A simple and fast atmospheric correction for spaceborne remote sensing of surface temperature

A.N. French a,*, J.M. Norman b, M.C. Anderson b

a Hydrological Sciences Branch, Code 974, NASA Goddard, Greenbelt, MD 20771, USA
b Department of Soil Science, University of Wisconsin-Madison, 1525 Observatory Drive, Madison, WI 53706, USA

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Abstract

Accurate surface temperature retrieval using thermal infrared observations from satellites is important for surface energy balance modeling; however it is difficult to achieve without proper correction for atmospheric effects. Typically the atmospheric correction is obtained from radiosonde profiles and a radiative transfer model (RTM). But rigorous RTM processing is impractical for routine continental scale modeling because of long computational times. An alternative, simpler, and faster approach for correcting observations in the 10–12.5 μm band is developed from a previously published water vapor continuum absorption function. Using the RTM program MODTRAN as a reference, the function is calibrated against 159 radiosondes, and then validated against the TIGR radiosonde (1761 profiles) data set. Implementation of the calibrated absorption function usually produced larger temperature corrections than without calibration, an effect due to water vapor band type absorption and to non-water vapor constituents. The resulting surface temperature estimates, within 0.8 °C of MODTRAN estimates, were achieved at 15 × less processing time than MODTRAN.

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1. Introduction

Land surface temperature is an important property for modeling hydrological processes within the surface boundary layer because of its close relation to evapotranspiration and other surface energy fluxes. Surface temperature can also be an indicator of soil or vegetation moisture status. Currently, spatially distributed estimates of surface temperature can be retrieved from several satellite platforms such as the Advanced Spaceborne Thermal Emission and Reflection radiometer (ASTER), the Moderate Resolution Imaging Spectroradiometer (MODIS), and the Enhanced Thematic Mapper+ (ETM+), at resolutions ranging between 60 m and 1 km. These instantaneous estimates can be augmented with frequent (1/2 hourly), but coarser (4–5 km) resolution observations from geostationary satellites such as GOES (Mecikalski, Diak, Anderson, & Norman, 1999).

However, land surface temperatures derived from remote sensing are not often used because accurate estimates are difficult to obtain. Significant obstacles include uncertain instrumental calibration and surface emissivity, but the most important limitation is the need to remove atmospheric effects. Atmospheric transmissivity and path radiance within the thermal infrared (TIR) window, 8–12.5 μm, can cause apparent surface temperatures to deviate from actual temperatures by 10 °C or more. Uncorrected, this level of deviation is certainly unacceptable for hydrological models, where accuracies on the order of 1–2 °C are desired. A good way to remove atmospheric effects upon TIR observations is to combine atmospheric profile observations, derived from radiosonde data, with a radiative transfer model (RTM; e.g. MODTRAN; Berk et al., 1998) to return estimates of band-averaged atmospheric transmissivity and path radiance. This method, however, may be impractical for real-time applications because an RTM can be slow, and the image data sets may be very large.

An example application is the Atmosphere Land Exchange Inverse (ALEXI) model, which can estimate surface energy fluxes at 4–5 km resolutions over entire continents (Anderson, Norman, Diak, Kustas, & Mecikalski, 1997; Mecikalski & al., 1999). Implementation of ALEXI at this scale requires daily corrections to millions of GOES pixels. Extension of ALEXI to high resolutions (30 m) using a
disaggregation technique (DisALEXI; Norman et al., 2003) increases the correction requirements to tens, even hundreds of millions pixels. Clearly, a fast and simple algorithm for TIR atmospheric correction would be a tremendous aid to operational implementation of models such as ALEXI and DisALEXI.

