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## 1. Introduction

The previous year saw a change in emphasis from purely modeling to combining modeling of surface energy fluxes with in situ measurement of surface parameters, specifically the surface sensible heat flux and the substrate soil moisture. In France we were able to incorporate a vegetation component into our atmospheric/substrate model and subsequently to show that the fluxes over vegetation can be very much different than those over bare soil for a given surface-air temperature difference. Results of this work are summarized in an article by Taconet et al. (1986b).

Because of my interaction with members of the CRPE/CNET group in France we feel that a high priority must be given to interpreting the temperature signatures measured by a satellite or airbourne radiometer in conjunction with surface measurements of modelled parameters, e.g., surface sensible or latent heat fluxes and soil moisture. Paradoxically, our analyses of the large-scale distribution of soil moisture availability show that there is a very high correlation between antecedent precipitation and inferred surface moisture availability, even when no specific vegetation parameterization is used in the boundary layer model (Carlson, 1985; Flores and Carlson, 1985; Flores and Carlson, 1987). A similar result was obtained by Wetzal and Woodward (1986).

Accordingly, we have continued to maintain an effort in modeling the surface canopy but with a greater emphasis on verification. One weak link in the theory pertains to the transfer of radiant and heat fluxes in vegetation canopies. The amounts of radiant energy incident on the vegetation and on the soil beneath depend on the density of vegetation, normally parameterized using the leaf area index or similar quantity. We have also begun to test simple models based on the full boundary layer model but which require a minimum of observations. We have begun to look closely at observations, including

### 3. Conclusion

Our major scientific findings are summarized in three publications attached to this report: the long-awaited review article in Remote Sensing Reviews, a discussion of preliminary results by E. Perry of our infrared/microwave measurement comparisons from the Beauce experiment and a summary of the results of modelling experiments done in France by the PI in conjunction with the CRPE/CNET boundary layer model. The latter paper is merely an extended abstract of a fourth publication, which is soon to appear in the Journal of Climate and Applied Meteorology (Taconet et al., 1986b), concerning the modelling done in France. Much of the time spent during the last year involved preparatory work in streamlining the present boundary layer model, developing better algorithms for relating surface temperatures to substrate moisture, preparing for participation in the French HAPEX experiment, and analyzing aircraft microwave and radiometric surface temperature data for the 1983 French Beauce experiments.

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16

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SOIL MOISTURE ESTIMATES FROM SATELLITE INFRARED TEMPERATURES  
AND THEIR RELATION TO SURFACE MEASUREMENTS

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## 1. INTRODUCTION

Spatial and temporal variations in soil moisture are reflected by variations in the surface temperature. Exact measurements of the soil moisture using remote measurements of thermal infrared temperatures of the surface are possible, in theory, but very difficult to make in practice because of the enormous complexity of the ground surface, especially in the presence of vegetation. Our approach is to invert a time-dependent, initial-valued, one-dimensional boundary layer model in conjunction with measured surface temperatures to obtain the surface energy balance, a soil moisture parameter, called the moisture availability, and the thermal inertia. For the theory behind the model, the reader is referred to papers by Carlson and Boland (1978) and Carlson et al. (1981; henceforth referred to as CM). Similar models have been used to obtain soil moisture values from satellite infrared measurements by Wetzel et al. (1984) and Price (1982) in the United States and by Taconet et al. (1986a,b; henceforth referred to as OT) and Abdellaoui et al. (1985) in France and by Nieuwenhuis et al. (1985) in Holland. Recently, modelers have begun to treat vegetated surfaces by including parameters that govern the fluxes between the soil/plant/atmosphere system. The vegetation component in both the CM and OT are nearly identical, being based on an earlier version by Deardorff (1978), but simplified and streamlined by Bernard et al. (1986) using data for sorghum.

The purpose of this paper is to report some of the recent results showing agreement between moisture values calculated with the model using satellite infrared temperatures as input and ground-based measurements of related parameters.

## 2. RESULTS

### 2.1 Soil Moisture Versus Antecedent Precipitation as Measured by GOES

Twenty days of satellite images were analyzed for a period during the summers of 1978 and 1980 over Kansas. The region and dates were chosen because of the extreme drought conditions that were affecting a wide region of the midwest extending from Kansas to Texas. We reasoned that if the infrared method was viable it should prove so under conditions of great horizontal variation in soil moisture and rainfall. Accordingly, the parameter called moisture availability ( $M$ ) was determined over a region a few hundred kilometers on a side using GOES daytime and nighttime infrared temperature measurements (Carlson et al., 1984). The vegetation model was not used for this study.

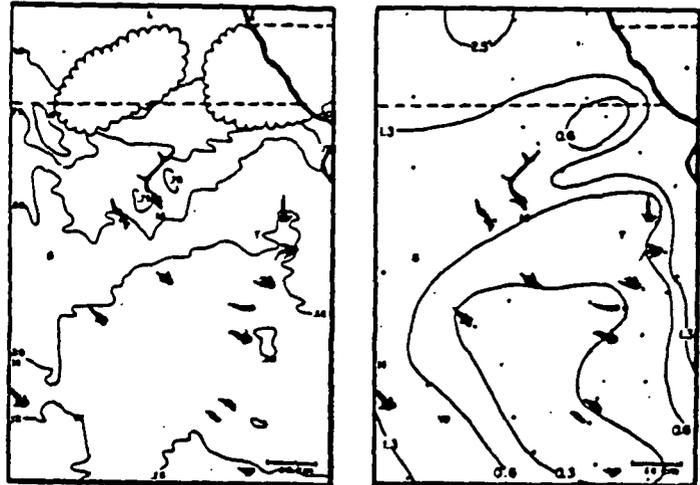
Moisture availability is defined as the ratio of evaporation to potential evaporation and is also equal in the model to the ratio of soil water content ( $\theta$ ) over a layer near the surface to the value of soil water at field saturation ( $\theta_{fg}$ ). Thus,  $M$  varies from 0 (for perfectly dry surfaces) to 1.0 for saturated surfaces. In the example shown in Fig. 1, the vegetation model was not used to derive the results. Consequently, the prairie surface is treated as bare soil. Contours of  $M$  in Fig. 1 closely reflect the surface temperature over the region of Kansas on the afternoon of the 14th of July, 1980. There is an overall similarity on all the days between the  $M$  field and antecedent precipitation, which is effectively a running mean of the rainfall. Low values of  $M$  (0.12-0.25) in the south correspond roughly to low rainfall and high values in the north to higher rainfall totals. The correlation between the log of  $M$  and the antecedent precipitation is about 0.7 for all cases. A similar relation was found by Flores and Carlson (1985) for scenes over Texas on three days during the same month of July, 1980.

### 2.2 AVHRR Images and Surface Measurements Made Over France

During a field measurement program conducted over the Beauce region of France (near the town of Voves) by scientists at two French national laboratories (INRA, CRPE), there occurred a remarkable increase in the surface sensible heat flux during a period of drying. Although the drying trend had been continuing a couple of weeks, the increase (measured by both the SAMER method of Itier (1981) and by sodar) began just after the 6th of July and was most abrupt between the 11th and 14th (Fig. 2).

Heat flux values derived from the CM and OT models (using the same set of initial conditions) using afternoon temperatures (approximately 1400 sun time) from NOAA AVHRR as input also show a similar trend, although the version of CM with bare soil was less able to capture the increase in sensible heat flux with time.

Fig. 1. Moisture availability (M) analysis over a portion of eastern Kansas and Nebraska for 14 July 1980 based on surface temperature measurements for a pair of GOES images (left). Dark, irregular shapes are bodies of water, including the Missouri River near upper right. Scalloping denotes regions obscured by cloud. Antecedent precipitation for 14 July 1980 is shown on the right. Dots indicate location of rainfall measurement stations.



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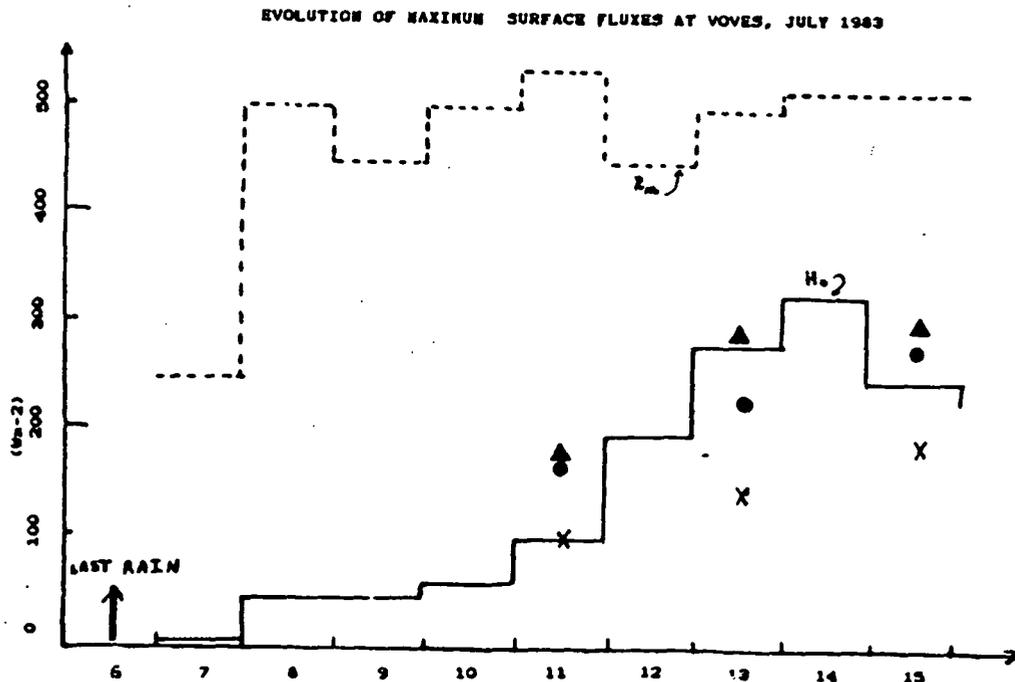


Fig. 2. Evolution of maximum net radiation (top curve dashed) and maximum surface sensible heat flux (bottom curve; in  $\text{Wm}^{-2}$ ) for Voves during during a period in July 1983. The last rainfall before the end of the record occurred on the 6th of July. Symbols refer to maximum heat flux values determined from the models using AVHRR data as surface temperature input: the CM for bare soil (X), the CM for vegetation (O) and OT for vegetation (A).

Soil water content ( $\theta$ ) measured with a neutron probe near the Voves site continued only up until the 12th of July. Averages of  $\theta$  over four depths (Fig. 3) show a decrease at all levels down to about 40 cm below the ground; the most rapid drying, of course, occurred in the top 30 cm, where levels of water content were approaching the wilting point of the soil (approximately  $0.15 \text{ cm}^3/\text{cm}^3$ ; field saturation was about  $0.35 \text{ cm}^3/\text{cm}^3$ ).

Model results also show a corresponding decline with time in soil water content, although more precipitiously than the measurements made earlier in the period suggest. Curiously, although the model

water contents pertain to the vegetation parameterization, the bare soil values for CM were virtually identical to those shown in Fig. 3. These values correspond to very low moisture availability (approximately 0.2 on the 15th), which are comparable to those in the dryer parts of the Kansas image of Fig. 1. The puzzling fact here is that the surface measurements seem unassailable and are supported by the model results. At the same time it is to be noted that the wheat crop over which these measurements were made did not appear to be highly water stressed.

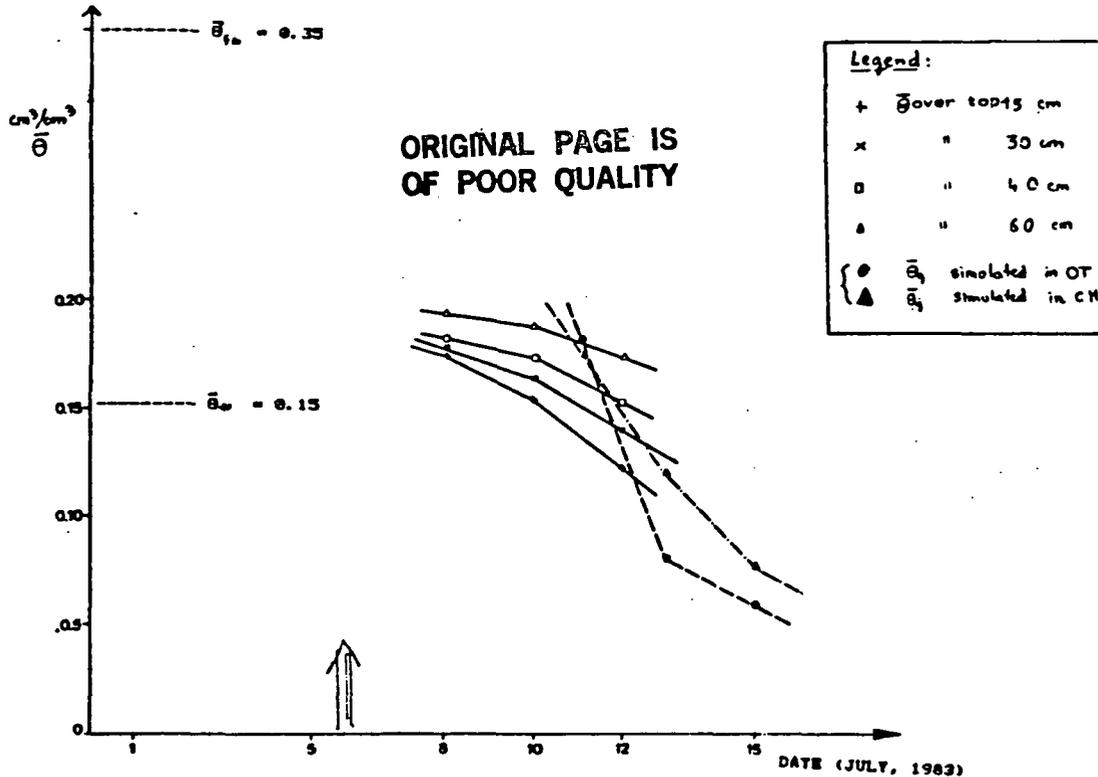


Fig. 3. Measured (solid lines) soil water content ( $\text{cm}^3/\text{cm}^3$ ) versus time for averages over four vertical soil depths as given in the key. Soil water content ( $\theta$ ) determined from the model using the 1346 GMT AVHRR surface temperature measurement is shown by the dashed and dot-dashed lines, respectively, for the OT and CM. The vertical arrow denotes the last rainfall during the period.

### 3. CONCLUSION

Infrared temperatures measured with the aid of satellite can reflect large spatial and temporal changes in soil moisture but not enough is known and the parameterization sufficiently undeveloped to arrive at a completely consistent measure of the real soil water content or the soil depth over which it applies.

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CORRELATIONS BETWEEN REMOTELY MEASURED SOIL WATER CONTENT AND  
SURFACE TEMPERATURES FOR THE BEAUCE REGION, FRANCE

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## 1. INTRODUCTION

Soil water content has been shown to be highly correlated with radiometric surface temperatures, and a number of studies have demonstrated that the surface temperature increases as the substrate soil water content decreases. (See Hatfield et al. (1983) and Carlson (1984) for a summary of articles on this research.) As part of a study to model the available water in the unsaturated zone using thermal infrared data, this paper is a first look at some aircraft measurements that show a correlation between microwave measured soil water content and radiometric surface temperatures.

During an experiment held in the summer of 1983 near Voves, France, active microwave and thermal infrared measurements were made during the early afternoon aboard a French aircraft flying a grid pattern over a flat, agricultural region south of Paris called the Beauce. The vegetation consisted mainly of wheat and corn, with some stands of trees. The infrared sensor was a Barnes PRT-5 radiometer which has a measurement accuracy to within .5 degrees celsius. The active microwave measurements were taken with the ERASME scatterometer (5.35 GHz (FM-CW) with an incidence angle of eleven degrees for minimizing the effects of roughness and therefore maximizing the accuracy of soil water measurement. Calibration missions of the microwave measurements were performed with reference to in-situ measurements of soil water content; the correlation coefficient was .89 between the microwave back scatter coefficients and the ground measurements of the soil water content in the first ten centimeters. Accordingly, soil water content was converted directly from backscattering cross section to soil water content. The scatterometer and radiometer were mounted so that the two instruments were taking measurements of the same ground area simultaneously, five measurements per second, with each having a surface sampling area of approximately 40 meters squared. The average flight altitude was about 400 meters.

