The Treatment of Long-Wave Radiation and Precipitation in Climate Files for Building Physics Simulations

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ABSTRACT

There is an increased demand for hourly heat and moisture simulations for buildings. In simulation programs aimed at energy calculations, it has been enough to include the parameters: temperature, relative humidity, solar radiation, wind speed, wind direction, and cloud cover (Crawley et al. 1999; Wilcox and Marion 2008). These parameters are often a mix of measured and calculated data. Long-wave radiation from the sky and precipitation are sometimes used as parameters but are not as frequently measured. When focusing on horizontal or tilted surfaces, long-wave radiation can not be neglected. In most climate files, the long-wave radiation is not present, so the simulation programs have to make an educated guess based on available parameters. One important secondary parameter is cloudiness, which is measured or calculated from solar radiation compared to maximal theoretic solar radiation. Precipitation is necessary when making moisture calculations for the building envelope. The precipitation is often measured every 6th or 12th hour, so the simulation programs must distribute this over a period hours in between. This paper presents some of the techniques for these calculations and compares the results with real, hourly measured data for four locations in Sweden. General results are that the existing investigated models for long-wave radiation give a root-mean-square accuracy between 24–27 W/m^2 and that the models for Sweden using parameters identified above give a root-mean-square error up to 23.2 W/m^2. For precipitation, the optimal hourly limit value for precipitation was 88% RH_c.

INTRODUCTION

When calculating temperature and moisture for buildings, some kind of hourly climate file must be used. The two most obvious possibilities are to use measured data from a specific location or constructed data as a base for the climate file. The constructed data can be, for example, hourly values for a typical year or a design (worst case) year for a location. In any case, the climate data must be based on measurements. Meteorological stations usually measure temperature, moisture content (measured as dew point or relative humidity), pressure, wind speed, and wind direction on an hourly basis. Precipitation and cloud cover are often measured more seldom; for example, cloud cover is measured every three and precipitation every twelve hours or daily (SMHI 1988). Global and diffuse solar radiation are measured in fewer locations. If the solar parameters are not measured, they must be modeled from latitude, cloud cover, etc. (Meyers and Dale 1983; Atwater et al. 1978). Note that there are, of course, large differences between countries and locations; the airports will always have climate stations but do not necessarily have automatic hourly stations. Temperature calculations for walls and windows can, in most cases, be made based on this data with reasonable accuracy. The sky temperature for these vertical cases typically is set equal to the air temperature minus around 10°C. For tilted or horizontal surfaces (roofs, glazed spaces, etc.) the calculations must also include the detailed long-wave thermal radiation from the sky to be accurate (Wall 1996). However, the long-wave radiation from the sky is often not measured at all. The reason for this is probably that the price of the measuring instrument has been higher than the expected use of the data. For moisture calculations, hourly values for precipitations are needed. This paper investigates

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some existing techniques for constructing hourly values for long-wave radiation and precipitation.

LONG-WAVE RADIATION

Long-wave radiation from the sky (L_w) is typically measured in W/m^2 or W·h/m^2·h. The measuring instrument is some kind of pyrgeometer, e.g., a Hukseflux IR02 with an accuracy of 10% for daily sums (Hukseflux manual). To make it clear that this radiation does not originate from the sun, it is sometimes called the *atmospheric* long-wave radiation. The long-wave radiation is in the order of 200–400 W/m^2 (Flerchinger et al. 2009; Crawford and Duchon 1999) and varies on a daily and seasonal basis (see Figure 1).

When formulating algorithms based on other meteorological data, it is natural to start with the Stefan-Boltzmann equation for thermal radiation from a surface with temperature T (K):

$$L_w = \varepsilon \cdot \sigma \cdot T^4 \tag{1}$$

where ε is the emissivity of the surface and σ is the Stefan-Boltzmann constant. Many authors have formulated algorithms for L_w mainly based on

- outdoor temperature T_o (K);
- moisture content in air described by outdoor vapor pressure e_o (kPa), dewpoint T_d (K), or precipitable water w (mm);
- some kind of cloudiness index, cx (dimensionless); and
- atmospheric pressure at the ground (more seldom).

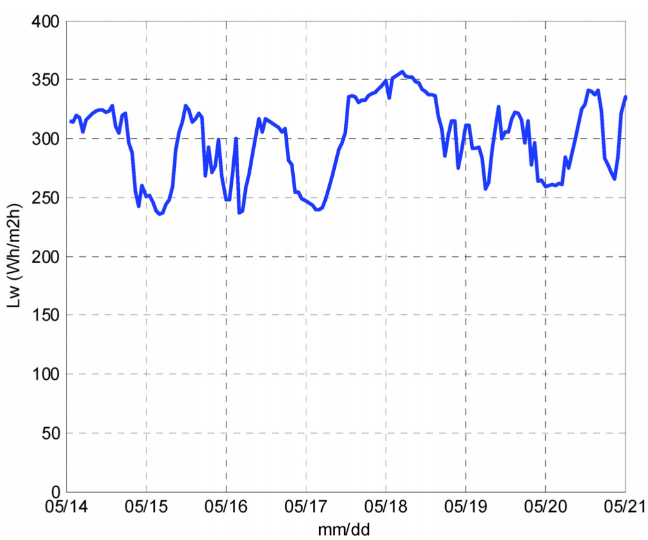

Figure 1 *Measured long-wave radiation from Lund (May 14, 1995 to May 21, 1995).*

