Tethys
Journal of Mediterranean Meteorology & Climatology

Tethys 2010, **7**, 67–74
www.tethys.cat
ISSN-1697-1523
eISSN-1139-3394
DOI:10.3369/tethys.2010.7.06
Journal edited by **ACAM**

Determination of the surface temperature by remote sensing

V. García-Santos, E. Valor and V. Caselles

Earth Physics and Thermodynamic Department, Universitat de València, C/ Dr. Moliner, 50, 46100 Burjassot, València

Received: 6-X-2009 – Accepted: 16-V-2010 – **Translated version**

Correspondence to: vicente.garcia-santos@uv.es

Abstract

The knowledge of land surface temperature (LST) is of great importance, because it helps us to understand processes such as energy exchange between surface and atmosphere, water requirements in agricultural soils, control and prevention of fires and the evolution of climate change. Also, we need to try to know its value with enough accuracy. Two important factors are taken into account: the surrounding environmental conditions on the surface and its own emissivity. When performing a measurement of the LST, two types of corrections should be applied: first, the atmospheric correction in order to eliminate its contribution to the satellite measurements and a second one due to the effect of surface emissivity when the measurement is done both from satellite and in the field. This work presents an exhaustive review of the methodology currently used to perform both corrections. In the case of surface emissivity, the techniques known to determine it and the methodologies used for its correction will be shown. For atmospheric correction, two widely contrasted methods are exposed: the single-channel method and the differential absorption of the split-window method, which avoid the need for radiosoundings, in order to correct the radiative effect of the atmosphere. The knowledge of the methodology shown can be of help for any study of the LST, either from satellite or on ground level.

Key words: temperature, emissivity, atmospheric correction, thermal infrared

1 Introduction

The temperature of all land surfaces (LST) is above absolute zero (T>0 K) and they therefore emit radiation according to Planck's law. If we consider the areas we are studying to have the radiative behavior of a black body, when measuring the radiation from the surface with a radiometer in the spectral interval called thermal infrared (TIR) interval, where there is maximum radiation of land surfaces, we could use that energy to reverse the Planck function and obtain the real LST, given that we are certain that the measured radiation is emitted only by the surface. In fact, almost no real surface can be considered a black body, and that the radiation measured on all of them has a double contribution: first, the radiation directly emitted by the surface, and on the other hand, the radiation reflected in it, coming from the surroundings and the atmosphere.

The radiance of a surface that reaches a sensor aboard a satellite has to go through the atmosphere that separates both. When a sensor aboard a satellite receives radiation from a surface, the effect that the atmosphere has on this radiation must be taken into consideration. Energy can usually undergo two phenomena of attenuation:

- Absorption due to the presence of components such as O_2, CO_2 and especially water vapor (H_2O). This absorbed energy is then re-emitted as electromagnetic radiation but at a different wavelength.
- Scattering of incident radiance due to its interaction with gaseous atmospheric components and aerosols. This process involves a change in direction of the incident radiance and does not involve a transfer of energy.

The atmosphere contributes not only by attenuating the radiance that reaches the sensor aboard the satellite, but also provides a certain amount of radiation as a direct result of the absorption process of the radiance from the surface. This is clearly understandable recalling the Planck law as the atmospheric temperature, like in any other element, is above 0 K and thus emits radiation that can be recorded by the sensor.

The radiance measured by the sensor aboard a satellite is different to that emitted by the surface. From the satellite measurement an apparent or brightness temperature is obtained, which must be corrected both from the effect of the atmosphere and the one of emissivity. These radiance measurements are performed at some intervals of wavelength, called atmospheric windows, where the radiance suffers less atmospheric attenuation. The possible absorption obtained here is almost entirely due to the H_2O. These windows are located mainly in the TIR between 3.7-4.1 μm, 8-9.5 μm and 10-12.5 μm.

To summarize, the radiance received by a sensor aboard a satellite is described by the following equation of balance, called radiative transfer equation (RTE):

$$L_\lambda(\theta) = \{\varepsilon_\lambda(\theta)L_\lambda^\circ(LST) + [1-\varepsilon_\lambda(\theta)]L_{atm,\lambda}^{\downarrow}\}\tau_\lambda(\theta) + \\ + L_{atm,\lambda}^{\uparrow}(\theta) \quad (1)$$

where $L_\lambda(\theta)$ is the radiance directly measured from a surface, λ is the wavelength, θ is the observation zenith direction of the surface, ε_λ is the surface spectral emissivity, $L_\lambda^\circ(LST)$ is the Planck's radiance of the surface, for a LST temperature, $L_{atm,\lambda}^{\downarrow}$ is the downward atmospheric radiance of the entire dome that covers the surface, τ is the transmissivity of the atmosphere and $L_{atm,\lambda}^{\uparrow}$ is the atmospheric radiance in the direction of the satellite sensor.

The term in curly brackets in Equation 1 is the radiance only from the surface and shows the importance of knowing the exact value of the emissivity of a surface, since errors in its determination can lead to significant errors in obtaining its temperature.

