An Improved Parameterization for Estimating Effective Atmospheric Emissivity for Use in Calculating Daytime Downwelling Longwave Radiation

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ABSTRACT

An improved parameterization is presented for estimating effective atmospheric emissivity for use in calculating downwelling longwave radiation based on temperature, humidity, pressure, and solar radiation observations. The first improvement is the incorporation of an annual sinusoidal variation in effective clear-sky atmospheric emissivity, based on typical climatological variations in near-surface vapor pressure. The second is the continuous estimation of fractional cloudiness by taking the ratio of observed solar radiation to a modeled clear-sky solar radiation. Previous methods employed observer-estimated fractional cloudiness. Data from the Atmospheric Radiation Measurement (ARM) program were used to develop these improvements. The estimation of cloudiness was then used to modify the effective clear-sky atmospheric emissivity in order to calculate 30-min averages of downwelling longwave radiation. Monthly mean bias errors (mbe's) of -9 to $+4$ W m^{-2} and root-mean-square errors (rmse's) of 11–22 W m^{-2} were calculated based on ARM data over a 1-yr period. These mbe's were smaller overall than any of the six previous methods tested, while the rmse's were similar to the best previous methods. The improved parameterization was then tested on FIFE data from the summer of 1987. Although the monthly mbe's were larger, the rmse's were smaller.

It is also shown that data from upper-air soundings can be used to calculate the effective atmospheric emissivity rather than specifying the aforementioned sinusoidal variation. Using ARM upper-air soundings, this method resulted in larger mbe's, -7 to $+11$ W m^{-2}, especially during the summer months, and similar rmse's. The success of the method suggests that it has application at any observing site within reasonable proximity of an upper-air sounding, while removing the empiricism used to specify the annual sinusoidal variation in emissivity.

1. Introduction

Attempts to investigate the energy balance at the earth's surface are often hindered by a significant uncertainty in the estimated magnitude of downwelling longwave radiation received at the surface (LW$_d$). Accurate estimates of LW$_d$ are vitally important in determining the radiation budget, which, in turn, modulates the magnitude of the terms in the surface energy budget (e.g., evaporation). Many reasonably successful techniques have been developed in recent decades that estimate LW$_d$ based on surface observations alone. These methods have had varying degrees of success, and new techniques continue to be developed. The main problem seems to be versatility; that is, the methods are developed empirically at one location and with one set of instruments. Often, those developing new methods later show that these methods do not fare well in other locations. In addition, previous methods were developed for daily or longer-term averages and therefore are usually much less accurate at shorter time intervals. Finally, practically all techniques are valid in clear skies only, greatly limiting their utility. We provide a formula for estimating atmospheric emissivity that can be used to calculate LW$_d$ under any daytime sky condition.

Brunt (1932), based on a perceived similarity between heat conduction and radiative transfer, theorized that LW$_d$ was related to the square root of near-surface vapor pressure, *e*. Using monthly averages of LW$_d$ and *e*, he developed the first empirical relationship between the two quantities. Use of this technique resulted in a correlation coefficient of 0.97. Three decades later, Swinbank (1963) argued that LW$_d$ was not related to *e* at all, but to the square of the temperature *T* alone and that the Brunt (1932) formula worked only due to the positive correlation between *e* and *T*. Swinbank's new formula resulted in a correlation coefficient of 0.99 between observed and estimated LW$_d$ and a root-mean-square error (rmse) of less than 5 W m^{-2}.

Idso and Jackson (1969) developed an equation, also dependent on T^2, which was tested against a much wider range of temperatures than the previous formulations. This formula was touted as being ''valid at any latitude and for any air temperature reached on earth,'' and had a correlation coefficient of 0.99 between measured and

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estimated LW_d. Staley and Jurica (1972) integrated emissivity over the entire atmosphere using vertical profiles of vapor pressure, carbon dioxide, and ozone and related it to *e*. This formula was much more theoretical than previous efforts and was successfully used by Deardorff (1978) in the development of a surface energy budget model. Finally, Satterlund (1979) found that previous formulations did not perform well in temperatures below 0°C and developed a new formulation that claimed to be more accurate at extreme temperatures and of similar accuracy at moderate temperatures.

