

ASSESSING INTEGRITY OF WEATHER DATA FOR REFERENCE EVAPOTRANSPIRATION ESTIMATION

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ABSTRACT: Procedures and guidelines are recommended for assessing integrity, quality, and reasonableness of measured weather data and equipment calibration for automated and electronic agricultural weather stations. The procedures include calculation of hourly and 24-h clear sky envelopes for solar radiation, validation of net radiation measurements using calculation equations, and evaluation of expected trends and relationships between air vapor content and air temperature. The procedures for creating clear sky solar radiation envelopes include equations to account for the effects of atmospheric water vapor content and sun angle. Procedures for adjusting air temperature and air vapor content data are introduced to compensate for the aridity of the weather station environment. All of the guidelines are simple and straightforward, and can serve as preliminary "filters" by which to scrutinize weather measurements and as near real-time data flagging procedures for agricultural weather networks.

INTRODUCTION

Estimates of reference evapotranspiration (ET_0) are no better than the weather data upon which they are based. Assessments of weather data integrity and quality need to be conducted before the data are utilized in ET_0 equations. When necessary, corrections to the data should be made to account for poor sensor calibration or effects of a poorly watered environment. Data integrity screening procedures should be a part of every agricultural weather network to insure that high quality and representative data are being measured and transferred to users.

Various techniques for assessing quality of measured weather data in near real-time have been implemented by the California Irrigation Management Information System (CIMIS) (1988). These guidelines include computation of extreme outliers for weather data measurements, but do not generally provide means for assessing quality of data which fall within extreme ranges. Ley (1995) applied a discrete Kalman filter/multivariate autoregressive (1) stochastic model to forecast daily weather data in an attempt to identify data outliers. Effects of the mechanics of data logger operation such as voltage measurement, equipment wiring techniques, and sampling and reporting frequency have been addressed by Tanner (1990) and Ley and Elliott (1993).

This paper presents guidelines to be used to calculate both extreme ranges for weather data measurements and also means to assess integrity of data which fall between the extremes. These guidelines are by no means a substitute for reliable, precision equipment, regular site visitation, proper equipment calibration and instrument replication, but address environmental and integrity issues and effects on recorded data.

When applying the Penman-Monteith, Penman or other evapotranspiration (ET) equations to predict reference evapotranspiration, it is desirable to measure wind, humidity and temperature over a "reference" surface. [Reference evapotranspiration was defined in the United Nations Food and Agriculture Organization (FAO) ID Paper 24 for a grass reference as "the rate of evapotranspiration from an extensive surface of 8 to 15 cm tall, green grass cover of uniform height, actively growing, completely shading the ground and not short

of water" (Doorenbos and Pruitt 1977).] In the case of grass reference ET_0 , this implies a uniform, well-watered clipped grass surface below and surrounding the weather equipment for a distance of 100 times the height of the wind, air temperature and relative humidity sensors in all directions. This requirement can be important for ET_0 calculations, as the rates of ET and sensible heat (H) from vegetation affect the shapes and magnitudes of vapor and air temperature profiles. In the same manner, the roughness of an evaporating surface affects the shape and magnitude of wind velocities above the surface. As ET from the surroundings of a weather station decreases, sensible heat transfer increases resulting in increased air temperature and a reduction in water content (humidity) above the surface.

SOLAR RADIATION

Clear Sky Comparisons

Pyranometer operation and calibration accuracy can be evaluated by plotting hourly or daily average pyranometer readings against computed short wave radiation expected under clear sky conditions (R_{so}). R_{so} can be computed as

$$R_{so} = K_T R_a \quad (1)$$

where R_a = extraterrestrial radiation; and K_T = "clearness" index. R_a can be computed for daily or hourly periods as a function of latitude, day of the year and hour of the day (see Appendix I). For 24-h data, K_T may range from 0.7 to 0.8, depending on atmospheric clarity (dust, pollution, humidity, etc.) and elevation and sun angle.

A simple prediction for K_T where only site elevation is considered was developed by Allen et al. (1994) by regression of the integrated Beer's radiation extinction function (Monteith and Unsworth 1990) with elevation between sea level and 3,000 m

$$K_T = 0.75 + (2E-5)z \quad (2)$$

where z = station elevation (m). Eq. 2 was developed for an average sun angle of 45° above the horizon.

Improved estimates of R_{so} for periods or days having low sun angles can be made using an expression determined by integrating Beer's law through the atmosphere and by considering the effect of sun angle on total atmospheric path length (Allen et al. 1994)

$$K_T = \exp \left(\frac{-0.0018P}{K_b \sin \phi} \right) \quad (3)$$

where K_b = clarity coefficient, $0.5 < K_b \leq 1.0$, where $K_b = 1.0$ for clean air and $K_b =$ about 0.5 for extremely turbid or

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dusty air. Generally K_b is taken as 1.0 to predict the upper limit of K_T for most agricultural areas. Variable P = mean atmospheric pressure (kPa), and ϕ = angle of the sun above the horizon (rad). The -0.0018 coefficient was found to produce best agreement with estimates by the more theoretical (4)–(7) introduced next. Allen et al. (1994) used a coefficient equal to -0.0021 .

A portion of the absorption or scattering of direct beam radiation in the atmosphere is caused by water vapor. (Direct beam radiation is defined here as the short wave radiation flux density coming directly from the sun's beam, incident to a plane parallel to the earth's surface. Diffuse radiation is defined here as the short wave radiation flux density coming from scattered sunlight.) Transferability of K_T functions can be improved by including precipitable water as a variable in the extinction function for direct beam radiation. A function proposed by Majumdar et al. (1972) and Boes (1981) predicts the clearness index for direct beam radiation, K_b , under clear sky conditions with low turbidity

$$K_b = 0.98 \exp \left[\frac{-0.00146P}{K_b \sin \phi} - 0.162 \left(\frac{W}{\sin \phi} \right)^{0.25} \right] \quad (4)$$

where P = atmospheric pressure (kPa); and W = precipitable water in atmosphere in cm. Eq. 4 was modified here by the addition of K_b . The 0.98 coefficient adjusts for the lower solar constant used by Majumdar et al. (1972). W can be approximated for most conditions using an equation by Garrison and Adler (1990) derived from data from 82 sites in the continental U.S. and Alaska

$$W = 0.014e_d P + 0.21 \quad (5)$$

where e_d = vapor pressure near surface (kPa); and W is in cm. Majumdar et al. (1972) estimated W in India as $W = 1.6e_d$.

A diffuse radiation index, K_D , can be estimated from K_b using the following expressions

$$K_D = 0.35 - 0.33K_b \quad \text{for } K_b \geq 0.15$$

$$K_D = 0.18 + 0.82K_b \quad \text{for } K_b < 0.15 \quad (6)$$

which were derived from a figure presented in Boes (1981), drawn using data from SOLMET (National 1977). Eq. 6 agrees with diffuse and total radiation relationships reported by Ruth and Chant (1976), Orgill and Hollands (1977), and Reindl et al. (1990), and predicts 20 to 30% higher than an earlier function reported by Liu and Jordan (1960).