Clear Sky

It is natural to start by deriving algorithms for clear-sky conditions. The clear sky gives a more stable long-wave radiation and is of special interest when the minimum long-wave radiation is the key parameter. Given the parameters in Equation 1, one possible choice is to use the outdoor temperature T_o as T in Equation 1 and have ε dependent on the other measured parameters. Examples of this are Equations 3–5:

$$L_{w,\,clr} = \varepsilon_{clr}(e_o, T_o, \ldots) \cdot \sigma \cdot T_o^4 \tag{2}$$

Ångström (1918):

$$\varepsilon_{clr} = 0.83 - 0.18 \cdot 10^{-0.067 e_o} \tag{3}$$

Berdahl and Martin (1984):

$$\varepsilon_{clr} = 0.738 + 0.61\left(\frac{T_d}{100}\right) + 0.20\left(\frac{T_d}{100}\right)^2 \tag{4}$$

Niemelä et al. (2001)

$$\varepsilon_{clr} = \begin{cases} 0.72 + 0.09(e_o - 2), & e_o \geq 0.2, \\ 0.72 - 0.76(e_o - 2), & e_o < 0.2 \end{cases} \tag{5}$$

Here T_d is the dewpoint temperature (°C), e_o is the vapor pressure (kPa), and $L_{w,clr}$ is the long-wave radiation from a clear sky. The Ångström equation is as cited by Flerchinger et al. (2009). Another possibility is to have $L_{w,clr}$ dependent on the parameters directly.

Dilley and O'Brien (1998):

$$L_{w,\,clr} = 59.38 + 113.7\left(\frac{T_o}{273.16}\right)^6 + 96.96\sqrt{\frac{w}{25}} \tag{6}$$

with w as the precipitable water in millimeters.

$$w = 4650\frac{e_o}{T_o} \tag{7}$$

Cloudy Sky

For hourly calculations in the context of buildings, the cloudy skies must also be accounted for. If the cloudiness is somehow directly measured, this value is of course to be used. If however the solar radiation is measured but not the cloudiness, the cloudiness can be calculated from the measured meteorological data, e.g., global solar radiation, direct solar radiation, or diffuse solar radiation. Solar data is only available during daylight hours so some strategy must be chosen for how to handle the whole day, e.g., a moving twenty-four-hour average or an average of a few hours close to sunset and sunrise. Since the calculated cloudiness should describe the whole sky, it is natural to choose the global solar radiation. The algorithms below for calculating the cloudiness use either the theoretical extraterrestrial global solar radiation on a surface parallel to the Earth's surface outside the atmosphere H_0 (Wh/m^2·h) or the theoretical global solar radiation on the Earth's surface S_0 (W/m^2·h). The formula for the extraterrestrial radiation takes into account the varying distance from sun to Earth, as well as the declination (El-Sebaii et al. 2010). In this paper, H_0 is the daily average extraterrestrial radiation (Wh/m^{2}h):

$$H_0 = \frac{I_{SC}}{\pi} \cdot \left[1 + 0.033\cos\left(2\pi\frac{d_{nr}}{365}\right)\right]$$

$$x(\cos(\varphi)\cos(\delta)\sin(\omega_s) + \omega_s\sin(\varphi)\sin(\delta)) \quad (8)$$

where φ (rad) is the latitude, δ (rad) is the solar declination, ω_s (rad) is the sunset hour angle, I_{SC} is the solar constant (1367 Wh/m^{2}h), and d_{nr} is the day number (1–365).

$$\delta = 0.4093 \cdot \sin\left(2\pi\frac{284 + d_{nr}}{365}\right) \quad (9)$$

$$\omega_s = \cos^{-1}(-\tan(\varphi)\tan(\delta)) \quad (10)$$

The clearness index K_0 (dimensionless) is defined as the quota between the measured global solar radiation on a horizontal surface on the ground I_G (W·h/m^2·h) and the extraterrestrial solar radiation H_0.

$$K_0 = \frac{I_G}{H_0} \quad (11)$$

Since the clearness index is the quota of two different entities, the actual cloud cover is not obviously linear dependent on K_0. The cloud cover c (dimensionless) is a value between 0 and 1 that describes the fraction of the sky that is covered by clouds. According to Flerchinger et al. (2009) c can be calculated as $c = 0$ for $K_0 > k_{clr}$ and $c = 1$ for $K_0 < k_{cld}$ and linearly interpolated in between. The values for

$$k_{clr} \approx 0.7$$

and

$$k_{cld} \approx 0.3$$

were investigated in detail by Flerchinger et al. (2009).