We can understand the concept of emissivity of a surface as an indicator of how good an emitter this is, that is to say, how close it is to a black body. Thus the emissivity is defined as:

$$\varepsilon_\lambda(\theta) = \frac{L_\lambda(T)}{L_\lambda^\circ(T)} \quad (2)$$

An added disadvantage that most surfaces have is their structural and composition complexity, that is to say, they show in all their extension a variability of elements and structures that make them heterogeneous, making it difficult to evaluate what is the relative weight of each of the elements that compose the surface. To interpret their radiative behavior, it is necessary to define an effective emissivity according to their characteristics. Another key issue is the problem of emissivity measurement by remote sensing, as a single measure of radiance is simultaneously influenced by both the emissivity and the LST, these being independent. This raises one of the basic problems of remote sensing of the TIR: how to break the indeterminacy mentioned by obtaining emissivity and temperature simultaneously (see 3.2).

On the other hand, the role of the atmosphere in the balance of energy exchange with the surface is essential, and knowledge and study of it is an important field of research in remote sensing. Observing the RTE at the height h of a sensor according to the Schwarzchild equation:

$$L_\lambda(h) = L_\lambda(0)\tau_\lambda(\theta,h,0) + \int_z^h L_\lambda^\circ(T_z)\frac{\partial\tau_\lambda(\theta,h,z)}{\partial z}dz \quad (3)$$

where $L_\lambda(h)$ is the radiance reaching the sensor located at a height h, $L_\lambda(0)$ the radiance at surface level, the term in curly brackets in Equation 1 and $\tau_\lambda(\theta,h,0)$ the transmissivity of the atmosphere from the height z to the position h defined as:

$$\tau_\lambda(\theta,h,0) = \exp\left[-\int_z^h \frac{\kappa_\lambda(z')\rho(z')}{\cos\theta}dz'\right] \quad (4)$$

In the right hand side of Equation 3 the two terms represent the fraction of surface radiation transmitted through the atmosphere and the fraction of radiation that the atmosphere emits to the sensor, respectively. It should be mentioned that in Equation 4, the argument of the exponent is characterized by the absorption coefficient (κ_λ) and the concentration of H_2O in the atmosphere (ρ). The optical thickness (u_λ), that denotes a section of the atmosphere (from a point z to the height h where the sensor is located) and determines how opaque the atmospheric element to radiation is, defined as:

$$u_\lambda = \int_z^h \kappa_\lambda\rho dz \quad (5)$$

To perform the atmospheric correction in the radiance measured by the sensor it is necessary to know the absorbing properties of the H_2O content in the atmospheric windows, information that is achieved by evaluating the coefficient of the continuous absorption of water vapor (French et al., 2003).

This paper reviews the concept of emissivity and the effect of the atmosphere from the point of view of thermal remote sensing. The following sections describe various techniques to determine emissivity and the methods to correct it (Section 2 and 3) and how to correct the atmospheric effect (Sections 4 and 5).

2 "In situ" measured emissivity

2.1 The box method

One way to obtain the emissivity of a surface is to isolate the sample from the surrounding environment through a box, eliminating the contribution of the environmental radiance to the measurement of the radiance from the surface. The inside of the box is covered with mirror of polished aluminum, with a very low emissivity ($\varepsilon_c \approx$ 0.03) and the outside is fully thermally isolated. Two lids are used, a hot cover with high emissivity ($\varepsilon_h \approx$ 0.98) and a cold lid, also made of aluminum, with a difference of temperature between the two of 20 K. A third lid is used to be substituted, in its due course, by the sample. The measurement configuration is: cold lid-sample, hot lid-sample, hot lid-cold lid and cold lid-cold lid.

2.1.1 Cold lid-sample

The sample is almost perfectly isolated from the outside, entirely measuring the radiance of the full sample (L^1), that will have black body behavior.

$$L^1 = L^\circ(T_m) \tag{6}$$

where T_m is the sample temperature.

2.1.2 Hot lid-sample

The sample is isolated from external environmental radiation, receiving only what is emitted by the hot lid, and therefore the radiance measured by the sensor will have a double contribution: the radiance directly emitted by the ground, that this time will be weighed by its emissivity (ε_s) and that radiance reflected by the ground from the lid weighed by the term (1 - ε_s), given that the surface is opaque (τ = 0) and its absorption is equal to its emissivity (Kirchhoff's law). Thus, the radiance becomes:

$$L^2 = \varepsilon_s L^\circ(T_m) + (1 - \varepsilon_s)L^\circ(T_c) \tag{7}$$

where T_c is the radiative temperature of the hot cover.