Brutsaert (1975, hereafter B75) was the first to develop a more physically rigorous parameterization of atmospheric emissivity. This parameterization is based on Schwarzschild's equation (Liou 1980, 22) and assumptions of standard atmospheric lapse rates of temperature and vapor pressure. Culf and Gash (1993) concluded that this method was superior to previous formulations since it is easily adjusted for locally measured lapse rates. Like so many of the other techniques, though, it was developed for clear skies alone. Deardorff (1978) developed a simple correction for cloudiness and applied it to the Staley and Jurica (1972) clear-sky parameterization for emissivity. When observations of fractional cloud cover are not available, however, the correction for cloudiness cannot be determined. Here, we have developed an improved method that incorporates the B75 clear-sky parameterization and the Deardorff (1978) cloudiness correction. A clear-sky model was used along with the observed magnitude of solar radiation to provide a proxy for fractional cloudiness. We have also added an annual sinusoidal modification to the clear-sky Brutsaert parameterization coefficient, which is representative of typical variations in vertical atmospheric profiles of water vapor pressure. We found that this new emissivity parameterization, when used to calculate LW_d, performs better than previous methods and performs well at any time of the year and in any sky condition, although its usefulness is restricted to daylight hours.

It also will be shown that atmospheric sounding data can be used, along with the derivations found in B75, to estimate the clear-sky coefficient directly. This method is even less empirical and shows great promise for use in estimating LW_d at any observing station collocated with an upper-air sounding system or any location at which vertical moisture profiles can be reasonably interpolated from nearby soundings.

2. Previous formulations

The amount of downwelling longwave radiation is determined by the bulk emissivity $\varepsilon_{\mathrm{atm}}$ and effective temperature T_{atm} of the overlying atmosphere according to

$$LW_d = \varepsilon_{\mathrm{atm}} \sigma T_{\mathrm{atm}}^4, \tag{1}$$

where σ is the Stefan–Boltzmann constant. Since it is difficult to specify $\varepsilon_{\mathrm{atm}}$ or T_{atm} for a vertical column of atmosphere, methods have been developed to parameterize LW_d from the measured temperature and/or vapor pressure near the surface during clear skies such that

$$LW_d = \varepsilon_c(T, e)\sigma T^4, \tag{2}$$

where ε_c is the effective clear-sky atmospheric emissivity, and T and e are the near-surface temperature and vapor pressure, respectively. In the past, empirical formulations for ε_c were developed based on least squares regression of observed LW_d during periods of clear skies.

Since the presence of clouds significantly increases the total effective emissivity ε of the sky, modifications must be made to the existing clear-sky formulations. Deardorff (1978) used a fairly simple cloud modification, which involves introducing a cloud fraction term (clf). In our study, clf was defined by

$$\mathrm{clf} = 1 - s, \tag{3}$$

in which s is the ratio of the measured solar irradiance to the clear-sky irradiance.

The clear-sky shortwave irradiance I at the ground was calculated using a previously developed model based on the results of Paltridge and Platt (1976) and Meyers and Dale (1983). This quantity was approximated by

$$I = I_o (\cos Z) T_R T_{\mathrm{pg}} T_w T_a, \tag{4}$$

where I_o is the effective solar constant, Z is the solar zenith angle, and T_i the transmission coefficients for Rayleigh scattering R, absorption by permanent gases pg and water vapor w, and absorption and scattering by aerosols a.

The effective solar constant (in W m^{-2}) is given by

$$I_o = 1370(\bar{r}/r)^2, \tag{5}$$

where $\bar{r}$ and r are the average and daily distances between the sun and the earth, respectively. The cosine of the solar zenith angle is represented by

$$\cos Z = \sin\gamma \sin\delta + \cos\gamma \cos\delta \cos H, \tag{6}$$

where γ is the latitude of the station, δ is the solar declination, and H is the hour angle. The hour angle is

$$H = (\pi/12)(t_{\mathrm{noon}} - t), \tag{7}$$

where t_{noon} is local solar noon ($\sim$12.5 in Oklahoma) and t is the local solar time (e.g., $t = 12.5$ and $H = 0$ at local solar noon). The empirical expression for the product of the first two transmission coefficients is (Atwater and Brown 1974)

