Parameterization of atmospheric long-wave emissivity in a mountainous site for all sky conditions

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Abstract

Long-wave radiation is an important component of the energy balance of the Earth's surface. The downward component, emitted by the clouds and aerosols in the atmosphere, is rarely measured, and is still not well understood. In mountainous areas, the models existing for its estimation through the emissivity of the atmosphere do not give good results, and worse still in the presence of clouds. In order to estimate this emissivity for any atmospheric state and in a mountainous site, we related it to the screen-level values of temperature, relative humidity and solar radiation. This permitted the obtaining of: (1) a new set of parametric equations and (2) the modification of Brutsaert’s equation for cloudy skies through the calibration of $C$ factor to 0.34 and the parameterization of the cloud index $N$. Both fitted to the surface high-resolution data measured at a weather station at a height of 2500 m a.s.l. in Sierra Nevada, Spain. This study analyzes separately three significant atmospheric states related to cloud cover, which were also deduced from the screen-level meteorological data. The validation of the expressions in two alternative sites shows that the superior accuracy in the new 3-state parametric equation is restricted to local use. On the other hand, parameterization of cloud influence in Brutsaert’s equation through the use of screen-level measurements of relative humidity and solar radiation can provide a simple expression to calculate instantaneous atmospheric emissivity of a broader applicability.

1 Introduction

Long-wave radiation has an outstanding role in most of the environmental processes that take place near the Earth's surface (e.g., Philipona, 2004). Radiation exchanges at wavelengths longer than 4 μm between the Earth and the atmosphere above are due to the thermal emissivity of the surface and atmospheric objects, typically clouds, water vapour and carbon dioxide. This component of the radiation balance is responsible for the cooling of the
Earth’s surface, as it closely equals the shortwave radiation absorbed from the sun. The modelling of the energy balance, and, hence, of the long-wave radiation balance at the surface, is necessary for many different meteorological and hydrological problems, e.g., forecast of frost and fog, estimation of heat budget from the sea (Dera, 1992), simulation of evaporation from soil and canopy, or simulation of the ice and snow cover melt (Armstrong and Brun, 2008).

Even though long-wave radiation instrumentation (pyrgeometer) is nowadays usually deployed at weather stations specifically designed for scientific purposes (e.g., Sicart et al., 2006), it is not so common in the most habitual automated weather stations. Hence, all energy balance models estimate long-wave components independently through different physical relations and parameterizations. Downward long-wave radiation is difficult to calculate with analytical methods, as they require detailed measurements of the atmospheric profiles of temperature, humidity, pressure, and the radiative properties of atmospheric constituents (Alados et al., 1986; Lhomme et al., 2007). To overcome this problem, atmospheric emissivity and temperature profile are usually parameterized from screen level values of meteorological variables. The use of near surface level data is justified since most incoming long-wave radiation comes from the lowest layers of the atmosphere (Ohmura, 2001).

It is relatively easy to create parameterizations to estimate emissivity under clear sky conditions. Several studies have compared the performance of different parameterizations over long-wave records (e.g., Sugitia and Brutsaert, 1993; Gabathuler et al., 2001) and for all cloudy sky conditions (Pluss and Omhura, 1996; Crawford and Duchon, 1999; Pirazzini et al., 2000; Kjaersgaard et al., 2007; Sedlar and Hock, 2009, Staiger and Matzarakis, 2010). But only a few of them were carried out on highland sites (Iziomon et al., 2003; Lhomme et al., 2007; Flerchinger et al., 2009). Besides, the effect of clouds and stratification on atmospheric emissivity is highly dependent on regional factors, which may lead to the need for local expressions (e.g., Alados et al., 1986; Barbaro, et al., 2010).