When (4)–(6) are used the clearness index is calculated as the sum of K_b and K_D

$$K_T = K_b + K_D \quad (7)$$

For hourly or shorter periods, ϕ in (3) and (4) is calculated as

$$\sin \phi = \sin \varphi \sin \delta + \cos \varphi \cos \delta \cos \omega \quad (8)$$

where φ = station latitude (rad); δ = solar declination (rad); and ω = solar time angle (rad) at midpoint of hourly or shorter period (see Appendix I for equations for calculating δ and ω). For application of (3) and (4) to 24-h or monthly data sets, a weighted 24-h average $\sin \phi$ can be estimated by weighting instantaneous $\sin \phi$ according to instantaneous R_a during the course of a day. The integrated and averaged result is

$$(\sin \phi)_{24} = \left\{ \sin^2 \varphi \sin^2 \delta \omega_s + 2 \sin \varphi \sin \delta \cos \varphi \cos \delta \sin \omega_s + \cos^2 \varphi \cos^2 \delta \left[\frac{\omega_s}{2} + \frac{\sin(2\omega_s)}{4} \right] \right\} / (\omega_s \sin \varphi \sin \delta + \cos \varphi \cos \delta \sin \omega_s) \quad (9)$$

where ω_s = sunset hour angle, defined in Appendix I. A simplification of (9) determined by regression is

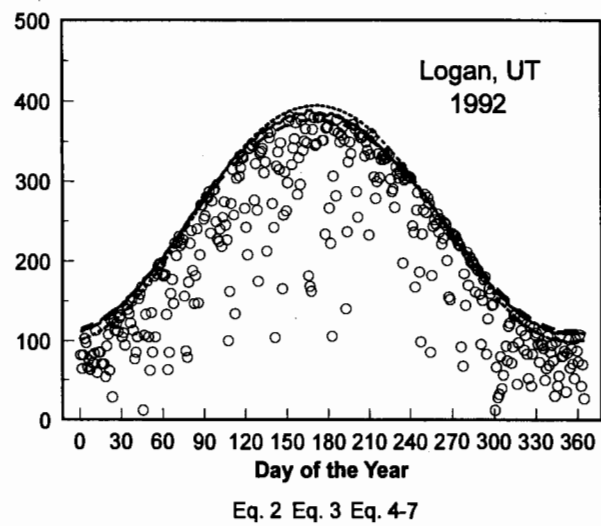
$$(\sin \phi)_{24} = \sin \left[0.85 + .3\varphi \sin \left(\frac{2\pi}{365} J - 1.39 \right) - 0.42\varphi^2 \right] \quad (10)$$

where J = day of year (1 to 365); and φ = latitude in radians. Eq. 10 estimates $(\sin \phi)_{24}$ with a standard error of estimate of about 4% for spring, summer, and fall periods over a range of -65 to 65° latitude and 6% for all months over a range of -60 to 60° latitude. This accuracy is adequate for predicting K_T for 24-h periods.

Example R_{so} Application

Examples of using (1)–(10) in estimating R_{so} envelopes are shown in Fig. 1 for daily (24-h) time steps and in Fig. 2 for 30-min time steps. Fig. 1(a) shows 24-h average R_a measured at Logan, Utah during 1992 and Fig. 1(b) shows 24-h average R_a measured at North Baltimore, Ohio during March–October, 1990. [All data from Ohio were obtained from the Ohio Agricultural Weather Network (courtesy of Dr. J. R. Holman, Ohio Agric. Res. and Dev. Ctr., Ohio State Univ., Wooster, Ohio, 1994).] The elevation of Logan is 1,350 m and the elevation of North Baltimore is 240 m. Mean minimum relative

Ave. 24-hour Solar Radiation, $W m^{-2}$



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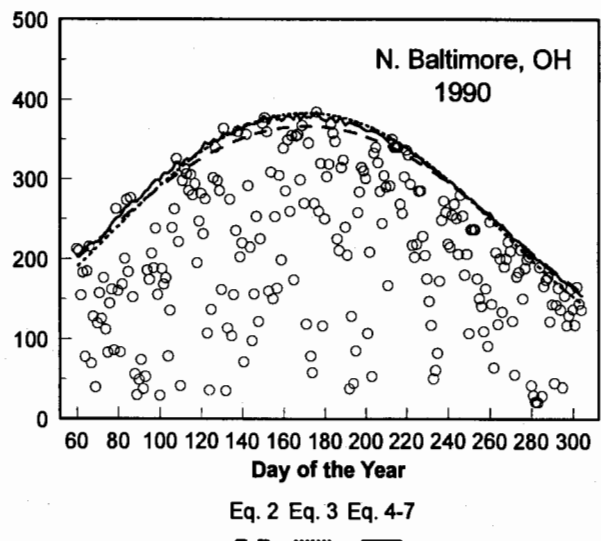


FIG. 1. 24-h Average R_a and Estimated R_{so} Envelopes at (a) Logan, Utah; and (b) North Baltimore, Ohio

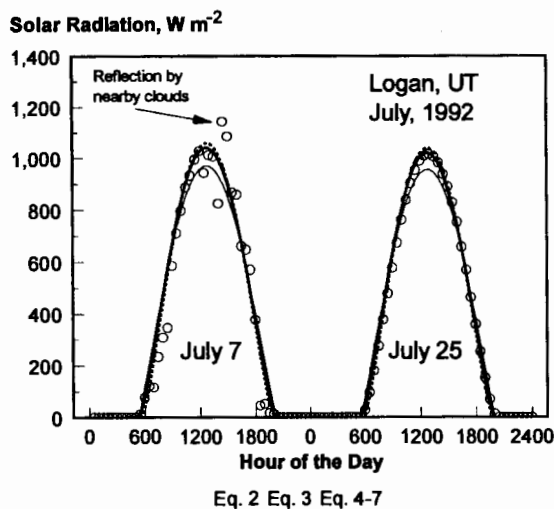


FIG. 2. 30-min Average R_s and Estimated R_{so} Envelopes at Logan, Utah during July 7 and July 25, 1992

humidity at Logan averages about 15% to 30% during summer months, whereas minimum relative humidity at North Baltimore averages about 50%. Latitudes of both locations are about 41° N. The pyranometer at Logan was a LiCor 200S which was field calibrated using an Eppley PSP pyranometer, which was in turn calibrated to the National Renewable Energy Laboratory (NREL) standard (1992a). Eq. 10 was used to calculate $(\sin \phi)_{24}$ for (3) and (4) in Fig. 1, and K_b was set equal to 1.0.

R_s measurements on clear-sky days in Fig. 1(a) and 1(b) in general followed the calculated clear-sky envelopes, especially for K_T calculated from (4)–(7), which predicted well during all months. K_T from (2) fit the clear-sky data closely during winter to early spring and late summer to winter periods, but underpredicted R_{so} during late spring and summer at both locations. K_T from (3) underpredicted measured R_{so} during winter to early spring and late summer to winter periods. The R_{so} envelope computed using (4)–(7) varied irregularly from day to day due to changes in atmospheric vapor content which were estimated using humidity measurements at 2 m using (5). Days with drier air had relatively higher values for K_T . Many of the clear sky days at North Baltimore [Fig. 1(a)] coincided with these higher values in K_T , which is expected, since when the atmosphere was cloud free, it probably contained less total precipitable water. Logan had a higher incidence of clear sky days than did North Baltimore due to the aridity of Utah. Near zero values of average daily R_s at Logan during winter periods were caused by snow deposition on the pyranometer.