This note demonstrates how the atmospheric correction process can be performed quickly, yet still have results comparable in quality to those obtained from the most rigorous RTMs. Although this demonstration is based on TIR corrections to band 4 (Fig. 1) on the GOES 8 satellite and on band 5 on the airborne TIMS sensor (both bands cover the range 10.2–11.2 μm), it is also valid for other satellite observations within the TIR window. Over 2000 atmospheric profiles, representing a wide-range of conditions, were used to calibrate a water vapor continuum function, described in Roberts, Selby, and Biberman (1976) and extended in Price (1983). Of these radiosonde profiles, 159 were from the Southern Great Plains 1997 experiment (SGP97, see http://hydrolab.arsusda.gov/sgp97 and http://daac.gsfc.nasa.gov/CAMPAIGN_DOCS/SGP97/sgp97.html). Calibration is based on a reference model, MODTRAN, which utilizes the same continuum function, but is more comprehensive, including water vapor band-type absorption, as well as non-water vapor aerosol effects. By comparing atmospheric property estimates made in the two different ways, we show how implementation of the continuum function alone can be adjusted to return atmospherically corrected observations that are nearly the same as those results returned by an RTM. The adjustment is made by adding an extinction coefficient correction term. This term minimizes systematic discrepancies between radiative transfer estimates and the simplified approach, as verified over a wide range of atmospheric conditions and surface temperatures.

2. Reference atmospheric correction

TIR observations of the earth’s surface are commonly corrected for atmospheric effects with the following model:

\[ L_{\lambda,\text{srf}} = \frac{L_{\lambda,\text{sns}} - L_{\lambda,\downarrow}}{\tau} - (1 - \varepsilon_{\lambda})L_{\lambda,\uparrow} \]

(1)

where \( L_{\lambda,\text{srf}} \) is surface radiance (W m⁻² sr⁻¹ μm⁻¹), \( L_{\lambda,\text{sns}} \) is radiance observed by the sensor, \( L_{\lambda,\downarrow} \) is the downwelling radiance, \( \tau \) is the atmospheric transmissivity, \( \tau \) is the atmospheric transmissivity, \( \varepsilon_{\lambda} \) is the soil emissivity, and \( L_{\lambda,\uparrow} \) is the upwelling radiance.

Fig. 1. Spectral properties for the TIR window, including upwelling atmospheric radiance (A), atmospheric transmissivity (B), and typical bare soil emissivity (C). GOES 8 response functions [thick lines in panel (B), from http://www.oso.noaa.gov/goes/goes-calibration/goes-sounder-srfs.htm] span TIR segments with relatively few band-type absorption features, as compared with the 8–9.5 μm segment. Bands 4 and 5 also correspond to high emissivity portions of the soil sample.

Fig. 2. Average directional downwelling radiance vs. clear sky, nadir view upwelling radiance (W m⁻² sr⁻¹ μm⁻¹) for GOES 8 band 4 derived from 159 (N) SGP97 radiosondes. The curve represents Eq. (2), accurately predicting \( L_{\lambda,\downarrow} \) from \( L_{\lambda,\uparrow} \) with a standard error (Sₐ) of 0.0457 W m⁻² sr⁻¹ μm⁻¹ and a coefficient of determination \( R^2 \) of 0.9976.
atmospheric radiance, $L_{\lambda,1}$ is the average directional downwelling atmospheric radiance, $\tau_{\lambda}$ is atmospheric transmissivity, and $\varepsilon_{\lambda}$ near-nadir surface emissivity. Examples of $L_{\lambda,\perp}$ and $\tau$ spectra (Fig. 1A and B) show that GOES band 4 spans a slightly more transparent portion of the TIR spectral window than does band 5, making band 4 preferable for surface observations. The solution to Eq. (1) requires three atmospheric properties ($\tau_{\lambda}$, $L_{\lambda,\perp}$, $L_{\lambda,1}$), and one surface property ($\varepsilon_{\lambda}$). Since our main interest is with the use of GOES band 4 under clear sky conditions, Eq. (1) can be simplified by reformulating the second term. This is achieved by either setting $\varepsilon_{\lambda}$ to an independently measured value for the scene in question, or by setting $\varepsilon_{\lambda}$ to a constant value, and then representing $L_{\lambda,1}$ as a function of $L_{\lambda,\perp}$. For GOES band 4, using constant emissivities for vegetated and bare soil surfaces is usually accurate over view angles between 0° and 45°. Common vegetated surfaces have nearly constant emissivities close to 0.99 between 10 and 12.5 μm. Bare soil emissivities over band 4, though relatively lower than vegetation, are also nearly constant. For example, the average band 4 emissivity obtained from laboratory measurement of a bare soil sample from central Oklahoma is 0.96 (Fig. 1C). With few exceptions, band 4 emissivities ranged between 0.95 and 0.99 for a collection of 124 soil, vegetation and water spectra taken from the ASTER spectral library.1 Vegetated surfaces, due to multiple scattering of emitted radiation, would have yet higher emissivities, near 0.98 or 0.99 between 10 and 12.5 μm. The second simplification is possible because $L_{\lambda,1}$ varies systematically with the nadir value of $L_{\lambda,\perp}$ under clear sky conditions. For GOES band 4 (using SGP97 radiosondes, which are further discussed in Section 4) $L_{\lambda,1}$ is well approximated (Fig. 2) by a power function of the upwelling atmospheric radiance:

