The Thermal Infrared Multispectral Scanner (TIMS) was flown on the NASA C-130 aircraft for series of 12 flights during the HAPEX experiment in the Southwest of France during 1986. TIMS provides coverage of the 8–12 μm thermal infrared band in six contiguous channels so that it is possible to observe the spectral behavior of the surface emissivity over this wavelength interval. The surfaces observed ranged from bare soils to coniferous forest. As expected the fully vegetated fields, a small lake, and a coniferous forest exhibited little or no spectral variation. However, the bare soil surfaces had a 5–10°C difference in radiant temperature over the TIMS channels with the three shorter wavelength channels (8.0 μm < λ < 9.6 μm) having lower temperatures than the three longest wavelength channels. This qualitatively agrees with laboratory measurements done by a group from the Université Louis Pasteur in Strasbourg on soil samples from this area. The quantitative agreement of the relative variation after correcting for atmospheric effects is good, but does depend on the day and location of the sample in the field. This spectral variation arises from the absorption of infrared radiation due to the stretching vibrations of the silicon-oxygen bonds of silicates in the soil. These results indicate that the longer wavelength channels in the 8–12 μm band may be preferable for surface temperature sensing for soils rich in silicates.

INTRODUCTION

The thermally emitted radiance from any material surface depends upon two factors: 1) the surface temperature, which is an indication of the equilibrium resulting from the energy balance of the fluxes between the atmosphere, the surface, and the subsurface soil; and 2) the surface emissivity, which is the efficiency of the surface for transmitting the radiant energy generated in the soil into the atmosphere. The latter depends on the composition, surface roughness, and physical parameters of the surface, such as its moisture content. Furthermore, the emissivity varies with wavelength for many natural surfaces. Understanding this spectral variation of the surface emissivity in the
thermal infrared portion, 8–14 μm, of the electromagnetic spectrum is important for the determination of its surface temperature. Over the ocean the emissivity is nearly constant over this wavelength band; however, for land there is considerable evidence of the spectral variation of the surface emissivity. Large spectral features in the 8–10 μm region are attributed to the fundamental vibrations of Si—O bonds (Hunt, 1980). They are most pronounced in quartz. An example of this feature is shown in Figure 1, where laboratory measurements of the emissivity for two soils from our test area in France are presented (Nerry et al., 1988). The emissivity varies from about 0.85 between 8 μm and 9 μm to about 0.95 at 10–12 μm. The substitution of other metals, for example, Al or Mg for Si, produces changes in the width, depth, and center wavelength of the absorption band and thus affords possible approaches for mapping silicate rocks (Gillespie, 1986). However, for the purpose of using thermal infrared radiation to estimate surface temperature, these variations in surface emissivity provide additional complications (Price, 1984).

Radiometers on board an aircraft or spacecraft measure not only the radiation emitted by the surface but also the contribution from the intervening atmosphere and in some cases measure only the latter, for example, clouds. Thus the problem of remotely sensing surface temperatures depends not only on the emissivity of the surface of interest but on the contribution of the atmosphere. In this paper we will discuss the possibility of obtaining the relative spectral emissivity of some agricultural surfaces using data acquired with the Thermal Infrared Multispectral Scanner (TIMS) during the HAPEX-MOBILHY (Hydrologic Atmospheric Pilot Experiment and Modélisation du Bilan Hydrique) program (André et al., 1986; 1988) in southwest France in June 1986. We will describe the sensor and its performance, the atmospheric corrections applied to the data, and two methods for retrieving the relative spectral emissivity from a remote platform.

TIMS

The data we present here were obtained with the TIMS sensor on board the NASA C-130 during the HAPEX aircraft campaign in 1986. HAPEX had as its primary objective the improved parameterization of the surface fluxes in atmospheric circulation models (André et al., 1986; 1988). It included an intensive program of ground and aircraft observations. The remote sensing of the surface was performed by a suite of sensors on the C-130 aircraft. These included a microwave radiometer for soil moisture observations, a visible and near IR scanner for vegetation, and albedo data and a thermal infrared scanner (TIMS) for surface temperature.

TIMS has six channels in the thermal infrared (8–12 μm) region of the electromagnetic spectrum. The filter functions of the channels are shown in Figure 1. From the superimposed emissivity measurements in this figure it is clear that for these soils, at least, the lower three channels will have lower emissivities.