2.1.3 Hot lid-cold lid

It is intended to turn into a black body the emission of the hot lid, as in 2.1.1, but instead of the sample, the only transmitter will be the lid. Thus, its radiance, after being reflected on the sides and the base, where the sample has been replaced by the aluminum lid with no hole, will reach the sensor. The remaining expression is:

$$L^3 = L^\circ(T_c) \tag{8}$$

The three equations above lead to an expression of the ground emissivity. This is the expression that Conaway and van Bavel (1967) and Dana (1969) achieved from the initial method proposed by Buettner and Kern (1965). The idea is to replace the measurements 2.1.1 and 2.1.3 in the radiance of the 2.1.2 configuration, reaching:

$$L^2 = \varepsilon_s L^1 + (1 - \varepsilon_s)L^3 \tag{9}$$

from where it is easy to get to:

$$\varepsilon_s = \frac{L^2 - L^3}{L^1 - L^3} \tag{10}$$

In fact the box does not have ideal behavior, as it has been indicated (ε_c = 0.03 and ε_h = 0.98), and therefore Equation 10 must be modified with a correction factor:

$$\varepsilon_s = \frac{L^2 - L^3}{L^1 - L^3} + \delta\varepsilon_s \tag{11}$$

2.1.4 Cold lid-cold lid

The objective of this configuration is to obtain the correction factor that allows obtaining a realistic value of the sample emissivity. What is measured is the radiance contribution of the walls and covers made of a mirror of polished aluminum, and the influence of the geometry of the box and the emissivity of the cold lids. This measurement is the corresponding radiance to the Planck function at the temperature of the aluminum sheets (T_f).

$$L^4 = L^\circ(T_f) \tag{12}$$

Combining the above expressions, Rubio et al. (1997) got the equation:

$$\varepsilon_s = 1 - \frac{(L^2 - L^1)(1 - \varepsilon_c)}{L^3 - L^1 - (L^3 - L^2)P + (L^1 - L^4)Q} \tag{13}$$

where ε_c is the known emissivity of the cold lid. P and Q are factors that depend on the geometry of the box and the emissivities of the cold lid and the hot lid, and the following values are obtained: P = 0.01460 and Q = 0.2921.

3 Measurement of the emissivity from satellite sensors

3.1 Measurement of the emissivity by the vegetation cover method

Natural land surfaces are usually heterogeneous and rough, composed of various elements with different properties and characteristics. When performing a measurement of radiance from the surface as a whole, what is obtained is the temperature and effective emissivity of all elements. Emissivity is expressed as:

$$\varepsilon = \varepsilon_0 + d\varepsilon \tag{14}$$

where ε_0 is the emissivity of the radiance directly emitted by the surface towards the sensor and $d\varepsilon$ is the emissivity due to the indirect radiance emitted by the surface due to reflections between floor and walls of the roughness. It is called the cavity effect.

In many situations the studied areas are composed of a given density of vegetation. Such coverage exerts the role of ceiling and walls of the surface roughness and the ground tends to be relatively homogeneous. It also has a geometric shape that will be responsible for the cavity effect (Colton, 1996). The elements of the first term on the right on Equation 14 can be redefined as:

$$\varepsilon_0 = \varepsilon_v P_v + \varepsilon_g(1 - P_v) \tag{15}$$

where ε_g and ε_v are the emissivities of ground and vegetation, that are to be measured independently, and P_v is the proportion of vegetation cover.

The cavity term is:

$$\begin{aligned} d\varepsilon = {} & (1 - \varepsilon_g)\varepsilon_v F(1 - P_v) + \\ & + P_g(1 - \varepsilon_v)(\varepsilon_g G + \varepsilon_v F') \end{aligned} \tag{16}$$

where P_g is the proportion of ground and F, G and F' are the form factors of the vegetation cover (Colton, 1996).

The complexity of the cavity term (Equation 16) makes difficult to use practically. To fix this, Valor and Caselles (1996) developed a model based on a simple idea suggested by van de Griend and Owe (1993), which linked the normalized difference vegetation index (NDVI, Rouse et al., 1974) with emissivity. An expression of the emissivity depending on the vegetation cover was obtained in the form:

$$\varepsilon = \varepsilon_v P_v + \varepsilon_g (1 - P_v) + 4 < d\varepsilon > P_v (1 - P_v) \tag{17}$$

where $< d\varepsilon >$ is the maximum value of the cavity term of the average vegetation structure. The terms ε_v, ε_g and $< d\varepsilon >$ are specified for each type of surface, and P_v is the parameter that defines the percentage of coverage vegetation from NDVI (Valor and Caselles, 1996):

$$P_v = \frac{1 - \frac{i}{i_g}}{1 - \frac{i}{i_g} - \kappa \left(1 - \frac{i}{i_v}\right)} \tag{18}$$

where i is the NDVI of the whole surface (ground and vegetation), i_g that of the ground and i_v that of the vegetation. This is called the vegetation cover method (VCM) and requires some knowledge of the geometry of the surface.