But mountainous catchments are very sensitive areas as they are greatly exposed to meteorological conditions. Here, the surface energy balance has the greatest influence on environmental processes, especially if snow is present. As existing measurements are scarce (e.g., Iziomon et al., 2003; Sicart et al., 2006), a correct parameterization of downward long-wave irradiance under all sky conditions is essential for these areas. Herrero et al. (2009) modelled the energy balance of the snowpack in Sierra Nevada Mountains (Spain), by the Mediterranean sea. Different parameterizations for atmospheric long-wave emissivity (Brunt, 1932; König-Langlo and Augstein, 1994; Prata, 1996) were tested for clear sky periods, and although the best model performance was obtained using Brutsaert (1975) (same as Kimball et
al., 1982; Kustas et al., 1994; Iziomon et al., 2003), the extension to cloudy conditions (e.g. with Crawford and Duchon, (1999)) turned into a global underestimation of incoming long-wave radiation. This underestimation prevented the model from reproducing the different winter snow melting cycles typical of this Mediterranean low-latitude area. This problem was overcome through the use of a simple parameterization for atmospheric emissivity based on 2-yr screen level values of solar radiation, temperature and relative humidity that greatly improved the simulation of the snow cover evolution (Herrero et al., 2009).

In this work, a deeper analysis of long-wave incoming radiation through measurements and its relation to other meteorological data in a high mountain site is presented. From this analysis, a local parameterization for atmospheric emissivity under all sky conditions, based on 5-min surface measurements of relative humidity, temperature, and solar radiation is proposed and validated against direct local measurements. For this purpose, two different approaches were performed: (1) a new empirical expression for Sierra Nevada from 5 yr of surface meteorological data furthering the results in Herrero et al. (2009); (2) a modification of Brutsaert’s equation (Brutsaert, 1982) by means of the parameterization of its cloudiness-related index, $N$.

2 Site description and instrumentation

The study site is the Southern slope of Sierra Nevada Mountain (Fig. 1), located 35 km north from the Mediterranean Sea in Southeastern Spain (37.5º N). This mountain range raises 3500 m a.s.l. and runs parallel to the sea for approximately 60 km. It is characterized by high altitudinal gradients and a heterogeneity produced by a high mountain climate influenced by the surrounding Mediterranean climate. The presence and influence of winter snow becomes important at above 2000 m a.s.l. The snowmelt season generally extends from April to June, even though the mild winter periods characteristic of the Mediterranean climate can melt most of the snow before the end of the snow season (especially during January and February). Typically, several consecutive accumulation/melting cycles take place during one year. Sublimation from the snow can also be very important, up to 40% of year snow precipitation, if the appropriate meteorological conditions prevail (Herrero et al., 2009). Sierra Nevada houses a Spanish National Park and one of the International Global Change Observatories in Mountain Areas because of its particular conditions and delicate environment.

An automatic weather station was operated in Refugio Poqueira (RP Station), at 2500 m a.s.l. (Herrero et al., 2011). Measurements of incoming shortwave and long-wave radiation (Kipp&Zonen SP-Lite pyranometer and CGR3 pyrgeometer), and 2-m air temperature and relative humidity (Vaisala HMP45), among others, have been conducted continuously since
November 2005. The CGR3 pyrgeometer has a spectral range comprised between 4.5 and 44 μm and an accuracy of 5 Wm⁻². A Campbell CR-510 datalogger recorded 5-min averages of 5s sampling rate observations. Additionally, for this study we have used the data recorded by two new weather stations installed in the proximity of RP Station in 2009 that were equipped with downward long-wave sensors: (1) EN2 Station, belonging to the Department of Agriculture, Fishing and Environment of the Regional Government of Andalusia, is located at only 4 km East from RP Station and at 2325 m a.s.l., within the same Southern slope of Sierra Nevada.

Radiation is measured by a NR01 Hukseflux 4-component net radiometer, while temperature and relative humidity are measured by a Vaisala HMP45. Data are recorded at 10-min intervals. (2) Contraviesa Station (C Station) is located 25 km South from RP Station at 1332 m a.s.l., on the ridge of Contraviesa mountain range, which is a lower range parallel to Sierra Nevada. It has the same configuration as RP Station, except from the radiation sensors, which, in this case, are an IR02 pyrgeometer and a LP02 pyranometer, both from Hukseflux.

3 Data analysis

3.1 Long-wave data

After the Stefan-Boltzmann Law for the radiation emission of any body at a temperature $T$ (K), downward long-wave radiation $L^\uparrow$ (Wm⁻²) coming from the near-surface layer of the atmosphere may be written as:

$$L^\uparrow = \varepsilon_a \sigma T_a^4$$  \hspace{1cm} (1)

where $\varepsilon_a$ is the apparent emissivity of the sky (Unsworth and Monteith, 1975), $\sigma$ (Wm⁻²K⁻⁴) is the Stefan-Boltzmann constant, and $T_a$ (K) is the air temperature near the surface (typically 2 m).