Data was collected during July and September of 1983; the dates in July were 8 and 12 July during a period of drying following a 6 July rainfall episode; the September dates were the 20, 23, 26, 28 and 29 September following a 19 September rainfall episode. Data segments along two of the aircraft legs were selected to discuss because they represent alternately the poorest and best correlation of surface temperatures with microwave-derived soil water content measurements, respectively on 8 July (1:45 LT) and 26 September (3:45 LT).

To examine the data through time and perform some descriptive statistics, a statistics/graphics package (SAS) was used to plot the microwave and thermal IR data, and to perform correlations between them. Figure 1 shows a plot of the temperature response as the sensor was flown over a west to east transect near the town of Voves on 8 July. Figure 2 shows a plot of the soil moisture values measured along the same transect. In order to gain a simple visual comparison between the temperature and microwave measurements, the two lines are overlain on the same plot, using a cubic spline to fit the points (see figure 3). Note the inverse relation between the microwave backscatter and the surface temperature at the points where the sensor was flown over road surfaces (A, B, D, L, M, N on figure 3) which are very dry and have a high surface temperature. Figure 4 is a transect flown on 26 September, over approximately the same area as in figure 3. Note that for the same roads (points B, D, L, M), the data again show that a low soil water content correlates with a high surface temperature. However, for the 26 September case, this inverse relation is demonstrated over the entire transect. For both of the cases, the subset area labelled on figures 3 and 4 were plotted to show the variations of the microwave and thermal infrared response on a field basis (see figures 5 and 6).

## 2. RESULTS AND DISCUSSION

The first and most important observation is the poor correlation between the microwave backscatter coefficients and the thermal infrared measured temperatures for the 8 July case (shown in figure 3). The correlation coefficient for the entire data segment is .3459. The correlation between these same data types is -.7795 for the 26 September case, which is what would be expected given previous research. Possible explanations (not yet confirmed) for this lack of correlation of the vegetated 8 July case are:

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i) For the July case, the plants may be in effect "integrating" the available water because they are using the water at depth (in the root zone). If the surface temperature was being influenced by the amount of water available at a depth of perhaps 45 cm or more, and the variability of this soil water content is not as great as that at 0-10 cm (being measured by the microwave), there would be a difference in the variability of the surface temperature and the microwave measurements. This difference in variability could decrease the correlation between the remotely measured temperature and soil moisture for any given point.

ii) There could be a difference in the "measurement depths" between the July and September cases. For the July case, the thermal infrared sensor is measuring temperatures that are largely influenced by the water available to the plants in the root zone (much deeper than 10 cm), while the microwave backscatter coefficients are most highly correlated with the top ten centimeters. For the September case, with nearly bare soil conditions, the thermal infrared temperatures are more influenced by the soil water content near the surface, and the microwave backscatter coefficients are again most correlated with the soil water content in the first ten centimeters.

iii) If the stomates were closed for some reason other than lack of available water, the plant temperatures measured by the thermal infrared sensor could be much higher than what would be expected for a given soil water content.

Another observation is the possible lack of a significant correlation between remotely measured soil water content and surface temperatures within an individual field. For one corn field (field 10) there is a correlation coefficient of  $-0.326$  for 8 July and  $-0.484$  for 26 September. For the forest stand (field 7), the correlations are  $+0.403$  for 8 July and  $-0.18$  for 26 September (see Table I).

Finally, it can be seen from figures 5 and 6 that there is a greater variation of microwave backscatter coefficients than surface temperature. This would tend to support the idea that the variability of soil moisture is greater in the surface layers of the soil.

Table 1. CORRELATION COEFFICIENTS BETWEEN TEMPERATURE AND SOIL MOISTURE

DATE	FIELD COVER	SURFACE TEMPERATURE	MICROWAVE	N	CORREL.
8 July	all data (mixed cover)	29.7 +/- 1.98	-2.10 +/- 2.26	9165	$-0.004$
	segment 3 (mixed)	30.1 +/- 1.98	-1.87 +/- 1.77	1004	$.346$
	segment 3 subset (mixed cover)	30.4 +/- 1.15	$-0.660$ +/- 1.52	218	$-0.170$
	field 10 (corn)	29.1 +/- $.308$	$-0.051$ +/- $.984$	68	$-0.326$
	field 7 (woods)	29.2 +/- 1.07	$-2.80$ +/- 1.23	29	$.403$
26 Sept.	all data (mixed cover)	31.3 +/- 2.79	$-3.98$ +/- 2.82	7556	$-0.712$
	segment 3 (mixed)	31.4 +/- 3.02	$-3.15$ +/- 2.82	926	$-0.780$
	segment 3 subset (mixed cover)	30.8 +/- 2.92	$-2.94$ +/- 2.69	232	$-0.722$
	field 10 (stubble)	26.8 +/- $.537$	$.258$ +/- $.755$	73	$-0.484$
	field 7 (woods)	32.4 +/- $.538$	$-4.61$ +/- $.642$	24	$-0.181$

Temperature (degrees celsius)

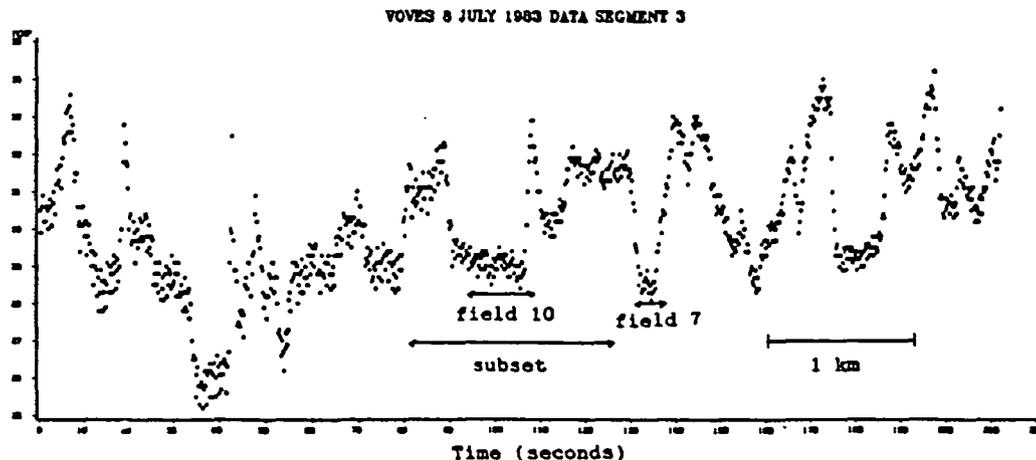


Figure 1. West to east tranverse showing the thermal infrared response. Time of day approximately 1:45 LT Note locations of fields 10 and 7, and the location of the subset to be examined later.

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Temperature (degrees celsius) - - - - -  
Soil water content (cm<sup>3</sup>/cm<sup>3</sup> x 10<sup>2</sup>) \_\_\_\_\_

VOVES 8 JULY 1983 SEGMENT 3  
SOIL WATER CONTENT AND TEMPERATURE VS TIME

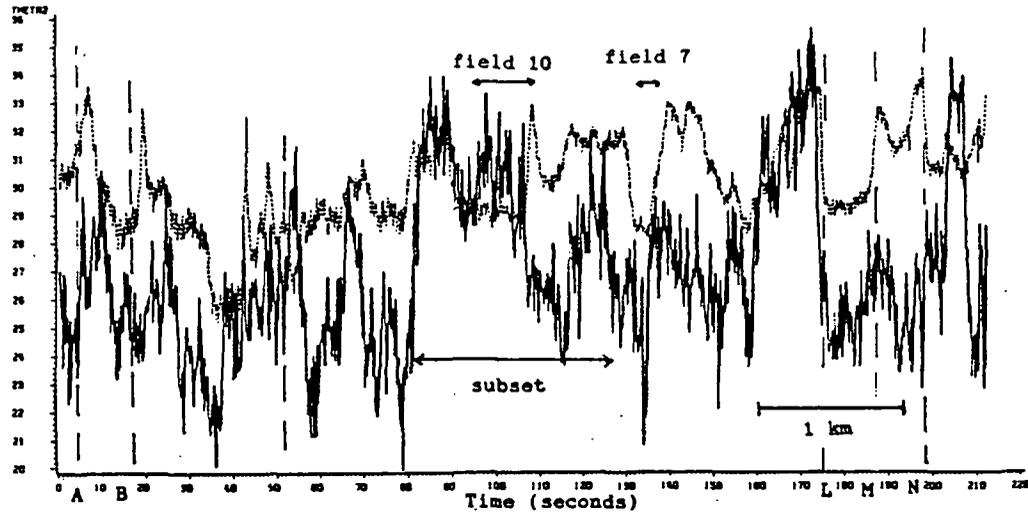


Figure 3. Combined plot for soil water content and temperature. Note the inverse relation of the responses for roads (points A,B,D,L,M,N).

Soil water content (cm<sup>3</sup>/cm<sup>3</sup>)

VOVES JULY 6 1983 DATA SEGMENT 3

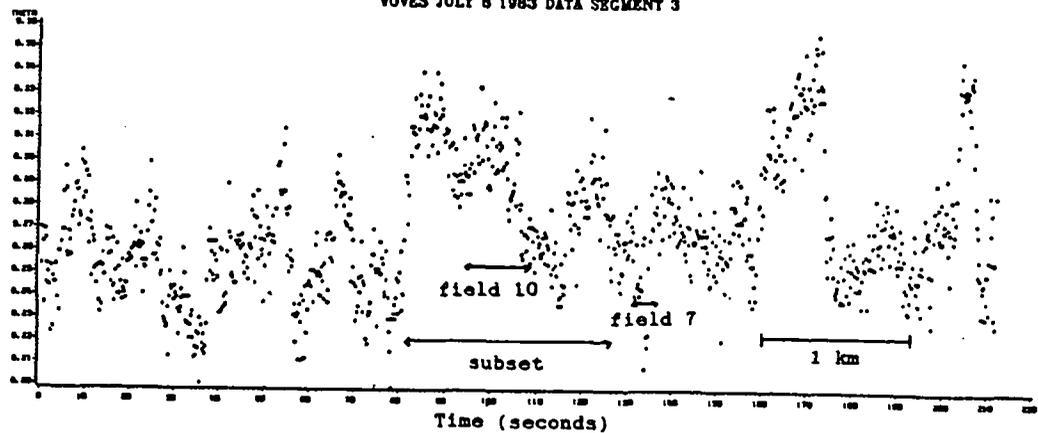


Figure 2. Soil water content measured with active microwave, same transverse as in figure 1.

Temperature (degrees celsius) - - - - -  
 Soil water content (cm<sup>3</sup>/cm<sup>3</sup> x 10<sup>2</sup>) ———

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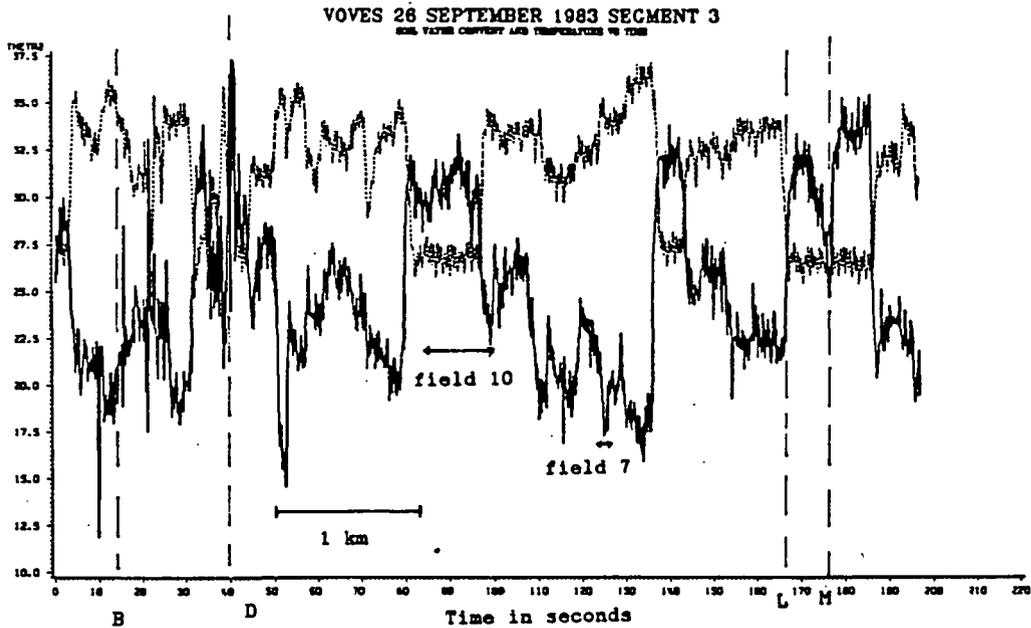


Figure 4. Combined remotely measured soil water content and surface temperature for approximately the same transect as in figure 3, data taken on 28 September at approximately 3:45. Note the inverse relation of the microwave and temperature responses. The road locations are the same as in figure 3.

Temperature (degrees celsius) - - - - -  
 Soil water content (cm<sup>3</sup>/cm<sup>3</sup>) ———

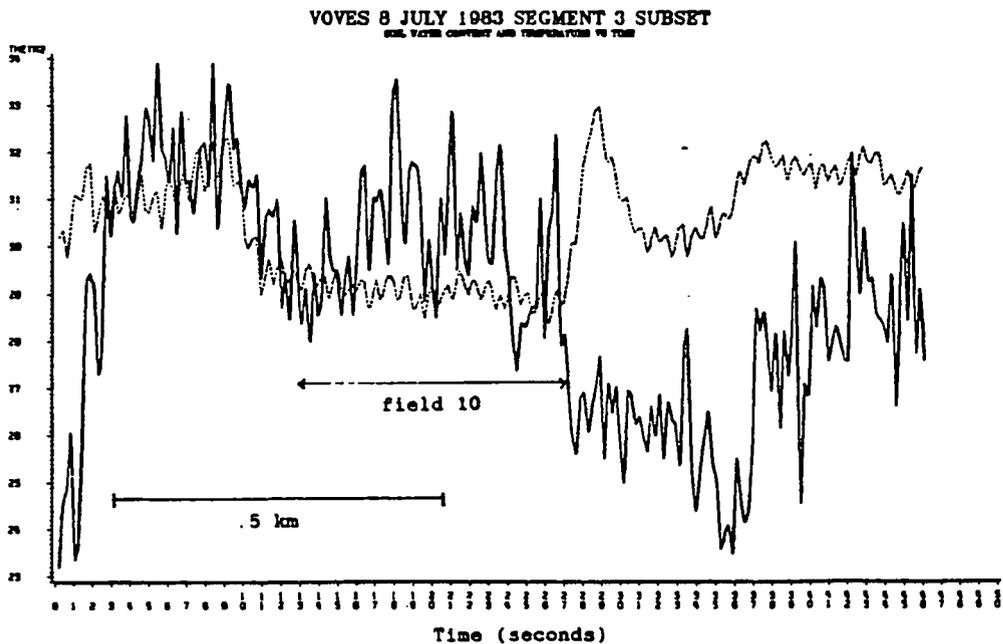


Figure 5. Combined temperature and soil water content responses for the subset of data segment 3. Note that the variation for soil water content within a field is much greater than the variation of temperature.

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Temperature (degrees celsius) - - - - -  
Soil water content (cm\*\*3/cm\*\*3) ———

VOVES 26 SEPTEMBER 1983 SEGMENT 3 SUBSET  
SOIL WATER CONTENT AND TEMPERATURE VS TIME

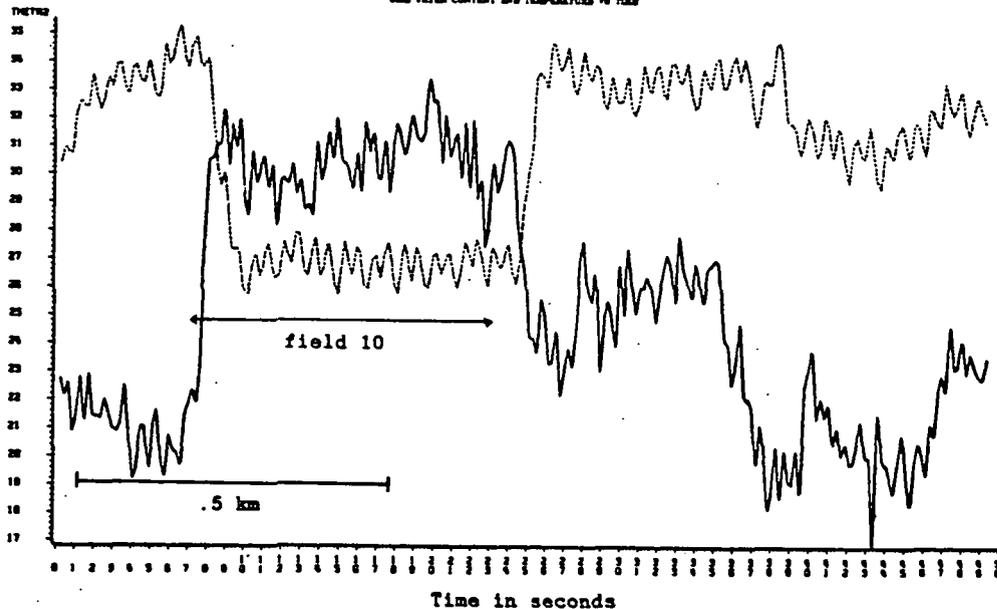


Figure 6. Combined temperature and soil water content for the subset of data segment 3, for the 28 September case. Note again the greater variability of the soil water content measured by the microwave.