$$T_R T_{\mathrm{pg}} = 1.021 - 0.084[m(0.000949p + 0.051)]^{1/2}, \tag{8}$$

where p is the pressure in millibars and m is the optical air mass at $p = 1013$ mb given by $m = 35 \cos Z(1224 \cos^2 Z + 1)^{-1/2}$. The third coefficient is (McDonald 1960)

$$T_w = 1 - 0.077(um)^{0.3}, \tag{9}$$

where u is the precipitable water given by $u = \exp[0.1133 - \ln(G + 1) + 0.0393T_d]$ (Smith 1966), T_d is the dewpoint (°F), and G is an empirical constant dependent upon time of year and latitude. The fourth transmission coefficient is (Houghton 1954; Meyers and Dale 1983)

$$T_a = 0.935^m. \tag{10}$$

Once I was calculated from (4), direct observations of solar irradiance were used to calculate s and clf in (3). Inclusion of the effects of clouds yields

$$\mathrm{LW}_d = [\mathrm{clf} + (1 - \mathrm{clf})\varepsilon_c]\sigma T^4 = \varepsilon\sigma T^4. \tag{11}$$

As clf increases from 0 to 1, ε proportionally increases between the clear-sky value ($\varepsilon = \varepsilon_c$) and the limiting (but unobserved) value ($\varepsilon = 1$). Calculated values of clf less than zero were adjusted back to zero so as to be physically realistic.

Six popular ε_c formulations, modified by the cloudiness correction and inserted into (11), were tested against LW_d data collected at the Department of Energy Atmospheric Radiation Measurement (ARM) Southern Great Plains Cloud and Radiation Testbed Central Facility (CF) in Lamont, Oklahoma (Stokes and Schwartz 1994):

$$\varepsilon = [\mathrm{clf} + (1 - \mathrm{clf})(0.68 + 0.036e^{1/2})], \quad \text{Anderson (1954);} \tag{12}$$

$$\varepsilon = [\mathrm{clf} + (1 - \mathrm{clf})(9.36 \times 10^{-6}T^2)], \quad \text{Swinbank (1963);} \tag{13}$$

$$\varepsilon = (\mathrm{clf} + (1 - \mathrm{clf}) \times \{1 - [0.261 \exp(-7.77 \times 10^{-4}) \times (273.15 - T)^2]\}), \quad \text{Idso and Jackson (1969);} \tag{14}$$

$$\varepsilon = \{\mathrm{clf} + (1 - \mathrm{clf})[0.67(1670q)^{0.08}]\}, \quad \text{Staley and Jurica (1972);} \tag{15}$$

$$\varepsilon = \{\mathrm{clf} + (1 - \mathrm{clf})[1.24(e/T)^{1/7}]\}, \quad \text{B75;} \tag{16}$$

and

$$\varepsilon = (\mathrm{clf} + (1 - \mathrm{clf})\{1.08[1 - \exp(e^{T/2016})]\}), \quad \text{Satterlund (1979)} \tag{17}$$

where T is in degrees kelvin and e is in millibars. In (15), q is the specific humidity, which is a function of e and the barometric pressure. Equation (12) is a modification of the formulation originally developed by Brunt (1932) and has been shown to provide acceptable clear-sky estimates of LW_d [using the value of ε in (11)] in Oklahoma (Arnfield 1979).

3. Results and discussion

The data needed for testing values of LW_d [using (12)–(17) in (11)] included near-surface observations of LW_d, downwelling shortwave radiation SW_d, barometric pressure, vapor pressure, and temperature.

At the ARM CF site (36.61°N, 97.49°W, altitude = 318 m), measurements of LW_d (SW_d) were made with an upward-looking hemispherical broadband Eppley pyrgeometer (pyranometer). The instruments were mounted at a height of 1.5 m and employed a 1-s sampling interval. Barometric pressure was measured at a height of 1 m at 1-min intervals and vapor pressure and temperature were measured at 2 m at 1-s intervals. All variables were averaged over 30 min.