The downward long-wave radiation measured for 5 consecutive years at RP Station, converted to $\varepsilon_a$ according to Eq. (1), is shown on Fig. 2a and summarized in the probability density function (pdf) in Fig. 3. The lower values of $\varepsilon_a$ belong to clear sky situations, and in the pdf they smoothly fit a Gaussian with a mean value of 0.68 and a standard deviation of 0.0565. During very clear days, with a low temperature and relative humidity, it exhibits values ranging from 0.5 to 0.6. In the pdf, 0.77 sets the limit between clear sky and partly covered situations; higher values of $\varepsilon_a$ denote the presence of clouds in the atmosphere. A seasonal pattern is easily observed in Fig. 2a, where the lowest emissivity values from clear skies are reached during winter. This emphasizes the importance of long-wave balance for cooling the soil and snow under high mountain clear skies. These measurements are similar to those found by Frigerio
(2004) in Argentina, at 2300 m a.s.l., with night values of atmospheric emissivity of under 0.7 with clear skies. Figure 2b represents daily variation of $\varepsilon_a$, that is, the difference between maximum and minimum daily values. It exhibits a marked seasonality, where wider daily variations of $\varepsilon_a$ in winter are in accordance with wider variations in temperature and relative humidity. Minimum instantaneous values of $\varepsilon_a$ during winter can be as low as 0.4, while in summer they rarely drop to under 0.6.

These measured values are lower than those estimated from the usual empirical expressions, which casts a doubt over the latter for their general use in the highland under any atmospheric state. Thus, the expression by König-Langlo and Augstein (1994), used by Jordan (1999) in the SNTHERM model, gives a minimum value for emissivity of 0.765, much higher than the real values measured in this site. Prata (1996) also overestimates the lower values found under clear skies. Only Brutsaert (1975) gives more realistic values of $\varepsilon_a$ for clear skies, and is capable of reproducing values of below 0.60 during cold days with a clear sky and low relative humidity.

3.2 Parameterizations from screen-level data

From the previous analysis of the data recorded by RP Station, it was found that relative humidity, $W_a$, exhibited more compact relations with $\varepsilon_a$ and $T_a$ than the water vapour pressure, $e_a$. So, despite $e_a$ being the variable commonly used in the calculation of $\varepsilon_a$ for clear skies, $W_a$ was chosen for the parameterizations because it seems to represent the variation in $\varepsilon_a$ better due to the presence of water in the atmosphere at high altitudes. Figure 4a shows the relationship between the measured values of $\varepsilon_a$, $T_a$, and $W_a$ for all sky conditions. That relationship is especially strong for clear and completely covered skies, as shown by the low magnitudes of the standard deviation (std) in Fig. 4b for the values of $\varepsilon_a$ under 0.7 and over 0.9, respectively. Partly covered skies appear as a transition zone between these two boundary situations. There are some differences in these relationships between daytime and night-time values, but they were not found to be significant for these particular data.

In order to evaluate the relationship existing between $\varepsilon_a$ and cloudiness, the Clearness Index CI has been used, as in Sugita and Brutsaert (1993), and equivalent to ratio $s$ in Crawford and Duchon (1999). CI is the ratio between the theoretical shortwave irradiance at the top of the atmosphere (extraterrestrial radiation) and the surface-measured solar radiation. By means of the CI, calculated with the topographical model described in Aguilar et al. (2010), it is possible to find out the degree of opacity of the atmosphere due to the concentration of aerosols and clouds during the hours with sunshine. Figure 5 shows how the states of clear sky (region A) and sky completely overcast (region B) are very well represented in the relation $W_a$-CI-$\varepsilon_a$. The transition area between both regions concentrates the dispersion of the values (a
The region of the completely covered skies has a very high emissivity, of above 0.95. This means that not only are there clouds but also that they are close to the surface, which is common in mountainous areas and the reason why the relative humidity of air is highly correlated with cloudiness.