$$L_{\lambda,1} = 1.744 * L_{\lambda,\perp}^{0.841}$$  (2)

The atmospheric correction task for bands within 10–12.5 μm can therefore be accomplished by estimating only two atmospheric parameters: $L_{\lambda,\perp}$ and $\tau_{\lambda}$. The most important conditions controlling $L_{\lambda,\perp}$ and $\tau_{\lambda}$ are the vertical distributions of pressure, temperature and humidity. Virtually all atmospheric water vapor resides within the lowest 5 km, so the profile should contain the most detail in this interval. Various RTMs, including line-by-line and correlated-k programs (Kratz & Rose, 1999), can then be used to determine band aver aged atmospheric transmittances and radiances. We use MODTRAN (Berk et al., 1998), which is a single-parameter band model. For broadband observations, surface temperature is conveniently calculated from atmospherically corrected surface radiances using a power function which is fit to a set of temperature/radiance pairs, generated from the forward

Planck function over the appropriate wavelengths (Fig. 1B) and surface temperatures between 25 and 60 °C:

$$L_\lambda = aT_\text{surf}^b$$

where $T_{\text{surf}}$ is temperature in Kelvin, and $a$ and $b$ are least-squares fit coefficients specifically for GOES 8 bands 4 and 5 (Table 1).

### 3. Water vapor continuum correction

The simplified TIR atmospheric correction technique proposed here is based upon a temperature-compensated water vapor continuum function described in Roberts et al. (1976). Henceforth, we refer to this technique as the ‘Roberts’ approach. The function estimates the water vapor absorption coefficient as a composite of two empirically derived exponential terms. In the following equations, coefficients found in Roberts et al. (1976) have been re-written in terms of wavelength and SI units. The base term describes water vapor absorption at 296 K for wavelengths between 8 and 13 μm:

$$C(\lambda) = c + d \exp \left(\frac{-\beta}{\lambda} \right)$$

where $c = 4.124 \times 10^{-3}$ m² kg⁻¹ (kPa)⁻¹, $d = 5.509$ m² kg⁻¹ (kPa)⁻¹, $\beta = 78.7$ μm and $\lambda$ is wavelength (μm). This base term is exponentially scaled to compensate for temperature effects:

$$C(\lambda, T) = C(\lambda) \exp \left[ T_o \frac{1}{T} - \frac{1}{296} \right]$$

where $C(\lambda, T)$ is the compensated absorption coefficient and $T_o$ is the temperature dependence parameter (≈ 1800 K).

Computing the absorption coefficients from Eq. (5), the water vapor continuum extinction coefficient, $\sigma_i$ (m⁻¹), for each atmospheric layer $i$ is:

$$\sigma_i = C(\lambda, T_i) \rho_i [e_i + \gamma (P_i - e_i)]$$

where $P_i$ is total pressure (kPa) of the specified layer, $\rho_i$ is water vapor density (kg m⁻³), $e_i$ is the water vapor pressure component (kPa), and $\gamma$ (= 0.002) is a relative measure of the ambient to self-broadened water vapor continuum.