The instantaneous field of view is 2.5 mrad, while the detector signals are sampled every 2.08 mrad along the scan (Palluconi and Meeks, 1985). Thus from an altitude of 1500 m the pixel size is 3.75 m and the cross track separation is 3.2 m at nadir.

For calibration the system is equipped with cold and hot reference sources or blackbodies, whose temperatures span the range of interest. For most of HAPEX the temperature separation between the two references was set at 30°C. The reference temperatures are known to better than 0.5 K (Palluconi and Meeks, 1985).

TIMS responds to the incident radiance (W/m² sr cm⁻¹) and is related to the brightness
Spectral Emissivity Variation in Surface Temperature

The temperature of the observed surface is obtained via the Planck equation for blackbody (BB) radiation. Thus, to calculate the apparent temperatures, it is necessary to obtain the BB radiance of the hot and cold references for each channel. Thus the spectral radiance at the aircraft is

\[ L_i(a/c) = (DN - DN_i) \times \text{SLOPE} + L_i(T_c), \]

where

\[ \text{SLOPE} = (L_i(T_h) - L_i(T_c)) / (DN_h - DN_c) \]

and the DNs are digital counts for the target, and the references. The radiances \( L \) are given by the Planck blackbody equation:

\[ L_i(T) = \text{BB}(k_i, T) = C_1 k_i^2 / \left[ \exp \left( C_2 k_i / T \right) - 1 \right], \]

where \( C_1 = 1.191 \times 10^{-8} \, \text{W/(m}^2 \, \text{sr} \, \text{cm}^{-1}) \), \( C_2 = 1.439 \, \text{cm K} \), and \( k_i \) is the weighted average wavenumber (\( \text{cm}^{-1} \)) for TIMS Channel \( i \). The \( k_i \) are derived so that the spectral radiance at this frequency is equal to the average spectral radiance over the entire band. This approach simplifies the inversion of the radiances to obtain a brightness temperature. The values of \( k_i \) depend slightly on temperature and are given in Table 1.

While studying the data from a small lake that was overflown at both mid (1.5 km) and high (6 km) altitudes, we detected a calibration problem in that the brightness temperatures observed over the lake at the high altitude were higher than those at the midaltitude. Similar behavior was observed during the FIFE experiment in 1987 over a reservoir in Kansas (F. Palluconi, JPL, personal communication). The apparent cause of the problem is that the emitting surfaces of the references bodies are cooled by the air flow and are cooler than the temperatures recorded by a thermistor (Hoover, 1990). Thus the actual radiating temperatures were

\[ T_{h,c} = T'_{h,c} - b \times (T'_{h,c} - T_{\text{air}}), \]

where \( T'_{h,c} \) are the nominal blackbody temperatures recorded in the data stream and \( T_{\text{air}} \) is the air temperature at the aircraft altitude. The value of \( b \) was estimated by comparing the TIMS temperatures over the lake with those obtained by a wide band PRT-5 radiometer on the aircraft and found to be 0.1 (Stoll et al., 1990). At the 6 km altitude where \( T_{\text{air}} = -15^\circ \text{C} \) and \( T_{\text{air}}' = 45^\circ \text{C} \), this cooling of the radiating surface can produce errors on the order of 5°C or 6°C in \( T_h \) with resulting high estimates of surface brightness temperatures. The corrected values of \( T_{h,c} \) were used in this paper.

**ATMOSPHERIC EFFECTS**

The temperatures given by the sensor represent the detected radiance at the aircraft altitude. In order to convert this result to the actual surface temperature, atmospheric effects must be taken into account. These include the absorption and emission by the atmospheric gases, primarily water vapor for this portion of the spectrum. A number of methods have been developed to calculate the atmospheric correction (Deschamps and Phulpin, 1980; Price, 1983; McClain et al., 1985). The method used here was to calculate the atmospheric transmission and relevant fluxes with the Lowtran 7 Atmospheric Transmittance/Radiance model developed at the Air Force Cambridge Geophysics Laboratory, (Kniezys et al., 1988) and to use these results to correct the radiances observed by TIMS.