Thus, a clear sky region (A in Fig. 5a) and a completely overcast region (B in Fig. 5b) were identified from the analyses of the mean values (Fig. 5a) and their std (Fig 5b). These regions were delimited by the following expressions as a function of $W_a$ and $CI$:

Region A: $CI > 0.25 W_a^2 + 0.025 W_a + 0.65$ \hspace{1cm} (2a)
Region B: $CI < 2.667 W_a - 1.867$ \hspace{1cm} (3)

where $W_a$ is expressed as a fraction of one. This partition was made on the basis of the relation between CI, $W_a$ and emissivity as shown in Fig. (5). Region A for clear skies defines the area in a CI-$W_a$ axes, where the mean value for the emissivity is lower than 0.7. Conversely, region B for completely covered skies delimits the area where emissivity is greater than 0.9. It must be emphasized that these two regions include most of the atmospheric states found, since 59% of all the daily states are clear skies and 14% are completely covered skies. The intermediate states correspond to partly cloudy skies or anomalies in the two previous regions, so that it is a zone with a great dispersion in the values of $\varepsilon_a$.

For “clear sky” conditions, the following expression for atmospheric emissivity $\varepsilon_a^{cs}$ was derived from a polynomial fit of the available screen-level measurements at daytime, where the non-significant terms have been neglected:

$$\varepsilon_a^{cs} = -1.17 + 0.16 W_a + 0.0062 T_a$$ \hspace{1cm} (4)

where $W_a$ is expressed again as a fraction of one and $T_a$ in K. In the case of the “completely covered skies”, the emissivity $\varepsilon_a^{ccs}$ does not show any relation to $T_a$ but it does to CI. Therefore, the following parametric function was fitted, the variables being expressed as before:

$$\varepsilon_a^{ccs} = 1 - 1.38 CI + 1.33 W_a CI$$ \hspace{1cm} (5)

For “partly covered skies”, the best fitted expression of the emissivity $\varepsilon_a^{pcs}$ obtained was:
Alternatively, a correction of the Brutsaert equation extended to cloudy conditions (Eq. 7), which had proven to be the expression for emissivity that performed best at this site (Herrero et al., 2009), has been developed. Brutsaert (1982) extended $\varepsilon_a^{cs}$ for all sky conditions by means of a factor $F$:

$$\varepsilon_a = \varepsilon_a^{cs} F = 1.72 \left( \frac{e_a}{T_a} \right)^{1/7} \left( 1 + C N^2 \right) \quad (7)$$

where $e_a$ is the vapour pressure near the surface in kPa, and $F \geq 1$ is the increase in the sky emissivity due to the presence of clouds. This factor is split in $N$, a cloud index varying between 0 for clear skies and 1 for totally overcast skies, and $C$, an empirical factor dependent on the cloud types. Since there are no direct measurements of cloudiness, $N$ has been parameterized using the actual screen-level values of $W_a$ and CI in Eq. (7). This was achieved by comparing measured and simulated $\varepsilon_a$. $C$ was also calibrated in the process, with a value of 0.34 being obtained.

$$N = 1 - 0.45 \ CI - 3.5 \ W_a \ CI + 4 \ W_a^2 \ CI \quad (8)$$

The value of $N$ obtained from Eq. (8) is never allowed to be lower than 0 or greater than 1.

Equations (2) to (8) have been obtained from a calibration dataset composed of all the 5-min data from November, 2004, to December, 2010, including daytime records for any cloudiness degree.

Crawford and Duchon (1999) developed a similar model to the modified Brutsaert equation proposed for $\varepsilon_a$ in Eq. 7 and 8. Also based on Brutsaert (1975), the modelling of the cloudiness relies upon screen-level measurements of temperature, humidity, solar radiation, and, in addition, atmospheric pressure. Their model includes two modifications to the original by Brutsaert (1975): (1) extension to cloudy conditions through a simple linear relation between $\varepsilon_a$ and the ratio of the measured solar irradiance to the clear-sky irradiance, $s$, in fact equivalent to the propagation of CI across the atmosphere; and (2) the substitution of the leading coefficient, $l_c$, (1.72 in Eq. 7) by:

$$l_c = (1.22 + 0.06 \sin[(\text{month}+2) \pi/6]) \times 10^{1/7} \quad (9)$$
where month is the numerical month starting in January (=1). This expression results in a leading factor ranging from 1.78 in January to 1.61 in July. Notice that lc is dimensional so the value of 1.72 in Eq. 7 is defined for $e_a$ in kPa and $T$ in K, this being 1.24 if $e_a$ is in hPa and $T$ in K. This model, CD99, was used for comparison with the two approaches presented so far: the 3-state parametric expressions, 3-sParam, and the modified Brutsaert equation, modB82. Besides, variable leading coefficient was tested in an alternative version of modB82, modB82-var, to assess its validity in the meteorological data from Sierra Nevada.