### 3. CONCLUSION

The data analyzed so far has indicated a strong negative correlation between remotely sensed soil water content and surface temperatures for the near bare soil conditions of 26 September, while for the same area on 8 July there is little correlation between the remotely sensed soil water content and surface temperature.

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# Regional-Scale Estimates of Surface Moisture Availability and Thermal Inertia Using Remote Thermal Measurements

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## 1 INTRODUCTION

Thermal infrared measurements (10–12  $\mu\text{m}$ ) made by satellites reveal temperature patterns which are closely related to surface properties, especially soil moisture and thermal inertia. In this paper, a variety of models are reviewed which have been developed to interpret thermal IR measurements and to determine these governing parameters. These models fall into four general classes: analytic, predictive, diagnostic and empirical, of which, the diagnostic models of Price (1982a, 1982b) and the predictive models of Carlson (Carlson *et al.*, 1981), and Wetzel and Atlas (1981, 1983) already have been used to obtain regional-scale soil moisture or evaporation patterns from satellite infrared measurements.

The purpose of this paper is to (1) present a brief summary of the various theoretical approaches (i.e. models) which have been developed specifically for determining the surface energy fluxes and governing surface parameters using remote thermal IR measurements, (2) present some recent results from one of these methods, that of Carlson *et al.* (1981; henceforth referred to as CM), (3) discuss limitations and uses in applying such methods by listing the various sources of errors and constraints inherent in theory and measurement and (4) conclude with an assessment of present capabilities of the thermal infrared method and its future potential.

## 2 FACTORS INFLUENCING SURFACE TEMPERATURE

The basis for determining surface parameters remotely is a variety of observations which show, for example, that evaporating surfaces are cooler than dry surfaces, and that as the soil moisture content decreases, the amplitude of the surface temperature wave becomes larger (Jackson, 1982). At an idealized surface, the surface temperature response is governed by a variety of surface, substrate, and atmospheric variables, most notably the ground moisture and thermal inertia (which will be shortly defined) and albedo (Carlson and Boland, 1978). The greatest temperature variation occurs at the ground-air interface but with increasing distance from the surface, in the ground or in the atmosphere, the diurnal temperature wave decreases. Well above the surface advection plays such an important role that the air temperature variation becomes responsive to surface heating over a relatively large area and may no longer reflect the local character of the substrate.

Over bare soils the temperature is known to vary rather sensitively with both the thermal inertial and volumetric moisture content (Myers and Heilman, 1969; Idso *et al.*, 1975d; Idso *et al.*, 1976; Idso *et al.*, 1977c). This sensitivity is reflected in both the ground surface and screen-level (1.3 m) air temperatures (Idso *et al.*, 1976), especially the day-night differences.

Over plant canopies the temperature of the leaf is known to rise in response to a reduction of available water in the root zone (Byrne *et al.*, 1979; Jackson, 1982). Severe reduction in available water to the plant canopy can lead to what is called "plant water stress", which has been shown by various workers (Cihlar, 1976; Idso *et al.*, 1977a; Ehrler *et al.*, 1978) to be related to crop yield. The extent of plant stress is highly correlated with moisture in the root layer of the soil, especially the top few centimeters. Radiometric measurements show that the leaf temperature can vary from one crop to another under identical meteorological conditions (Blad and Rosenberg, 1976; Heilman *et al.*, 1976) and that such temperature variations partially reflect differences in evapotranspiration between the canopies (Wiegand *et al.*, 1968; Wiegand, 1971; Schmutge, 1978; Heilman and Moore, 1980).

The observation that leaf temperature becomes elevated when evaporation diminishes has led to the development of methods for remotely determining soil moisture from radiometric surface temperature measurements. Empirical results by Idso *et al.* (1975a, 1976, 1977a)

relating radiant energy available from the sun to soil moisture clearly demonstrate that algorithms can be formulated for the purpose of calculating plant moisture availability from canopy temperatures. The development of a reliable method for interpreting satellite blackbody temperature measurements within the framework of a soil-atmosphere model was accelerated by the advent of meteorological satellites such as TIROS-N (currently NOAA-7), the Heat Capacity Mapping Mission (HCMM) satellite, and GOES, all of which are capable of making surface temperature measurements close to the optimum times.

The value of being able to determine surface evaporation patterns is obvious in agriculture and hydrology. Techniques for combining a numerical or analytical surface layer or boundary layer model with satellite IR measurements for the purpose of determining the governing surface parameters and surface energy fluxes originated with geologists, who were interested in mapping the thermal inertia of mineral outcroppings and other geological structures (Watson, 1971; Pohn *et al.*, 1974; Watson, 1975; Kahle, 1977; Price, 1977; Watson and Miller, 1981). Price (1977) and Kahle (1975) both refer to the techniques as thermal inertia mapping, following an idea previously suggested by Pohn *et al.* (1974) and Blanchard *et al.* (1974). Indeed, the latter article plainly addresses the idea of soil moisture measurement by satellite.

Recent efforts to use thermal inertia mapping to determine moisture availability and the surface energy fluxes from satellite IR measurements have been made by Carlson *et al.* (1981, 1984), Wetzel and Atlas (1981, 1983) and by Price (1982a, 1982b) in the United States and by Rosema *et al.* (1978) and Soer (1980; see also Dejace *et al.*, 1979; Reiniger and Seguin, 1985—this issue RSR) in Europe. Although Carlson *et al.* (1981) restricted their investigation to an analysis of the urban heat island, they showed that thermal inertia mapping affords the opportunity to determine the surface moisture availability over region-scale areas including both urban and vegetated types of surfaces. The fact that moisture availability may be the single most important parameter governing the partition between sensible and latent heat flux at the surface (Carlson and Boland, 1978; McCumber and Pielke, 1981), suggests that thermal inertia mapping may contribute to improved initialization of atmospheric prediction models.

The relatively large fluctuation in the surface temperature, compared with that in the atmosphere, and the strong sensitivity of that surface

temperature to surface and substrate parameters, such as soil moisture and thermal inertia, suggests that remote measurements can be used in conjunction with an atmosphere-substrate model to infer values for the surface fluxes and the substrate and surface parameters. In the case of remote measurements, such as those derived from satellite, the information available is a blackbody temperature measurement integrated over the area of one pixel.

Models of differing complexity have been developed for the purpose of calculating surface fluxes of heat and moisture and deriving the surface and substrate parameters using remote surface temperature measurements. While complexity in a model is advantageous in research for studying the sensitivity of surface temperature to atmospheric and surface conditions, the lack of knowledge of initial or in-situ atmospheric and substrate conditions and the errors of measurement produce uncertainties in the results which effectively limit the value of any model and thereby lessen the advantages of complex parameterizations.

A strong impetus for current research into remote determination of the surface parameters was the Heat Capacity Mapping Mission (HCMM) satellite which was launched in early 1978. The relatively high resolution (500 m) of the HCMM radiometers and its favorable orbital period, which allowed measurements to be made near 2 AM and 2 PM local times, afforded an opportunity to study the surface properties on a scale small enough to resolve details of the land use pattern but broad enough to permit relatively large areas to be examined. Essentially, the problem was to derive information about the surface from a pair of temperatures, one made during the early afternoon (local sun time) and the other at night. Day-night pairs of temperatures were obtained by HCMM 12 hours apart every 16 days and 36 hour pairs about every five days. HCMM made it necessary to develop models relating the surface temperature response to the surface parameters as measured at two times during the daily temperature cycle, via the use of atmospheric-substrate models.

### 3 MATHEMATICAL PROCEDURES FOR INTERPRETATION OF THERMAL INFRARED DATA

#### (a) Analytic approximations

The familiar daily variation of surface temperature  $T_o$  results from the balance between incoming and outgoing energy fluxes at the earth's surface. This balance is expressed through the equation

$$R_n = G_o + H_o + L_e E_o = R_s + R_L - R_I \quad (1)$$

where the net radiant energy ( $R_n$ ), consisting of the absorbed solar flux ( $R_s$ ), the downward longwave flux ( $R_L$ ), and the outgoing longwave flux ( $R_I$ ), is balanced by the upward fluxes of sensible heat ( $H_o$ ) and latent heat ( $L_e E_o$ ) into the atmosphere and the flux of sensible heat into the ground ( $G_o$ ). The simplest solution to the surface energy balance equation is to consider a time-dependent energy flux  $F \cos \omega t$  incident on a uniform homogeneous material satisfying the heat flow equation

$$\rho' c' \frac{\partial T}{\partial t} = \frac{\partial}{\partial z} \left( \lambda \frac{\partial T}{\partial z} \right) \quad (2)$$

where  $\rho'$  is the density,  $c'$  the heat capacity, and  $\lambda$  the thermal conductivity;  $\lambda/\rho' c'$  is the thermal diffusivity of the substrate, referred to as  $\kappa'$ .

Classical solutions to the heat transfer Eq. (2), as presented by Jaeger (1953) and by Carslaw and Jaeger (1959), inevitably led to adaption of the simple cases to which the solutions are appropriate for application to boundary meteorology (Sellers, 1965) and subsequently to their use in remote sensing of surface temperature and surface characteristics. In regard to the latter, the earliest theoretical developments were made with geological applications in mind, the thermal inertia being a measurable property of soils and rock. Watson (1973) recognized the uses of the thermal inertia method in identifying minerals and his theoretical papers (1973, 1975; see also Watson and Miller (1981) and Watson, 1982) were soon followed by those of Price (1977, 1980, 1982a, 1982b) and Pratt and Ellyett (1979; see also Pratt (1980) and Pratt *et al.* (1980) in which thermal inertia was related directly through analytic expressions to the maximum and minimum surface temperatures, which could be measured by a satellite. Fourier models have also been

formulated in Europe (Viellefosse and Favard, 1979; Hechinger *et al.*, 1982).

Equations (1) and (2) can be analyzed in order to derive formulas pertaining to the 24-hour mean surface temperature and the amplitude of the diurnal temperature cycle, the latter being obtainable from a day-night pair of satellite temperature measurements. These quantities are shown to be closely associated with the evaporative flux at the surface and with the diurnal heat capacity (or thermal inertia), which is also closely related to the soil moisture in the near-surface layer (Pratt and Ellyet, 1979; Dejace *et al.*, 1979; Idso *et al.*, 1975a, 1975b, 1975c, 1977a, 1977b, 1977c).

By time averaging Eq. (1), a simple result may be obtained. Assuming that no net heating or cooling of the earth occurs over a 24-hour period, so that the average of  $G_o$  vanishes over a complete solar cycle, one finds

$$\langle R_m \rangle = \langle H_o \rangle + \langle L_e E_o \rangle \quad (3)$$

where the brackets  $\langle \rangle$  represent a 24-hour average.

Now, using simple expressions for the solar and longwave fluxes,

$$R_s = SV(1 - A_o) \cos Z_o \quad (4a)$$

$$R_t = \epsilon_g \sigma T_o^4 \quad (4b)$$

$$R_L = \epsilon_a \epsilon_g \sigma T_o^4 \quad (4c)$$

the net available radiant energy  $R_n$  can be calculated from available measurements of temperature and humidity and a knowledge of the solar geometry and atmospheric transmission. Here,  $V$  represents the effective atmospheric transmission for direct and diffuse solar radiation,  $A_o$  is the surface albedo and  $Z$  is the solar zenith angle. For longwave radiation  $\sigma$  is the Boltzmann constant,  $T_o$  the near-surface air temperature,  $\epsilon_g$  the emissivity of the ground (usually taken as 1.0), and  $\epsilon_a$  is the emissivity of the atmosphere, usually calculated as a function of surface specific humidity (Brunt, 1932) or as a function of precipitable water (Monteith, 1961). Similarly, simple expressions for surface heat and moisture fluxes can be written in the following form (Sellers, 1965)

$$H_o = c_p f_h(\bar{u})(T_o - T_a) \quad (5a)$$

$$L_e E_o = L_e f_e(\bar{u})(q_o - q_a) \quad (5b)$$

where  $c_p$  is the specific heat of air at constant pressure,  $L_e$  is the latent heat of evaporation,  $q_o$  and  $q_a$  are specific humidity of the surface and air, and for the sake of brevity, we write the atmospheric transfer coefficients for heat and water vapor, respectively, as  $f_h(\bar{u})$  and  $f_e(\bar{u})$  to signify the dependence of these coefficients on the ambient wind speed  $\bar{u}$ ; in general,  $f_e(\bar{u}) = f_h(\bar{u})$ . Accordingly, it is possible to write an explicit expression for  $L_e E_o$  by averaging the various quantities over a 24-hour day, by making use of the constant term in the Fourier expansion of the angle of solar incidence ( $C_o$ ), and by imposing some additional simplifications concerning the averaging procedure such as in the linearization of Eqs. (4) and (5). Thus, Price (1980) obtains

$$\begin{aligned} \langle L_e E_o \rangle = & SV(1 - A_o)C_o + \epsilon_g \sigma \langle T_a^4 \rangle (a + b \sqrt{\langle q_a \rangle}) \\ & - \epsilon_g \sigma \{ \langle T_o \rangle^4 + 0.75 \langle T_o \rangle^2 (T_{o_{max}} - T_{o_{min}})^2 \} \\ & - \rho k^2 c_p / [\ln z_o / z_a]^2 \{ \langle u \rangle \langle T_o \rangle - \langle u T_o \rangle \} \end{aligned} \quad (6)$$

where  $a$  and  $b$  are constants in Brunt's (1932) equation for the emissivity of the atmosphere,  $k$  is the Von Karman's constant, and  $\rho$  is the air density (see also Sellers, 1965). The roughness length and height of the atmospheric measurements (nominally 1.3 m) are expressed, respectively by  $z_o$  and  $z_a$ . The daily maximum and minimum temperatures are indicated by the appropriate subscripts. The result is that the mean daily evaporation can be calculated from mean daily values of ground temperature ( $T_o$ ), the near-surface temperature ( $T_a$ ), specific humidity ( $q_a$ ) and wind speed ( $\bar{u}$ ), which can be obtained from conventional meteorological measurements. Solar geometry is used to provide estimates of  $S$  for clear skies. Surface roughness, albedo and emissivity are determined from a knowledge of the surface characteristics.

Because of the linearization and averaging of the long wave fluxes, which contain a  $T_o$  dependence, the maximum and minimum ground temperatures also appear in the equation for  $L_e E_o$ . These temperatures can be measured by satellite, although the mean daily evaporation in this formulation is not very sensitive to the diurnal temperature amplitude at the ground. Similar equations for the mean daily evaporation fluxes have been presented by Pratt *et al.* (1980).

Both Price (1980) and Pratt *et al.* (1980) explicitly treat surface evaporation although only Price considers moisture flux explicitly as a variable to be determined from the temperature wave. Pratt *et al.* (1980) solve for two surface parameters simultaneously, thermal inertia and albedo, following the graphical models. So far, only Price (1980) has derived patterns of a surface parameter (evapotranspiration) from satellite data using an analytical model.

An approximate description of the diurnal temperature cycle may be obtained through consideration of a periodic energy flux incident on a uniform material. The analytic solution for the temperature is readily obtained: its value at the surface  $T_o(t) = T(z = 0, t)$  is given by

$$T_o(t) = F \cos(\omega t - \pi/4)/(\omega \rho' c' \lambda)^{1/2} \quad (7)$$

Further, the heat flux across the ground-air interface ( $G$ ) can be represented by the equation

$$G_o = \lambda(\partial T/\partial z)_{z=0} = \lambda(T_o - T_{-1})/\Delta Z_{-1} \quad (8)$$

where  $T_{-1}$  is the temperature of the first substrate level, a distance  $\Delta Z_{-1}$  below the surface in the finite-difference notation on the right-hand side of Eq. (8).