Having solved for water vapor absorption in each layer, the transmittance of each layer, $\tau_i$, can now be determined as a

<table>
<thead>
<tr>
<th>Band</th>
<th>Power function coefficients for GOES bands 4 and 5</th>
</tr>
</thead>
<tbody>
<tr>
<td>4</td>
<td>$a = 1.745 \times 10^{-10}$, $b = 4.338$</td>
</tr>
<tr>
<td>5</td>
<td>$a = 1.902 \times 10^{-9}$, $b = 3.905$</td>
</tr>
</tbody>
</table>

$L_j = aT^n$, with $L_j$ in W m⁻² sr⁻¹ μm⁻¹ and $T$ in Kelvin.

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1 Data courtesy Jet Propulsion Laboratory, California Institute of Technology, Pasadena, CA, USA ©1999.
function of three quantities: water vapor absorption, \( \sigma_i \); nadir view path length, \( z_i \); and sensor view angle, \( \theta \):

\[
\tau_i = \exp(-\sigma_i z_i \sec \theta) \tag{7}
\]

Knowing \( \tau_i \) for each layer, atmospheric upwelling radiance received by the sensor is a cumulative summation of radiance from all layers. The blackbody radiance of each atmospheric layer (\( L_{BB,i} \)) is calculated from the Planck equation.

Hence, the upwelling radiance at a layer, \( L_{\text{atm}}(i) \), is:

\[
L_{\text{atm}}(i) = e_i L_{BB}(i) + \tau_j L_{\text{atm}}(j-1) \tag{8}
\]

where contributions from the current layer (the first term on the right hand side of Eq. (8)) are added to the net contribution from the underlying layers (the second term).

Representation of the ‘Roberts’ approach can be in terms of pathlength, or as demonstrated by Price (1983), in terms of pressure. If extinction is measured by pathlength, Eq. (6) is used, along with coefficients set for Eq. (4). If extinction is measured by pressure, the hydrostatic equation is applied to change variables:

\[
dP = -\rho_{\text{air}} g dz \tag{9}
\]

where \( \rho_{\text{air}} \) is density of moist air, \( g \) is gravitational acceleration, and \( dz \) is the distance differential. Water vapor absorption in pressure terms (kPa\(^{-1}\)) is:

\[
\sigma_i = [0.004123844 + 5.509455 \times \exp(-78.7/\lambda)] \\
\times \exp\left[1800 \left(1 - \frac{1}{T_i} - \frac{1}{296}\right)\right] \times \frac{q_i}{g} [e_i - 0.002(P_i - e_i)] \tag{10}
\]

where \( q_i \) is specific humidity in kg kg\(^{-1}\). Eq. (10) is then applied in the same way as before, except that path length \( z_i \) in Eq. (7) is replaced by a pressure differential.

### 4. Calibration with ARM-SGP97 data

Calibration of the ‘Roberts’ modeling approach with respect to MODTRAN used 159 radiosondes collected during the Southern Great Plains 1997 field experiment (SGP97, see http://www.arm.gov/docs/sites/sgp/sgp.html). These radiosondes represent detailed atmospheric profiles at 3-h intervals from sites over central Oklahoma and southern Kansas from 29 June–2 July 1997. Each profile was resampled to 60 layers and modeled with MODTRAN. Non-water vapor constituents were defaulted to the standard ‘mid-latitude summer’ profiles. The resulting GOES band 4 transmissivities and upwelling radiances, as a function of columnar water vapor, are shown in Fig. 3. The range of atmospheric water vapor is broad, ranging from 1.4 to 5.2 cm precipitable water. Modeling the SGP97 profiles illustrates the large variation between columnar water vapor and observed atmospheric transmissivity and upwelling radiance. For example, the range in transmissivity for 3 cm columnar water vapor spans at least 0.6–0.7, with upwelling radiances respectively ranging from 2.2 to 3.0 W m\(^{-2}\) sr\(^{-1}\) \(\mu\text{m}^{-1}\). The equivalent uncertainty in surface temperature, using Eq. (1), is large, ~7 °C for an actual surface temperature of 40 °C. This result shows that columnar water vapor is not an accurate predictor of transmissivity, nor path radiance, over the sub-humid Oklahoma environment. Consequently, estimation of atmospheric correction terms using surface observations along lines described by Qin, Karniel, and Berliner (2001) is generally an unsuitable approach, except for arid conditions.

