ASTEROBSERVATIONS FOR THE MONITORING OF
LAND SURFACE FLUXES

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ABSTRACT -- This paper presents a review of how data from the Advanced Spaceborne Thermal Emission radiometer (ASTER) can be used to estimate the energy fluxes from the land surface. The basic concepts of the energy balance at the land surface are presented along with an example of how remotely sensed surface brightness temperatures can be used to estimate the sensible heat. The example is from the Monsoon 90 experiment conducted over an arid watershed in the state of Arizona in the United States.

INTRODUCTION

The monitoring of the land surface fluxes at regional spatial scales is recognized as important for applications such as the modeling of atmospheric behavior and the monitoring of water resources. When one looks at a thermal infrared image, it is clear that the large variations in surface brightness temperatures, $T_B$, which are seen arise from differences in the surface energy balance for the land surfaces. Recall that $T_B$ is a measure of the emitted radiation from the surface that is directly related to the temperature of the surface and its emissivity. The temperature contrasts seen between fields with different vegetation conditions imply a different partition of the incoming solar energy into latent and sensible heat components. Cooler temperatures usually indicate that there is sufficient moisture available so that most of the incoming energy goes into latent heat or evaporation, while hotter temperatures indicate that most of the incoming energy goes into the sensible or convective heating of the atmosphere. However there is another aspect of the problem of determining the surface fluxes which arises due to the differences in the heat transfer coefficients arising from the various types of vegetation and their heights. Thus a 20 m tall pine forest can be just as cool as a well-watered field but transpiring much less because of its greater heat transfer capability. The problem is to quantify these fluxes in terms of the remotely sensed $T_B$ and other observables from the Advanced Spaceborne Thermal Emission Radiometer, ASTER, to be flown on NASA’s Earth Observing System (EOS) first spacecraft to be launched in 1998. There is a long history on the use of $T_B$ to monitor these surface fluxes and in this paper we will describe the contributions remotely sensed data from ASTER can make towards quantifying these fluxes.

SURFACE ENERGY BALANCE

To estimate the land surface fluxes it is necessary to determine:
1) The energy driving forces, i.e. the incident solar energy (insolation), surface albedo and resulting net radiation;
2) The moisture availability or status in the soil and the vegetation/soil interaction; and
3) The capacity of the atmosphere to absorb the flux, which depends on surface air temperature, vapor pressure gradients, and surface winds.

There has been considerable work recently on methods for estimating factors relating to the first two items from remotely sensed data. It is possible to estimate from remotely sensed data the surface parameters related to the soil/vegetation system (leaf area indices and surface soil moisture), radiation forcing components (essentially incident solar radiation and albedo) and indicators of the surface response to them (surface temperature). However there is no remote sensing method for estimating the surface atmospheric parameters. Therefore conventional surface measurements will be required for this factor.

To understand better how ASTER observations can contribute to the determination of the surface fluxes, let us consider the basic energy and moisture balance equations. In the absence of advection or precipitation the energy balance at the land surface is given by:

$$R_n + G + H + LE = 0$$  \hspace{1cm} (1)

where $R_n$ is the net radiation, $G$ the soil heat flux, $H$ the sensible heat flux and LE the latent heat or moisture flux into the atmosphere. The net radiation is the sum of the incoming and outgoing short and long wave radiation fluxes:

$$R_n = (1 - \alpha) \cdot R_s + (1 - \epsilon) \cdot R_{LI} - \epsilon \sigma T^4$$  \hspace{1cm} (2)

where $\alpha$ is the surface albedo, $R_s$ is the incoming solar radiation, $R_{LI}$ is the incoming long-wave radiation, $\epsilon$ the surface emissivity and $T$ the surface temperature in Kelvins.

There has been considerable progress in estimating $R_s$ and $\alpha$ from geostationary satellite data. The basis of the method is that the major modulator of surface insolation is cloudiness. The information contained in the satellite radiances is interpreted in terms of scattering, reflection and absorption parameters which are subsequently used in radiative transfer model calculations for the atmosphere [4]. The albedo $\alpha$, can also be estimated from these data. $R_{LI}$ can be estimated from the atmospheric sounders or empirically from surface conditions. The surface temperature $T$ can be estimated from the thermal channels of ASTER. The surface albedo can also be estimated from multi-spectral data such as that which will be available form the VNIR channels of
The latent heat flux involves the vapor pressure gradient. Thus for the ground heat flux we have:

\[ G = \lambda dT_{soil} / dz \]  

(3)

where \( \lambda \) is thermal conductivity of the soil. While the soil temperature gradient cannot be determined from remotely sensed data the temperature profile can be modeled with sufficient accuracy to estimate \( G \). There is a reasonable empirical relationship between the ratio \( G/R_n \) and vegetation indices such as and NDVI.

The sensible heat flux into the atmosphere is:

\[ H = \rho c_p \left( T_{aero} - T_a \right) / r_s \]  

(4)

where \( \rho \) is the air density, \( c_p \) is the specific heat of air at constant pressure, \( T_{aero} \) is the aerodynamic temperature in the canopy, \( T_a \) is the air temperature just above the canopy, and \( r_s \) is the aerodynamic resistance. The latter, \( r_s \), is a rather complex function of various geometrical factors such as roughness lengths, displacement heights, etc., and the wind speed and is usually empirically determined. \( T_{aero} \) is the temperature of the source for the convective heat transfer and can be determined from the profiles of temperature and windspeed in the boundary layer. For an aerodynamically smooth surface, \( T_{aero} \) and \( T_B \) are equivalent since such a surface is the source for both the radiative and convective sensible heat fluxes. However for most natural surfaces, \( T_{aero} \) and \( T_B \) are not equivalent as demonstrated by Hall et al. [6] working with data from FIFE. This difference between \( T_{aero} \) and \( T_B \) is thoroughly discussed in the recent paper by Norman et al. [7].