Often, the most difficult part of obtaining an accurate measurement of downwelling longwave radiation is overcoming the problems associated with solar heating of the instrument dome, which can result in spuriously high LW_d. The upward-looking pyrgeometer at the CF site was shaded and ventilated to reduce these dome heating effects, and a correction was also made based on the measured dome and case temperatures (M. Splitt 1997, personal communication).

Calculations of LW_d using (11) and the six parameterizations of ε (12)–(17) were first compared to observed LW_d ARM data over four 1-month periods: November 1995, February 1996, May 1996, and August 1996. By doing this, seasonal biases in the performance of any of the formulas could be detected. Since the clear-sky model could be used only during daylight hours, the total amount of data available was restricted. For fall and winter (spring and summer), ARM data from 1400–2230 (1300–2330) UTC were used. Table 1 shows the results of these comparisons.

The mean bias error (mbe) is given by

$$\mathrm{mbe} = \frac{1}{n}\sum_{i=1}^{n} [\mathrm{LW}(p)_{d,i} - \mathrm{LW}(o)_{d,i}], \tag{18}$$

where $\mathrm{LW}(p)_{d,i}$ and $\mathrm{LW}(o)_{d,i}$ are the parameterized and observed values, respectively, and n is the total number of half-hourly averaged observations for the month. A positive value means that the parameterization overestimates LW_d. Only the Anderson (1954) and the B75 schemes had absolute mbe's that were consistently less than 12 W m^{-2} throughout the year.

The rmse is given by

$$\mathrm{rmse} = \sqrt{\frac{1}{n - 1}\sum_{i=1}^{n} \{[\mathrm{LW}(p)_{d,i} - \mathrm{LW}(o)_{d,i}] - \mathrm{mbe}\}^2}. \tag{19}$$

Assuming a normal error distribution means that 68% of the individual errors are within a range bounded by mbe ± rmse. Values of the rmse's were consistently higher in fall and winter than in spring and summer for all schemes. The Anderson (1954) scheme had the lowest rmse's over the course of the year. Statistics from a

TABLE 1. Comparative LW_d statistics for the six ϵ formulations (12)–(17).

Method	Parameters	Nov 1995 (fall)	Feb 1996 (winter)	May 1996 (spring)	Aug 1996 (summer)
Anderson (1954)	mbe (W m^{-2})	6.6	7.2	1.7	−0.3
	rmse (W m^{-2})	13.0	18.1	10.6	11.4
	best-fit R	0.94	0.93	0.94	0.83
	best-fit y intercept (W m^{-2})	12.7	−3.4	31.8	82.6
	best-fit slope	0.98	1.04	0.92	0.80
Swinbank (1963)	mbe (W m^{-2})	12.4	12.1	11.7	4.7
	rmse (W m^{-2})	19.9	25.0	16.7	12.5
	best-fit R	0.90	0.90	0.87	0.83
	best-fit y intercept (W m^{-2})	−15.2	−50.8	31.6	38.7
	best-fit slope	1.10	1.23	0.95	0.92
Idso and Jackson (1969)	mbe (W m^{-2})	16.3	18.8	14.1	7.4
	rmse (W m^{-2})	18.4	20.7	17.8	12.9
	best-fit R	0.89	0.90	0.86	0.82
	best-fit y intercept (W m^{-2})	15.0	−4.3	29.7	37.7
	best-fit slope	1.00	1.09	0.96	0.93
Staley and Jurica (1972)	mbe (W m^{-2})	18.3	14.3	15.7	12.2
	rmse (W m^{-2})	14.7	20.2	13.2	10.8
	best-fit R	0.93	0.91	0.91	0.84
	best-fit y intercept (W m^{-2})	14.3	−22.7	60.4	114.8
	best-fit slope	1.01	1.14	0.88	0.75
B75	mbe (W m^{-2})	0.2	3.9	−10.0	−11.9
	rmse (W m^{-2})	14.9	22.2	11.9	10.5
	best-fit R	0.95	0.92	0.94	0.85
	best-fit y intercept (W m^{-2})	−36.6	−48.7	11.9	104.7
	best-fit slope	1.13	1.16	1.00	0.77
Satterlund (1979)	mbe (W m^{-2})	19.7	15.8	14.9	10.3
	rmse (W m^{-2})	14.9	20.2	13.4	10.9
	best-fit R	0.93	0.91	0.91	0.84
	best-fit y intercept (W m^{-2})	19.2	−19.6	65.4	108.6
	best-fit slope	1.00	1.13	0.87	0.76