The quantity  $(\rho' c' \lambda)^{1/2}$  is frequently referred to as the thermal inertia ( $P$ ) or sometimes the conductive capacity, thermal property or thermal admittance, having a value of approximately 500 TIU (joules  $m^{-2} C^{-1} s^{-1/2}$ ) or 0.0125 (cal  $cm^{-2} C^{-1} s^{-1/2}$ ) for dry sand and about 4000 TIU for solid quartz. Because these units are somewhat clumsy to work with it is sometimes convenient to incorporate the earth's rotation rate ( $\omega = 2\pi/24$  hours) and define the diurnal heat capacity

$$D = (\omega \rho' c' \lambda)^{1/2} = (\omega \lambda / \kappa')^{1/2} \quad (9)$$

The factor  $\omega$  brings  $D$  to the typical range of values 4–35  $W m^{-2} C^{-1}$ , which are simpler units. The advantage in using the thermal inertia is that it contains two surface parameters, conductivity and diffusivity (or heat capacity) in one parameter. As Carlson and Boland (1978), Blackadar (1978a) and Price (1982a, 1982b) have shown, most of the variance in the surface temperature wave is contained in the first

Fourier harmonic and, consequently, differing values of  $\lambda$  and  $\kappa'$  will yield almost identical results in a model for a fixed value of  $P$  or  $D$ .

Using Eq. (7) as the thermal forcing at the ground-air interface yields a very rough estimate of the temperature variation at the earth's surface:

$$T_{o_{max}} - T_{o_{min}} = 2SV(1 - A_o)/D \quad (10)$$

For typical values of the solar amplitude, atmospheric transmission, albedo and diurnal heat capacity, one finds that the diurnal temperature range ( $T_{o_{max}} - T_{o_{min}}$ ) is rather too large. It is evident, therefore, that energy exchanges at the earth's surface must be considered, although Eq. (10) properly identifies the solar forcing, surface albedo and thermal inertia ( $D$ ) as important factors influencing the variations in surface temperature. Both Pratt *et al.* (1980) and Price (1982a, 1982b) consider more realistic boundary conditions which require additional simplifying assumptions (Watson, 1975; Price, 1977). Price represents the solar flux as an explicit function of time, while the diurnal variation of all other fluxes exchanged with the atmosphere is regarded as linear in the surface temperature. Thus the surface boundary condition Eq. (1)

$$R_s = R_s(t) = A + BT_o + G_o \quad (11)$$

The constants  $A$  and  $B$  are evaluated by averaging over 24 hours and by averaging the appropriate partial derivative with respect to surface temperature. Realizing that  $\langle G_o \rangle = 0$ ,

$$\langle R_s(t) \rangle = A + B \langle T_o \rangle \quad (12)$$

and

$$B = \left\langle \frac{\partial}{\partial T_o} (R_i - R_L + H_o + L_o E_o) \right\rangle \quad (13)$$

where the averages are readily computed from the expressions for the solar and surface heat and evaporation fluxes (Price, 1980). The first equation is simply the linearized version for the expanded form of Eq. (1). The second equation defines a quantity which is significant through its effect on the day-night temperature variation. Because  $H_o$  and  $R_L$  depend on air temperature, it is appropriate to recognize that

air temperature is strongly affected by surface temperature. For these terms the approximation

$$\frac{\partial}{\partial T_o} = \frac{\partial T_o}{\partial T_o} \frac{\partial}{\partial T_a} = \frac{[T_o(\text{day}) - T_o(\text{night})]}{[T_o(\text{day}) - T_o(\text{night})]} \frac{\partial}{\partial T_a} \quad (14)$$

is used.

Given these assumptions, it was shown by Price (1980) that the 11-hour surface temperature variation, which was to be measured by HCMM, could be approximated by the expression.

$$T_o(1330\text{LST}) - T_o(0230\text{LST}) = \frac{2SV(1 - A_o)C_1}{[D^2 + B^2 + DB]^{1/2}} \quad (15)$$

where  $C_1$  is the first Fourier coefficient in the expansion of the insolation. In Eq. (15) the times of measurement represent the northern hemisphere mid-latitude ( $45^\circ\text{N}$ ) overpass times for the HCMM. However, it is possible to adjust the theory by changing the phase differences between the two temperatures in order to accommodate surface temperatures measured at two different times; here it is assumed that those temperatures represent the maximum and minimum values.

Price (1980) obtained a simple approximation for the thermal inertia. Defining a parameter  $Q$  which is equal to

$$\frac{2SV(1 - A_o)C_1}{[T_o(1330\text{LST}) - T_o(0230\text{LST})]} \quad (16a)$$

he obtained the result that

$$D \cong Q - 0.9B \quad (16b)$$

With  $B$  set equal to zero Eq. (16) reduces to an expression for "apparent" thermal inertia which is a parameter provided by the HCMM program (HCMM Data Users Handbook, 1978).

Price (1982a, 1982b) used the results of a prognostic model (the European TELL-US predictive model referred to in the next section) to tune his own model results, specifically by adjusting his expressions for  $L_e E_o$  and  $D$ . In this manner Price was able to eliminate what he felt was internal bias resulting from the limiting assumptions and simplifi-

cations specific to the analytic model. Pratt *et al.* (1980) have suggested a similar direct approach for determining the parameter  $B$  from ground measurements, a procedure necessitated by the uncertainty in evaluating this parameter theoretically.

A primary virtue of the analytic models is their simplicity, which allows for a high computational efficiency without seriously degrading the results. They are somewhat limited, however, by the approximations involved in obtaining an expression for the thermal forcing at the surface and by the time averaging of meteorological quantities. Perhaps the most serious limitation is the simplified manner in which atmospheric processes are treated, particularly those changes in temperature, wind, and moisture which are occurring in the surface and mixing layers. In the following section we treat a physically more elaborate model in which some of these simplifications are addressed.

#### (b) Combined ground/atmosphere simulation models

Time-dependent, initial value prediction models are widely used in meteorology and a variety of boundary layer models exist as separate one-dimensional components for diagnostic purposes (e.g. Myrup, 1969; Sasamori, 1970) or as integrated routines in two- and three-dimensional numerical atmospheric models (Blackadar, 1978a, 1978b). The function of such sub-components in atmospheric models is of course, to predict changes in the substrate temperature and, in the atmospheric temperature, wind speed and moisture as a function of time. Carlson and Boland (1978) and Rosema (1978; see also Dejace *et al.*, 1979) proposed similar methods in which a predictive model can be inverted, given the measured surface temperatures, to yield unique values of two surface parameters, either thermal inertia moisture availability (or surface relative humidity) or thermal inertia and albedo.

Sensitivity tests by Saltzman and Pollack (1977) and by Carlson and Boland (1978) demonstrate that the three most important surface parameters which govern the surface temperature response are the moisture availability, thermal inertia and albedo. However, within the normal range of satellite pixel brightnesses encountered over heterogeneous terrain, such as a city or a vegetation canopy, the horizontal variations in surface albedo are relatively small. The relative importance of the moisture availability and thermal inertia, as opposed to surface roughness or albedo, can be seen by inspection of Figure 1.

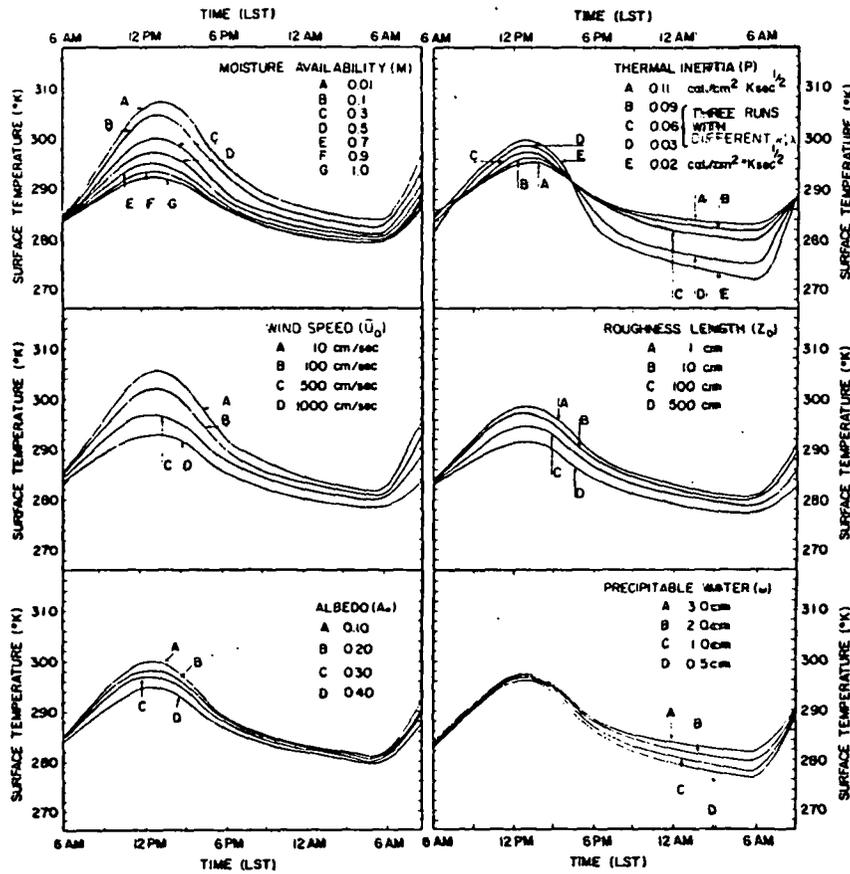


FIGURE 1 Model sensitivity tests varying moisture availability, thermal inertia, wind speed at  $z_0$ , roughness length, precipitable water and albedo.

Moisture availability exerts a large influence on the temperature variation during the heated portion of the day whereas thermal inertia variations produce their greatest effect on the night-time temperatures. The remaining parameters not only exert a smaller effect on the surface temperature wave but such properties of the ground as albedo or roughness can be estimated from an approximate knowledge of the terrain to within an acceptable accuracy. Initial wind speeds at the top

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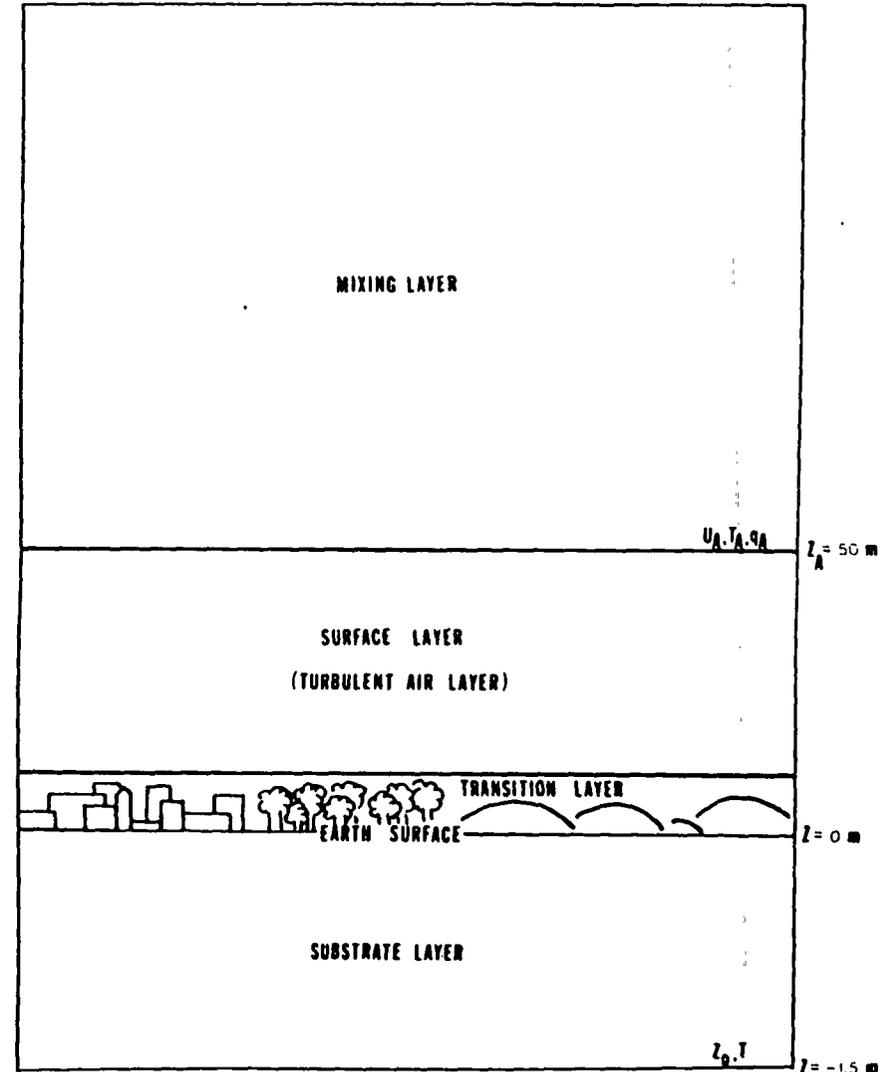


FIGURE 2 Basic model framework of the CM.

and above the atmospheric surface layer (Figure 2) can be provided by conventional large-scale weather observations and thereafter predicted internally in the model.

### The Carlson simulation method

The underlying constraint in virtually all predictive models is the surface radiation balance, as expressed by Eq. (1). The basic structure of the Carlson model (Figure 2) consists of four layers, an atmospheric mixing layer, a surface layer, a shallow transition layer between the surface air layer and the substrate. Fluxes in the atmosphere are specified by equations similar to those used by Price (1982a, 1982b) and by Pratt *et al.* (1980). To clarify the definition of the important parameters, moisture availability and thermal inertia, and in order to discuss the predicted variables, the flux equations are presented, using the resistance notation of Monteith (1975), as

$$H_o = \frac{\rho c_p (T_o - T_a)}{R_a + R_{ch}} \quad (17a)$$

$$L_e E_o = \frac{\rho L_e (q_{os}(T_o) - q_a)}{R_a + R_{cv}} \cdot M \quad (17b)$$

where  $T_o - T_a$  and  $q_{os} - q_a$ , respectively refer to the vertical temperature and specific humidity differences between a saturated surface (subscript zero) and the top of the surface layer (subscript *a*), nominally chosen to be 50 m above the ground. The resistance for eddy conduction  $R_a$  refers to that in the atmospheric surface layer for heat and moisture fluxes between the top of the transition layer at height  $z_t$  and at the top of the surface layer at  $z_a$ . The subscripts *ch* and *cv* refer, respectively, to resistances for heat and moisture fluxes through the transition layer, which is envisioned to be a narrow region over which a combination of eddy, molecular and radiative fluxes coexist in some fashion.  $L_e$  is the heat of vaporization for water.

The above flux equations are more elaborate versions of those used by Price (1982a, 1982b) and Pratt *et al.* (1980) in their analytic models and referred to in Eqs. (5) and (6). The factor  $M$  in Eq. (17b) is called the moisture availability and is defined as

$$M \equiv E/E_o(T_o) \equiv \frac{R_b}{R_b + R_s} \quad (17c)$$

where  $R_b = R_a + R_{cv}$  and  $R_s$  is a surface resistance, which over vege-

tation functions as a bulk stomatal resistance parameter, accounting for the resistance for water vapor between the canopy and the atmosphere. Generally,  $M$  represents the fraction of potential evaporation ( $E/E_o(T_o)$ ) at the temperature of the surface ( $T_o$ ). Under saturated conditions  $R_s$  is essentially zero and evaporation is equal to the potential value. For a completely dry surface  $M$  is zero. Below some low value of  $M$  soil or plant wilting occurs (Tanner and Pelton, 1960), the vegetation may begin to suffer physiological stress (Monteith, 1975) and the soil approaches a limiting dryness (Jackson, 1973; Idso *et al.*, 1974).