These MODTRAN results are then compared with ‘Roberts’ model results (left-hand column in Fig. 4, Fig. 3. Atmospheric transmissivity (top) and upwelling radiance (bottom) derived from SGP97 radiosonde profiles and MODTRAN.
entitled ‘Unadjusted’). The results are strongly linearly correlated. \( R^2 = 0.998 \) for both transmissivity and path radiance. However, the comparisons show systematic bias, with \( \sim 0.06 \) overestimated transmissivity and \( \sim 0.500 \, \text{W m}^{-2} \, \text{sr}^{-1} \, \mu\text{m}^{-1} \) underestimated path radiance. The source of bias is due to water vapor band type absorption as well as absorption by aerosols, CO\(_2\), and O\(_3\), constituents considered by MODTRAN models, but not by the ‘Roberts’ water vapor continuum model. The consequence of this bias could be significant underestimation of surface temperature. For nadir view of an atmosphere containing 3 cm of precipitable water vapor, the ‘Roberts’ based estimate would be \( \sim 1.1 \, ^\circ\text{C} \) too low for a surface at 40 \(^\circ\text{C}\). The underestimation, however, is likely to be even greater for higher true surface temperatures, and for GOES mid-latitude view angles (\( \sim 40^\circ \)). Considering the increased atmospheric path length for such views, as well as the likelihood of uncertain atmospheric profile estimates, there is good reason to take advantage of the good correlation just observed in Fig. 4, and apply an appropriate correction to the ‘Roberts’ model.

Close inspection of the comparisons in Fig. 4 shows that prediction biases, though nearly constant, also diminish with decreasing transmissivity and increasing path radiance. These relations suggest a \( \sigma \) correction term be added to Eq. (10), dependent upon water vapor concentration. The required functional form should be important when water vapor pressure is low, and less important when water vapor pressure is high. The correction form we choose alters the water vapor absorption in Eq. (10) by

![Modeled SGP97 Radiosonde Data](image)

Fig. 4. ‘Roberts’ model results, without adjustment (left column) and with adjustment (right column) for non-water vapor constituents, compared with MODTRAN results for SGP97 radiosonde data.
adding an absorption term containing two empirical components $k$ and $h$:

$$
\sigma_{adj} = \sigma_i + \left[ \frac{k}{N} - \frac{e_i}{h} \right] 
$$

The $k$ term, $(\text{kPa})^{-1}$, is the most important correction term and represents the total absorption due to non-water vapor constituents. The vertical distribution of these constituents is unknown, so a uniform distribution is assumed by equally partitioning the $k$ term amongst the ‘N’ atmospheric layers. The $h$ term, $(\text{kPa})^{-2}$ is less important and diminishes the absorption compensation as water vapor density increases. A series of approximations determined by trial and error produced acceptable values for coefficients $k$ and $h$ (Table 2).

The adjustment term components reduce prediction bias by $\sim 90\%$ (Table 3). For the SGP97 data set (right column, Fig. 4), overall bias in total atmospheric transmissivity, $\tau z$, is reduced to 0.007 from -0.060. Bias in total upwelling atmospheric path radiance, $L_{z, \tau}$, is reduced to 0.041 W m$^{-2}$ sr$^{-1}$ $\mu$m$^{-1}$ from 0.478 W m$^{-2}$ sr$^{-1}$ $\mu$m$^{-1}$. The adjustment term does not remove all of the systematic bias, but the discrepancies are typically no greater than 0.2 W m$^{-2}$ sr$^{-1}$ $\mu$m$^{-1}$. Furthermore, the adjustment term does not reduce estimation precision. Standard errors ($\sigma_s$) for $\tau z$ and $L_{z, \tau}$ change minimally before and after adjustment. Unadjusted and adjusted $\sigma_s$ values for transmissivity are respectively 0.006 and 0.005. For path radiance, the unadjusted and adjusted values are 0.051 and 0.043 W m$^{-2}$ sr$^{-1}$ $\mu$m$^{-1}$.