An index describing the cloud cover can also be based on theoretic global solar radiation on the Earth's surface S_0 (W·h/m^2·h). The solar index s (dimensionless) is defined as the quota between the measured global solar irradiance I_G and the theoretical value S_0 (Flerchiner et al. 2009; Crawford and Duchon 1999):

$$S = \frac{I_G}{S_0} \quad (12)$$

This is, however, much more complicated than calculating the extraterrestrial radiation, since the influence of the atmosphere must be taken into account. This includes the length of the solar rays in the atmosphere (air mass), Rayleigh scattering, aerosol extinction, water vapor absorption, and permanent gas absorption. Numerous models exists to describe these phenomena Yang et al. (2006), Atwater and Ball (1978), Meyers and Dale (1983), López et al. (2007). Some are based on physical interpretations and some are more direct parameter fitting. Flerchinger et al. (2009) used a formula presented by Crawford and Duchon (1999) that was an adaptation of a formula presented by Atwater and Ball (1978). We will present the model from Crawford as presented by Flerchinger to calculate S_0 for every hour:

$$S_0 = I_{SC}\sin(h) \cdot \tau_R \cdot \tau_{pg} \cdot \tau_w \cdot \tau_a \quad (13)$$

where τ_R, τ_{pg}, τ_w, and τ_a denote the transmission coefficients for Rayleigh scattering, absorption by permanent gases, absorption by water vapor, and absorption and scattering by aerosols, respectively. The parameter h is the height of the sun over the horizon (solar altitude). The product of Rayleigh scattering and absorption by water vapor was calculated as follows:

$$\tau_R \cdot \tau_{pg} = 1.021 - 0.084[m(0.00949P + 0.051)]^{0.5} \quad (14)$$

Here m is the optical air mass at 101.3 (kPa) and P is the pressure (kPa). The air mass can be calculated with different formulas of increasing complexity and accuracy. Atwater and Ball (1978) and, successively, Crawford and Duchon (1999) and Flerchinger et al. (2009), calculated m as (even if the formula itself is erroneously described by Crawford and Duchon and Flerchinger et al):

$$m = \frac{35}{(1224 \cdot \sin^2(h) + 1)^{0.5}} \quad (15)$$

A more accurate formulae from Young (1994) (originally expressed in zenith angle) is

$$m = \frac{1.002432\sin^2(h) + 0.148386\sin(h) + 0.0096467}{\sin^3(h) + 0.149864\sin^2(h) + 0.0102963\sin(h) + 0.000303978} \quad (16)$$

Young claims this formula has an error less than 0.0037 air masses close to the horizon, where the air mass is 31.7. The formula from Atwater and Ball gives an air mass at the horizon of 35.

The transmission coefficient for absorption in water τ_w is from Flerchinger et al. (2009)

$$\tau_w = 1 - 0.077\left(w\frac{m}{10}\right)^{0.3} \quad (17)$$

where w is the precipitable water (mm). The transmission coefficient for aerosol extinction τ_a is from Meyers and Dale (1983) and, successively, Flerchiner et al.:

$$\tau_a = 0.935^m \quad (18)$$

The number 0.935 is very much dependent on the amount of particles in the atmosphere. Another way of expressing τ_a is from Louche et al. (1987):

$$\tau_a = 0.1082 + 0.878 e^{-1.60\beta m} \quad (19)$$

Here β is the Ångström turbidity. To have a similar behavior as Equation 18, the turbidity should be about $\beta = 0.06$. Sinus of the solar height can be calculated as

$$\sin(h) = \sin(\varphi)\sin(\delta) + \cos(\varphi)\sin(\delta)\cos\left(\frac{\pi(t-12)}{12}\right) \quad (20)$$

Equation 13 can be made more accurate to include the Earth's elliptic path around the sun, as in Equation 8. The equation then becomes

$$S_0 = I_{SC}\sin(h)\left[1 + 0.033\cos\left(2\pi\frac{d_{nr}}{365}\right)\right] \cdot \tau_R \cdot \tau_{pg} \cdot \tau_w \cdot \tau_a \quad (21)$$

In Flerchinger et al., the twenty-four-hour average of S_0 was used; it was also used in the investigation presented below. With the clearness index K_0, the cloud cover c, or the solar index s it is now possible to formulate long-wave radiation from a cloudy sky in many different ways.

Aubinet (1994):

$$L_{w,\,cld} = \varepsilon_{cld} \cdot \sigma \cdot T_o^4 \quad (22)$$

$$\varepsilon_{cld}(K_0) = 0.93 - 0.139\ln(1 - K_0) \quad (23)$$

Aubinet (1994):

$$L_{w,\,cld} = \sigma \cdot T_{cld}^4 \quad (24)$$

$$\begin{aligned} &T_{cld}(e_o, T_o, K_0) \\ &= 94 + 12.6\ln(1000 e_o) - 13K_0 + 0.341 T_o \end{aligned} \quad (25)$$

In Equation 25, T_{cld} (K) is the temperature of the cloud cover and, the emissivity is set to 1.0 in Equation 24.