Moisture availability can be expressed in various other forms than that of Eq. (17b). Nappo (1975), for example, explored an alternate parameterization in which the moisture parameter is expressed as a surface relative humidity ( $h_o$ ), a form which both Price (1982a, 1982b) and Pratt *et al.* (1980) prefer (i.e., in Eq. 17b the quantity  $(q_{os}(T_o) - q_a)M$  is replaced by  $(h_o q_{os} - q_a)$ ). Both  $M$  and  $h_o$  vary between 0 and 1.0 and can be expressed as mathematical transforms of one another. In most cases  $L_e E_o$  depends on saturation specific humidity at the surface, and must vary with surface temperature and, consequently, its value may change rapidly with time for a constant value of  $M$ . The virtue in using  $M$  is that it represents an intrinsic property of the soil such as the ratio of volumetric water content to that at field saturation for bare soil. If so, it must vary over a small fraction of its maximum range during the day when averaged over a typical rooting depth (Jackson, 1973). Horizontal variations in  $M$  or temporal changes over a period of several days should, therefore, reflect real changes in the near-surface or vegetation moisture content.

The resistance terms are calculated as follows. For heat

$$R_{ch} + R_a = \int_0^{z_a} dz / (K_h(z) + C_h) \quad (18a)$$

$$R_{cv} + R_a = \int_0^{z_a} dz / (K_h(z) + C_c) \quad (18b)$$

where  $K_h$  is the eddy diffusivity for both heat and moisture fluxes in the surface layer. These integrals depend upon the wind speed, roughness length, and static stability (surface heat flux) and thus require iteration to achieve solution. The integrals in Eq. (18) are obtained in two parts, by considering that molecular and radiative processes are unimportant

above the transition layer and that stability corrections to the logarithmic profile (Panofsky, 1974) are negligible in the transition layer. Solutions are obtained for both heat and moisture conduction by insertion of the appropriate constants for  $C_h$  and  $C_v$ ;  $R_{ch}$  and  $R_{cv}$  are assumed to be equal. For want of better values, the molecular conduction coefficients were used for  $C_h$  and  $C_v$ . Analogous expressions for the momentum eddy conduction  $K_m$ , evaluated between the limits of the roughness height and  $z_a$  are also determined. The eddy conduction integrals such as Eq. (18) are solved by using analytic expressions for  $K_h$  and  $K_m$  (Panofsky, 1974) based on solutions presented by Paulson (1970) and Benoit (1977) for the case of an unstable temperature profile.

Tests with the Carlson model show that the results are somewhat sensitive to the values of the transition layer depth. Large vertical gradients are generated by the inclusion of small values of the transition layer coefficients and for that reason the molecular values may be rather too small, at least over rough terrain where surface elements may protrude into the airstream.

**Solution** Equations (1), (2), (4), (8), (17) and (18) form a complete set which can be solved to obtain the variables  $T_0$ ,  $L_e E_0$ ,  $H_0$ ,  $G_0$  and  $T_{-1}$  in the atmospheric surface layer and substrate given the measured temperature, specific humidity and wind speed at two levels (e.g. the surface and at  $z_a$ ). In the Carlson model, the temperature, specific humidity and wind speed at  $z_a$  and in the mixing layer are predicted from a set of initial conditions, as is the temperature in the ground. The diffusion Eq. (2) is used to predict the temperature variation below the ground. In the mixing layer, temperature and specific humidity are calculated from a model originally formulated by Tennekes (1973) for conditions of free convection during the day. Wind speed is calculated from the time-dependent momentum equations, including the effects of coriolis force and vertical mixing, the latter being determined by specifying the vertical distribution of the mixing coefficient ( $K_h$ ) in the mixing layer as a function of  $z$  (O'Brien, 1970).

At night the critical Richardson number formulation of Blackadar (1978a) is used to calculate the temperature and wind speed tendencies in the surface and turbulence layers with an additional equation for temperature tendency imposed near the surface. The surface temperature is determined as a quasi-equilibrium value at each time step from the forementioned set of equations. Solutions quickly approach radia-

tive equilibrium after sunset with the vanishing of turbulence, except under windy conditions when turbulence episodes may still occur. Under radiative equilibrium the equations closely resemble those of Outcalt's (1972) model. Advection is completely neglected.

An advantage in using a predictive model is that it permits a fuller use of the governing equations than does a simpler model. In allowing solutions to proceed from a single set of initial conditions, which may be determined from large-scale meteorological data, the results do not depend upon continuous in-situ measurements of air temperature, humidity and wind speed. Moreover, the vertical profiles of the atmospheric variables in the surface layer are internally consistent with the surface fluxes and with the surface energy balance. The Carlson Model (CM) treats the surface and mixing layers in detail, accounting for the temperature, wind and moisture variations throughout the lower atmosphere.

As an example, consider the daily temperature wave generated by the CM in Figure 3. The figure suggests that the temperature wave is highly sensitive to the moisture availability during the day and to the thermal

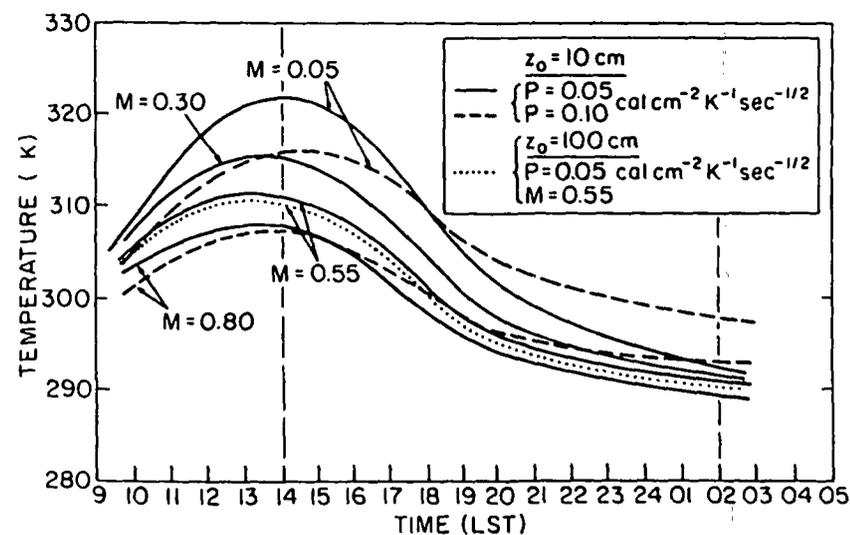


FIGURE 3 Simulated surface temperatures for different values of moisture availability ( $M$ ) thermal inertia ( $P$ ), and surface roughness ( $z_0$ ) starting at 0600 LST. The vertical dashed lines refer to the times of HMM satellite data measurement.

inertia at night, as is also indicated in Figure 1. It is not obvious, however, that unique values for the surface parameters can be determined from an inversion of the temperature information contained in Figure 3. Carlson and Boland (1978) and Rosema (1978) suggested a method whereby two ground parameters, thermal inertia and moisture availability (or surface relative humidity), can be determined uniquely in most cases from two or more measurements of surface temperature. Other parameters are held fixed at their assumed values during the calculations.

To illustrate the inversion method for recovering the two governing surface parameters, thermal inertia and moisture availability, consider the values of temperature at 1400 and 0200 local time in Figure 3, approximately the times corresponding to those of a day-night HCMM image pair. If the temperatures are then replotted for both the day and night-time values in a graph of  $M - P$  space, one obtains a pair of intersecting contours, as is shown in Figure 4 for the data used to construct Figure 3. It is evident from inspection of Figure 4 that unique values of  $M$  and  $P$  are obtainable for this case within the range of temperatures shown in the figure. Provided that the measured ground temperatures fall into the range of temperatures shown in Figure 4 a map of thermal inertia and moisture availability (and also the surface heat fluxes) can be generated from two satellite temperature images, one at 0200 and the other at 1400 local time. Bi-valued solutions for the surface parameters are possible under certain conditions but these situations appear to be rare and occur when extraordinary initial conditions are chosen. Ambiguous solutions for  $M$  and  $P$  may occur, however, if the temperature pairs do not differ from each other by a large amount. A more complete discussion of the uncertainties in this method, including alternate combinations of satellite imagery, other than those at 1400 and 0200 local time, is presented in Section 5.

In practice, images of the transformed quantities,  $M$ ,  $P$ ,  $H_o$ ,  $L_e E_o$  and  $G_o$ , are not obtained graphically but by means of regression equations fitted to the model output from a succession of model runs with differing values of the parameters  $M$  and  $P$ . Unlike the energy fluxes, which correspond to the time of the afternoon image, the intrinsic terrain parameters,  $M$  and  $P$ , are considered averages for the entire day. Recently, Polansky (1982), Wetzel (1983) and Wetzel and Atlas (1981, 1983) have shown that image pairs, other than those for the day and night, and image triplets can be used effectively to derive unique values for the governing surface parameters.

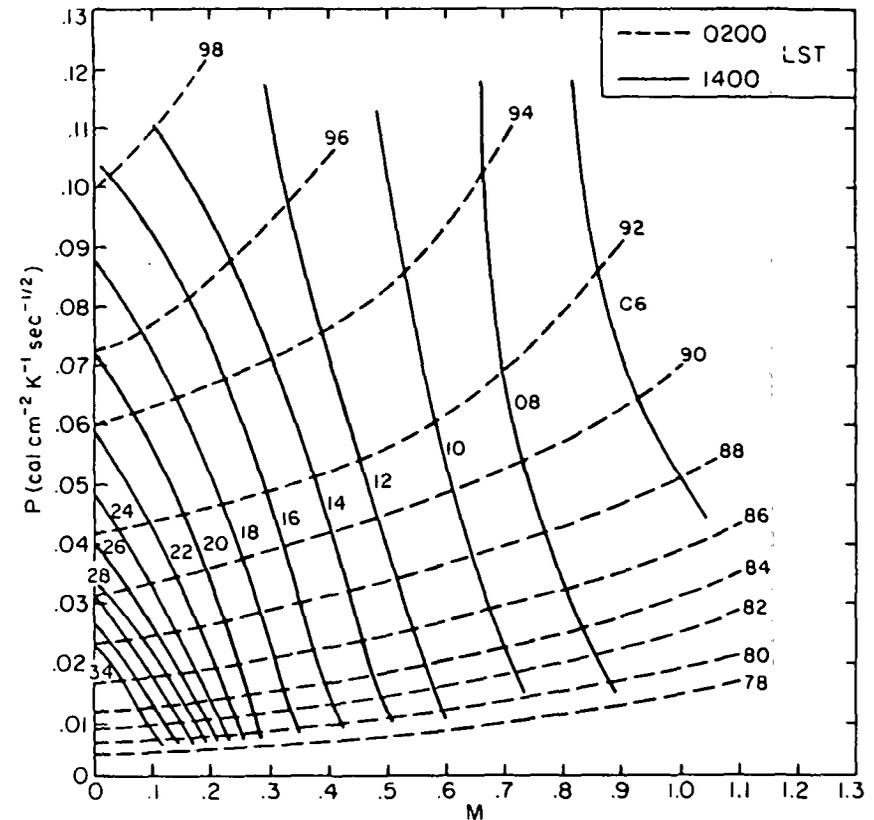


FIGURE 4 Variation of surface temperature as a function of  $M$  and  $P$  for daytime temperatures at times of HCMM measurements ( $\sim 1400$  LST solid lines and 0200 LST dashed lines labeled in  $^{\circ}\text{C}$  above 200 K). Initial conditions for model same as those for Figure 3. The lead digit has been deleted from the temperature values.

*Other models* In Europe a parallel effort has been made in developing predictive models for the remote determination of the governing surface parameters. Such models, as reported by Rosema *et al.* (1978), Nieuwenhuis and Klaassen (1978) and Soer (1977), are variants of the TELL-US model (Reiniger *et al.*, 1982) which is similar to that of CM, except the former does not account for a turbulent interaction between the surface layer and the free atmosphere above, nor does it possess a detailed predictive capability except in the substrate. It also requires meteorolog-

ical observations at 2 m as hourly input. The TELL-US model does, however, attempt to describe some detailed aspects of the crop-atmosphere exchange in a manner more closely tailored to the character of the crop canopy than does the model of CM. The models resemble each other with regard to the formulation of a crop-atmospheric resistance parameter (i.e.  $R_{cv}$  in Eq. 18), which is specified by Soer (1980) as a generalized canopy resistance which may depend upon the canopy geometry (Klaassen, 1979). A further discussion of the need for an improved vegetation parameterization is contained in Section 7.

Wetzel and Atlas (1981) and Wetzel *et al.* (1984) describe the use of a predictive boundary layer model of Wetzel (1978) for obtaining the governing surface parameters from measurements made from geostationary satellites such as GOES. Soil moisture is included in their model explicitly as a fraction of soil saturation, a similar parameter to  $M$  in Eq. (3); the model is otherwise very similar to that of CM. Application of the model to satellite measurements is made not using the full model but, instead, the morning rise in temperature in the model is fitted to a simple polynomial, which relates wind speed and normalized rate of rise of morning temperature to the soil moisture parameter; the relationship is assumed to be applicable to a wide range of meteorological conditions, latitudes and seasons. This method requires only two temperatures made about two to four hours apart during the morning but explicitly uses only one temperature difference.

The predictive model of Kahle (1977; see also Kahle *et al.*, 1975) also employs similar equations to those referred to above and has been used to infer the thermal inertia of geological formations. This model does not account for an exchange with the free atmosphere above the surface layer, nor does it specifically deal with latent heat flux.

#### (c) Diagnostic models

A class of models which can be referred to as diagnostic are the so-called combination models which make use of the flux relationships, such as Eqs. (17) and (18), combined with the surface energy balance Eq. (1). An example of a combination model is the Penman-Monteith equation (Monteith, 1975), which, as formulated does not lend itself to direct satellite measurement of surface temperature since it is expressed in terms of air temperatures. In using a combination model to derive surface moisture availability, concurrent surface and air temperature,

wind speed and humidity are required, the surface temperature to be provided by remote observations. In effect, Price (1982a, 1982b) uses a form of combination equation in Eq. (6) to determine the evaporation. Various combination models have been derived for agricultural purposes, such as those by Black *et al.* (1970), Brun *et al.* (1972), Ritchie (1972), Brown and Rosenberg (1973), Stone and Horton (1974), Heilman and Kanemasu (1976), Verma *et al.* (1976), and Byrne *et al.* (1979). These authors discuss various forms of combination equations as applied to vegetated canopies where detailed measurements of air temperature and humidity may be available.

Combination models are, in principle, easy to apply but they require a knowledge of the atmospheric parameters immediately above the surface. Consequently, there is a danger that the measured air temperature and that for the ground derived from satellite may be incompatible, yielding an erroneous or inconsistent result when vertical differences are formed between values at the canopy surface and in the atmosphere. In this respect, the prediction methods avoid the problem by determining internally consistent vertical profiles of temperature, wind, and humidity.

#### (d) Empirical models

A fourth class of model can be called empirical because the physical relationships are replaced by empirical equations which are based either on direct measurements or on the results of more complicated models. Underlying the empirical approach is the observation that leaf-air temperature differences (or soil-air temperature differences) are dependent upon soil moisture. Such models have been developed largely by the Arizona group (Idso *et al.*, 1975a, 1975b, 1975c; Idso *et al.*, 1976; Idso and Ehrler, 1976; Idso *et al.*, 1977a, 1977b, 1977c; Jackson *et al.*, 1977; Millard *et al.*, 1977; Ehrler *et al.*, 1978). Daily sums of leaf-air temperature differences, called the "accumulated stress", are shown to be related not only to soil moisture but to crop yield. Based on simple formulae, the measured canopy temperature during the afternoon or the measured day-night canopy temperature difference are related to measurements of soil moisture, evapotranspiration, or crop yield. For example, Jackson *et al.* (1977) gives the following equation which is based on an expansion of the surface energy budget Eq. (1) but with measurements used to determine values for the

coefficient  $B$

$$L_e E_o = R_m - B(T_c - T_a) \quad (19)$$

where  $T_c$  is the canopy temperature, analogous to  $T_o$ .

Idso and his group at Arizona have endeavored to utilize the empirical relationships between the temperature response, as measured by a radiometer, to derive expressions for evaporation and crop yield. These formulae, while of practical significance, suffer from being sensitive to ambient meteorological conditions and are appropriate for use only over the region and type of surface for which the expressions were derived.

So far, empirical models have not been applied directly to satellite measurements, although Heilman and Moore (1980) derive their empirical equations from an analysis of in-situ soil moisture data and corresponding HCMM satellite infrared surface temperature measurements. Though not universally applicable to satellite measurements, such empirical expressions provide further evidence that the satellite may be capable of determining soil moisture patterns (Seguin and Itier, 1983).