linear regression of the comparisons were also analyzed. The correlation coefficient is denoted by ''best fit R.'' The B75 and Anderson (1954) schemes had the highest values throughout the year, and every scheme had its smallest value in summer. The y intercept and the slope of the regression lines are also given in Table 1. A perfect scheme would produce a y intercept of zero and a slope of one. The Idso–Jackson (1969) scheme performed best in these two categories but had the largest rmse's of all the equations tested.

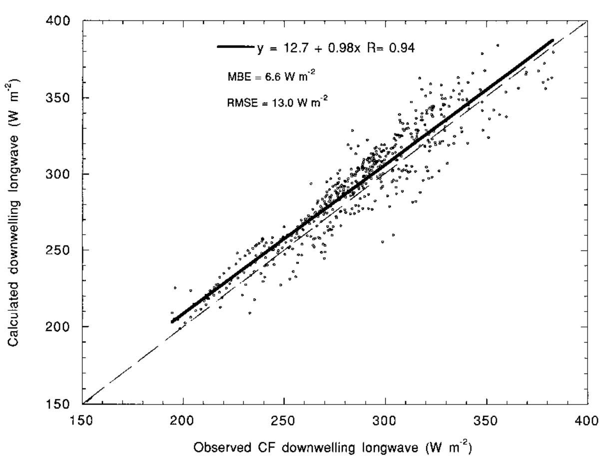

FIG. 1. Comparison between observed and calculated LW_d using the Anderson (1954) emissivity scheme from the daylight hours (1400–2330 UTC) in November 1995. The data were obtained from the ARM CF site. The solid line represents the results of the linear regression, while the dashed line represents a ''perfect-fit'' line.

Based on their comparatively low mbe and rmse values, the Anderson (1954) and the B75 formulations were considered to be superior to the other four. Scatterplots of the two schemes for November 1995 are depicted in Figs. 1 and 2. One can see that the distributions of data points about the linear fits are similar. What distinguishes the formulations is the slope of each regression curve. Since the B75 equation has a physically based derivation while the Anderson (1954) equation is strictly empirical, the B75 scheme was chosen for further investigation.

4. Seasonal adjustment to Brutsaert coefficient

Equation (16) was derived using Schwarzschild's radiative transfer equation, in which standard atmosphere vertical profiles of temperature and vapor pressure were used to calculate the leading coefficient (1.24) [see (1)–(11) in B75]. Using measured profiles of water vapor pressure and temperature in Niger, Culf and Gash (1993)

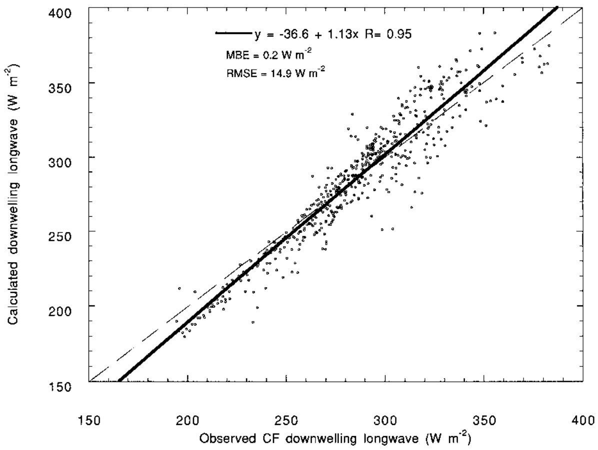

FIG. 2. Same as Fig. 1 except using the B75 scheme.