These four models were tested against the calibration dataset in RP Station and against three validation datasets: (1) 2011 measurements in RP Station, which approximately represent 15% of the whole 5-yr dataset, (2) whole record in C Station (august 2009 – April 2012) and (3) whole record in EN2 Station (October 2009 – March 2012). The goodness of agreement of each model was valued by the common statistics Mean Absolute Error MAE and Root of the Mean Square Error RMSE.

4 Results and discussion

Figure 6 shows the comparison between daytime $e_a$ measurements and values estimated by the different models for the calibration period at RP Station. Figure 7 shows the same comparison but for the validation at C Station, the lower study site. The complete results from the statistical analysis of all four models for the calibration and the three validation datasets are shown in Table 1. There, the results for the complete daytime data for each case along with the separation for each of the three atmospheric states (clear, totally covered and partly cloudy skies) are presented.

The results from the calibration and validation tests at RP Station agree, so calibration is confirmed for this site. The performance of the 3-sParam model stands out over the rest of models, especially for clear and completely covered skies. Partly cloudy skies are also best represented by 3-sParam, even though the differences in this state are lower. The graphical representation of these transition states in Figs. 6 and 7 shows a greater scattering, while measurements and predictions for clear and overcast states clearly fit more tightly. Brutsaert’s equation improves when the variable leading coefficient is used (modB82-var), especially for clear skies. CD99 exhibits an overall good performance, very similar to modB82 and modB82-var models, even though it fails to reproduce higher values of emissivity with completely covered skies. In this atmospheric state, measurements of $e_a$ clearly meet at 1, while CD99 never reaches that value.

The results of the validation at the lower site of C Station show an outstanding loss of performance of the 3-sParam model, particularly for the lower values of emissivity for clear
skies, which are vastly underestimated by this model. The transition state is drawn with much
more scattering for this model (Fig. 7c). For this dataset, the variable leading coefficient in
modBrut82-var and CD99 is much less effective than the constant coefficient, as opposed to
what happened at RP Station. CD99 is also still penalized by its incorrect simulation of higher
emissivities, whose measurements are very close to the unity for this site too. modB82 has
improved substantially for every atmospheric state and exhibits an outstanding performance
(without calibration). \( \varepsilon_a \) measurements are steadier in this lower site compared to what
happened at very high altitudes in RP Station.

Finally, the validation at EN2 Station, located at a very high altitude, displays a very
similar behaviour and statistics for models 3-sParam and modB82 to that found at RP Station,
even though measurements are even more unsteady here than in RP site. However, models
modB82-var and CD99 clearly get worse for all atmospheric states. The variable leading
coefficient makes both models underestimate emissivity for clear skies, while covered skies
with emissivities very close to 1 again are not captured by CD99. 3-stateParam is still the most
efficient model, followed by modB82.

The classification of the data set in 3 atmospheric states, clear, completely covered, and
partly cloudy skies, allows a better adjustment and analysis of the performance of the models.
The highest error is concentrated in the intermediate atmospheric states, those with partial
cloud cover, where the surface measurements are not capable of representing by themselves the
complex state of the atmosphere and the presence of clouds and aerosols in it.

From Fig. 6, it can be seen that the lowest values for measured \( \varepsilon_a \) at RP Station, those
between 0.4 and 0.5, are grouped in a scattered cloud of points with an estimated value
between 0.6 and 0.7. They are overestimated by all the models. In fact, these measurements are
taken under similar atmospheric states, corresponding to sunny winter days with low wind
speeds (< 1m s\(^{-1}\)), and this overestimation may be caused by the overheating of the
pyrgeometer dome by solar radiation under insufficient ventilation. This effect has already
been reported (e.g. Weis, 1981), but it is normally not accounted for as the induced errors are
low (Lhomme et al., 2007). However, in this work the errors in measured long-wave radiation
may be important for these specific meteorological conditions, with an absolute overestimation
in measured \( \varepsilon_a \) up to 0.2.