### 5. Validation with TIGR profiles

Although the SGP97 radiosonde data set represents a wide range of atmospheric conditions, calibration verification requires a more general data set. For this purpose we used 1761 profiles from the TOVS Initial Guess Retrieval (TIGR) data set (Chedin, Scott, Wahucle, & Moulinier, 1985). These profiles, each specifying 40 levels and representing conditions spanning tropical to sub-arctic latitudes, are a mix of actual and synthetic radiosondes derived from over 3000 actual radiosonde observations.

To assess the calibrated values for coefficients $k$ and $h$ in Eq. (11), the TIGR data set was processed with MODTRAN and the ‘Roberts’ approach in a similar way to the SGP97 data set. The most significant processing change was adaptation of modeled non-water vapor constituent profiles to the most representative of five standard models (tropical, mid-latitude summer, mid-latitude winter, sub-arctic summer and sub-arctic winter). The practical effect of this change upon atmospheric transmissivity and path radiance, however, was not important. Seasonal and geographical variations in aerosols, CO$_2$ and O$_3$ modeled by MODTRAN have negligible effects within the 10–12.5 $\mu$m band. A complication with the TIGR data was the frequent occurrence of supersaturated layers, a condition unconsidered by either MODTRAN or the ‘Roberts’ approach. In MODTRAN simulations, supersaturated layers were automatically reset to 100% saturation. To ensure compatibility, the Roberts processing used a similar reset.

Using the established correction coefficients, $k = 2.36 \times 10^{-2} (\text{kPa})^{-1}$, $h = 6000.0 (\text{kPa})^2$, the ‘Roberts’ approach once again was in excellent agreement with MODTRAN results (Table 4). Standard error ($\sigma_s$) and $R^2$ values were nearly the same as for the adjusted SGP97 radiosonde data (Table 3), indicating that the correction procedure has general applicability.

### 6. Practical considerations

Implementation of the adjusted water vapor continuum approach shows surface temperature retrieval accuracies comparable to MODTRAN with the benefit of greatly increased computer processing speeds.

To demonstrate retrieval accuracy we used high-resolution (12 m) TIR observations from the aircraft-based TIMS instrument (Palluconi & Meeks, 1985) in comparison with ground-based, 0.5 m footprint, TIR measurements. GOES observations could not be used in this case because spatial resolution differences were too large for meaningful comparison. Comparative observations over a thick grazing land site with emissivities $\sim 0.98$ (SGP97 El Reno field ER01; French, Schmugge, & Kustas, 2000) showed that surface temperatures agreed within 0.2–0.6 $^\circ$C of ground observations for both early morning and mid-morning times (Table 5).

**Table 3** Differences between atmospheric parameter estimates without and with adjustment to the ‘Roberts’ approach

<table>
<thead>
<tr>
<th></th>
<th>Unadjusted</th>
<th>Adjusted</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\tau$</td>
<td>-0.060</td>
<td>0.007</td>
</tr>
<tr>
<td>$L_{z, \tau}$</td>
<td>0.478</td>
<td>0.041</td>
</tr>
</tbody>
</table>

**Table 4** Validation of absorption adjustment with TIGR data

<table>
<thead>
<tr>
<th>$\tau$</th>
<th>$S_d$</th>
<th>$R^2$</th>
<th>Slope</th>
<th>Intercept</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.013</td>
<td>0.006</td>
<td>0.999</td>
<td>1.0166</td>
<td>0.026</td>
</tr>
<tr>
<td>0.049</td>
<td>0.058</td>
<td>0.997</td>
<td>1.0358</td>
<td>-0.046</td>
</tr>
</tbody>
</table>