Crawford and Duchon (1999):

$$L_{w,\,cld} = \varepsilon_{cld} \cdot \sigma \cdot T_o^4$$

$$\varepsilon_{cld}(s) = (1 - s) + s \cdot \varepsilon_{clr} \quad (26)$$

where ε_{clr} can be calculated by the different equations above. Flerchinger et al. used Dilley and O'Brien (1998) (Equation 6) with good results.

Kimball et al. (1982):

$$L_{w,\,clr} = L_{clr} + \tau_8 c f_8 \sigma \cdot T_c^4 \quad (27)$$

$$\tau_s = 1 - \varepsilon_{8z}(1.4 - 0.4\varepsilon_{8z})$$

$$\varepsilon_{8z} = 0.24 + 2.98 \cdot 10^{-6} \cdot e_o^2 \cdot e^{\frac{3000}{T_o}}$$

$$f_8 = -0.6732 + 0.6240 \cdot 10^{-2} T_c - 0.9140 \cdot 10^{-5} \cdot T^2$$

Here T_c (K) is the cloud temperature. Flerchinger et al. used $T_c = T_o - 11$ K. Obviously, the long-wave radiation $L_{w,clr}$ from the clear sky must be calculated by other means. Flerchinger et al. used the Dillay and O'Brian (1998) formula (Equation 6), which is also used in this paper. They performed an investigation of the optimal choices for k_{clr} and k_{cld} for this formula which gave

$$\begin{aligned} k_{clr} &= 0.8 \\ k_{cld} &= 0.25 \end{aligned} \quad (28)$$

Problem and Hypothesis

The need for long-wave radiation data in Sweden has increased since a greater accuracy in energy, temperature, and moisture calculations is in demand. This data is, however, lacking, except for a few locations, so data must be calculated from equations as described above. The commercial program, Meteonorm, is used by many engineers to produce this data based on Aubinet (1994) (Equation 25). This paper investigates three of the existing formulas above (Equations 25–27) as well as six new formulas (Equations 29–34) for reference. These new formulas are formulated from a parameter estimation view and not necessarily the physical interpretation. The equations are described below with denoted names for simpler reference. The hypothesis is that there exists one or more formula that gives results with acceptable accuracy for the long-wave radiation. Acceptable accuracy is not exactly defined—it depends on the particular usage—but since the stated accuracy of one pyrgeometer (Hukseflux) is around 10% for daily sum, this indicates an accuracy of 30 W/m^2.

For reference, a parameter fit of Equation 25 from (1994) was done:

Aubinet (1994) identified

$$T_{cld}(e_o, T_o, K_0) = \theta_1 + \theta_2\ln(1000 e_o) + \theta_3 K_0 + \theta_4 T_0 \quad (29)$$

Two equations were formulated to test the sensitivity to the clearness index K_0 and the solar index s:

Clearness identified

$$\varepsilon_{cld} = \theta_1 + \theta_2\left(\frac{T_d}{100}\right) + \theta_3\left(\frac{T_o}{273.15}\right) + \theta_4 K_0 \quad (30)$$

Solar index identified

$$\varepsilon_{cld} = \theta_1 + \theta_2\left(\frac{T_d}{100}\right) + \theta_3\left(\frac{T_o}{273.15}\right) + \theta_4 s \quad (31)$$

Note that ε_{cld} describes the total emissivity and should be used as in Equation 22. The method of Crawford and Duchon (1999) (Equation 26), together with a parameter fit based on Berdahl and Martin (1984) (Equation 4), gave the following:

Berdahl 1 identified

$$\varepsilon_{clr} = \theta_1 + \theta_2\left(\frac{T_d}{100}\right) + \theta_3\left(\frac{T_d}{100}\right)^2 \quad (32)$$

During the investigation, the accuracy of θ_2 was low, so another equation was also tested:

Berdahl 2 identified

$$\varepsilon_{clr} = \theta_1 + \theta_2\left(\frac{T_d}{100}\right) + \theta_3\left(\frac{T_o}{273.15}\right)^2 \quad (33)$$

Finally, a formula based on Crawford and Duchon (1999) (Equation 26) and Dilley and O'Brien (1998) (Equation 6) was tested.

Crawford Dilley identified

$$L_{w,\,clr} = \theta_1 + \theta_2\left(\frac{T_o}{273.16}\right)^6 + \theta_3\sqrt{\frac{w}{25}} \quad (34)$$

Method

At four locations in Sweden, the hourly long-wave radiation, temperature, relative humidity, and global solar irradiance have been measured from 1990–1998. Over this period, the equipment has sometimes failed. Two longer periods with valid data have been chosen for each location. The locations and the periods are shown in Table 1.

The hourly data have been used to identify the parameters in Equations 29–34 and to verify these identifications. The identification was made using multiple linear regressions based on the least squares error function:

$$V(\theta) = \frac{1}{2}\sum_{i=1}^{N}(L^i_{w,\,measured} - L^i_{w,\,calculated}(\theta))^2 \quad (35)$$

Here θ is the parameter vector. The estimate of θ denoted $\hat{\theta}$ minimizes the error function $V(\bar{\theta})$. The data from Period A was used for identification, and the data from Period B for verification.