were able to rederive the original B75 equation and get a slightly different value of the leading coefficient. By doing this they were able to reduce the rmse by 50%. They also found that, during the dry season in Niger, the lapse rate of vapor pressure was significantly smaller than the standard atmospheric lapse rate assumed in B75. This means that for the given surface conditions there was more water vapor aloft, and thus higher emissivity, than expected using a standard atmosphere vapor pressure profile. Because of this, the leading coefficient was increased to 1.31 in order to properly represent LW_d in these conditions. On the other hand, for wet surface conditions, the vapor pressure lapse rate would be expected to be larger than in a standard atmosphere since vapor pressure varies much more near the surface than aloft. In this case, the B75 coefficient would have to be reduced to compensate. It is important to note that it is the magnitude of the lapse rate that determines the correct value of the leading coefficient, not the magnitude of the measured vapor pressure at the surface.

Since the Culf and Gash (1993) results showed a variation in the leading coefficient between the dry season and the wet season, it was hypothesized that this coefficient may undergo an annual sinusoidal variation similar to that of other meteorological variables (temperature, solar radiation, and vapor pressure). In Oklahoma, the lowest value was expected to occur in July (more humid, larger vapor pressure lapse rate) and the highest value in January (less humid, smaller vapor pressure lapse rate). Analysis of a 12-month period of ARM CF data (November 1995–October 1996) was performed in order to find the best-fit sinusoidal variation. The resultant leading coefficients ranged from 1.28 in January to 1.16 in July according to

$$LW_d = \{\mathrm{clf} + (1 - \mathrm{clf})(1.22 + 0.06 \cdot \sin[(\mathrm{month} + 2) \cdot \mathrm{pi}/6])(e/T)^{1/7}\}\sigma T^4, \quad (20)$$

where month is the numerical month (e.g., January = 1). The fact that the values of the leading coefficient were largest in winter and smallest in summer may seem counterintuitive since LW_d is least in winter and greatest in summer. However, the relatively low e during the winter months more than offsets the larger leading coefficient, with the opposite relationship during the summer months.

TABLE 2. Statistics for the improved formulation based on a sinusoidal variation of the leading coefficient.

Month	mbe (W m^{-2})	rmse (W m^{-2})	R	y intercept (W m^{-2})	Slope
Nov 1995	−1.6	15.0	0.94	−34.6	1.13
Dec 1995	−4.3	16.6	0.93	−3.6	1.00
Jan 1996	0.2	19.6	0.91	−9.2	1.04
Feb 1996	−0.5	22.3	0.91	−55.4	1.21
Mar 1996	−1.1	17.9	0.95	−13.6	1.05
Apr 1996	1.9	12.2	0.94	10.3	0.97
May 1996	0.3	10.6	0.95	−3.1	1.01
Jun 1996	−2.6	11.6	0.94	28.7	0.92
Jul 1996	−7.3	12.1	0.87	65.7	0.82
Aug 1996	3.7	12.0	0.82	72.9	0.81
Sep 1996	−8.7	16.2	0.88	31.7	0.89
Oct 1996	−6.4	15.7	0.90	13.8	0.94

Table 2 reveals that using (20) rather than (16) resulted in fairly uniform and small absolute mbe's (less than 10 W m^{-2}) throughout the year. It is immediately apparent that the new scheme had a smaller mbe than the other five schemes (Table 1), with the exception of the Anderson (1954) scheme in August. The rmse's are similar to the original B75 scheme, while correlation coefficients were slightly smaller. The y intercept and slope of the regression lines were also mostly similar to B75. The new scheme appears to be valid for all seasons and sky conditions with rmse's less than 23 W m^{-2} for 30-min averages. This scheme will not work at night since there is then no way to ascertain cloud fraction.