3.2 Decoupling between emissivity and temperature

In the measurement of the radiance of a surface (term in curly brackets in Equation 1), there is an inherent coupling between emissivity and temperature. Gillespie (1986) proposed a method that allowed their separation, the NEM method (Normalized Emissivity Method), whose theoretical basis is:

- A sensor with n-spectral channels (point 4 explains the concept of spectral channel), will measure a radiance L_i for each channel i from the surface according to Equation 1. Gillespie introduced a constant effective emissivity close to the real constant known as ε_{NEM}.
- In Equation 1 the downward atmospheric radiance $L^{\downarrow}_{atm}$ is also known. Thus the term in curly brackets in that equation is cleared and inverted to obtain the temperature in each channel i:

$$T_i = L^{0-1}_{\lambda} \left[\frac{L_i(\theta) + (\varepsilon_{NEM} - 1) L^{\downarrow}_{atm}}{\varepsilon_{NEM}} \right] \tag{19}$$

  Once the temperatures for each channel are calculated, the maximum value for each channel (T_{max}) is chosen, as it is considered the closest to the actual surface temperature value.
- Returning to the curly bracketed term in Equation 1, this time clearing the emissivity term, the value of T_{max} is introduced as a constant value, thus calculating n emissivities, one for each spectral channel. If ε_{NEM} matches with the actual ε_{max} surface then the ε_i and T_{max} obtained are correct, otherwise the spectral variation of emissivity (spectral curve) is correct but not the amplitude or the T_{max} of the surface.

In this way it is possible to obtain spectral emissivity values for each channel and to know the temperature value (T_{max}). This method provides a good estimate of the spectral variation of emissivity; however, it does not always correctly set the position of the spectrum. Gillespie et al. (1998) developed a new method that is a bit more complex than the NEM, called TES (Temperature Emissivity Separation), which is developed in modules, with the first being the NEM.

To solve the problem of the spectral displacement of emissivity, an improvement to the NEM is applied through a new method called ANEM (Adjusted Normalized Emissivity Method), Coll et al. (2003). The aim is to choose the initial value of the emissivity (ε_{NEM}) properly for it to be close to the ε_{max} value in each type of surface. To do this:

- A map of $\varepsilon = \varepsilon_{max}$ for the entire image of the studied area is previously generated by the method of Valor and Caselles (1996), the VCM, Equation 17:

$$\varepsilon = \varepsilon_{VCM} \tag{20}$$

- This emissivity map is the map introduced as emissivity in the NEM, $\varepsilon_{NEM} = \varepsilon_{VCM}$ in the i point, and from here the NEM is developed as before.

4 Single-channel method

Once the radiative behavior of the atmosphere is known, its contribution to the energy measured by a satellite sensor can be eliminated, thus obtaining the energy emitted by the surface, and from this the LST. This is the atmospheric correction by the single-channel method, which requires the calculation of transmissivity and atmospheric radiance (see Equation 1) through the knowledge of the vertical profiles of temperature and humidity obtained by radiosounding of the area and introduced in a radiative transfer model (RTM). This is usually a drawback of this method, as radiosoundings of the area and the time of the sensor passing are not always available.

The atmospheric correction by the single channel starts in Equation 1, specifically focusing on the term in curly brackets:

$$\varepsilon_{\lambda} L^{\circ}_{\lambda}(LST) + [1 - \varepsilon_{\lambda}] L^{\downarrow}_{atm,\lambda} \tag{21}$$

The magnitude $L^{\downarrow}_{atm,\lambda}$ is defined as:

$$L^{\downarrow}_{atm,\lambda} = \int_0^{2\pi} d\varphi \int_0^{\pi/2} L^{\downarrow}_{\lambda}(\theta') \sin\theta' \cos\theta' d\theta' \tag{22}$$

The atmospheric radiance emitted upwards is already defined in the second term of Equation 1, which, as previously discussed, is the atmospheric contribution to the radiance recorded by satellite. It is defined as:

$$L^{\uparrow}_{atm,\lambda}(\theta) = \int_0^h L^{\circ}_{\lambda}(T_z)\frac{\partial \tau_{\lambda}(\theta,h,z)}{\partial z}dz \tag{23}$$

In reality, sensors aboard satellites perform radiometric measurements in spectral channels of a certain width, characterized by a filter function, $f_i(\lambda)$. The signal recorded by the sensor is the following:

$$L_i = \int_0^{\infty} f_i(\lambda)L_{\lambda}d\lambda \tag{24}$$

where L_i is the channel radiance obtained from the convolution of $f_i(\lambda)$ with the monochromatic radiance L_{λ} measured by the sensor.

After applying the new terms and transforming them into channel measurements according to Equation 24, Equation 21 becomes:

$$\begin{aligned} L^{\circ}_i &= L^{\circ}_i(T_i) = \tau_i(\theta,h,0)[\varepsilon_i L^{\circ}_i(LST) + (1-\varepsilon_i)L^{\downarrow}_{atm,i}] + \\ &+ L^{\uparrow}_{atm,i}(\theta) \end{aligned} \tag{25}$$

For the sensor this radiance is a Planck function, because according to it the whole surface-atmosphere radiative whole behaves like a black body radiating at a T_i temperature, which is not the intended LST. This equation shows the relationship between brightness temperature (T_i) measured by the channels of the sensor on the satellite and the LST. In the case of heterogeneous and rough surfaces, we must remember that ε_i and T_i would be effective emissivity and temperature.

A RTM that can be used is MODTRAN 4.0 (Berk et al., 1999) to estimate the atmospheric magnitudes that appear in Equation 25. This model requires, as input, vertical profiles of temperature and humidity obtained by radiosounding. This equation forms the basis of the single-channel atmospheric correction method, so it is possible to obtain LST from T_i through inversion.