**Table 5** Estimated radiometric surface temperatures ($^\circ$C) at El Reno, OK, for three cases: no atmospheric correction (at sensor), atmospheric correction without adjustment for non-water vapor constituents, and atmospheric correction with adjustment

<table>
<thead>
<tr>
<th>Time (CST)</th>
<th>Actual</th>
<th>At sensor</th>
<th>Un-adjusted</th>
<th>Adjusted</th>
</tr>
</thead>
<tbody>
<tr>
<td>6:15</td>
<td>21.2</td>
<td>21.3 (0.1)</td>
<td>21.0 (−0.2)</td>
<td>21.8 (0.6)</td>
</tr>
<tr>
<td>10:30</td>
<td>32.1</td>
<td>28.5 (−3.6)</td>
<td>30.6 (−1.5)</td>
<td>31.9 (−0.2)</td>
</tr>
</tbody>
</table>

Differences from ground-based surface temperature are shown in parenthesis.
With no atmospheric correction the mid-morning radiometric surface temperature was underestimated by 3.6 °C. Correction using the unadjusted ‘Roberts’ approach reduced the underestimation to 1.5 °C. But by using the adjustment for non-water vapor continuum constituents, surface temperature estimation was within 0.2 °C of ground-based observations. Hence the temperature correction was significantly greater (1.3 °C) than would be estimated by Price (1983), who did not consider absorption by non-water vapor constituents. The early morning results, on the other hand, showed that when surface temperatures were cooler, atmospheric correction can be small. The temperature correction difference between early and mid-morning observations highlights a sometimes neglected fact: magnitude of atmospheric correction is not only a function of the atmospheric profile itself, but is also a function of the difference between sensor radiance and upwelling atmospheric radiance (Eq. (1)).

The adjusted water vapor continuum algorithm was also computationally fast. Multiple time trials showed that it computed atmospheric correction values at least 15 times faster than MODTRAN using the same computer platform (Intel Xeon 1700 MHz, 256 KB cache, 3.8 GB RAM, Linux 2.4.19), and identical atmosphere profile data. For example, MODTRAN analyses of 100 atmospheric profiles with 60 levels took ~ 1.57 s, while the identical profiles took ~ 0.10 s when analyzed by our correction routine; a difference of 4 h of computer time for 1 million pixels. This is a significant time saving for operational use.

7. Conclusions

Initial modeling of the water vapor continuum within a 10–11.2 μm window, using an approach based upon work by Roberts et al. (1976) and by Price (1983), shows close agreement with radiative transfer computations from MODTRAN. The agreement between transmissivity and path radiance is adequate to produce corrected surface temperature estimates that are typically within 1–1.5 °C. But agreement can be improved. There are systematic reducible biases between the ‘Roberts’ and MODTRAN approaches due to absorption by non-water vapor constituents. Improvement is achieved by adjusting the absorption coefficient (σi) estimates for each layer, significantly reducing discrepancies in transmissivity, τ, and upwelling radiance, Lz.

The benefits of applying this correction to the ‘Roberts’ algorithm are demonstrated in Fig. 5, which shows that adjustment for non-water vapor continuum constituents can reduce typical prediction bias from ~ 1.6 °C to less than 0.8 °C.

This calibration and validation work shows that for single-band TIR observations within the 10–12.5 μm band, atmospheric correction based upon known profiles can be done in a simpler and more accessible way than resorting to a rigorous RTM. Without question, modeling with programs such as MODTRAN remains the most reliable TIR correction approach (Ellingson, Ellis, & Fels, 1991). But as a practical matter optimal benefits of RTM’s are almost never obtained. Typically, very little site specific information is known about non-water vapor distribution; consequently no amount of computational effort via RTM’s can obscure the use of default distributions for constituents such as aerosols, CO2 and O3. Hence the difference between rigorous radiative transfer and water vapor continuum models is frequently small. This result, combined with the significantly faster processing times achieved from the implementation shown here, greatly benefits large scale hydrological modeling efforts. Calibration of the water vapor continuum function for other bands, such as GOES 8 band 5 and ETM+ band 6, should return similarly good results.

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References