Because of the empiricism involved in fitting the sinusoid to the CF data, the broader applicability of (20) may be in question. For this reason, data from the First International Satellite Land Surface Climatology Project (ISLSCP) Field Experiment (FIFE; Sellers et al. 1992) were acquired and used to test the new formulation. These data were available as 30-min averages. Barometric pressure and the wet- and dry-bulb temperatures were measured at 2 m at 10 different locations within the FIFE domain (39.05°N, 96.53°W, altitude = 410 m). These 10 values were averaged to obtain representative site values at each half-hourly observation time. The site-average mixing ratio was then calculated from the site-average barometric pressure and wet-bulb temperature data and converted to vapor pressure. The LW_d and SW_d data were measured by Eppley pyranometers and pyrgeometers, respectively, at two different sites located 14 km apart, then averaged to obtain representative site values for the area at 30-min intervals. These observations were compared to the parameterization represented by (20), and the error statistics are shown

TABLE 3. Comparisons of ARM and FIFE LW_d datasets using the sinusoidal variation in effective atmospheric emissivity in (20). ARM data are from 1996; FIFE data are from 1987. Units in W m^{-2}.

	ARM	FIFE		ARM	FIFE
Jun MBE	−2.6	−6.7	Jun rmse	11.6	11.5
Jul MBE	−7.3	−14.9	Jul rmse	12.1	9.2
Aug MBE	3.7	−14.5	Aug rmse	12.0	9.7
Sep MBE	−8.7	−11.9	Sep rmse	16.2	10.8

in Table 3. The FIFE comparisons show larger mbe's but smaller rmse's than the ARM data. Considering that the method was developed empirically for ARM data, the FIFE comparisons are very encouraging.

The results obtained in this study compare favorably to recent parameterizations of LW_d in all sky conditions. Sugita and Brutsaert (1993) were able to empirically calibrate the adjustable parameters in their formulation using 1987 FIFE data (summer and fall only) to reduce their rmse's to around 15–17 W m^{-2}, which are larger than the summer and fall errors found by our improved technique. Moreover, they used only 325 data points and did not test the effectiveness of their method in winter and spring. Culf and Gash (1993), by tuning the Brutsaert coefficient to 1.31 during the dry season in Niger in 1990, were able to get an rmse of 12 W m^{-2}, but again, this is a limited dataset from summer and early fall and was only valid in clear skies.

5. Using sounding data to adjust Brutsaert coefficient

The Culf and Gash (1993) results also suggested that the leading coefficient could be adjusted, based on local sounding data, in order to provide accurate estimates of LW_d. To test this idea, soundings from 1730 UTC at the ARM CF were analyzed from November 1995–October 1996. The sounding time was chosen to represent the middle of daylight hours. Monthly averaged water vapor lapse rates were calculated, and the B75 derivation was reworked to provide a monthly averaged leading coefficient. A comparison of the leading coefficients derived from the soundings and those from the sinusoidal hypothesis are shown in Fig. 3. It is apparent that the most significant disagreement between the two methods occurred during the spring and summer months. Since the sinusoid represents a ''best fit'' to the data, the largest mbe's will occur during these months if the sounding method is used, as seen in Fig. 4. Nevertheless, the absolute mbe's were still no greater than 15 W m^{-2} using the sounding-derived coefficients, and Fig. 5 shows that the rmse's were actually smaller. It is also interesting to note in Fig. 3 that the sounding-derived coefficients were close to the original 1.24 coefficient derived from the standard atmosphere assumptions employed by Brutsaert to derive (16).

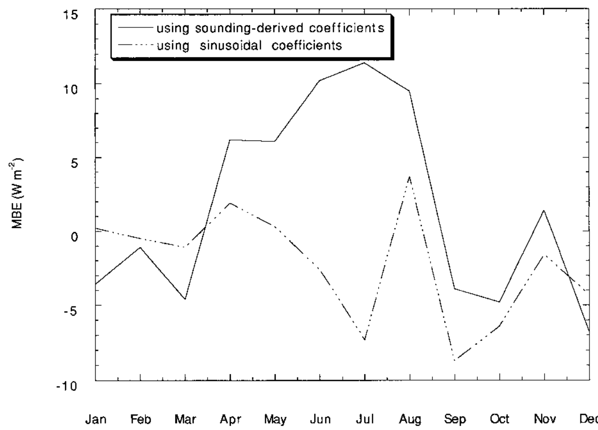

FIG. 4. Comparison of the mbe's of both variations of the improved method.