Under cloud-free conditions the radiance, \( L_i(a/c) \), received by an airborne radiometer is related to the brightness temperature \( T_i \) by

\[ L_i(a/c) = (1 / \Delta k_i) \int f_i(k) \times \text{BB}(k, T_i) \, dk = \text{BB}_i(T_i) \]

and to the surface radiance by

\[ L_i(a/c) = \tau_i \times L_i(\text{surf}) + L_i(\text{atmU}), \]
where \( r_i \) is the transmission of the atmosphere for channel \( i \), \( f_i(k) \) is the normalized spectral response of the radiometer for channel \( i \) as shown in Figure 1, \( \text{BB}(k, T) \) is the Planck radiation given by Eq. (2). \( L_i(\text{atm} U) \) is the upwelling atmospheric radiance, \( L_i(\text{surf}) \) is the upwelling radiance at the surface given by

\[
L_i(\text{surf}) = \varepsilon_i \cdot \text{BB}(k_i, T_{\text{grd}}) + L_i(\text{ref}) = \text{BB}(T_{\text{grd}})
\]  

(6)

\( T_{\text{grd}} \) is the brightness temperature measured at the ground, \( T_{\text{grd}} \) is the physical temperature of the surface, \( \varepsilon_i \) is the surface emissivity for channel \( i \), and \( L_i(\text{ref}) \) is the reflected atmospheric radiance given by

\[
L_i(\text{ref}) = \int \rho_{bi}(\theta, \phi, \theta', \phi') \cdot L_{\text{atm} i}(\theta', \phi') \cdot \cos(\theta', \phi') \, d\Omega,'
\]  

(7)

where \( \rho_{bi} \) is the bidirectional reflectance and \( L_{\text{atm} i} \) is the downwelling radiance. If we use the assumption that the surface is a Lambertian reflector, this reduces to

\[
L_i(\text{ref}) = \rho \cdot L_i(\text{atm} D) = (1 - \varepsilon_i) \cdot L_i(\text{atm} D),
\]  

(8)

where \( \rho \) is the hemispherical reflectivity and \( L_i(\text{atm} D) \) is the hemispheric downwelling atmospheric radiance given by

\[
L_i(\text{atm} D) = \left(1/\pi \right) \int L_i(\theta, \phi) \cdot \cos \theta \, d\Omega.
\]  

(9)

As pointed out in Becker and Li (1990), the Lowtran model can be used to calculate directly the transmittance \( \tau_i \), the upwelling radiance \( L_i(\text{atm} U) \), and \( \text{BB}(k_i, T) \) but not \( L_i(\text{atm} D) \). In order to calculate this term, Eq. (9) must be used with the Lowtran results at various angles, the approximate result is \( L_i(\text{atm} D) = 1.6 \cdot L_i(\theta = 0^\circ) \). The exact value of the factor depends on the channel and the atmospheric profile but it was found to range from 1.5 to 1.7. Then

\[
L_i(\text{surf}) = \varepsilon_i \cdot \text{BB}(T_{\text{grd}}) + (1 - \varepsilon_i) \cdot L_i(\text{atm} D).
\]  

(10)

It can be shown that the error introduced in determining \( L_i(\text{ref}) \) by assuming a Lambertian surface is on the order of \( 0.1 \Delta \rho / \langle \rho \rangle \), where \( \Delta \rho \) is the range of \( \rho_{bi} \) for \( \theta \) between \( 0^\circ \) and \( 90^\circ \) and \( \langle \rho \rangle \) is its hemispheric average. For soil Nerry et al. (1988) have shown that \( \Delta \rho \) is \( \approx 0 \) while for vegetation, it is about 0.2. Thus the magnitude of the error introduced by using Eq. (8) instead of (7) is 5% or less.

The Lowtran calculations were performed using atmospheric profiles obtained from radiosoundings released at the central site within 1 h of the pass (Parlange and Brutsaert, 1989). The resulting radiances and transmissions as functions of frequency were integrated over the TIMS filter functions to give the radiance and transmission for each channel. The channel radiance was divided by the integral of the filter functions \( \Delta k_i \) to yield the average spectral radiance for each channel. An example of the results is given in Table 1 for 16 June 1986 (Julian day 167) from an altitude of 1500 m. For this clear day the values of \( \tau_i \) ranged from a low of 0.69 for Channel 1 to 0.83 for Channels 4 and 5. The low value for Channel 1 results from the increased water vapor absorption at the 8 \( \mu \)m edge of the window, and this is also seen in the larger atmospheric radiances for this channel. The upwelling radiance is for the 1500 m