## 4 RESULTS

### (a) The urban heat island

One of the most notable meteorological anomalies produced by human activity is the urban heat island. This phenomenon is readily distinguishable on satellite imagery (Rao, 1972; Carlson and Augustine, 1977; Matson *et al.*, 1978) and is indisputably linked to human alterations of the surface, specifically the removal of vegetation and its replacement with non-evaporating and non-porous materials.

Recent studies by Landsberg (1979) for Washington, D.C. and by Changnon (1978) for St. Louis show that the precipitation patterns downwind of urban areas are significantly influenced by the heat island. The cause of this weather change in the urban areas appears to be due to differential heating at the surface and its effects on the boundary layer (Kropfli and Kohn, 1978; Wong and Dirks, 1978). Significant inflow of surface-level air into the warmer parts of the city and the subsequent lifting of the converging warm air leads to the formation of an elevated warm plume downwind of the urban centers.

In Figure 5, taken from Carlson *et al.* (1981), the warmest temperatures appear over downtown St. Louis (the location marked D), East St. Louis (E) and over Granite City (G). The temperature pattern in Figure 5 is very similar to those obtained for other HCMM scenes (CM) and for TIROS-N measurements (unpublished data) over this city. Moreover, the overall character of the urban temperature anomaly also resembles those for other urban centers, such as Los Angeles (Carlson and Augustine, 1977; CM), Houston, TX and Washington, D.C. (unpublished analyses).

A close relationship is evident between the afternoon temperature and moisture availability upon inspection of Figures 5 and 6. Temperatures exceeding 40°C over the downtown portion of the city correspond to relatively low values of  $M$  (less than 0.2) in Figure 5 and to relatively high surface heat fluxes (in excess of 200 Wm<sup>-2</sup>) in Figure 7. Interestingly, despite the pronounced urbanization over these cities, the Bowen ratio ( $H_o/L_e E_o$ ) is near 1.0 on a scale of a satellite pixel, a situation which attests to the importance of existing trees and grass in the evapotranspiration process over cities. A similar conclusion was reached by Oke (1982) for Vancouver, B.C.

Another curious result obtained by CM was that the distribution of thermal inertia, once thought to be the important parameter in forming the nocturnal urban heat island, exhibits a rather flat pattern with no pronounced maximum over the city. It appears that the daytime heat island, in summer at least, is the driving mechanism behind the nocturnal temperature anomaly, rather than the distribution of absorbing materials in the substrate. The city remains relatively warm at night because the surface is hotter during the day and more heat is transferred into the substrate because of the greater surface-substrate temperature difference. The urban summertime nocturnal temperature maximum therefore is ultimately the result of vegetation removal in the urbanization process. A similar conclusion regarding the genesis of urban heat islands was reached by Oke (1982) except that he regards canopy geometry to be as important as moisture.

### (b) Soil moisture over vegetated areas

It is obvious from examination of Figure 6 that HCMM satellite imagery is capable of detecting significant variations in ground moisture over urban areas. Over vegetation the response of the thermal

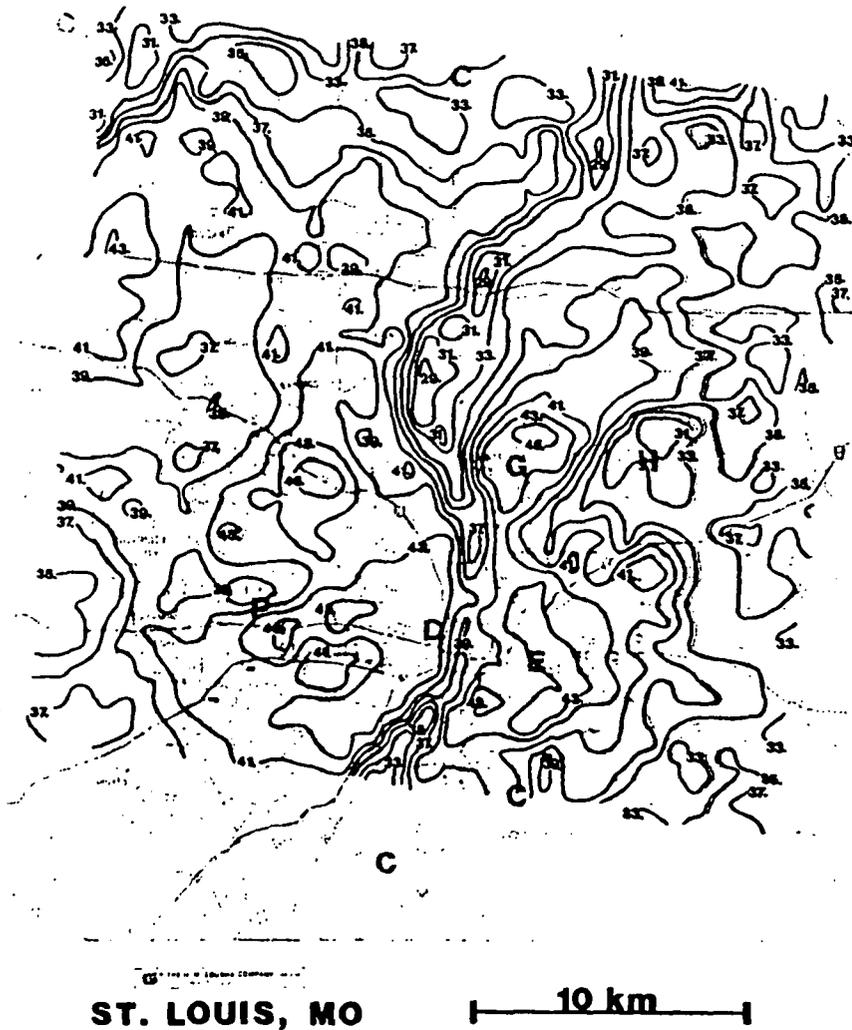


FIGURE 5 Surface temperature analysis in °C over St. Louis, MO, at approximately 1330 LST, 23 August 1978. Bold-faced capital letters refer to surface types or sites (C: croplands; D: downtown St. Louis; E: east St. Louis; G: Granite City; BE: Horshoer Lake; P: park).

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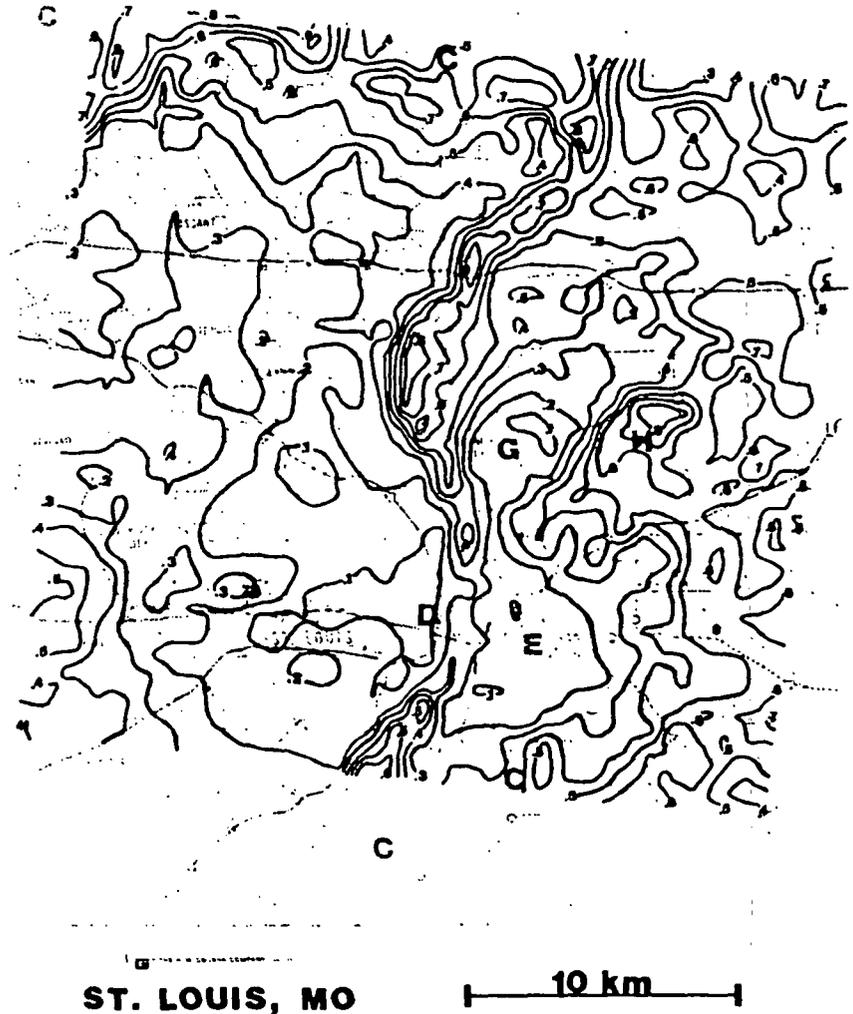


FIGURE 6 Surface moisture availability analysis over St. Louis, MO, 23 August 1978 based on a day-night HCMM image pair.

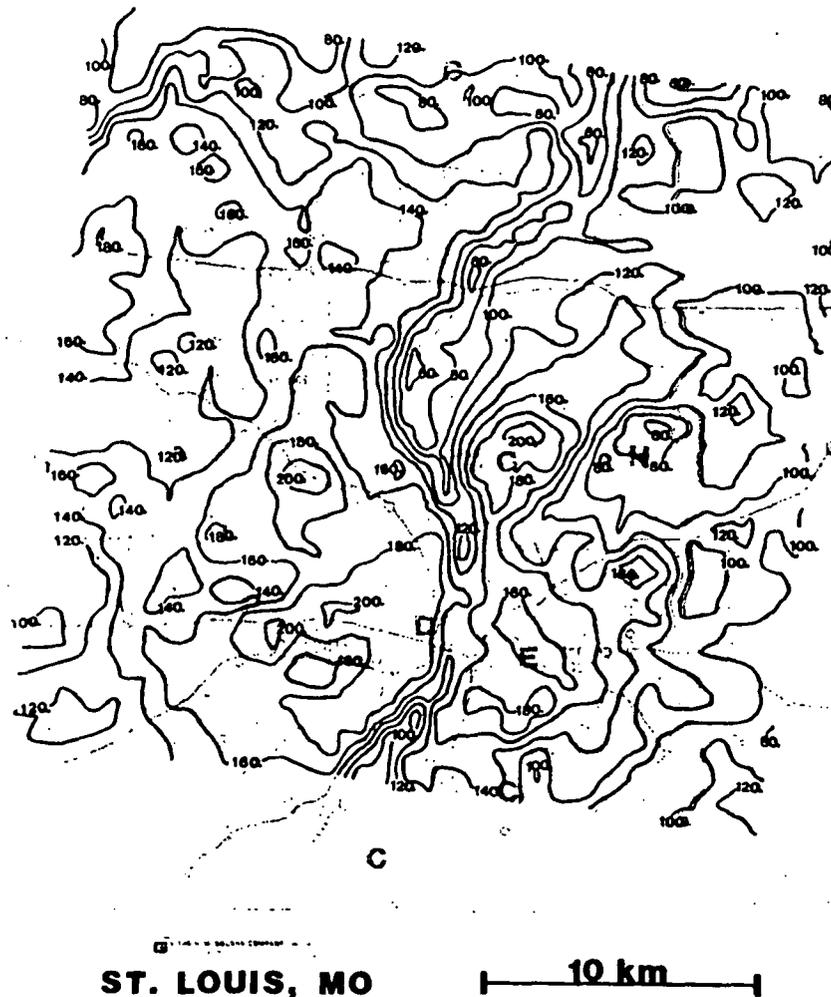


FIGURE 7 Surface heat flux analyses ( $W m^{-2}$ ) for 1330 LST 23 August 1978 based on day-night HCMM image pair.

pattern to soil moisture is much more complex than over bare soil or urban areas. While temperature variations between crops can reveal significant differences in evapotranspiration or soil moisture, such

differences may be no greater than a few degrees or less (under identical meteorological conditions) between extremes of dryness. Because vegetation type and geometry play such an important role in plant response to thermal stress, it is still not clear to what degree quantitative measurements of soil moisture can be made over vegetated surfaces. Indeed, Monteith (1981) has voiced reservations concerning the capability of satellites to measure soil moisture over vegetated surfaces.

In the face of such pessimistic evaluations, recent evidence cited in this paper suggests that the satellite possesses a modest capability of measuring a soil moisture parameter with some degree of definition over short vegetation such as grass. Evidence for this positive assessment is provided by earlier studies over vegetated regions by Kocin (1979), Heilman and Moore (1980) and Harlan *et al.* (1980).† In this paper we present further evidence of a relationship between the derived  $M$  parameter and antecedent precipitation over regional-scale areas covered by short grassy vegetation.

An analysis of both HCMM and GOES satellite temperature measurements was made on 27–28 July 1978 over a region in eastern Kansas (Figure 8). This area was chosen because of the extreme variation in precipitation and in the crop moisture index over eastern and central Kansas during the summer of 1978 (Figure 9). HCMM images 36 h apart on July 27 and 28 and GOES images on July 27 were obtained over the target areas (the solid and dot-dash rectangles shown in Figure 8). During this two-day period clear skies and light winds characterized the weather over Kansas. Over the target area the terrain consists largely of unirrigated range and crop land with corn and winter wheat the dominant cash crops. Very light rainfall (less than 0.5 inches) had fallen during the previous three-week period over the southwestern corner of the target area. Low values of the crop moisture index ( $-2$  to  $-3$ ) occurred over the southwestern corner of the domain with relatively higher indices ( $-1$  to  $+1$ ) over the extreme northern and eastern portions of the working area.‡ Similarly, the three-week cumulative rainfall totals were less than 1.0 over the southwestern

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†Harlan, J. C. (1980). Dryland pasture and crop conditions as seen by HCMM. Report to HCMM Investigators Meeting, Nov. 24–26, 1980, NASA/GSFC, Greenbelt, MD. Unpublished and undated manuscript.

‡U.S. crop moisture index published in the Weekly Weather and Crop Bulletin, U.S. Department of Commerce.

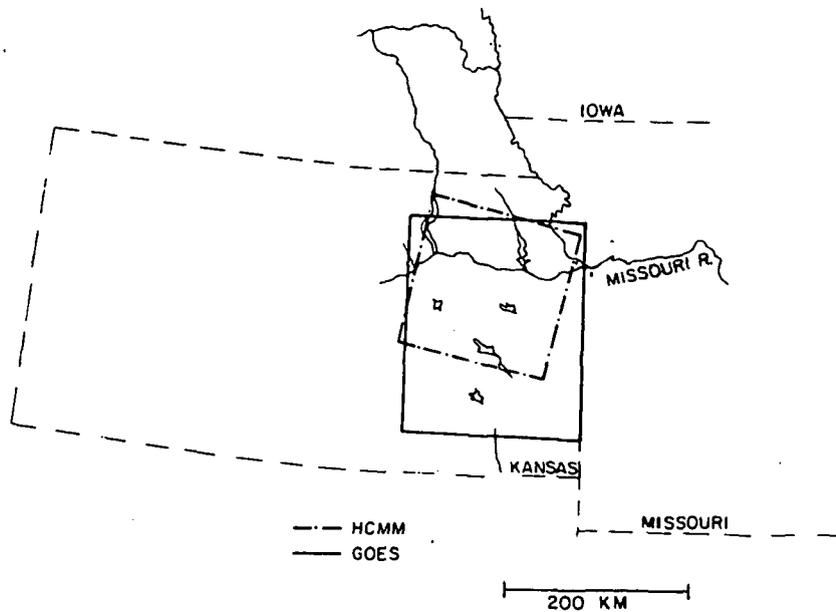


FIGURE 8 Sketch of Kansas and adjoining state boundaries, showing analysis areas for HCMM and GOES.

quadrant (Category 3 in Figure 9) with amounts above 2.0 inches (Category 4) over eastern and northern portions of the domain.†

Clearly, if soil moisture variations associated with rainfall anomalies are to be detected by satellite, they will certainly be most noticeable in situations where there are large horizontal gradients of rainfall, such as appeared over Kansas during July 1978. Moreover, errors in measurement and model are less serious where  $M$  is small (Section 5). The surface temperature analysis derived from HCMM for the afternoon of July 28 appears to confirm this expectation (Figure 10). Thus, in comparing Figure 10 with Figure 9, high temperatures (generally

†Cumulative precipitation is here defined as the total precipitation over a three-week period preceding the date of the analysis. Antecedent precipitation ( $API$ ) is defined according to Saxton and Lang (1967) and Blanchard *et al.* (1981) using formula  $API = kAPI_{i-1} + P_i$  in which the precipitation  $P_i$  on day  $i$  is added to a weighted value of the existing  $API$  on day  $i - 1$ . The result is a precipitation amount which decays toward zero when no new rainfall occurs. The weighting factor ( $k$ ) was taken as 0.92.