The identification gives

- fitted parameters $\hat{\theta}$ and
- standard deviation σ (%) in the parameter fitting.

The verification gives

- root mean square error *RMS*, (W·h/m^2·h);
- mean bias *MB* (W·h/m^2·h) (positive when giving too high a value); and
- regression value *r* (dimensionless).

Results for Long-Wave Radiation

The results from the verification of Equations 25–29 and the identification and verification of Equations 29–34 are presented in Table 2. The *RMS MB* and *r* are based on 64 204 hourly values (Table 1, Period A) from the four locations in Sweden, and the identified parameters θ are based on different 121 924 hourly values (Table 1, Period B)from the same location.

The results in Table 2 for the existing Equation 25, Equations 26 plus Equation 6, and Equation 27 plus Equation 6 are similar to the results calculated by Flerchinger et al. (2009), based on seven locations in United States, including Alaska. These values are compared in Table 3.

Given that the long-wave radiation varies between 200–400 Wh/m^2·h, the RMS is about 10%. From Tables 2 and 3, it is clear that the equations do not give exceptionally different results. The equation that gives the best results is the one denoted "Clearness Identified," which uses outdoor (T_o), dewpoint temperature (T_d), and the clearness index (K_0) as parameters. The Solar Index identification equation gives a slightly worse result. Of the already existing equations, the Kimball et al. (1982) and Dilley and O'Brien (1998) equations gave a surprisingly similar result (given that there was no fitting to the Swedish climate).

During the analysis of the results, some issues was especially noted:

- There was a tendency for higher latitudes to give a larger error.
- There was a tendency for the estimated global solar irradiance S_0 to be too low for higher latitudes, especially for Luleå at 65.55°, where twenty-four-hour average S_0 was lower than the 24h average measured global solar irradiance on numerous days.

This led to a more detailed analysis. The simple formula (Equation 15) for calculating air mass reduced S_0, compared to the more exact formula (Equation 16). Close to the horizon, the difference is 35 versus 31.7.

The formula for the aerosol extinction (Equation 18) was equivalent to an Ångström turbidity of about 0.06. The Ångström turbidity will strongly affect the global solar irradiance, especially for high latitudes (~55°), where the sun is low and, consequently, the air mass high. It is a well-known fact

Table 1. Locations and Periods with Valid Data

Location	Latitude	Period A	Period B
Lund	55.72	90.06.18–94.06.05	94.07.09–95.11.29
Stockholm	59.35	92.07.10–95.08.01	95.08.04–98.07.02
Borlänge	60.48	92.11.10–95.10.30	97.08.19–98.11.30
Luleå	65.55	93.03.03–97.01.31	91.05.23–93.02.21

Table 2. Results from Tested Equations

Equation	*RMS* (*MB*)	*r*	θ (σ %)
Aubinet (1994) (Equation 25)	27.0 (–10.8)	0.861	—
Crawford and Duchon (1999) (Equation 26) + Dilley and O'Brian (1998) (Equation 6)	25.4 (–9.3)	0.878	—
Kimball et al (1982) (Equation 27) + Dilley and O'Brian (1998) (Equation 6)	24.0 (4.8)	0.892	—
Aubinet Identified (Equation 29)	23.4 (–3.9)	0.898	34.17 (9.9) 7.18 (4.0) –21.35 (1.8) 0.700 (2.6)
Clearness Identified (Equation 30)	23.2 (–3.0)	0.900	1.547 (0.8) 0.598 (0.9) –0.569 (2.2) –0.280 (0.4)
Solar Index Identified (Equation 31)	23.5 (–3.3)	0.897	1.684 (0.7) 0.631 (0.9) –0.706 (1.8) –0.194 (0.4)
Berdahl 1 Identified (Equation 32)	23.8 (–3.1)	0.894	0.747 (0.04) 0.456 (0.8) 1.789 (1.8)
Berdahl 2 Identified (Equation 33)	23.5 (–2.7)	0.898	1.068 (0.7) 0.679 (0.9) –0.294 (2.3)
Crawford Dilley Identified (Equation 34)	23.3 (–2.5)	0.899	59.00 (0.8) 114.37 (0.6) 111.29 (0.8)

Table 3. Comparison between Results from This Investigation and Flerschineg et al. (2009) for Cloudy Skies (W·h/m²·h)

Equation	*RMS* (*MB*) Sweden	*RMS* (*MB*) Flerchinger
Aubinet (1994) (Equation 25)	27.0 (–10.8)	30.1 (–15.7)
Crawford and Duchon (1999) (Equation 26) + Dilley and O'Brian (1998) (Equation 6)	25.4 (–9.3)	25.8 (–1.3)
Kimball et al (1982) (Equation 27) + Dilley and O'Brian (1998) (Equation 6)	24.0 (4.8)	26.7 (1.0)

that turbidity varies strongly, with a tendency to be highest close to the equator (~0.08) and lower at the poles (~0.01), (Yang et al. 2001). The value also has a tendency to be lower in winter and higher in the summer with a variation of +0.02–0.06 (Yang et al. 2001):

$$\beta = 0.025 + 0.1\cos(\varphi) \cdot e\frac{0.7z}{1000} \pm (0.02 - 0.06) \qquad (36)$$

Here z (m) is the height over water level. Fox (1994) estimated the Ångström turbidity for Fairbanks, Alaska, (latitude 64.82°) for the longest consecutive period (Table 4). The yearly turbidity is about 0.04±0.02. It must also be stated that the turbidity did not vary smoothly over all the measured periods.