<table>
<thead>
<tr>
<th>Channel</th>
<th>Lake</th>
<th>Oats</th>
<th>Corn</th>
<th>Forest</th>
<th>Bare</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.059</td>
<td>20.7</td>
<td>0.061</td>
<td>22.7</td>
<td>0.078</td>
</tr>
<tr>
<td>2</td>
<td>0.068</td>
<td>21.2</td>
<td>0.071</td>
<td>23.5</td>
<td>0.081</td>
</tr>
<tr>
<td>3</td>
<td>0.075</td>
<td>21.3</td>
<td>0.078</td>
<td>23.9</td>
<td>0.100</td>
</tr>
<tr>
<td>4</td>
<td>0.093</td>
<td>21.6</td>
<td>0.063</td>
<td>23.9</td>
<td>0.120</td>
</tr>
<tr>
<td>5</td>
<td>0.102</td>
<td>21.4</td>
<td>0.106</td>
<td>23.6</td>
<td>0.135</td>
</tr>
<tr>
<td>6</td>
<td>0.113</td>
<td>21.4</td>
<td>0.117</td>
<td>23.5</td>
<td>0.144</td>
</tr>
</tbody>
</table>
layer below the aircraft. These are the values for the nadir view.

The average brightness temperatures computed with Eqs. (2) and (4), for four different land surface conditions and a small lake are given in Table 2. The fields were in a large clearing in the Les Landes forest in southwestern France near the small town of Lubbon. The lake is about 30 km south of the clearing. Figure 2 is a brightness temperature map of the clearing for day 167 from an altitude of 1.5 km.

The data for the oat field were from the coolest portion of the field where presumably the plants were transpiring at close to the potential rate. The corn field was just across the road and the plants were approximately 60–70 cm high with an LAI of 1.8 (Carlson et al., 1990). The data for the forest were from one of the cooler portions where we presume there was minimal bare ground showing. The trees were maritime pine about 20 m high.

The warmest portion of a rough plowed bare field was chosen under the assumption that it had the smallest amount of vegetation (weeds) or shadows. The data for the oat and bare fields were for a view angle of about 20° while the others were at approximately nadir. There are the values at the aircraft altitude of 1500 m and have to be corrected for the effects of the intervening atmosphere. As seen in Eq. (5), this is given by

\[ L_i(\text{surf}) = \left[ L_i(\text{a/c}) - L_i(\text{atmU}) \right] / \tau_i \]  

The Lowtran values of \( \tau_i \) and \( L_i(\text{atmU}) \) at an angle of 20° were used for the oat and bare field while the nadir values from Table 1 were used for the other fields. The results of this correction are given in Table 3. We see that the brightness temperatures of the vegetated surfaces are almost the same for all six channels whereas those for the bare field show considerable variation. The surface
Table 3. Spectral Radiances and Temperatures at the Ground $L_{surf}$ in 
$\text{W/m}^2\text{sr cm}^{-1}$, $T$ in $^\circ\text{C}$

<table>
<thead>
<tr>
<th>Channel</th>
<th>Lake $L$</th>
<th>Oats $T$</th>
<th>Lake $L$</th>
<th>Oats $T$</th>
<th>Lake $L$</th>
<th>Oats $T$</th>
<th>Lake $L$</th>
<th>Oats $T$</th>
<th>Lake $L$</th>
<th>Oats $T$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.060</td>
<td>22.0</td>
<td>0.064</td>
<td>25.0</td>
<td>0.089</td>
<td>42.8</td>
<td>0.067</td>
<td>27.6</td>
<td>0.103</td>
<td>51.6</td>
</tr>
<tr>
<td>2</td>
<td>0.069</td>
<td>22.0</td>
<td>0.072</td>
<td>24.9</td>
<td>0.098</td>
<td>42.4</td>
<td>0.076</td>
<td>27.6</td>
<td>0.113</td>
<td>50.8</td>
</tr>
<tr>
<td>3</td>
<td>0.076</td>
<td>21.9</td>
<td>0.080</td>
<td>25.1</td>
<td>0.107</td>
<td>42.1</td>
<td>0.083</td>
<td>27.5</td>
<td>0.120</td>
<td>49.7</td>
</tr>
<tr>
<td>4</td>
<td>0.060</td>
<td>22.2</td>
<td>0.094</td>
<td>25.0</td>
<td>0.128</td>
<td>44.5</td>
<td>0.112</td>
<td>27.5</td>
<td>0.148</td>
<td>55.2</td>
</tr>
<tr>
<td>5</td>
<td>0.103</td>
<td>22.0</td>
<td>0.108</td>
<td>24.6</td>
<td>0.143</td>
<td>44.6</td>
<td>0.112</td>
<td>27.5</td>
<td>0.165</td>
<td>55.4</td>
</tr>
<tr>
<td>6</td>
<td>0.114</td>
<td>22.1</td>
<td>0.119</td>
<td>24.9</td>
<td>0.154</td>
<td>43.9</td>
<td>0.124</td>
<td>27.5</td>
<td>0.175</td>
<td>54.5</td>
</tr>
</tbody>
</table>