Measured R_s and calculated R_{so} are shown in Fig. 2 for 30-min periods during two days at Logan during 1992. Data for July 25 represent a completely clear day. Measured R_s followed the R_{so} curves computed using (3) and using (4)–(7) [with $\sin \phi$ calculated using (8)] very well. The R_{so} envelope computed using (2) was below R_s measurements during peak radiation hours. Data for July 7 are included to show that it is possible for R_s to exceed the predicted R_{so} envelope during partly cloudy conditions when no clouds shade the pyranometer. The increase in measured R_s to above the R_{so} envelope is caused by reflection of direct beam radiation from nearby clouds. This phenomenon can be common in mountainous regions where the sky above a pyranometer located in a mountain valley may be generally clear, but where clouds occur along nearby mountain ridges. The clouds reflect additional short wave radiation toward the pyranometer, increasing R_s readings over what would be experienced on a completely cloud-free day or in more flat terrain.

In summary, it appears that use of (4)–(7), which accounts for the effects of both sun angle (and therefore atmospheric path length) and atmospheric moisture content on R_{so} , predicts better than the more simple (2) and (3) during all times of the year. The equations are valid for predictions of R_{so} for hourly or shorter periods and for 24-h periods. Eq. 2 may be useful for making quick and simple estimates of R_{so} and for estimating the ratio of R_s/R_{so} when calculating net radiation (Wright 1982).

A once-per-day humidity reading can be used to estimate W when investigating hourly data, since e_d and precipitable moisture in the atmosphere do not change substantially during the course of a day. In addition, as is discussed in a later section, when humidity data are lacking, e_d can be approximated as saturation vapor pressure at minimum daily air temperature.

Agreement between R_{so} predicted using (4)–(7) and measured values does not in itself confirm correct calibration of pyranometers, due to uncertainties in atmospheric turbidities and in the prediction algorithms themselves. It is recommended that a long series of R_s measurements be compared to R_{so} estimates so that clear sky days having low atmospheric turbidity ($K_b = 1$) can be found. These instances will generally occur following precipitation events which scrub the air.

Occasionally, biases in sensor calibration occur due to background voltage or ground biases or gradients in microdata-loggers. These voltage biases are sometimes manifested as negative or positive recordings during periods of darkness and should be considered when interpreting and using pyranometer readings.

Monthly Solar Radiation Data

The R_{so} curves cannot be applied with monthly R_s data since all months contain some days having some cloud cover. Therefore, monthly average R_s will always fall below a monthly average R_{so} envelope. For monthly data, then, the user must use general equations which estimate R_s from air temperature data. An example of a simple equation is that presented by Hargreaves and Samani (1982)

$$R_s = K_r(T_{\max} - T_{\min})^{0.5} R_e \quad (11)$$

where $T_{\max}$ and $T_{\min}$ = mean maximum and minimum daily air temperature ($^\circ\text{C}$) for period (generally one month); and K_r =

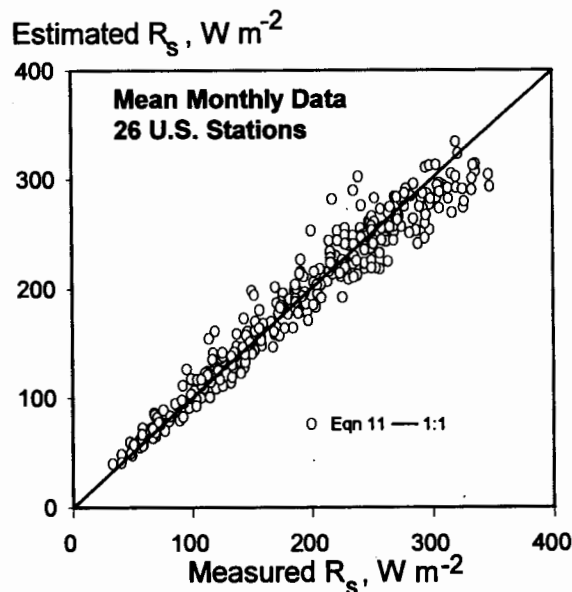


FIG. 3. Monthly Mean Solar Radiation Estimated Using (11) versus Measurements at 26 U.S. Weather Service Stations across U.S.

empirical coefficient. Hargreaves (1994) has recommended using $K_r = 0.16$ for "interior" regions and $K_r = 0.19$ for coastal regions. Allen (1995) suggested using $K_r = 0.17 (P/P_o)^{0.5}$ for interior regions and $K_r = 0.20 (P/P_o)^{0.5}$ for coastal regions to account for elevation effects on the volumetric heat capacity of the atmosphere, where P = mean atmospheric pressure of site, kPa; and P_o = mean atmospheric pressure at sea level (101.3 kPa).

Eq. 11 with the Hargreaves coefficients provides reasonable estimates of R_n for weekly or monthly periods as indicated in Fig. 3, where estimates of R_n are compared with long term monthly mean measurements for 26 weather stations across the U.S. (National Renewable 1992b). Some underestimation occurred at the highest levels of monthly R_n .

NET RADIATION

Comparing Measurements with Computation Equations

Equations for estimating hourly and 24-h average net radiation (R_n) using R_s measurements are accurate under most conditions. Therefore, measured R_n data should always be plotted against R_n estimated using equations based on measured R_s , air temperature and vapor pressure [see Brutsaert (1982); Wright (1982); and Monteith and Unsworth (1990)]. Most of these equations can be applied on monthly, daily and hourly or shorter time steps.

If measured values of R_n chronically deviate from estimated values by more than 3–5%, then the calibration or operation of the R_n device (radiometer) should be scrutinized or the estimate or calculation of surface albedo should be confirmed. Comparing calculated to measured R_n can readily point out days or periods during which the radiometer device has malfunctioned due to effects of dust, bird droppings, moisture condensation inside domes, lack of levelness, or lack of green vegetation beneath the sensor. The value for albedo (α) used in the R_n estimating equation should represent conditions of the surface beneath the radiometer.

The user of net radiometer data must be aware that net radiometers manufactured by different companies may not yield the same measurements of net radiation even when placed over the same surface. Differences may be caused by differences in sensitivities of the various radiometers to long wave and short wave radiation and by variations among methods used for calibrating sensors during manufacturing. Differences in calibrations and responses between net radiometer brands can be substantial.

The type, density and height of vegetation beneath the net radiometer and relative soil moisture content should be monitored and reported with the data. Care should be exercised when positioning the radiometer to avoid shading the sensed vegetation with other instruments or structures and to insure that the radiometer is not shaded by other instruments or structures at any time of the day or year.