A \( C \) coefficient in the extended Brutsaert equation (Eq. 7) of below 0.34 prevents the
high values of \( \varepsilon_a \) which are measured in very cloudy states, from being reached by models
modB82 and modB82-var. This is a much higher value than the 0.22 originally proposed by
Brutsaert (1982). This reflects the fact that, in mountainous areas, the interaction of the clouds
with the surface of the terrain and, therefore, their effect on $\varepsilon_a$ is much more intense than in valley areas.

Clear sky data are well predicted in this mountainous site using the original coefficient of 1.72 in Eq. (7) suggested by Brutsaert (1975). The seasonally variable leading coefficient suggested by Crawford and Duchon (1999) (Eq. 9) causes the Brutsaert equation to underestimate emissivity more than its original formulation in two of the three tested sites, which is the same result found by Kjaersgaard et al. (2007). Consequently, there was no need to correct this coefficient, as was already pointed out by Flerchinger et al (2009).

5 Conclusions

The high resolution long-wave measurements recorded in a weather station at an altitude of 2500 m in a Mediterranean climate are not correctly estimated by most of the existing models and frequently used parameterizations. These measurements show a very low atmospheric emissivity for long-wave radiation values with clear skies (up to 0.5) and a great facility for reaching the theoretical maximum value of 1 with cloudy skies. Despite the good behaviour of Brutsaert (1975) for clear skies, the cloudiness effect considered in Brutsaert (1982) cannot be effectively added because of the lack of any cloud index $N$ measurements. The relationships between the screen-level values of temperature, relative humidity, and solar radiation by means of the clearness index with the emissivity under clear and cloudy skies, allows one to define two parametric approaches with good results and a different applicability for estimations of the instantaneous values of the atmospheric emissivity: (1) a complete parametric expression, split into three atmospheric states parametrically regionalized (clear, completely covered and partly covered skies), with an outstanding performance at a very local scale even with the unsteady measurements at high altitude mountainous sites; and (2) a modification of Brutsaert (1982) by means of a parameterization of $N$ from the screen level measurements of humidity and solar radiation and a calibration of $C$ index, set to 0.34. This model has proven to have an overall good performance for all atmospheric states and, more important, a broader scope of applicability at different sites without further calibration.

The use of a seasonally variable leading coefficient for clear sky emissivity in Brutsaert (1972), as proposed by Crawford and Duchon (1999), was rejected because it underestimates emissivity for clear skies.

As a result, it is now possible to obtain atmospheric emissivity series in stations without any long-wave direct measurements, with a direct applicability in the surroundings of Sierra Nevada. Complete parametric expressions should have, in general, a very local scope of applicability, as the validity of these fits is linked to their ability to characterize the state of the
atmosphere, with regard to the presence of clouds, only with surface measurements of
temperature, humidity, and solar radiation.

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Table 1. Summary of the goodness of agreement for the new 3-state parameterization (3-sParam, Eqs. (4) to (7)), the modified Brutsaert’s equation (modB82, Eqs. (7) and (8)), the same modB82 with a variable leading coefficient (modB82, with Eq (9)) and Crawford and Duchon (1999) (CD99) for different atmospheric states for the calibration and validation datasets. MAE: Mean Absolute Error; RMSE: Root Mean Square Error.