brightness temperatures are shown in Figure 3 for the five surfaces. Using the radiosonde data the average air temperature of the $0-1.5$ km atmospheric column is calculated to be $19^\circ\text{C}$, which implies that the atmosphere will reduce the intensity observed at the aircraft level. The corrections for the cooler surfaces, for example, water or the oat field, are $1-2^\circ\text{C}$ whereas for the bare soil the corrections are $6-10^\circ\text{C}$. From the calibration point of view, it is reassuring to see that the range of temperatures among the six channels is $0.5^\circ\text{C}$ or less for the water or vegetated surfaces. The latter results imply that the emissivity for the heavily vegetated surfaces is virtually the same for all channels.

For comparison, the air temperature just above both the corn and oat fields was $27.7^\circ\text{C}$ and at the top of a similar stand of trees about $1$ km away was $25.9$. Thus the brightness temperature of the oat field was about $2.5^\circ\text{C}$ cooler than the air whereas for the forest it was $1-2^\circ\text{C}$ warmer. If we assume that the radiating surfaces of the plants are in thermal equilibrium with the surrounding air, these results imply that the emissivity of these vegetated surfaces is close to 1. Note that the intensities and thus the temperatures in Table 3 were not corrected for the reflected downwelling atmospheric radiation. This is not significant for the vegetated surfaces where $\varepsilon$ is close to 1, but as we will see later is significant for the bare soil.

The corn field with a mixture of vegetation and hot bare soil in the field of view displays a smaller range of intensity over the six channels. But this difference may be exploited to estimate the fractional vegetation cover, if we assume that the corn is radiating at the same intensity as the oats across the road and the bare soil between the rows is radiating at the same intensity as the bare field. Thus

$$L_i(\text{corn}) = f \cdot L_i(\text{oats}) + (1-f) \cdot L_i(\text{bare}). \quad (12)$$

Using the values in Table 3 yields a range of $f$'s between 0.34 and 0.39 for the six channels. This is in qualitative agreement with a ground photo taken 3 days earlier. A nearby corn field with shorter crops yielded a fraction between 0.15 and 0.18.

### EMISSIVITY

If the emissivity curves shown in Figure 1 are integrated over the filter functions for the TIMS, we obtain the emissivities for each channel. The results are given in Tables 4 and 5 along with the
Table 4. Lubbon Bare Soil Results for Days 167 and 178

<table>
<thead>
<tr>
<th>Channel</th>
<th>( \beta_i \cdot 167 )</th>
<th>( E_i \cdot 167 )</th>
<th>( \epsilon_i \cdot 167 )</th>
<th>( \epsilon_i \cdot 178 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.868</td>
<td>0.310</td>
<td>0.906</td>
<td>0.810</td>
</tr>
<tr>
<td>2</td>
<td>0.875</td>
<td>0.210</td>
<td>0.900</td>
<td>0.820</td>
</tr>
<tr>
<td>3</td>
<td>0.892</td>
<td>0.210</td>
<td>0.888</td>
<td>0.800</td>
</tr>
<tr>
<td>4</td>
<td>0.961</td>
<td>0.278</td>
<td>0.966</td>
<td>0.946</td>
</tr>
<tr>
<td>5</td>
<td>0.963</td>
<td>0.221</td>
<td>0.971</td>
<td>0.963</td>
</tr>
<tr>
<td>6</td>
<td>0.958</td>
<td>0.259</td>
<td>0.963</td>
<td>0.949</td>
</tr>
</tbody>
</table>

Table 5. Castelnau Emissivity Results

<table>
<thead>
<tr>
<th>Channel</th>
<th>( \epsilon_i \cdot 167 )</th>
<th>( \epsilon_i \cdot 167 )</th>
<th>( \epsilon_i \cdot 178 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.853</td>
<td>0.801</td>
<td>0.817</td>
</tr>
<tr>
<td>2</td>
<td>0.856</td>
<td>0.826</td>
<td>0.844</td>
</tr>
<tr>
<td>3</td>
<td>0.859</td>
<td>0.811</td>
<td>0.886</td>
</tr>
<tr>
<td>4</td>
<td>0.932</td>
<td>0.925</td>
<td>0.947</td>
</tr>
<tr>
<td>5</td>
<td>0.946</td>
<td>0.946</td>
<td>0.946</td>
</tr>
<tr>
<td>6</td>
<td>0.946</td>
<td>0.944</td>
<td>0.923</td>
</tr>
</tbody>
</table>

Aircraft estimates of \( \epsilon_i \) for the soils from the two sites.