Hourly Calculations at Night

Equations by Wright (1982) [also referenced in Jensen et al. (1990) and Allen et al. (1995)] use the ratio of R_n/R_{so} to estimate the cloudiness of the atmosphere and its effect on the emissivity of the atmosphere. When applying these equations to hourly or shorter calculations during nighttime, the ratio of R_n/R_{so} for nighttime periods can be estimated by using the R_s/R_{so} ratio determined a few hours before sunset. This is necessary since R_{so} (and R_s) are by definition zero during nighttime periods.

Example Application

Fig. 4 shows measured and estimated R_n for cattail vegetation near Logan, Utah during 1993. A REBS Q-6 net radi-

Estimated R_n , $W m^{-2}$

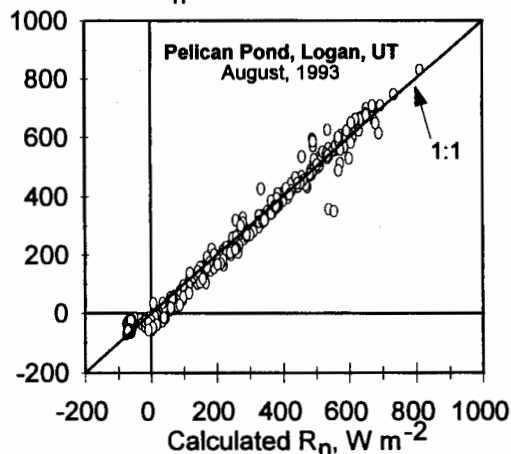


FIG. 4. Measured and Estimated R_n during 20 min Periods over Mature Cattail Vegetation near Logan, Utah during August, 1993 from Allen et al. (1994)

ometer with factory calibration was employed. R_n and albedo equations from Wright (1982) were used. The measurement and calculation time step was 20 min. The agreement between measurements and equation estimates was good, with some undermeasurement by the sensor during low radiation and nighttime conditions. A few outliers caused by measurement errors in R_s or R_n can be observed. Perfect agreement between the R_n measurements and R_n equations should not be expected, due to limitations of assumptions used in the equations (e.g., the value for albedo, means for estimating the net long wave radiation component, etc.).

SOIL HEAT FLUX

A relationship proposed by Choudhury et al. (1987) and Choudhury (1989) for predicting soil heat flux density (G) under daylight conditions is:

$$G = 0.4 e^{-0.5LAI} R_n \quad (12)$$

where LAI = leaf area index; e = natural number; and G has the same units as R_n . These estimates can be used to compare with measured G . Eq. 12 predicts $G = 0.1 R_n$ for $LAI = 2.8$, which is typical for clipped grass and is in agreement with findings by Clothier et al. (1986) and Allen et al. (1994). Soil heat flux under forage grass during nighttime hours was found by Allen et al. (1994) to be about 0.4 to 0.5 R_n . Pruitt (1995, personal communication) observed $G = 0.3 R_n$ during nighttime hours under clipped grass at Davis, Calif.

The G versus R_n relationships can be used to check the functioning and relative accuracy of soil heat flux plates after correcting measurements for the temperature change of soil above the plates. These relationships do not hold for 24-h data, as a positive 24-h soil heat flux estimate would always result. Eq. 12 is only approximate and does not consider effects of plant spacing, sun angle, soil color, soil moisture, or soil texture, nor the sensible heat balance at the surface on the ratio of G to R_n .

WIND

Accuracy of wind measurements is difficult to assess unless duplicate instruments are used. One should always scan wind records for the presence of consistently low wind recordings. For electronic instruments, these recordings may represent a numerical "offset" in the anemometer calibration equation which is intended to represent a constant turning friction and

initial turning threshold of the instrument. The consistent recording of these offsets indicates either the presence of exceptionally calm conditions [wind speeds less than about 0.5 m s^{-1} during the entire sampling period (which is rare)] or a malfunctioning of the wind speed sensor due to electrical shorting or fatigue of bearings. These problems may not be noticed by the station operator.

When paired checks of anemometers are not available, one may use a mass balance type of approach where the cumulative wind run for the anemometer is plotted against cumulative wind run for a nearby station. The trend in slope between the two accumulations should not change over time.

DC-Generators

When DC-generator-type anemometers have been used, the data user should confirm that large capacitors (for example, $1000 \mu\text{F}$) were installed between the anemometer leads or that 30 or more high frequency samples were averaged each sampling interval in order to dampen or average natural voltage spikes occurring in the anemometer signal. This is necessary, since voltage spikes from generator-type anemometers can often be of longer duration than the integration time of modern data loggers.

RELATIVE HUMIDITY AND VAPOR PRESSURE

Vapor pressure of air is difficult to measure accurately. Electronic relative humidity sensors are commonly plagued by hysteresis, nonlinearity and calibration errors. Some of these errors are inherent in the sensor design. Others result from dust, moisture, insects, pollution, and age.

Replication of RH Instruments

It is strongly recommended that duplicate relative humidity (RH) and air temperature sensors be employed in electronic weather stations. This practice is costly, but can save expense in troubleshooting time and worry concerning integrity of RH data, and is a relatively small cost when compared to the value of weather data used in irrigation scheduling and design, in water resources planning and modeling, and in global warming studies. When duplicate RH (and air temperature) sensors yield similar signals, then it is likely that both are functioning properly, provided proper calibration equations have been employed. Significant differences between sensors should flag some type of problem with one or both of the sensors.

Trends in Computed Dew-Point Temperature with Time

When relative humidity is measured, vapor pressure of the air (e) is calculated as

$$e = \frac{RH}{100} e^{\circ}(T) \quad (13)$$

where $e^{\circ}(T)$ = saturation vapor pressure at air temperature T ; and RH is in %. RH and T must be taken for the same time period.

Hourly (or shorter) measurements of RH , dew point temperature (T_d) or vapor pressure (e_d) can be preliminarily assessed by plotting hourly measurements of computed T_d or vapor pressure with time. Relative humidity will vary significantly with time of day, inversely with air temperature. However, both T_d and vapor pressure, either measured directly, or computed using RH and T measurements, should remain relatively constant throughout a 24-h period when the air mass is stable and advection of dry air from outside the area does not occur. During these stable periods, one should expect a