6. Summary and conclusions

An improved, practical technique for calculating effective atmospheric emissivity for use in estimating

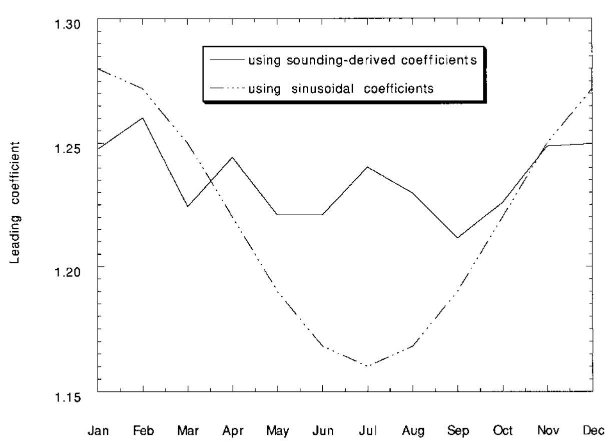

FIG. 3. Comparison of the calculated leading coefficients of both variations of the improved method. The dashed line represents the original B75 leading coefficient (1.24).

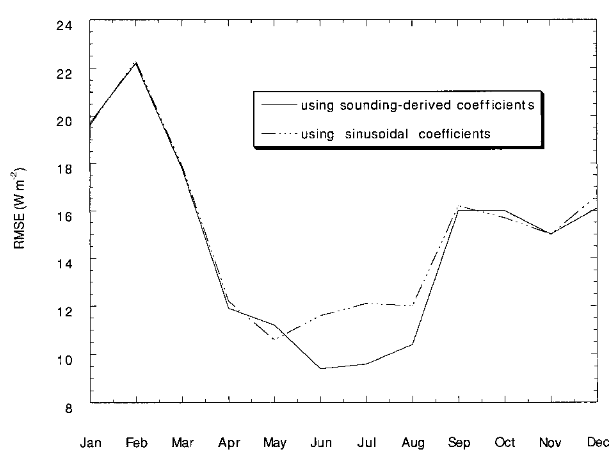

FIG. 5. Comparison of the rmse's of both variations of the improved method.

downwelling longwave radiation during the daytime has been investigated. The physically based model given in B75 has been modified using ideas from Deardorff (1978) and Culf and Gash (1993). Two improvements have been developed and tested. The first employs an empirical sinusoidal annual variation in the leading coefficient. This method was applied to two datasets with slightly different results. For the ARM data, the method resulted in absolute mbe's less than 9 W m^{-2} and rmse's less than 23 W m^{-2}. For the FIFE data, absolute mbe's less than 15 W m^{-2} and rmse's less than 12 W m^{-2} were found over a 4-month warm season period. The second improvement uses an objective method for determining the cloud fraction using observed solar radiation and modeled clear-sky radiation.

The use of upper-air sounding data to calculate the leading coefficient in B75 was also studied. The yearly variation of these coefficients hinted at a sinusoidal variation of similar phase but much smaller magnitude than the ''best fit.'' Although mbe's were consequently larger using the sounding-derived coefficients, rmse's were slightly smaller, and the empiricism of the best-fit sinusoid was removed. Because of this, the sounding method may be more portable than the sinusoid method, with the prerequisite that the site of interest have upper-air data available.

We recommend that if the climatology at a location where LW_d is needed but not measured is similar to that at another location where LW_d is measured, that the best-fit sinusoidal variation of the leading coefficient be employed. For example, the best-fit sinusoidal variation developed in this study based on observations of LW_d at the ARM sites in Oklahoma can be used to accurately estimate LW_d at many Oklahoma Mesonet (Brock et al. 1995) sites. In general, if the climatologies are sufficiently different, we suggest using the sounding-derived variation of the leading coefficients if there is a nearby upper-air station.

Future studies should include comparative analyses at the two other ARM sites in northern Alaska and the tropical western Pacific, where all components of the radiation budget in addition to standard meteorological variables are measured. Since these two climates are at opposite ends of the meteorological spectrum, these data would provide rigorous tests of the improved method. In addition, more tests using the sounding-derived variation of the leading coefficient should be performed using data from other field experiments (e.g., FIFE). These tests would act to verify or negate the promising results found in this study using ARM CF data.

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