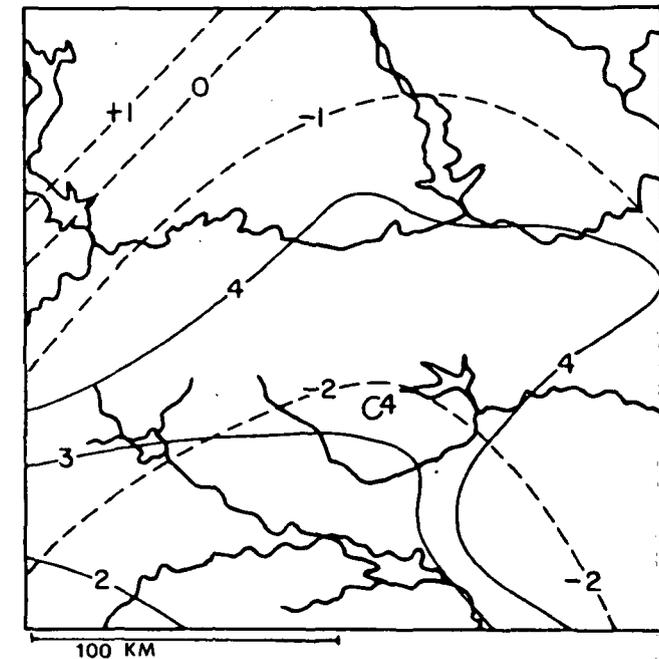


FIGURE 9 Three-week cumulative rainfall (solid lines) and crop moisture index (dashed lines) for the Kansas case over the HCMM domain of Figure 8, 29 July 1978. Contours of precipitation represent logarithmic scale: 1=0.25 inch; 2=0.5 inch; 3=1.0 inch; 4=2.0 inch. Background shows streams and lakes.

40–50°C) are found over the arid southwestern sector and relatively cool temperatures (about 35°C) over the north and east. Similar patterns were obtained for analyses based on GOES infrared satellite measurements over the larger rectangle in Figure 8.

Not surprisingly, these satellite temperature patterns translate into similar patterns of the moisture availability  $M$ , which is less than 0.25 over the southwestern portion of the working area and greater than 0.75 but less than 1.0 over the east and north of the area, as shown in Figure 11. Similarly, the patterns of surface heat flux (Figure 12) resemble both the daytime temperature and moisture availability patterns with high values over the southwest corner of the area and low fluxes, corresponding to the relatively moist terrain, over the eastern

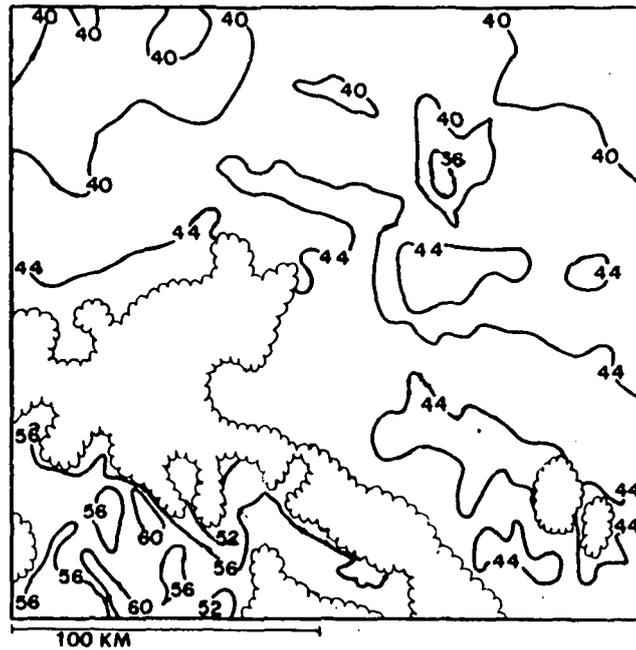


FIGURE 10 Surface temperature analyses ( $^{\circ}\text{C}$ ) over the smaller rectangular area of Figure 8 at approximately 1330 LST, 27 July 1978.

and northern quadrants of the figure. Contours of total daily evaporation (not shown) closely resemble those of  $M$ .

The correspondence between  $M$  (or  $E_0$ ) and precipitation is not an exact one. There are obviously some areas where agreement between rainfall and moisture availability is lacking, as might be expected since precipitation is only one component in the hydrological budget, which depends upon evaporation, runoff, and local water sources such as underground aquifers and irrigation. Nevertheless, the linear correlation coefficient ( $R$ ) between the moisture availability and precipitation fields (Figure 9 Versus 11) exceeds 0.6, indicating a significant degree of sensitivity of  $M$  to precipitation. In a similar case study, for an area over southern Illinois and Indiana (not presented), the correlation between  $M$  and precipitation was found to be about 0.6 for both HCMM and GOES analyses. Wetzel and Atlas (1983) found a log-linear correlation exceeding 0.8 between antecedent precipitation and the morning rise in

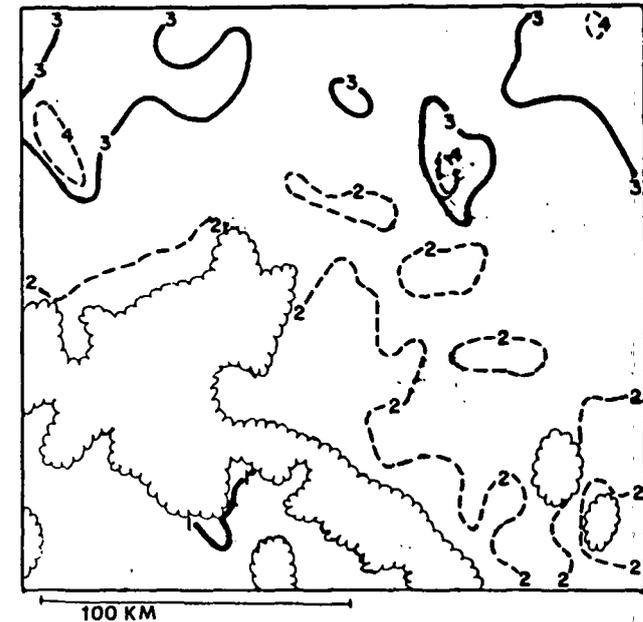


FIGURE 11 Moisture availability ( $M$ ) 23 August 1978 determined from a day-night HCMM image pair over the area of Figure 10.

temperature for the same July 1978 period over the Great Plains. Recently, we reworked all of the 1978 GOES satellite measurements over the Kansas target area, adding 10 more cases from the dry summer of 1980 (Carlson *et al.*, 1984). The mean correlation coefficient for all 12 cases for a log-linear relationship between  $API$  and  $M$  is about 0.7, but the relationship is clearly based on agreement between  $API$  and  $M$  on the larger scales, the fine details of the patterns showing relatively poor agreement. The results were better for a log-linear relationship (rather than a linear one) because  $M$  becomes relatively insensitive to changes in precipitation when it approaches 1.0 or as the soil reaches field saturation. Wetzel and Atlas (1983) also analyzed a much larger area than that shown in Figures 9–12 and derived a relatively smooth pattern. Inspection of the moisture availability pattern (Figure 11) indicates that the poorest correlation between  $M$  and rainfall is on the smaller scales, possibly because such scales are inherently more noisy or because the rainfall observations are not sufficiently dense to capture

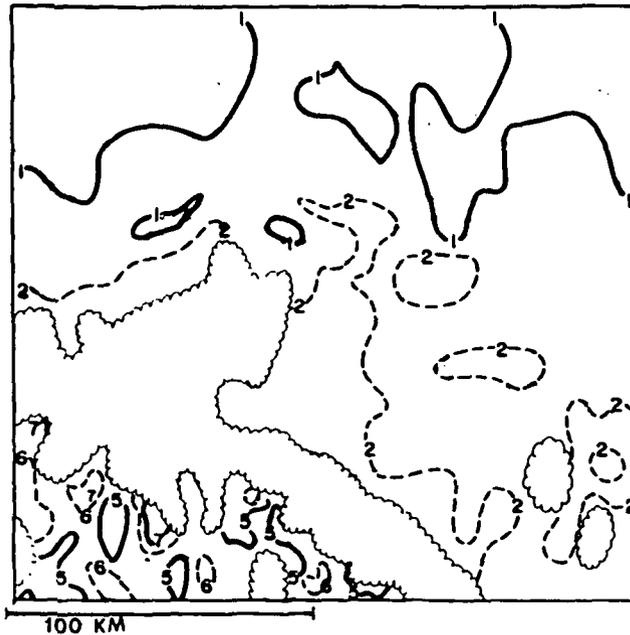


FIGURE 12 Surface heat flux analysis ( $H_s$ ) determined from a day-night HCMM image pair over the area of Figure 10 at approximately 1330 LST 23 August 1978. Contours have the following values; 1 =  $105 \text{ W m}^{-2}$ ; 2 =  $140 \text{ W m}^{-2}$ ; 3 =  $210 \text{ W m}^{-2}$ ; 5 =  $245 \text{ W m}^{-2}$ ; 6 =  $270 \text{ W m}^{-2}$ ; 7 =  $305 \text{ W m}^{-2}$ .

small-scale variations in  $M$ . Our results show an improvement in correlation when the data is smoothed. Wetzel and Atlas' analyses were derived from GOES but there is no evidence to suggest that this satellite is superior to HCMM in this regard; on the contrary, we have found slightly better correlations between  $M$  and precipitation with HCMM than those of GOES.

The presence of relatively high temperatures over the arid portion of the working area is somewhat puzzling. It is unlikely that the vegetation itself can reach temperatures as high as those shown in Figure 10. Our experience with the operation of manual radiometric measurements over vegetation during periods of low rainfall indicates that, while leaf temperatures may rise slightly above air temperature, the temperatures of bare soil and short grass between the plant rows or surrounding the cultivated fields may rise by a considerable amount during periods of

drying (Myers and Heilman, 1969; Cooper, 1981). Although a single satellite pixel may contain both vegetation and bare soil, remote temperature measurements made over areas the size of a satellite pixel appear to respond rather sensitively to the elevated warmth of bare soil and grass clumps which are interspersed throughout the taller vegetation. In the case of Kansas, the vegetation consists largely of grassland and unirrigated crops. On the other hand, the vegetation cover is more luxuriant over southern Indiana and Illinois where a significant correlation between  $M$  and rainfall was also found using both HCMM and GOES measurements (Polansky, 1982).

The response of moisture availability to precipitation over a regional-scale area was noted by Kocin (1979) in a study involving the analysis of HCMM temperatures over the Goodwater Creek watershed in Missouri (Figure 13). Approximately eighty-five per cent of the watershed is cultivated, three-quarters of which consist of corn, wheat, soybeans and various other grains and grasses. HCMM measurements were made for a pair of orbits 36 h apart on 9–10 June 1978, a period

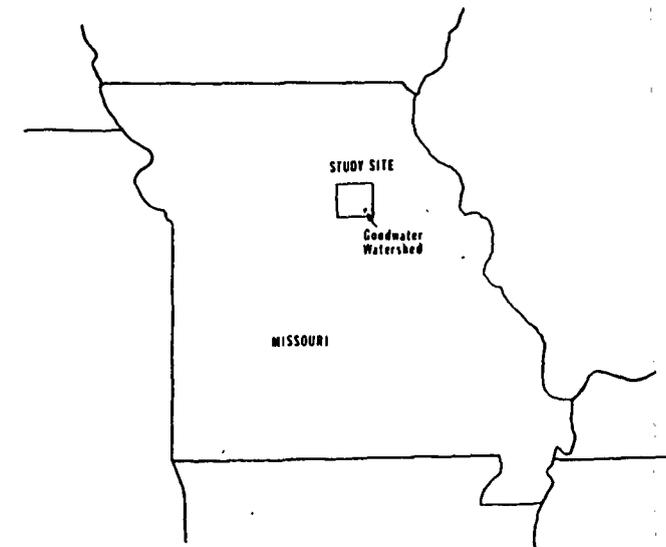


FIGURE 13 Location of Goodwater Watershed and surrounding countryside.

when the fields would have been relatively bare. The weather during this period was clear and cool.

The moisture availability pattern for this case (Figure 14) contains a considerable amount of small-scale detail, some of which must be related to the type of land surface and crop cover. Nevertheless, the correlation between  $M$  and cumulative precipitation is about 0.6. Kocin's results, however, show virtually no correlation between  $M$  and rainfall for analyses made for a case during late September 1978. Although it is likely that the vegetation would have been relatively dense and perhaps at least partially senescent, the crucial factor in the lack of correlation is that the horizontal rainfall variations over the watershed were quite small in September.

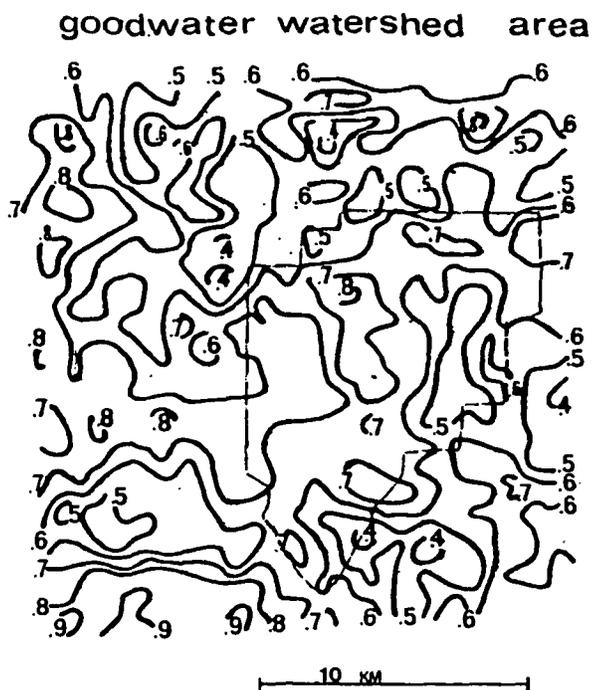


FIGURE 14 Enlargement of unsmoothed moisture availability analysis over the Goodwater Watershed area (outlined by thin dashed border), a sub-set of the rectangular area in Figure 13. Analysis for 10 June 1978, as determined from day-night HCMM image pair.

## 5 AN EXAMINATION OF ERRORS AND UNCERTAINTIES IN THE INFRARED METHOD

### (a) Model uncertainties

Model uncertainties can be described as (1) conceptual in origin, (2) the result of empirical physics and (3) due to limiting assumptions. Conceptual errors are related to the fact that a model represents a naive idealization of the ground surface, representing it as flat and homogeneous, overlying a purely diffusing substrate. Most boundary layer models currently in operation inadequately represent complex surfaces. Those which attempt to incorporate details of a vegetation canopy, such as the model of Deardorff (1978), tend to be based on untested relationships and, in any case, require more information than can be supplied or specify parameters whose values are virtually unknown and perhaps unknowable. There has been some attempt by European workers (e.g. Nieuwenhuis and Klaassen (1978) and Klaassen (1978)) to parameterize explicitly the soil and leaf resistances within a canopy-air layer corresponding closely in concept and formulation to the transition layer of CM. Thus, coefficients similar to  $C_a$  or  $C_s$  in Eqs. (18a) and (18b), are determined from direct aerodynamic measurements over leaves. Deardorff's model has been tested against real data using a boundary layer model (Jersey, 1982), which is that of Blackadar (1978a, 1978b). The results of this study were inconclusive regarding the possible improvements in accuracy to be made by inclusion of the Deardorff vegetation formulations. Klaassen (1978), however, found that the European TELL-US model yielded relatively accurate evaporation values, when compared to measurement made under controlled conditions with carefully chosen resistance parameters.

If the inclusion of an explicit vegetation parameterization for a well-known canopy can sometimes result in improvements in predicting surface evaporation, it does not solve the problem of defining the real nature of the surface under general conditions. That surface can consist of vegetation and other complex surface features, such as buildings or irregular terrain, which possess a bulk surface temperature which may vary with both the viewing angle of the sensor and the geometry of the sun and the surface. A model temperature, however, is obtained by solution of equations for some idealized surface. Both the predictive and analytic models suffer from their dependence upon a solution of the substrate diffusion Eq. (2), which is often applied indiscriminately to

any type of surface, even that which is not a truly diffusing one. Deardorff (1978) eliminates the need for using the diffusion equation within a vegetation canopy by explicitly parameterizing the transfer of heat between the canopy surface and the ground. He also deals separately with ground covered by vegetation by specifying a fractional coverage of biomass, an approach also used by Wetzal (1978).