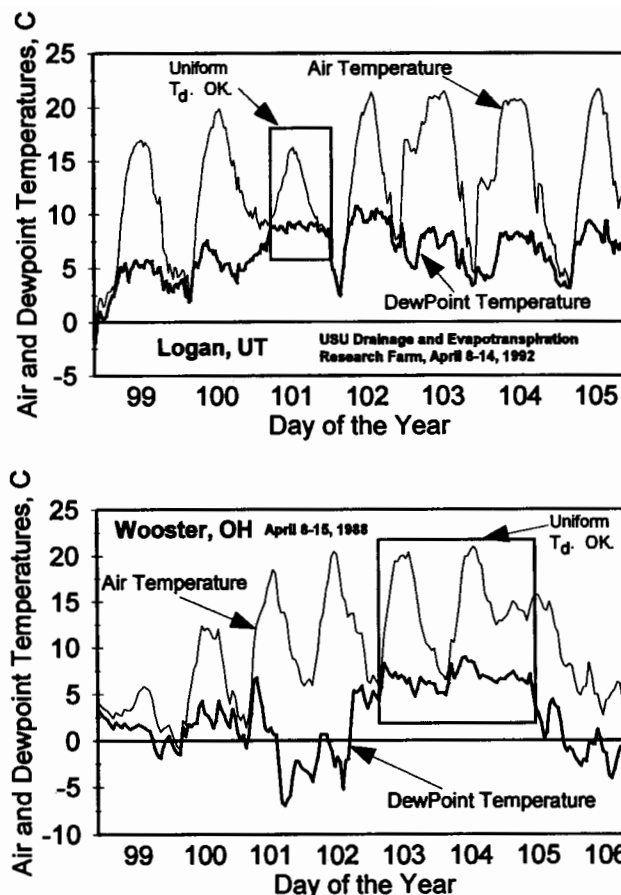


FIG. 5. T_d Computed from RH Measurements for April 8–15 Periods at: (a) Logan, Utah during 1992 (30-min Data); and (b) Wooster, Ohio during 1988 (hourly data)

small rise in T_d and vapor pressure during daytime periods, when ET fluxes humidify the equilibrium boundary layer. However, this increase is usually less than about 10 to 20%. Variation in T_d may increase substantially when a weather front passes overhead. Since vapor pressure is calculated as the product of RH and saturation vapor pressure at air temperature, any error in the RH calibration tends to cause false variation in T_d with changing air temperature. Visual observation of these “false” trends in T_d can flag a problem in RH readings.

Example

Fig. 5 shows T_d computed from measurements of RH and air temperature at weather stations in Logan, Utah (30-min data) and Wooster, Ohio (hourly data). T_d generally varied from hour to hour for both locations due to air mass instability or possibly due to advection of heat or dry air from adjacent, nonplanted fields during this early spring period (April 8–15). However, the data sequences at each station included periods of relatively constant calculated T_d during the complete course of a day, even though air temperature varied substantially [see the “uniform T_d ” boxed areas in Fig. 5(a,b)]. This is a good indication that the RH sensors were probably functioning correctly during the entire data sequence, and that the calibrations were probably valid.

Observations during Periods of Dew and Rainfall

In many climates, especially those where nighttime dew occurs, T_d during early morning hours coincides closely with recorded minimum daily air temperature, and RH approaches 100%. If no dew occurs, then minimum daily air temperature

may remain a few degrees above recorded T_d . The coincidence of T_d and air temperature during early morning is apparent for nearly all days at both stations shown in Fig. 5.

For automatic recording weather stations where tipping bucket or other recording rain gauges are used, one should expect RH recordings during periods of rain or light drizzle to exceed 95%. Relative humidity recordings which exceed 100% by more than 3–5% during early morning hours or during precipitation events indicate a need for recalibration and numerical adjustment of collected data.

Maximum Daily Relative Humidity

When humidity data are measured in a reference setting (over well-watered clipped grass which is actively transpiring), early morning RH will often approach 100%, even in semiarid areas if measurements are taken inside an irrigated region. As discussed in the following section on station aridity impacts, values of maximum relative humidity (RH_{max}) which consistently fall below 80 to 90% indicate problems in RH sensor calibration or functioning or indicate aridity of the measurement site and deviation from reference conditions.

Example

An example showing the impacts of sensor calibration and functioning and site aridity is shown in Fig. 6, where RH_{max} data measured at the USU Drainage and Evapotranspiration Research Farm at Logan, Utah [Fig. 6(a)] represent measurements from a semiarid region, but where the weather station is in the center of an irrigated area 15 km wide and 50 km long. Values for RH_{max} in Fig. 6(a) consistently remained above 80% even during the very dry, arid months of July and August (a period of zero precipitation), since the region was irrigated. Irrigation and the presence of transpiring vegetation draws the air temperature down to near the dew point each night (the reader is referred to the section on aridity impacts for a discussion on the physics of this process).

RH_{max} is shown in Fig. 6(b) for Grouse Creek, Utah, which is 160 km west of Logan, but is in a narrow, irrigated valley (<2 km wide) surrounded by arid rangeland. At this location, RH_{max} is commonly below 40% due to advection of dry, warm air from surrounding dry rangeland where the evaporative cooling of the air to the dewpoint is inhibited. RH_{max} near 100% occurs occasionally at Grouse Creek during the summer months on days when air movement is low, so that cool, humidified air from adjacent irrigated fields remains near the sensor. The consistent daily readings of minimum RH (RH_{min}) near 10% indicate that the RH sensor is operating consistently over time and is probably not malfunctioning. The calibration may or may not be valid. The data in Fig. 6(b) should not be used in reference ET calculations since the environment is not representative of a reference setting, and reference ET would be overestimated.

Fig. 6(c) shows daily measurements of RH_{max} from an electronic agricultural weather station located near North Baltimore, Ohio over a five year period. One would expect RH_{max} to approach 100% in this semihumid setting. However, one can see clear evidence that the RH sensor was undermeasuring RH_{max} during several years, with decreasing trends in measured RH_{max} during these years. The values measured for RH_{max} were consistent from one day to the next within each year, creating the trend lines. This indicates that the RH sensor was functioning properly, electronically, except during the first half of 1988. However the calibration of the sensor element was too low for 1988, 1990, and 1992, and drifted (fell) with time. The RH sensor element was replaced each fall (J. R. Holman, 1994, personal communication). RH data for 1990 and 1992 could be corrected by multiplying the RH measurements by a