The remaining problem is to relate the temperatures and intensities in Table 3 to these laboratory emissivity measurements without direct knowledge of the soil temperature, \( T_{grd} \). From Eq. (10) we obtain

\[
\epsilon_i = \frac{E_i}{(1 - \beta_i)}, \quad (13)
\]

where

\[
E_i = \frac{L_i(\text{surf})}{BB(k_i, T_{grd})}
\]

is an apparent emissivity and

\[
\beta_i = \frac{L_i(\text{atmD})}{BB(k_i, T_{grd})}
\]

is the correction factor for the downwelling radiation. It ranges from 0.2 to 0.5 in our data. To obtain an estimate of \( T_{grd} \), Eq. (10) is solved using \( \epsilon_i \cdot \text{lab} \) for the channel with the highest laboratory emissivity, that is, Channel 5. This yielded a \( T_{grd} = 57.7^\circ C \), and the resulting values of \( \beta_i \), \( E_i \), and \( \epsilon_i \) can be calculated. The results are given in Table 4 and the values of \( \epsilon_i \) are plotted in Figure 4.

A comparison of the values of \( E_i \) and \( \epsilon_i \) in Table 4 show the importance of including the effects of the downwelling atmospheric radiation for the channels with the lower emissivities. The values of \( \epsilon_i \) should be compared with the measurements for the Lubbon soil in Table 4 and we see that the agreement is reasonably good with average difference of 0.01 for the hot spot. Recall that these values are relative to that for Channel 5. However, when the analyses were repeated for 27 June 1986 (day 178), which had similar weather conditions, the agreement was poorer. The shorter wavelength channels were lower by about 0.04. Similarly, when the analysis was done for a cooler portion \( (T_{grd} = 53.8^\circ C) \) of the field, which was more representative of the entire field, the results were closer to those for day 178. For that day there was little difference in emissivity between the two portions of the field despite a 3°C difference in \( T_{grd} \) between the two portions of the field.

The analyses were also performed for a corn field near the town of Castelnau which is about 65 km south of the Lubbon clearing. There may be some difficulty in using the same atmospheric profile at this separation, particularly at the lower levels. The emissivity results for the days 167 and 178 are given in Table 5 and in Figure 5. From the table we see the expected good agreement for the upper three channels. However, with exception of Channel 2, the agreement is not as good between the two days for the lower channels. A portion of this could be due to atmospheric effects because the sounding used was taken at the Lub-
bon site 65 km north, and we expect that air was drier at the southern site.

We also have applied the procedure described in Becker and Li (1990), which also uses the value of emissivity in one of the channels as a reference. Again the value for Channel 5 was used and the results are in good agreement with those presented in Tables 4 and 5.

As noted above, these values of observed emissivity depend on the atmospheric values used, that is, transmission and the up- and downwelling radiations. Thus uncertainties in these quantities will cause variations in the resulting emissivities. Also taking data from different portions of the bare field affects the emissivity values. This may be the result of differences in the roughness, shadowing, and vegetation cover. It is not clear at the present time why these factors should affect the spectral variation so strongly.

**APPRAOXIMATE METHODS**

For small temperature variations, Slater (1980) has shown that the radiance \( L_i \) in Channel \( i \) can be approximated by

\[
L_i = \text{BB}_i(T) = a_i \cdot T^{n_i}.
\]

Values of \( n_i \) were obtained for the six TIMS channels over the temperature interval 285–325 KI are given in Table 6. This power law relation can be used to obtain relationships among the emissivities for the six channels. Thus, if we temporarily neglect the reflected downwelling atmospheric radiation

\[
\text{BB}_i(T_{g_i}) = E_i \cdot \text{BB}_i(T_{g_{rd}}),
\]

which with Eq. (14) leads to

\[
T_{g_{rd}} = E_i^{-1/n_i} \cdot T_{g_i}.
\]