From these facts, the conclusion was drawn that a turbidity equivalent to 0.06 for all locations and periods would decrease the global solar radiation too much in Sweden. From Persson (1999), the turbidity in Lund (latitude 55.72°) was around 0.07 and for Kiruna (latitude 67.83°) around 0.045. Therefore, a dependency on both year and latitude was used to capture this behavior more strongly than in Equation 36:

$$\beta = 0.025 + 0.1 \cdot \cos^2(\varphi) - 0.03\cos\left(2\pi\frac{d_{nr}}{365}\right) \qquad (37)$$

There is a strong tendency that increasing latitude gives larger error than in Flerchinger et al. (2009). This is also valid in this investigation.

The last investigated parameter was the albedo. The formulas for S_0 presented above do not take into account the fact that the Earth's albedo, which is about 0.2, will reflect part of the global solar irradiance back to the atmosphere where a smaller part will be reflected back down again due to the atmosphere albedo, which is about 0.0685 for clear skies, according to Atwater and Ball (1978). Atwater and Ball suggested the following formula:

$$S_0^a = S_0\frac{1}{1 - r_s r_a} \qquad (38)$$

Table 4. Ångström Turbidity from Fox (1994) for Fairbanks Alaska, Latitude 64.82°

Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Annual
0.023	0.024	0.027	0.04	0.059	0.048	0.044	0.028	0.029	0.026	0.027	0.022	0.03

Here, r_s (dimensionless) is the surface (Earth) albedo, r_a (dimensionless) is the atmosphere albedo and S_0^a is global solar irradiance corrected for albedo (W·/m^2·h). The influence of the Earth's albedo is mostly important when there is a snow layer. The albedo can then be around 0.8–0.9.

In the results shown in Table 2, all the following corrections were made in comparison to Flerchinger et al. (2009):

1. improved air mass calculation (Equation 16)
2. improved turbidity calculation (Equations 19 and 37)
3. albedo (with snow cover during December to February) (Equation 38)

This also explains part of the differences in Table 3.

Given all these corrections the equations above do not give a very accurate reproduction of the long-wave radiation on a detailed level. Figures 2 and 3 show a comparison between the presented formula for 45 days and 5 days, respectively, in Borlänge, which shows the variation between the equations.

PRECIPITATION

Precipitation can be measured on an hourly up to a daily basis in Swedish meteorological stations. It is common to measure every 12 h. Blocken and Carmeliet (2008) showed that to properly model the driving rain on a wall, the hourly driving rain should be based on nonarithmetic weighted average of 10 min measurements of rain and wind. Unfortunately this is not usually available data. Most building simulation programs need hourly values, so the rain must be distributed over the previous 12 h by some strategy. Harderup (1988) suggested the use of hourly relative humidity measurements to do this distribution. Other techniques include the use of statistical methods (Günter et al. 2001; Meteonorm handbook). These latter techniques are more aimed at hydrological applications and not, for example, mold estimates in building structures. There is also a risk that purely statistical methods will give conflicting data, for example low humidity and rainfall. As in the case of reconstructing long-wave radiation, it is not possible to accurately reconstruct rain based on 12 h values.

Method

Reconstructed hourly values based on 12 h measurements were compared with hourly measurements from the period of January 10, 1996 to October 27, 1997 in Stockholm Sweden. The distribution strategy that was chosen was that the hour received rain if the relative humidity exceeded a critical value RH_C = 88%. If no hours in the period exceeded this value, the hour with the maximal hour received the rain. This limit value was chosen since it maximized the fitted r value or quality of fit (0.486), as well as minimized the mean absolute hourly error (0.0565 mm/h).

Results

Reconstructed and measured hourly data for five days in June are shown in Figure 4. It is clear that the difference between the reconstructed and measured values is high. The standard deviation for reconstructed and measured data was: 0.198 mm/h and 0.310 mm/h, so the reconstructed data had a lower variation, which can be expected given the simple choice of reconstruction method.

Even if the choice of RH_C minimizes the mean absolute error, it is not a deep minimum. Table 5 shows the r value and mean absolute error for different choices of RH_c.

Figure 5 shows the measured rain versus the relative humidity. It is clear that the reconstruction of the hourly rainfall from 12 h values is a crude process at best.