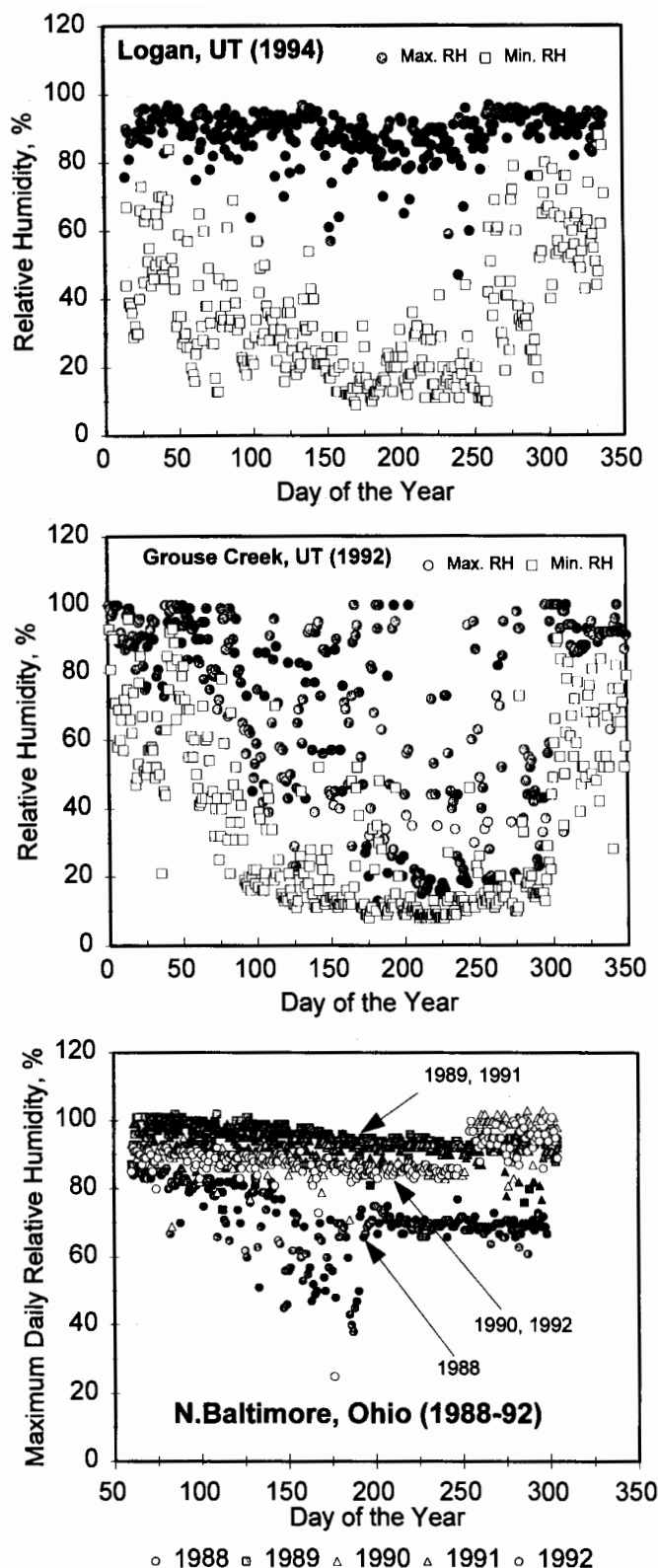


FIG. 6. Daily Values of Measured RH_{max} at (a) Logan, Utah during 1994; (b) Grouse Creek, Utah during 1992; and (c) North Baltimore, Ohio during 1988–92

correction factor or by adding an offset if it appears that the errors were due to an offset bias rather than a multiplier shift.

The type of plotting and screening demonstrated in Figs. 5 and 6 shows the simple types of integrity assessments that can be used in near-real-time or with historical data. These types of assessments should be applied to all weather data used in evapotranspiration estimation and should be adopted by operators of agricultural weather networks.

Computing 24-h Average Vapor Pressure

Often, ET_0 is calculated using 24-h time steps. With modern electronic weather stations, 24-h average vapor pressure can be calculated by averaging hourly values. However, if errors are present in the RH readings, then the user may benefit by selectively using portions of the data stream that are perceived to have higher levels of accuracy.

In general, smaller errors in vapor pressure computations occur during periods of high RH as compared to periods of low RH . At high RH , a 5% absolute error in RH will cause a smaller relative percentage error in total RH and therefore in calculated vapor pressure (13) than would a 5% absolute error in RH at low RH . In addition, during afternoon periods of high temperature when RH is lowest, the saturation vapor pressure at ambient temperature is at a maximum value at the same time that RH is at a minimum. Therefore, an error in daily minimum RH is further compounded by the relatively larger magnitude of the $e^o(T)$ multiplier.

Consequently, when integrity of RH data is in doubt, one may consider using RH during nighttime or early morning hours to represent the average daily vapor pressure of the air for the 24-h period for use in Penman-type evapotranspiration equations, rather than using RH measurements during high temperature periods of the day or rather than using an average vapor pressure based on computations of hourly RH measurements, due to greater errors with afternoon readings. Early morning readings were utilized by Wright (1982) as a more-or-less standard input to the Kimberly Penman equation. Comparison of vapor pressures independently calculated using alternate methods (RH_{max} and T_{min} , RH_{min} and T_{max} , or an average of hourly) can indicate the general quality of the RH data and perhaps the linearity of the RH instrument if vapor pressure is known to remain relatively constant throughout the day. Use of mean daily RH and mean daily air temperature to estimate 24-h average vapor pressure of the air is not recommended due to the nonlinearity of the vapor pressure-air temperature curve.

ARIDITY IMPACTS ON TEMPERATURE AND VAPOR PRESSURE

In arid and semiarid climates, T and RH measurements may misrepresent reference evapotranspiration environments if a fully vegetated, fully transpiring ground cover is not present. Evapotranspiration humidifies and cools the equilibrium boundary layer. The absence of ET therefore results in a drier and warmer equilibrium boundary layer and subsequent RH and T measurements.

Allen et al. (1983) and Allen and Pruitt (1986) developed empirical procedures for adjusting air temperature measurements from weather stations located in environments where ET was less than maximum rates due to limited soil moisture or lack of green vegetation. The objective of the adjustments was to create air temperature data sets reflective of well-watered environments over clipped grass for use in ET_0 equations. An aridity rating procedure was used to characterize effects of dryness of the local station environment as well as the region on required air temperature adjustments. In the application to weather stations in Idaho, daily maximum and minimum air temperatures were adjusted downward by as much as 4.5°C for stations having nonirrigated, nontranspiring surfaces surrounded by similar conditions. The magnitudes of adjustments by Allen and Pruitt were based on paired observations of air temperatures in southern Idaho (Allen et al. 1983).

Allen and Gichuki (1989) and Ley and Allen (1994) applied more complex methods for adjusting air temperature and humidity data from nonreference stations using energy balance and water balance relationships. These methods required mea-

surement of R_s , T , humidity, wind, and precipitation on a daily basis.

Background to Simplified Aridity Adjustment Approach

One simple way to assess the impact of "aridity" of a weather station where both air temperature and humidity are measured is to compare daily minimum air temperature, T_{min} , with dewpoint temperature, T_d . In an environment having healthy vegetation and adequate soil moisture, T_{min} usually approaches T_d , especially if the wind dies down by early morning. This is true even in arid and semiarid regions if the weather sensor environments are irrigated. The reasons for this follow.

The temperature (T) of the lower boundary layer above a surface decreases during nighttime hours due to surface cooling during the night and aerodynamically induced contact between the boundary layer and the surface. Surface cooling in a "reference" environment is caused by two primary processes: net outward long-wave radiative emission from the surface and evaporative cooling by vegetation and soil. Evaporation occurs whenever moisture is available at the soil or plant surface and the boundary layer has a positive vapor pressure deficit (VPD). The two processes of net long wave emission and evaporative cooling continue to reduce T during the course of the nighttime period. Under well-watered conditions where evaporable water is present, T will often decrease until it reaches T_d , provided that wind speed is relatively calm so that large amounts of warm, dry air (in an arid region) are not transported to the surface.

Near-surface T is prevented from decreasing below T_d due to condensation of vapor from the air and the corresponding latent heat released. Under these conditions, the relationship $T_{min} \approx T_d$ generally holds and can be used to assess the relative impact of station environment on humidity and air temperature and their representativeness of a "reference" environment for ET_0 calculations. When the environment is not actively evaporating due to low soil moisture or lack of vegetation, T_{min} may remain well above T_d .