We can rewrite Equation 25 as follows:

$$L^{\circ}_i(T_i) = L^{\circ}_i(LST) - \Delta I_{ai} - \Delta I_{ei} \tag{26}$$

with ΔI_{ai} as the attenuation of the radiance by the effect of atmospheric absorption and ΔI_{ei} the radiance decreasing due to the effect of emissivity. Their expressions are (Caselles et al., 1991):

$$\Delta I_{ai} = L^{\circ}_i(LST)[1-\tau_i(\theta,h,0)] - L^{\uparrow}_{atm,i}(\theta) \tag{27}$$

$$\Delta I_{ei} = (1-\varepsilon_i)\tau_i(\theta,h,0)[L^{\circ}_i(LST) - L^{\downarrow}_{atm,i}] \tag{28}$$

In the TIR, the Planck function is approximately linear with the temperature, hence we can develop to first-order the Taylor series of the above expression around LST.

$$L^{\circ}_i(T) = L^{\circ}_i(LST) + \left(\frac{\partial L^{\circ}_i(T)}{\partial T}\right)_{LST}(T - LST) \tag{29}$$

an acceptable approximation if $T - LST \leq$ 10-15 K. Using Equation 12 with $T = T_i$ in Equation 25 we get:

$$\begin{aligned} &L^{\circ}_i(LST) + \left(\frac{\partial L^{\circ}_i(T)}{\partial T}\right)_{LST}(T_i - LST) = \\ &= L^{\circ}_i(LST) - \Delta I_{ai} - \Delta I_{ei} \end{aligned} \tag{30}$$

an expression that slightly modified leads to:

$$LST - T_i = \frac{\Delta I_{ai}}{\left(\frac{\partial L^{\circ}_i(T)}{\partial T}\right)_{LST}} + \frac{\Delta I_{ei}}{\left(\frac{\partial L^{\circ}_i(T)}{\partial T}\right)_{LST}} \tag{31}$$

The expression of atmospheric correction by single-channel approach is obtained in terms of temperatures, the first addend of the second term represents the atmospheric correction due to atmospheric absorption and the second is the emissivity correction. Note that when the sensor gives an apparent temperature T_i it is possible to correct the temperature and obtain the actual temperature of the surface by using the approach:

$$LST = T_i + \frac{\Delta I_{ai}}{\left(\frac{\partial L^{\circ}_i(T)}{\partial T}\right)_{LST}} + \frac{\Delta I_{ei}}{\left(\frac{\partial L^{\circ}_i(T)}{\partial T}\right)_{LST}} \tag{32}$$

4.1 Single-channel correction equation

From Equation 25 it is possible to derive an expression that directly relates atmospheric correction and emissivity, therefore obtaining the temperature (T_i) provided by the sensor, the LST.

If we focus first on the atmospheric correction, as a first step we define a brightness temperature at surface level (T^*_i) corresponding to a radiance associated with the term in square brackets in Equation 25.

$$L^{\circ}_i(T^*_i) = \varepsilon_i L^{\circ}_i(LST) + (1-\varepsilon_i)L^{\downarrow}_{atm,i} \tag{33}$$

where $L^{\downarrow}_{atm,i}$ is the downward atmospheric radiance throughout the hemisphere, which can be defined as follows:

$$L^{\downarrow}_{atm,i} = L^{\circ}_i(T^{\uparrow}_a)(1-\tau_i) \tag{34}$$

where $T^{\uparrow}_a$ is the average value of the atmospheric temperature in the upward direction (McMillin, 1975) and τ_i the total atmospheric transmittance.

By replacing Equation 34 in Equation 33 and then this in Equation 25, we obtain the expression of temperature corresponding to the atmospheric correction term:

$$T^*_i - T_i = \frac{1-\tau_i}{\tau_i}(T_i - T^{\uparrow}_a) \tag{35}$$

By linearizing the Planck function in Equation 33, the emissivity correction complementary to Equation 35 is obtained:

$$LST - T^*_i = \frac{1-\varepsilon_i}{\varepsilon_i}b_i \tag{36}$$

where b_i is a parameter with temperature dimensions given by Coll and Caselles (1997):

$$b_i = \frac{T_i^*}{n_i} + \gamma_i \left(\frac{n_i - 1}{n_i} T_i^* - T_a^{\downarrow} \right) [1 - \tau_i(0°)] \tag{37}$$

with n_i as a radiometric parameter that depends on the measurement channel of the satellite sensor and on the temperature range considered, γ_i is a parameter also dependent on the measurement channel and on the atmosphere (Schmugge et al., 1991), $T_a^{\downarrow}$ is the radiometric temperature of the atmosphere in the downward direction and $\tau_i(0°)$ the atmospheric transmittance in the nadir direction.