<table>
<thead>
<tr>
<th>Atmospheric state</th>
<th>3-sParam MAE/RMSE</th>
<th>modB82 MAE/RMSE</th>
<th>modB82-var MAE/RMSE</th>
<th>CD99 MAE/RMSE</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Calibration. RP Station (Nov2004-Dec2010)</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Daytime. All data</td>
<td>0.045/0.066</td>
<td>0.060/0.078</td>
<td>0.056/0.076</td>
<td>0.058/0.080</td>
</tr>
<tr>
<td>- Clear skies</td>
<td>0.037/0.055</td>
<td>0.058/0.073</td>
<td>0.049/0.069</td>
<td>0.049/0.069</td>
</tr>
<tr>
<td>- Covered skies</td>
<td>0.025/0.040</td>
<td>0.042/0.057</td>
<td>0.042/0.056</td>
<td>0.069/0.084</td>
</tr>
<tr>
<td>- Partly cloudy</td>
<td>0.070/0.092</td>
<td>0.075/0.096</td>
<td>0.077/0.096</td>
<td>0.075/0.098</td>
</tr>
<tr>
<td><strong>Validation. RP Station (Jan2011-Dec2011)</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Daytime. All data</td>
<td>0.049/0.068</td>
<td>0.070/0.087</td>
<td>0.064/0.084</td>
<td>0.065/0.086</td>
</tr>
<tr>
<td>- Clear skies</td>
<td>0.045/0.062</td>
<td>0.073/0.088</td>
<td>0.061/0.082</td>
<td>0.060/0.081</td>
</tr>
<tr>
<td>- Covered skies</td>
<td>0.031/0.048</td>
<td>0.048/0.062</td>
<td>0.056/0.067</td>
<td>0.078/0.095</td>
</tr>
<tr>
<td>- Partly cloudy</td>
<td>0.067/0.087</td>
<td>0.075/0.094</td>
<td>0.077/0.095</td>
<td>0.071/0.094</td>
</tr>
<tr>
<td><strong>Validation. C Station (Aug2004-Apr2012)</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
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<tr>
<td>Daytime. All data</td>
<td>0.071/0.084</td>
<td>0.041/0.054</td>
<td>0.052/0.065</td>
<td>0.053/0.067</td>
</tr>
<tr>
<td>- Clear skies</td>
<td>0.084/0.092</td>
<td>0.041/0.050</td>
<td>0.059/0.068</td>
<td>0.057/0.068</td>
</tr>
<tr>
<td>- Covered skies</td>
<td>0.027/0.038</td>
<td>0.026/0.040</td>
<td>0.024/0.039</td>
<td>0.049/0.064</td>
</tr>
<tr>
<td>- Partly cloudy</td>
<td>0.072/0.087</td>
<td>0.047/0.062</td>
<td>0.054/0.069</td>
<td>0.049/0.066</td>
</tr>
<tr>
<td><strong>Validation. EN2 Station (Oct2009-Mar2012)</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Daytime. All data</td>
<td>0.043/0.055</td>
<td>0.060/0.077</td>
<td>0.075/0.088</td>
<td>0.074/0.088</td>
</tr>
<tr>
<td>- Clear skies</td>
<td>0.041/0.053</td>
<td>0.049/0.060</td>
<td>0.067/0.076</td>
<td>0.068/0.077</td>
</tr>
<tr>
<td>- Covered skies</td>
<td>0.024/0.033</td>
<td>0.049/0.063</td>
<td>0.058/0.068</td>
<td>0.070/0.081</td>
</tr>
<tr>
<td>- Partly cloudy</td>
<td>0.057/0.069</td>
<td>0.092/0.111</td>
<td>0.103/0.119</td>
<td>0.090/0.113</td>
</tr>
</tbody>
</table>
**Fig. 1.** Location of Sierra Nevada in Andalusia, Spain, and weather stations on Southern slope used.

**Fig. 2.** Atmospheric emissivity measured at RP station from 2005 to 2011. (a) Complete dataset with 5-min frequency and the 5-weeks moving average in white. (b) Daily variation (difference between maximum and minimum daily values).

**Fig. 3.** Pdf of the atmospheric emissivity 5-min values from 2005 to 2011 with a Gaussian fit for clear sky conditions, b exponential fit for completely covered data and c residual corresponding to partly covered sky situations.
Fig. 4. (a) Mean value and (b) standard deviation for relative humidity $W_a$ measurements as a function of temperature $T_a$ and atmospheric emissivity $\varepsilon_a$.

Fig. 5. (a) Mean value and (b) standard deviation for atmospheric emissivity measurements as a function of CI and $W_a$.

Fig. 6. Atmospheric emissivity measurements versus estimation obtained for the calibration at RP Station (2500 m a.s.l.) using the four different models.
Fig. 7. Atmospheric emissivity measurements versus estimation obtained for the validation at C Station (1332 m a.s.l.) using the four different models.