All other conditions being equal, T_{min} is more likely to approach T_d during winter, spring, and fall months than during summer months, since the length of the night is a factor in that the longer the nighttime hours, the longer the period during which radiative cooling of the surface can lower T to T_d . Because it is possible for T_{min} to approach T_d even when daytime $ET < ET_0$ if the nighttime is long enough for cooling to occur, one can not always conclude that the occurrence of $T_{min} \approx T_d$ indicates that reference conditions have occurred. However, this will usually be the case.

Under most evaporative conditions, T_d is higher during daytime hours than during nighttime hours due to humidification of the boundary layer by daytime ET . Consequently, daytime measurements of absolute humidity may exceed nighttime values by 5 to 20%. Therefore, the relationship between T_{min} and T_d may not be strictly one-to-one when T_d is based on 24-h or daylight averages. In these cases, T_{min} measurements can be expected to go a few degrees below 24-h average T_d measurements.

Example

An example of the effect of dryness of weather station environment on proximity of T_{min} to T_d is shown in Fig. 7 where mean monthly differences between T_{min} and T_d are plotted against ratios of mean monthly precipitation and ET_0 for three data sets [(a) 62 weather stations in Sudan, Africa; (b) 26 Weather Service locations across the U.S.A.; and (c) two stations in southern Idaho]. The ratio of P/ET_0 provides an in-

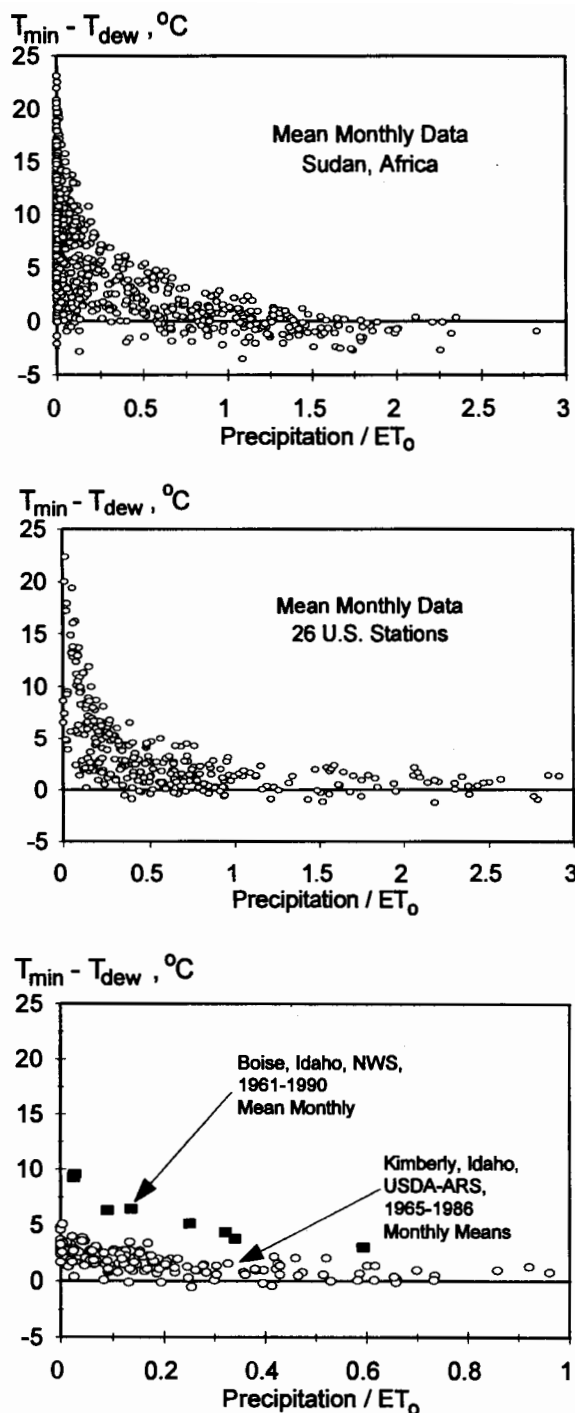


FIG. 7. Monthly Mean T_{min} - Monthly Mean T_d for: (a) Weather Stations in Sudan, Africa; (b) 26 U.S. Weather Service Stations across U.S.; and (c) Two Weather Stations in Southern Idaho

dication of the relative wetness of the station during a particular month and the likelihood of healthy, transpiring vegetation. Data for Sudan were taken from Smith (1993).

The increases in Fig. 7 for the differences $T_{min} - T_d$ at low values of P/ET_0 are caused by the effects of reduction in regional and possibly local ET on evaporative cooling. The similarity between Fig. 7(a) and 7(b) is surprising, given the general aridity of Sudan relative to most of the U.S. The similarity suggests that aridity effects on T_{min} and T_d may be similar over a wide range of climates and regions. The occurrence of $T_{min} - T_d \approx 0$ when $P/ET_0 \approx 0$ for some stations may indicate the effect of irrigation or carryover of soil moisture from a previous month.

Figs. 7(a) and (b) indicate that T_{min} approaches T_d when vegetation of the weather sites is in a reference (well-watered)

condition (e.g., monthly $P/ET_0 > 1$). This would also include conditions within most large irrigation projects. The average absolute deviation between T_{min} and T_d when $P/ET_0 > 1$ is about 2°C. When measured in a dry environment (monthly $P/ET_0 \approx 0$), T_d can be as much as 20°C less than T_{min} . Daily differences between T_{min} and T_d were similar to the monthly averages shown.

Fig. 7(c) demonstrates the effect of irrigation on $T_{min} - T_d$, where the two locations Boise and Kimberly, Idaho have similar regional climates and elevations, but where the Kimberly weather station is surrounded by about 150,000 ha of irrigation and the Boise station is located at the Boise Airport which is surrounded by about 5 to 10 km of nonirrigated rangeland with irrigated farmland beyond that. The effect of surrounding the weather station with well-watered vegetation is obvious. This "paired" comparison supports the need to "adjust" T_{min} and T_d to represent reference environments if ET_0 calculations are to represent irrigated or fully rainfed agricultural environments. Values for $T_{min} - T_d$ would increase for both of these locations if they were surrounded by large expanses of rangeland, only.

Adjustment Procedures

When weather data are measured in a nonreference environment, the relationship $T_d \approx T_{min}$ can be employed to assess the T and RH data and to potentially adjust these data for use in ET_0 calculations. "Adjusted," "reference" values for T_{max} , T_{min} and T_d can be approximated using the $T_d \approx T_{min}$ relationship as

$$T_{min_o} = T_{min} - K_{ar}\Delta T \quad (14)$$

$$T_{max_o} = T_{max} - K_{ar}\Delta T \quad (15)$$

$$T_{d_o} = T_d + (1 - K_{ar})\Delta T \quad (16)$$

for $\Delta T > 0$, where

$$\Delta T = T_{min} - T_d - \Delta T_{climate} \quad (17)$$

and where the "o" subscript is used to indicate data values that have been adjusted to represent the reference environment. K_{ar} is an empirical aridity proportionment coefficient equal to about 0.5, based on analysis of changes in T_{min} , T_{max} and T_d with ratios of P/ET_0 for weather stations in the U.S., Europe, Africa, and Asia. The assumption in (14) and (15) is that T_{max} is affected by station aridity in a similar magnitude as is T_{min} . $\Delta T_{climate}$ represents the $T_{min} - T_d$ "window" below which no adjustments should be made. Because T_{min} may often vary $\pm 3^\circ\text{C}$ of T_d under reference conditions due to sensor errors and due to changes in T_d during the day, it is recommended that $\Delta T_{climate}$ be set equal to 2 to 3°C, especially for arid and semiarid climates. For semihumid to humid climates, $\Delta T_{climate}$ should be set to 0°C.