The latent heat flux into the atmosphere is given by:

\[ \text{LE} = \rho c_p \left( e_a - e_s \right) / \gamma \left( r_s + r_g \right) \]  

(5)

where \( \gamma \) is the psychrometric constant, \( e_a \) is the atmospheric vapor pressure in the boundary layer, \( e_s \) is the saturation vapor pressure at the temperature \( T \), and \( r_g \) is the stomatal resistance to water vapor transport. To get around the absence of these temperature and vapor pressure gradients in the soil and lower atmosphere 1-D models have been developed of the heat transfer in the soil and from the land surface into the atmosphere [8]. These models use the remotely sensed surface temperature as a boundary condition and model parameters are varied to obtain the best fit between the predicted and observed surface temperatures.

From (1) to (5), we see that the available energy (\( R_n - G \)) is partitioned into latent and sensible heat components. The approach that is frequently used is to estimate the \( H \) using the remotely sensed surface brightness temperature (\( T_B \)) in (4) and to determine LE as a remainder term. This is the approach we will consider here.

Equation (4) implicitly assumes that the canopy is a thin layer with a single temperature, which is clearly not the case for a vertically developed canopy exhibiting variations in temperature. While both \( T_{aero} \) and \( T_B \) result from contributions of the surface temperatures of the canopy elements, they do so in different ways and as a result are not equivalent. The problem is that \( T_{aero} \) in (4) being the effective temperature for the canopy heat transfer process is not a measurable quantity and is not equal to \( T_B \). The thermal radiation observed by a radiometer originates from the soil and the vegetation elements having various temperatures and orientations which leads to variations in \( T_B \) with viewing angle. If the classical value of \( r_s \) is used in (4) with \( T_B \), it is found to give poor agreement with observed fluxes [6]. However Stewart et al. [9] have shown that if an additional resistance, \( r_p \), is added the agreement improves substantially. This additional resistance is necessary because \( T_B \) is larger than \( T_{aero} \) and is needed to account for the increased resistance to heat flow compared to that for the momentum flux.

RESULTS

In view of these difficulties in determining the appropriate surface resistance for heat transfer, it is clear that at the present time it may not be possible to estimate \( H \) from remotely sensed observations alone. However if we have a reference point where flux measurements are made it is possible to determine the spatial variation of \( H \) from TIR data. One example this approach is available from the Monsoon'90 experiment [10] conducted over the arid Walnut Gulch watershed in Arizona during the summer (Monsoon season) of 1990. A summary of the conditions from that experiment is given in Fig. 1, which presents box plots of the surface air temperature and surface radiation temperature for 8 sites over the watershed. The data are for 3 days which represent a range of conditions. The TIR radiation temperatures (\( T_{surf} \)) are from the NS001 sensor on board the NASA C-130 aircraft. The bandpass for the TIR channel is approximately 10 to 12 \( \mu \)m. The data are from 2 flight lines at an altitude of 2400m above the ground, yielding a pixel size of 6.3 m. The data were acquired between 10:00 and 10:30 local standard time, i.e., close to the time of the ASTER overflights. The temperatures were corrected for atmospheric effects using radiosoundings launched at the site, and the results were found to agree with ground radiometer measurements to within 1 K. The boxes show the median and range of \( T_{surf} \) and \( T_{air} \) values for each day. The air temperatures were relatively consistent for the 3 days about 25 C while the temperatures over the watershed. The data are for 3 days which represent a range of conditions. The TIR radiation temperatures (\( T_{surf} \)) are from the NS001 sensor on board the NASA C-130 aircraft. The bandpass for the TIR channel is approximately 10 to 12 \( \mu \)m. The data are from 2 flight lines at an altitude of 2400m above the ground, yielding a pixel size of 6.3 m. The data were acquired between 10:00 and 10:30 local standard time, i.e., close to the time of the ASTER overflights. The temperatures were corrected for atmospheric effects using radiosoundings launched at the site, and the results were found to agree with ground radiometer measurements to within 1 K. The boxes show the median and range of \( T_{surf} \) and \( T_{air} \) values for each day. The air temperatures were relatively consistent for the 3 days about 25 C while the surface temperatures show considerable variation resulting from the predicted and observed surface temperatures.
Figure 1. Boxplot of flux and temperature data from Monsoon 90 experiment. The vertical axis is in degrees C.

top for each day. It is clear that the magnitude of $H$ is proportional to the difference between the average $T_{surf}$ and the average $T_{air}$ values. Maps of these surface temperatures were used with an average value of $r_f$ derived from measured values of $H$, $T_B$, and $T_{air}$ at the 8 ground flux stations. The spatial variations of $r_f$ were used to derive the spatial variation of $H$ over the watershed. The reader is referred to [10] for the details of the procedures and the maps of the sensible heat flux.

CONCLUSIONS

In this paper we have reviewed the physics of the energy balance at the land surface and the factors that can be accessed through remote sensing. These include the incoming solar radiation, $R_s$, and the surface albedo, $\alpha$, from the VNIR channels of ASTER. The longwave radiation components, $R_L$ and $\epsilon\sigma T^4$, can be estimated from temperature sounding instruments. The primary contribution that ASTER can make to surface flux determinations will be through the thermal infrared channels. The problems with estimating the sensible heat flux were discussed and an example from the Monsoon 90 experiment was presented. By making use of a reference flux site a map of the spatial variation of $H$ was generated from the $T_B$ map and the range of values is in agreement with ground observations. These results indicate that at the present it may not be possible to determine the surface energy fluxes from remotely sensed data alone but that the spatial variation can be determined from $T_B$ maps.

REFERENCES