Once the two terms of radiometric correction are obtained, the atmospheric term (Equation 35) and the emissivity term (Equation 36), by replacing the former in the latter through T_i^* we obtain the single-channel correction equation that relates the radiometric temperature measured by the channel i with LST:

$$LST = T_i + \frac{1 - \tau_i}{\tau_i}(T_i - T_a^{\uparrow}) + \frac{\varepsilon_i - 1}{\varepsilon_i} b_i \tag{38}$$

5 Split-window method

The use of radiosoundings in the single-channel method is a disadvantage, because they are not always available. The split-window method uses the measurement of two channels within the 8-13 μm window where the atmospheric attenuation of ground radiance is proportional to the difference between the radiance measurements made in these two channels. Avoiding the use of radiosounding is an operational advantage in the calculation of the temperature of the surface, when the apparent temperatures recorded by the two channels, T_1 and T_2, are known.

Coll and Caselles (1997) proposed a split-window algorithm relating the temperature of a surface with the temperatures and emissivity measured in the spectral channels 4 and 5 of the AVHRR-NOAA 11 sensor, comparing it to other split-window algorithms such as that proposed by Becker and Li (1995), Prata (1993), or François and Ottlé (1996). The expression that came was:

$$T = T_1 + A(T_1 - T_2) + \Delta + B(\varepsilon) \tag{39}$$

where the terms A and Δ depend entirely on weather conditions and are completely independent on the effect of the emissivity, which is corrected by $B(\varepsilon)$, which is in turn dependent on the atmosphere.

5.1 Terms of atmospheric correction (A and Δ)

The coefficient A depends only on atmospheric conditions, specifically the atmospheric transmittance between sensor and surface, expressed as:

$$A = \frac{1 - \tau_1(\theta)}{\tau_1(\theta) - \tau_2(\theta)} \tag{40}$$

where $\tau_1(\theta)$ and $\tau_2(\theta)$ are atmospheric transmittances measured on channels 1 and 2 of the sensor.

The Δ coefficient corrects the effect of atmospheric emission and is expressed as:

$$\Delta = -[1 - \tau_2(\theta)]A(T_{a1}^{\uparrow} - T_{a2}^{\uparrow}) \tag{41}$$

where $T_{a1}^{\uparrow}$ and $T_{a2}^{\uparrow}$ are effective atmospheric temperatures upstreaming from channels 1 and 2 (McMillin, 1975). Equations 40 and 41 represent the classical coefficients derived from a black body (Maul, 1983). Note the initial negative sign in Equation 41, which indicates that the temperature value should be substracted, thus eliminating the atmospheric radiative contribution.

In fact, for the huge amount of data obtained by simulation or field measurements, the calculation of these coefficients, A and Δ, is quite nonfunctional; Coll and Caselles (1997) evaluated both terms by finding a surface that minimizes the effect of emissivity, and chose the sea surface, as it is very close to unity and the term $B(\varepsilon)$ can be neglected, with Equation 39 as follows:

$$T = T_1 + A(T_1 - T_2) + \Delta \tag{42}$$

That can written:

$$T - T_1 = A(T_1 - T_2) + \Delta \tag{43}$$

If simulated values of $T - T_1$ versus $T_1 - T_2$ are represented (Galve et al., 2008), the result is a graphic representation that can be fitted to a quadratic equation expressed as:

$$T - T_1 = a_0 + a_1(T_1 - T_2) + a_2(T_1 - T_2)^2 \tag{44}$$

Comparing Equation 44 to Equation 43 it is observed that A has a linear dependence on the temperature difference between channels and that Δ is constant:

$$A = a_1 + a_2(T_1 - T_2) \tag{45}$$

$$\Delta = a_0 \tag{46}$$

The coefficients a_0, a_1 and a_2 are obtained by linear regression using either matching measurements of the sensor and the surface, or through a simulated database.

It is surprising to see that A depends linearly on the temperature difference between two channels. Note that A and Δ depend on the zenith angle of observation, Equations 40 and 41. In this regard Niclòs et al. (2007) conducted a study of the variation in temperature of the sea with the observation angle using Equation 39. Through simulations it was observed how Equation 44 varied for four different zenith angles: 0°, 47.5°, 60° and 65°, noting that the coefficients a_0, a_1 and a_2 are consistent with a function like:

$$a_0 = a_{01}[\sec(\theta) - 1] + a_{02} \tag{47}$$

$$a_1 = a_{11}[\sec(\theta) - 1] + a_{12} \tag{48}$$

$$a_2 = a_{21}[\sec(\theta) - 1] + a_{22} \tag{49}$$

where a_{01}, a_{02}, a_{11}, a_{12}, a_{21} and a_{22} are obtained by linear regression. The values obtained by Niclòs et al. (2007) are given in Table 1.

Table 1. Tabulated coefficients of equations 77-79 (Niclòs et al., 2007) for the MODIS sensor aboard the Terra and Aqua platforms.

EOS platform	a_{01}	a_{02}	a_{11}	a_{12}	a_{21}	a_{22}
Terra	0.466 ± 0.012	0.392 ± 0.006	0.03 ± 0.02	2.57 ± 0.02	0.359 ± 0.011	0.427 ± 0.009
Aqua	0.466 ± 0.012	0.396 ± 0.006	0.02 ± 0.02	2.54 ± 0.02	0.357 ± 0.011	0.419 ± 0.009

Table 2. Coefficients of equations 83 and 84 (Niclòs et al., 2007) for the MODIS sensor aboard the Terra and Aqua platforms.