Preliminary analyses indicate that adjustments by (14)–(17) can be applied to both monthly and daily weather data. Use of the proposed modifications to T_{min} and T_d have reduced reference ET estimates using the Penman-Monteith and Hargreaves equations by up to 20% when applied to data sets from arid environments in Africa and in the Western U.S.

"DOUBLE MASS ANALYSIS" TECHNIQUES

All meteorological recordings can be periodically compared to recordings from nearby stations to assess whether shifts in equipment functioning or calibration have occurred over time. One technique which is used for this purpose is the double mass analysis technique where cumulative sums of a parameter from two locations are plotted against one another (for example, cumulative wind at site 1 versus cumulative wind at

site 2). A shift in the slope of the cumulative curve over time can indicate a change in functioning or calibration of one of the instruments. This technique is commonly used in hydrology to assess effects of changes in location or operation of precipitation gauges or stream flow gauges, or to estimate missing values. The double mass technique was used by Dutcher and Hubbard (1994) to conform a temperature bias induced following a change in instrumentation and station relocation.

The required proximity of two stations to be compared is dependent on the meteorological variable. Solar radiation and net radiation over similar vegetation should correlate well over distances of hundreds of km in nonmountainous areas. Relative humidity (or computed vapor pressure) and air temperature should correlate well over distances up to 100 km, provided that changes in topography or weather are not abrupt and that general vegetation and aridity of the two locations are similar. Wind speeds are the least likely to correlate over large distances due to local site effects which may change with wind direction and due to differences in local weather and frontal patterns and associated time lags. However, on a weekly or monthly basis, correlations may be strong.

SUMMARY AND CONCLUSIONS

Procedures have been introduced for assessing integrity and reasonableness of weather data for use in reference evapotranspiration equations. The procedures include using clear-sky envelopes for assessing calibration and functioning of solar radiation sensors and using net radiation estimation equations to assess net radiation measurements. Guidelines have been presented for assessing quality of relative humidity data by viewing daily maximum RH values and expected trends and relationships between air vapor content and air temperature.

When weather measurements have been affected by the dryness of the weather sensing environment, procedures for adjusting air temperature and air vapor content data by relying on the T_{min}/T_d relationship are employed. All of the recommended guidelines are simple and straightforward to use and serve as preliminary "filters" by which to scrutinize weather measurements.

APPENDIX I.

The following equations for calculating extraterrestrial radiation are based on Duffie and Beckman (1991). All trigonometric functions and angles are expressed in radians with the exception of L_z and L_m .

For 24-hour periods

$$R_a = \frac{G_{sc}}{\pi} d_r (\omega_s \sin \varphi \sin \delta + \cos \varphi \cos \delta \sin \omega_s) \quad (18)$$

where R_a = daily average extraterrestrial radiation ($W m^{-2}$); G_{sc} = solar constant ($1367 W m^{-2}$); d_r = relative distance factor from Earth to Sun; δ = solar declination (rad); φ = latitude (rad) (φ is negative for southern hemisphere); and ω_s = sunset hour angle (rad).

$$\omega_s = \arccos(-\tan \varphi \tan \delta) \quad (19a)$$

or as

$$\omega_s = \frac{\pi}{2} - \arctan \left[\frac{(-\tan \varphi \tan \delta)}{(1 - \tan^2 \varphi \tan^2 \delta)^{0.5}} \right] \quad (19b)$$

$$d_r = 1 + 0.033 \cos \left(\frac{2\pi}{365} J \right) \quad (20)$$

$$\delta = 0.409 \sin \left(\frac{2\pi}{365} J - 1.39 \right) \quad (21)$$

where J = number of day in year. For daily values, J can be determined by

$$J = \text{integer} \left(275 \frac{M}{9} - 30 + D \right) - 2 \quad (22)$$

provided that the following corrections are made: if $M < 3$, then $J = J + 2$, and during leap years when $M > 2$, then additionally $J = J + 1$, where M = month number (1–12); and D = day in month (Craig 1984).

For winter months in latitudes greater than 55° or less than -55° , the 24-h equations for R_a have a limited validity due to the shortness of daylength. In these situations, (19b) should be used, and the $(1 - \tan^2 \varphi \tan^2 \delta)$ expression, when it becomes negative, should be set equal to a very small number (i.e., 0.0001). This produces the product $\omega_s = 0$. Reference should be made to the Smithsonian tables (List 1984) to assess possible deviations.

For hourly or shorter periods

$$R_{ah} = \frac{12}{\Delta t \pi} G_{sc} d_r$$

$$[(\omega_2 - \omega_1) \sin \varphi \sin \delta + \cos \varphi \cos \delta (\sin \omega_2 - \sin \omega_1)] \quad (23)$$

where R_{ah} = extraterrestrial radiation during period ($W m^{-2}$); ω_1 = solar time angle at beginning of period (rad); and ω_2 = solar time angle at end of period (rad).

$$\omega_1 = \omega - \frac{\pi \Delta t}{24} \quad (24)$$

$$\omega_2 = \omega + \frac{\pi \Delta t}{24} \quad (25)$$

where ω = solar time angle at midpoint of period (rad); and Δt = length of calculation period (h), for example, $\Delta t = 0.5$ h for a 30-min period.

$$\omega = \frac{\pi}{12} \{ [t + 0.06667(L_z - L_m) + S_c] - 12 \} \quad (26)$$

where t = standard clock time at midpoint of period (h) (for example, for a period between 1400 and 1500 h, $t = 14.5$); L_z = longitude of center of local time zone (in degrees west of Greenwich, England); L_m = longitude of measurement site (in degrees west of Greenwich, England) and S_c = seasonal correction for solar time (h). Note that L_z and L_m are both in degrees, rather than radians. For example, $L_z = 75, 90, 105$, and 120° for Eastern, Central, Rocky Mountain, and Pacific time zones, respectively, in the U.S.A. $L_z = 0^\circ$ for Greenwich, England, 330° for Cairo, Egypt, and 255° for Bangkok, Thailand. From Duffie and Beckman (1980)

$$S_c = 0.1645 \sin(2b) - 0.1255 \cos(b) - 0.025 \sin(b) \quad (27)$$

$$b = \frac{2\pi(J - 81)}{364} \quad (28)$$

Eq. (27) is a simplification of the equation for S_c given in Duffie and Beckman (1991), but produces equivalent results.

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