CONCLUSIONS AND DISCUSSION

The long-wave radiation can be calculated from meteorological data to an accuracy of about 10%, which, incidentally, is the same size as the error reported for a pyrgeometer. The hypothesis that there exists one formula that give the most accurate values cannot be fully confirmed in this study. The existing formula from Kimball et al. (1982), together with Dilley and O'Brian, gave a good fit. This formula has the advantage of using the extraterrestrial global irradiance as reference, which makes it less sensitive to calculation of air mass and turbidity. For Sweden the best formula was a new formula (Clearness identification), which was similar to the one above since it also used the extraterrestrial global solar irradiance as reference. This formula was, however, much more simple. The formula by Aubinet (1994) (Equation 25) gave the worst result and, even when the parameters were identified to fit the Swedish data (Equation 29), had the lowest accuracy. Equation 26, using the global solar irradiance on the Earth surface as a reference, was very sensitive to how the air mass, turbidity, and Earth's albedo were made, especially for latitudes above 60°N. For the Equation where the parameters were identified to the Swedish data, the difference in performance was remarkably low—almost all had a root-mean-square error close to 23.5 Wh/m^2·h.

The reconstruction of rain from 12 h to 1 h values based on relative humidity gave a crude result, even if the chosen limit value for the humidity minimized the r value and mean absolute error. Perhaps estimation of the cloud cover could be included to increase accuracy, with the assumption that rain does not fall from a clear sky.

For the reconstruction of both rain and long-wave radiation to hourly values, the "true" values can never be achieved

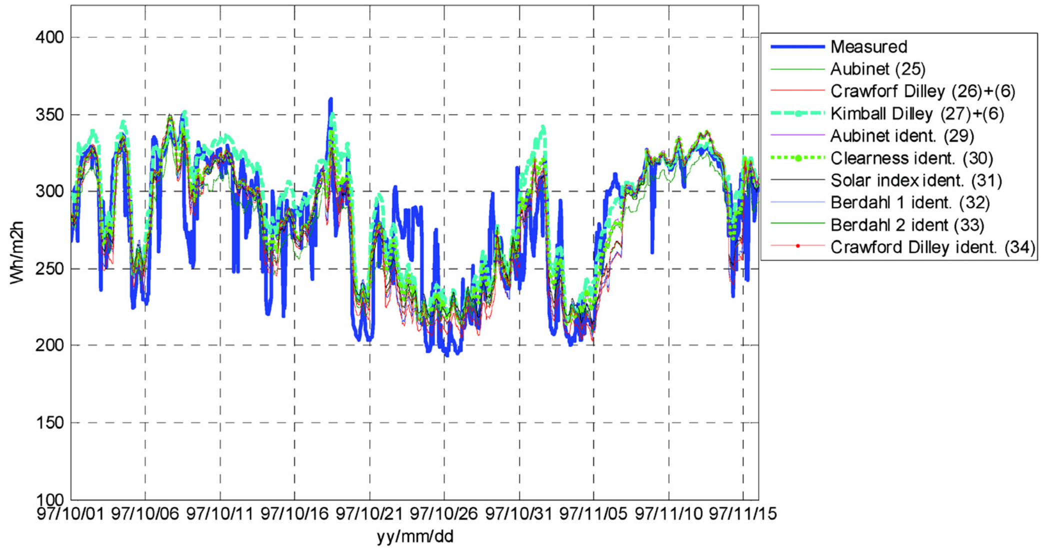

Figure 2 *Measured and calculated long-wave radiation in Borlänge (latitude 60.78°).*

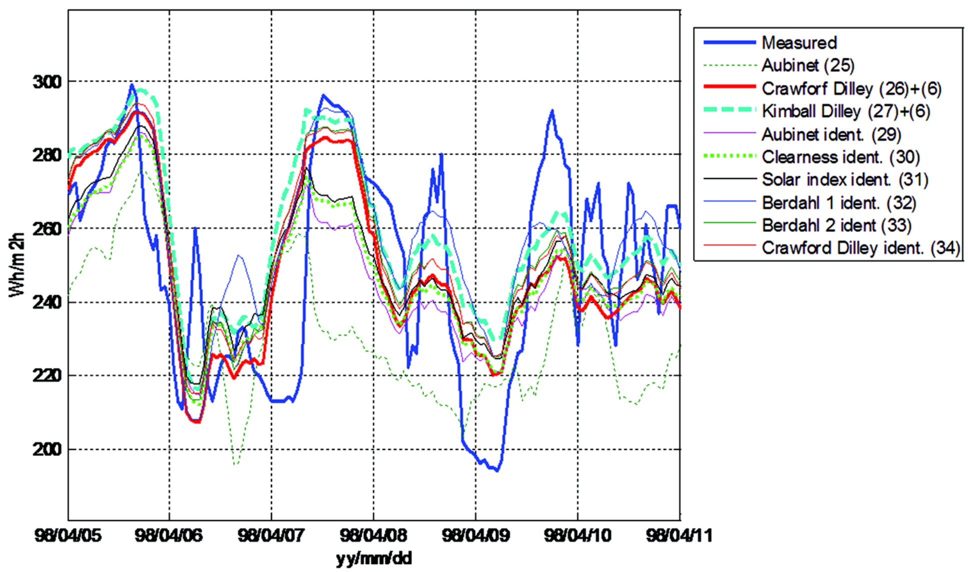

Figure 3 *Measured and calculated long-wave radiation in Borlänge (latitude 60.48).*

since the measured data do not have enough information. One can only create reasonable climate files with these methods. The choice of method should also be dependent on how these climate files would be used, i.e., if the aim is calculating a good average value or if the aim is to simulate worst case scenarios. Worst case for rain and mold can be a lower intensity of rain over a longer period or a short rain with high intensity. The limit factor should be reduced to produce rain with a lower intensity over a longer period and vice versa. Worst case for long-wave radiation can be to underestimate this value. A horizontal surface will then be colder, for example, an attic roof.

