	0.026
Christchurch	1.08	0.65	0.64	0.035
Invercargill	1.04	0.58	0.53	0.057
Auckland (MS)	0.67	0.40	0.42	—
Ohakea (MS)	1.08	0.59	0.53	—
Wellington (MS)	0.88	0.43	0.38	—
Christchurch (MS)	1.08	0.61	0.61	—
Dunedin (MS)	1.20	0.73	0.70	—
Invercargill (MS)	1.03	0.57	0.51	—
Brownsville	1.70	1.00	0.46	0.058
Phoenix	0.92	1.66	1.78	0.049
New York	0.87	0.50	0.51	0.035
Bismarck	1.66	0.86	0.75	0.036
Hojbakkegaard	0.91	0.35	0.29	—
Average[a]	1.10	0.71	0.64	0.042

[a]Values from duplicate locations (Auckland, Wellington, Christchurch, and Invercargill) were weighted by 0.5.

would expect this model to work less well near the poles when days or nights are much longer than those represented by our comparison data sets.

Vapor pressure. The comparisons of the observations of vapor pressure against the daily averages are presented in the right hand column of Table 3 and at the bottom of Fig. 1. The rms difference averaged over all the locations was only 0.042 kPa, so simply assuming that vapor pressure is constant each day is actually quite a good model. Geiger (1959) and Campbell (1977) discussed how vapor pressure decreases at night and increases during the day over a wet surface, and the Wellington data for June (Fig. 1) exhibit this pattern as strongly as any location for any month considered in this study. Even though the pattern is discernable, the amplitude is small, so for many energy balance studies of structures, the simple constant vapor pressure model is probably adequate. Considering also the Phoenix data where the surface could not be considered wet all the time, the diurnal pattern, if any, is opposite to that of Wellington with a decrease during the daytime. Overall, therefore, simply using a constant vapor pressure value for each day should be an adequate model.

If a particular location has a wet surface with dew fall at night, then the average daily vapor pressure will be higher than the nighttime vapor pressure. Under this circumstance, relative humidity values slightly higher than 100% will be predicted. However, the error generally will not be very large, and if desired, the user could constrain the relative humidity not to exceed 100%. Alternatively, the saturation vapor pressure at the minimum temperature could be used for the daily vapor pressure, thus making it unnecessary to obtain the actual average vapor pressure which is not as widely published as are values of minimum temperatures.

CONCLUSIONS

For modeling the diurnal pattern of solar radiation from daily totals, the half-cosine model works well. However, the value of the daylength should be adjusted using equation (7) to obtain a better fit, particularly at midday. The model of Parton and Logan (1981) modified to include a displacement generally works well to predict diurnal temperature patterns from daily maximums and minimums. Some caution should be exercised, however, when applying it to desert areas, because it did not fit the Phoenix data very closely. Vapor pressure changes very little during a day, so a constant daily average can be used for it. Relative humidity can be calculated using the constant vapor pressure value and the predicted temperature (equations 11—13) in standard psychrometric equations.

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