By taking the ratio of two channels, we obtain

\[
E_i^{1/n_i} / E_j^{1/n_j} = T_{g_i} / T_{g_j},
\]

which yields

\[
E_j = E_i^{n_j/n_i} \cdot (T_{g_j} / T_{g_i})^{n_j},
\]

where \( E_i \) and \( E_j \) are the apparent emissivities and related to the emissivities \( \varepsilon_i \) by Eq. (13), which effectively takes into account the reflected downwelling atmospheric radiation. The results from Eq. (17) are given in Table 6 along with those from the exact method described earlier and are seen to be in very good agreement. Similar agreement was found for other cases. A graphical comparison of the results from the two methods can be seen in Figure 4 for the Lubbon hot spot.

Equation (16) can be used to show that the average brightness temperature of several pixels can be used to determine emissivities. For the same type of surface in an image, the emissivity \( \varepsilon_i \) may be constant but a large range of surface temperature \( T_{g_{rd}} \) can exist which leads to the variation of brightness temperature \( T_{g_i} \) at the ground. For pixel \( k \) in channel we can write

\[
T_{g_{rd,k}} = \varepsilon_i^{-1/n_i} \cdot T_{g_i,k}
\]
By summing over $N$ pixels, we can define average ground and brightness temperatures:

$$\langle T_{grd} \rangle = \left(1/N\right) \sum T_{grd,k}$$

and

$$\langle T_{gi} \rangle = \left(1/N\right) \sum T_{gi,k},$$

yielding

$$\varepsilon = \varepsilon' - \delta' \left(\frac{T_{gi}}{T_{gi,j}}\right)'q.$$  \hspace{1cm} (18)

Thus, if an image has been well-classified \textit{a priori}, we can retrieve the spectral emissivity information from the average data according to Eq. (18).

**DISCUSSION AND CONCLUSIONS**

We have shown that it is possible to observe the spectral behavior of the emissivity for various surface using the TIMS data from an aircraft platform. As expected, fully vegetated and water surfaces show little or no emissivity variation while our "bare" soil cases had considerable variation. The observed variations were at least as large as those expected from laboratory measurements made on the same soils and usually larger. The relative values observed for the shorter wavelength channels, 1–3, were on the order of 10% below the laboratory measurements for the same soils. This large a difference was unexpected. Our results have shown a large sensitivity to the values used for the atmospheric corrections; this aspect of the problem requires further study.

During the course of this data analysis, we switched from using Lowtran-6 to Lowtran-7 for the atmospheric corrections. In doing this we noticed a significant reduction in the range of temperatures ($\Delta T$) for the six TIMS channels over surfaces where we expect to have little or no spectral emissivity variation, that is, the lake or heavily vegetated fields. For the lake $\Delta T$ was reduced from 0.75°C to 0.33°C, with similar reductions noticed for the oat field and the forest. We believe that most of this reduction was due to improved corrections for Channel 1. This switch in Lowtran versions also had a small effect on the resulting emissivities for the bare soils. The value of $\varepsilon$ for Channel 1 increased by up to 5% and by $\approx 1\%$ for Channel 2. The effect on the other channels was insignificant.

Earlier we had noted that error introduced in making the assumption that the surfaces were Lambertian reflectors was on the order of 5%. This is comparable to the variation in using the approximation $L_i(atmD) = 1.6 \pm 0.1 \cdot L_i(\theta = 0')$ in Eq. (8). Allowing for this variation in $L_i(atmD)$ resulted in a $\pm 1\%$ variation in the emissivity values.

To study the effect of errors in the values of $T_{gi}$, the surface brightness temperatures, we studied the effect of a $1\degree C$ variation in Eq. (17) on the emissivity. We found that temperature error of this magnitude would produce a 3% uncertainty in the emissivity values.

Our bare soil results indicate that a 8.5–9.5 $\mu$m channel would be very useful for discriminating various targets in the 8–14 $\mu$m window. Such a channel would greatly improve the ability to discriminate bare soils in the thermal infrared. This agrees with the conclusion of Hovis et al. (1968). Also the relative flatness of the emissivity in the 10–12 $\mu$m region may make possible the use of split window techniques for eliminating atmospheric water vapor effects in the estimation of land surface temperatures.

**REFERENCES**


Hoover, G. (1990), Temperature errors in TIMS caused by air blast, Internal note, Jet Propulsion Laboratory, Pasadena, CA.