EOS platform	α_0	α_1	α_2	β_0	β_1	β_2
Terra	53.23 ± 0.05	-1.27 ± 0.02	-0.210 ± 0.002	196.1 ± 0.2	-35.74 ± 0.10	1.785 ± 0.010
Aqua	53.36 ± 0.05	-1.27 ± 0.02	-0.211 ± 0.002	194.9 ± 0.2	-35.56 ± 0.11	1.779 ± 0.010

5.2 Correction term of the emissivity effect $B(\varepsilon)$

The correction of emissivity in the split-window algorithm is affected by atmospheric conditions. Coll and Caselles (1997) offer the following expression of the term:

$$B(\varepsilon) = \alpha(1 - \varepsilon) - \beta\Delta\varepsilon \tag{50}$$

where $\varepsilon = (\varepsilon_1 + \varepsilon_2)/2$ is the average emissivity of the two channels, $\Delta\varepsilon = \varepsilon_1 - \varepsilon_2$ their difference in emissivity and α and β two coefficients that are defined as:

$$\alpha = (b_1 - b_2)A\tau_2(\theta) + b_1 \tag{51}$$

$$\beta = A\tau_2(\theta)b_2 + \frac{\alpha}{2} \tag{52}$$

where b_i (i = 1, 2) is the channel parameter defined in Equation 37 that sees differently the behavior of emissivity at sea or on land, through its reflective character (parameter $T_{ai}^{\downarrow}$) because the angular dependence of transmittance must be known in the case of the sea; it being enough to know the transmittance at nadir in the case of earth.

The emissivity correction term is obtained by calculating Equations 50 to 52, but as with the terms of atmospheric correction, this term is rather impractical. A solution was proposed by Coll and Caselles (1997), who, after calculating the term $B(\varepsilon)$ by Equations 50 to 52, observed that α and β varied with the amount of atmospheric water vapor (W). Therefore, they can be adjusted by regression and a simple expression of α and β can be obtained according to W. Niclòs et al. (2007), considering both the specular nature of the sea surface and its properties, made a quadratic adjustment of these parameters with W:

$$\alpha = \alpha_0 + \alpha_1 W + \alpha_2 W^2 \tag{53}$$

$$\beta = \beta_0 + \beta_1 W + \beta_2 W^2 \tag{54}$$

Galve et al. (2008) also made the adjustment but for land surface data, limiting the β coefficient to the lineal term. The quadratic fit did not mean an improvement in the accuracy of the value of surface temperature. Both works show tabulated values of α_i, β_i (i = 0, 1 and 2) obtained by linear regression. Table 2 shows the values obtained by Niclòs et al. (2007).

In this way, a simple split-window equation (Equation 39) is obtained, whose terms A, Δ and $B(\varepsilon)$ are conveniently calculated using Equations 45, 46 and 50, respectively, taking into consideration in Equation 50 the terms dependent on W, $\alpha(W)$ and $\beta(W)$, Equations 53 and 54.

For further information, the works on this method by Wan (1999) made for the MODIS sensor and Prata (2002) made for the AATSR are recommended.

6 Conclusions

In this article we have achieved two key points in determining the LST from sensors aboard satellites: firstly, the need to know accurately and precisely the emissivity of land surfaces, and secondly, to take into account the radiative contribution of the atmosphere introduced between the sensor and the surface. Regarding emissivity, some important methods used today in its determination and application to temperature measurement algorithms have been shown, in order to correct possible errors in their value. To avoid the effect of the atmosphere in the determination of temperature, two methods widely known in the literature of thermal remote sensing have been proposed. The first one, called single-channel, although mathematically simple, has the disadvantage of requiring atmospheric values provided by radiosounding, which are not always available. The second one, called split-window, does not depend so much on atmospheric parameters, there is a need to know the amount of water vapor in the atmosphere in order to obtain the emissivity correction term of the surface studied if necessary (in marine areas it is not necessary, for example). The fact of dealing with a simple mathematical formula, composed of easy-access terms, today, makes the split-window method one of the most suitable for atmospheric correction in the radiance measurements done by satellite. This article is a review of the determination of temperature in thermal infrared, as well as a handy tool for the scientific communities that come to the field from other specialties, as well as for new readers who want to begin learning about remote sensing.

Acknowledgements. This work has been made possible thanks to the funding from the Spanish Ministry of Science and Innovation (FPI scholarship of V. Garcia, projects CGL2007-64666 and CGL2007-29819-E, co-financed by FEDER) and the *Conselleria d'Educació de la Generalitat Valenciana* (PROMETEO/2009/086). The authors acknowledge the valuable comments of Prof. César Coll, Dr. Juan Manuel Sánchez, Joan Miquel Galve and María Mira from the *Universitat de València*.

References

Becker, F. and Li, Z.-L., 1995: *Surfaces temperature and emissivity at various scales: Definition, measurement and related problems*, Int J Remote Sens, **12**, 225–253.