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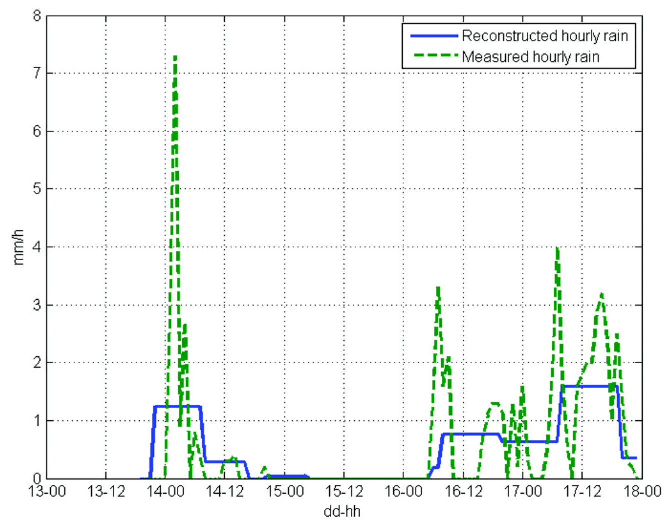

Figure 4 *Reconstructed and measured hourly rain data (mm/h) from June 15, 1996 to June 18, 1996 in Stockholm.*

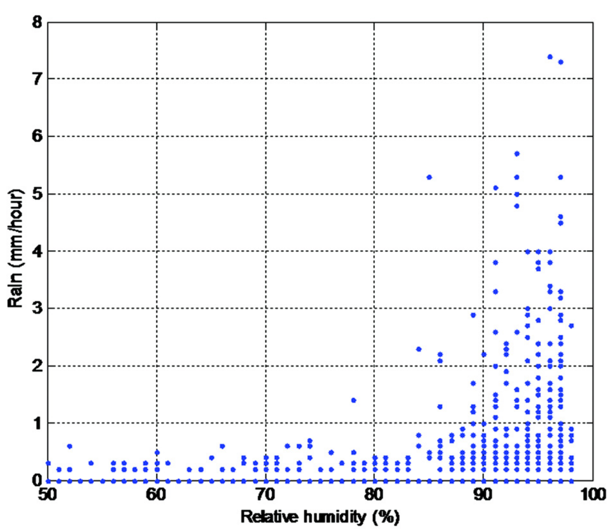

Figure 5 *Measured hourly rain as a function of the relative humidity October 1, 1996 to October 27, 1997.*

NOMENCLATURE

β = Ångström turbidity, dimensionless
δ = declination of sun, rad
ε = emissivity
ϕ = latitude, rad
θ = parameter, ( )
σ = standard deviation, %, or Stefan Boltzmann constant in T^4 law
τ = transmission coefficient

Table 5. The *r* Value and Mean Absolute Error for Some Different RH_c

RH_c, %	*r*, dimensionless	Mean Absolute Error, mm/h
85.0	0.485	0.0583
86.0	0.458	0.0582
87.0	0.421	0.0580
88.0	0.486	0.0565
89.0	0.473	0.0565
90.0	0.412	0.0569
91.0	0.132	0.0591

ω_s = sunset hour angle, rad
c = cloud cover, dimensionless
d_{nr} = day number, 1–365
e_o = vapor pressure, kPa
h = solar height over horizon, rad
H_0 = calculated daily average extraterrestrial radiation, $W{\cdot}h/m^2{\cdot}h$
I = solar irradiance, $W{\cdot}h/m^2{\cdot}h$
I_{SC} = solar constant, 1367 $W{\cdot}h/m^2{\cdot}h$
k = limit for clear and cloudy sky when calculating cloud cover, dimensionless
K_0 = clearness index, dimensionless
L_w = long-wave radiation from the sky
m = air mass
MB = mean bias
P = pressure, kPa
r = goodness of fit, dimensionless
r_a = albedo for atmosphere, dimensionless
r_s = albedo for Earth surface
RH_C = critical relative humidity for rain distribution
RMS = root-mean-square error
s = solar index, dimensionless
S_0 = calculated global solar irradiance on Earth surface, $W{\cdot}h/m^2{\cdot}h$
S_0^a = albedo corrected global solar irradiance on Earth surface, $W{\cdot}h/m^2{\cdot}h$
T = temperature, K, if nothing else stated
T_d = dewpoint temperature, °C
$V()$ = least squares error function
w = precipitable water, mm
z = height over water level, m

Subscripts

a = aerosol
pg = permanent gases

R = Rayleigh scattering

clr = clear sky

cld = cloudy sky

G = global sun

w = water

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