Berk, A., Anderson, G. P., Acharya, P. K., Chetwynd, J. H., Bernstein, L. S., and Shettle, E. P., 1999: MODTRAN 4 user's manual, MA: Air Force Research Laboratory, Space Vehicles Directorate, Air Force Material Command, Hascom AFB, 95.

Buettner, K. J. and Kern, C. D., 1965: *The determination of infrared emissivities of terrestrial surfaces*, J Geophys Res, **70**, 1324–1337.

Caselles, V., Coll, C., and Sobrino, J. A., 1991: La Teledetección en el seguimiento de los fenómenos naturales, Servicio de publicaciones de la Universitat de Valencia, València, 95-140, 149-182.

Coll, C. and Caselles, V., 1997: *A split-window algorithm for land surface temperature from adnvanced very high resolution radiometer data: Validation and algorithm comparison*, J Geophys Res, **102**, 1697–16, 713.

Coll, C., Valor, E., Caselles, V., and Niclòs, R., 2003: *Adjusted Normalized Emissivity Method for surface temperature and emissivity retrieval from optical and thermal infrared remote sensing data*, J Geophys Res, **108**, 12.1–12.14.

Colton, A. L., 1996: *Effective thermal parameters for a heterogeneous land surface*, Remote Sens Environ, **57**, 143–160.

Conaway, J. and van Bavel, C. H. M., 1967: *Evaporation from a wet surface calculated from radiometrically determined surface temperatures*, J Appl Meteorol, **6**, 650–655.

Dana, R. W., 1969: Measurements of 8-14 microm emissivity of igneous rocks and mineral surfaces, NASA science Report NSG-632, Goddard Space Flight Center, Greenbelt, Maryland (USA).

François, C. and Ottlé, C., 1996: *Atmospheric corrections in the thermal infrared: Global and water vapor dependent split-window algorithms- Applications to ATSR and AVHRR data*, IEEE Trans Geosci Remote Sensing, **34**, 457–469.

French, A. N., Norman, J. M., and Anderson, M. C., 2003: *A simple and fast atmospheric correction for spaceborne remote sensing of surface temperature*, IEEE Trans Geosci Remote Sensing, **87**, 326–333.

Galve, J. M., Coll, C., Caselles, V., and Valor, E., 2008: *An atmospheric radiosounding database for generating Land Surface Temperature algorithm*, Remote Sens Environ, **46**, 1547–1557. Doi: 10.1109/TGRS.2008.916 084.

Gillespie, A. R., 1986: Lithologic mapping of silicate rocks using TIMS, in the TIMS data user's workshop, Publ. Pasadena Calif, 29-44.

Gillespie, A. R., Rokugawa, S., Matsunaga, T., Cothern, J. S., Hook, S., and Kahle, B., 1998: *A Temperature and Emissivity Separation algorithm for Advanced Spaceborne Thermal Emission and Reflection radiometer ASTER images*, IEEE Trans Geosci Remote Sensing, **36**, 1113–1126.

Maul, G. A., 1983: *Zenith angle effects in multichannel infrared sea surface sensing*, Remote Sens Environ, **13**, 313–329.

McMillin, L. M., 1975: *Estimation of sea surface temperatures from two infrared window measurements with different absorption*, J Geophys Res, **36**, 5113–5117.

Niclòs, R., Caselles, V., Coll, C., and Valor, E., 2007: *Determination of sea surface temperature at large observation angles using an angular and emissivity-dependent split-window equation*, Remote Sens Environ, **111**, 107–121. Doi:10.1016/j.rse.2007.03.014.

Prata, A. J., 1993: *Land surface temperatures derived from the advanced very high resolution radiometer and the along track scanning radiometer. Theory*, J Geophys Res, **98(D9)**, 689–16, 702.

Prata, A. J., 2002: Land surface temperatures measurements from space: AATSR algorithm theoretical basis document, CSIRO Atmos. Res., Aspendale, Australia, tech. Rep. 34.

Rouse, J. W., Haas, R. H., Schell, J. A., Deering, D. W., and Harlan, L. C., 1974: Monitoring the vernal advancement of retrogradation of natural vegetation, NASA/GSFC, Type III, Final Report, Greenbelt, MD, p 371.

Rubio, E., Caselles, V., and Badenas, C., 1997: *Emissivity measurements of several soils and vegetation types in the 8-14 μm wave band: analysis of two field methods*, Remote Sens Environ, **59**, 490–521.

Schmugge, T. J., Becker, F., and Li, Z.-L., 1991: *Spectral emissivity variations observed in airbone surface temperature measurements*, Remote Sens Environ, **35**, 95–104.

Valor, E. and Caselles, V., 1996: *Mapping land surface emissivity from NDVI: application to european, african and south american areas*, Remote Sens Environ, **57**, 167–184.

van de Griend, A. A. and Owe, M., 1993: *On the relationship between thermal emissivity and the normalized difference vegetation index for natural surfaces*, Int J Remote Sens, **14**, 1119–1131.

Wan, Z., 1999: MODIS land surface temperature. Algorithm theorethical basis document, NAS5-31370.