Diagnostic models, in depending on a bulk canopy temperature alone, are independent of any substrate formulation, although they nevertheless suffer from conceptual ambiguities regarding the nature of the surface and from an inability to determine representative and internally consistent atmospheric properties. CM suggests that the conceptual problem of incompatibility between model temperatures and those measured remotely can be circumvented partially by considering the model parameters as *effective* values which are required to yield the correct (i.e. measured) surface temperatures in a similar boundary layer model. Since it is clear that the measured surface temperatures do vary in some systematic fashion with the moisture availability and since the physical relationships used in most models are very similar to each other, the model parameters should possess a physical meaning, although they do not necessarily lend themselves easily to direct measurement.

In regard to the model physics, the general laws governing the vertical turbulent transfer of heat from the surface to the atmosphere are fairly well-known (Panofsky, 1974). Nevertheless, the transfer coefficients for the eddy, molecular and radiative diffusion of heat, for example, are based on empirical relationships derived from an average of measurements made over relatively homogeneous terrain. Underlying these diffusion laws is the concept of the logarithmic profile, which applies to temperature, moisture and momentum. These profiles are subject to considerable deviation from a pure logarithmic behavior when the static stability (the change in potential temperature with height) deviates from its neutral (isentropic) state or when the terrain is inhomogeneous. Corrections to the logarithmic profile can be made for stability but since the stability is intimately dependent upon the heat flux and wind speed, the non-neutral temperature, moisture, momentum profiles require a knowledge of that which one is attempting to calculate. In practice, a solution to this problem is achieved by a process of iteration, as in CM, or by simply neglecting the stability corrections.

Immediately adjacent to the surface, eddy transfer of atmospheric

quantities becomes quite inefficient and the process of diffusion is carried out by a combination of eddy, molecular and (in the case of heat) radiative processes. This near-surface transfer, discussed by Klaassen (1978) and others, is itself highly dependent upon the nature of the surface, particularly the roughness. Large values of the near-surface resistance allow large gradients of the temperature and humidity to be established at the ground-air interface. The correct values for these near-surface resistances are unknown in most cases, being highly dependent on the details of the surface canopy. In most models the near-surface resistances are simply included in the bulk atmospheric resistance to eddy diffusion.

Various limiting assumptions are made in all models, the most serious of which, in the case of one-dimensional models, is the neglect of advection. Advection can be considered to be of two types, large-scale and local. Large-scale advections occur as the result of gradients (temperature, moisture and wind) which exist over a broad area. Local advection occurs as the result of horizontal gradients in atmospheric properties which exist as the result of spatial variations in the surface heating imposed by variations at the surface. Although it is possible to minimize the effects of large-scale advections by simply choosing to investigate situations when the meteorological conditions are appropriate, the small-scale advections are inherent in the nature of inhomogeneous surfaces and their neglect will tend to underestimate the magnitudes of the atmospheric fluxes in the model and may, in some instances, when vegetation is surrounded by expanses of arid terrain, result in the calculated fluxes being of the wrong sign (Brakke *et al.*, 1978). Our own experiences suggest that neglect of advection does not seriously degrade the results.

Another limitation which is particularly important in predictive and analytic models lies in the specification of various internal parameters, such as roughness, albedo, and the ambient meteorological conditions. In order to achieve a tractable solution a relatively few number of unknowns are allowed. In analytic models, such as that of Price (1982a, 1982b), the atmospheric parameters, temperature, moisture and wind, are determined as daily averages. In truly predictive models atmospheric parameters are specified as initial conditions, whereas in analytic models a single set of daily averages must be provided from large-scale data. These data must be obtained from soundings and weather maps, and may not precisely represent the real atmospheric conditions at any

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given pixel point. Even more difficult to specify is the initial temperature profile in the ground including the substrate reservoir temperature which is presumed to remain a constant lower boundary condition. Tests with the CM show that the night-time temperature is sensitive to the value of temperature at the lower substrate boundary although the initial temperature profile in the layer between the substrate reservoir and the surface is not a significant source of uncertainty. Deardorff (1978) shows, in fact, that the depth of the substrate reservoir is not an arbitrary choice. Moreover, the calculated surface temperatures are somewhat affected in predictive models by the vertical grid spacing, particularly in the upper substrate layer. Analytic or diagnostic models do not suffer from such uncertainty, being independent of an explicit substrate formulation, but they are more afflicted by uncertainties in the atmospheric conditions. In addition to the somewhat arbitrary manner in which albedo, roughness and other model parameters are assigned, the primary model unknowns, moisture availability and thermal inertia, are presumed to be invariant in time; Jackson's (1973) data indicates that the fractional amount of soil moisture may possess significant diurnal variation at the surface but it is relatively constant when averaged over typical vegetation rooting depth.

A further source of error, which is numerical in origin and peculiar to time-dependent models, is due to the method of inverting the model to calculate  $M$ ,  $P$  and other parameters. In the CM, the information contained in Figure 4 is expressed by a series of second-order multiple regression equations. Regardless of how the model output is expressed, however, a certain amount of inexactness occurs in deriving the governing parameters. Raffy and Becker (1984) address the problem of model inversion, maintaining that with proper constraints, an accuracy of a few percent is achievable. For the CM a 5 to 10 percent inversion error is typical for determining  $M$ .

#### (b) Measurement error

Measurement error can be divided into two classes; that arising from sensor error and that from uncertainties introduced by the intervening atmosphere and by the substrate itself. Sensor error is often quoted as a noise equivalent  $T$  (called the  $NE\Delta T$ ), which is usually much less than  $1^\circ\text{C}$ . Sensor drift, which is often unknown, may be quite large. HCMM,

for example, was found to have a  $5.5^\circ\text{C}$  calibration error which is thought to be sensor drift which occurred shortly after launch. To a great extent, however, absolute calibration errors can be eliminated by considering temperature differences rather than absolute temperatures in the model, as in analytic models, in the three-image method of Polansky (1982) or in the method of Wetzel and Atlas (1983) who treat the rate of morning temperature rise as their measured variable.

Atmospheric attenuation can introduce a more serious error in the temperatures than thermal noise, even in the water vapor window region customarily used in thermal scanners. The relationship between the water vapor correction and surface temperature is almost linear with surface temperature over the range of temperatures normally measured and it is exceedingly sensitive to the static stability and moisture content near the ground. The correction may be as much as several degrees centigrade or more during the day but can be negative under stable night-time conditions (Nieuwenhuis, 1979; Price, 1983). Since the atmosphere is seldom homogeneous with regard to moisture, sizeable temperature errors, possibly amounting to a degree or more, can arise from uncertainties in the water vapor correction. These errors are magnified in moist atmospheres. Error also can be introduced by neglect in horizontal variations in the surface emissivity. Although emissivities of most natural surfaces such as vegetation are very close to 1.0, an error of 0.01 corresponds to an error in the measured blackbody temperature of up to  $0.75^\circ\text{C}$ . Oke (1972) suggests that there may be an effective decrease in ground emissivity and in albedo as the result of the blocking of outgoing long and short wave radiation over rough surfaces such as those covered by buildings. Over certain types of dry surfaces, the ground emissivity may be significantly lower than 1.0.

Perhaps the greatest measurement problem is due to the presence of cloud, both seen and unseen. Figure 15 is an illustration of how a predictive model behaves when a cloud, defined as a reduction in the net surface radiation ( $R_n$ ) of 60 per cent, is imposed suddenly in the model for one hour between 1100 and 1200 local time. Total recovery of the surface temperature to the cloud-free case never quite occurs, although by 1500 local time, 3 hours after the cloud has vanished, the residual error in temperature is less than a degree. It is evident that even if cloud is absent at the time of satellite observations, the presence of cloud at any other time can influence the surface temperature.

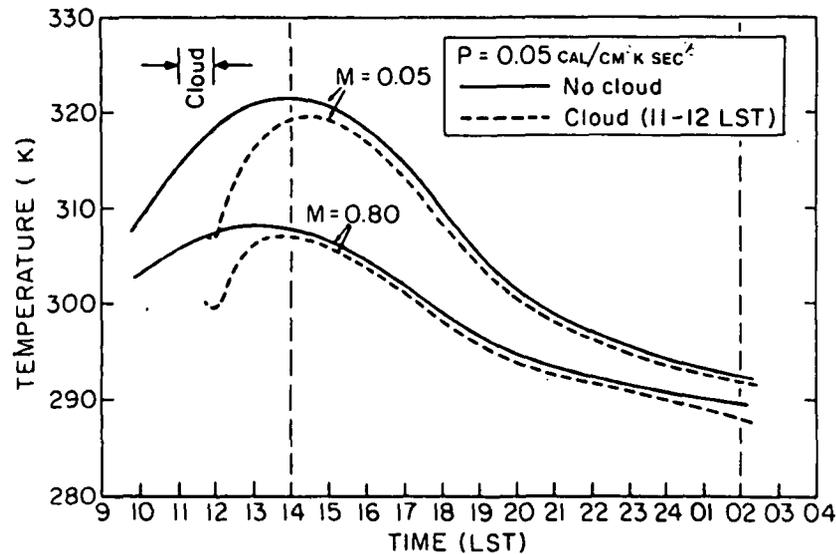


FIGURE 15. Simulated surface temperature for two values of  $M$  and a  $P$  of  $0.05 \text{ cal cm}^{-2} \text{ K}^{-1} \text{ s}^{-1/2}$  for clear skies throughout the day (solid curves) and for a case when a cloud decreases the net radiative flux by 60% between 1100 and 1200 LST (dashed lines branching after 1100 LST). The vertical lines at 1400 and 0200 LST denote the approximate HCMM image times. Model initial conditions same as for Figure 3.

Moreover, the presence of undetected cloud in the form of thin cirrus streaks or small cumulus can lead to anomalously cool temperatures which bias the results in a manner which is not necessarily uniform in time or space. To some extent, obviously cloudy areas can be ignored, as in the scalloped regions of Figures 10–12. The presence of cloud at time other than those of the satellite measurements cannot be determined easily, however. In our experience, it is very rare to find a regional-scale scene completely free of clouds over an entire day. Wetzel and Atlas' (1983) use of the morning rate of temperature rise minimizes but by no means eliminates, the problem of cloud. Satellites such as GOES afford a greater freedom in the choice of measurement times but they suffer more from cloud contamination because of their lower resolution with respect to that of HCMM or the NOAA vehicles. Diagnostic models, of course, are least affected by cloud but more sensitive to the need for an accurate water vapor correction.

One inherent error in all satellite measurements is that due to averaging of radiances within individual pixels. Price (1982a) addresses the problem of deriving large-scale averages of surface parameters from an array of individual pixel measurements. He reasons that averages of the derived surface parameters may not necessarily yield the correct average surface fluxes in a model. Similarly, individual pixels contain a multitude of surface elements, whose total emittance is integrated in some fashion by the sensor to yield a single temperature measurement. Although CM maintained that the integrative nature of a pixel measurement is an advantage because it allows one to define an effective radiating temperature of the surface, there are instances when that temperature can be misleading and lead to erroneous values in a model when large variations in the surface temperature and surface properties exist within a pixel as, for example, when a body of cool water lies adjacent to hot, dry ground. Errors generated by surface inhomogeneities are generally not large, however.

Registration errors in satellite images can produce serious pattern distortion in the vicinity of large surface temperature gradients and can lead to fictitious patterns of the derived model parameters where there are strong gradients in surface character. A careful analysis of ground control points can minimize registration error but it is unlikely that a pair of images can be rectified to identical ground coordinates to within an average error of one or two pixels. Registration error most seriously affects models which make use of more than one image but this error probably becomes relatively less important as the scale of analysis and the degree of image smoothing is increased.

Finally, a potentially serious source of error arises from the presence of non-uniform terrain height and terrain slope. While it is theoretically possible to correct for incident solar radiation on a sloping surface, knowing the elevation and azimuth angles of the slope at each point (Watson, 1975), the implementation of such a correction is cumbersome and perhaps futile in view of the equally deleterious effects of terrain slope on the validity of the flux equations. Moreover, since temperature, moisture and wind vary with height in some fashion, the ambient meteorological parameters used in a model would need to be adjusted accordingly in order to be consistent with the slope corrections. Sensitivity tests performed with solar models over sloping terrain suggests that the temperature errors due to terrain slope become crippling only when the slope angle exceeds about ten degrees.

### (c) Overall impact of error on the results

It is clear that errors in the derived moisture availability and thermal inertia are dependent on the type of model and the scale and degree of smoothing of the thermal images. Although no definitive assessment of the total error on the results can be made for this reason, a reasonable estimate for the net error of predictive or analytic models is likely to lie between one and two degrees centigrade. It is quite probable that the net random error in satellite measurement and image rectification is also between one and two degrees centigrade, a value of which has also been suggested by Boldyrev and Kharmarink (1973) as a representative lower limit for satellite temperature sensor errors. Consequently, an effective but probably irreducible error in the derived values of  $M$  and  $P$  is about plus or minus 0.1–0.2 and 0.01–0.02  $\text{cal cm}^{-2} \text{ s}^{-1} \text{ K}^{-1}$ , respectively.

The effects of a random 2°C error on  $M$  and  $P$  can be seen by inspection of Figure 4, in which the isotherms have been drawn at two degree intervals. Solenoids formed by the intersection of day and night temperatures therefore form a typical "error space" for both  $M$  and  $P$ . Evidently, serious or crippling errors can arise under relatively moist conditions, especially where  $P$  is large. Errors are relatively small over dry terrain where  $P$  is small. Under some meteorological conditions, however, the error solenoids become very large near values of  $M = 1.0$  so that slight temperature errors can produce unreasonably large values of  $M$  where the terrain is moist.

The host of model and measurement error limit the accuracy of the derived parameters but also justify many simplifying assumptions made in the various methods, such as the use of large-scale meteorological data for initial conditions, the neglect of advection when the atmospheric gradients are small, the neglect of higher order components in the Fourier expansion, the choice of a lower substrate boundary temperature from climatological data, etc. Similarly, no single source of the two degree temperature measurement error can be identified as being the main contributor.

Both the predictive and analytic models require either two or three satellite temperature measurements over the duration of one solar cycle. The relative importance of any error magnifies error in the results, however, as the difference in temperature between image pairs decreases; the effect of model and measurement error on the results become overwhelming when these temperature differences are small.

Consequently, as Wetzel and Atlas (1981) and Polansky (1982) demonstrate, there are relatively narrow time (and perhaps seasonal) windows within which satellite measurements can prove useable. Sensitivity tests with the model CM (see Polansky, 1982) show that a suitable image pair can be made only when the daytime image is made close to solar noon (1130–1500 local time) and the other image made either the following night (1100–0300 local time) or during the early morning, preferably one to two hours after sunrise (0730–0900 local time for midwestern summertime images). The three-image method of Polansky (1982), while capturing more of the diurnal temperature variation than the two-image method used by Wetzel and Atlas (1983) with GOES and by CM with HCMM, is also subject to greater uncertainties due to measurement and image registration error.

## 6 CONCLUSIONS

It is clear that despite limitations of the IR method, reasonable patterns of surface moisture and thermal inertia can be determined. Patterns of surface moisture or heat flux are probably most valid over flat, arid, sparsely vegetated terrain under conditions of weak large-scale advection, few clouds and strong sunshine. A primary model limitation is the inability to represent the vertical transfer of heat, momentum and moisture properly over complex terrain, such as forests or cities, and over deep water bodies. In regard to measurement, the greatest source of error lies in the interference by clouds, both seen and unseen, and the presence of other moisture inhomogeneities in the atmosphere.

Both the predictive and analytical techniques offer the greatest promise for use in remote soil moisture analysis. Because of the various sources of error, it is unlikely that predictive models are demonstrably superior to analytic ones, even with the fuller treatment of the atmosphere (Hechinger *et al.*, 1982). Regardless of the technique used to derive soil moisture or thermal inertia, it is unlikely that the derived moisture values are correct to within plus or minus 10–20 per cent. Thus, moisture availability, which varies in nature between 0 and 0.1, cannot be determined to within about plus or minus one-half to one category in a range of four or five categories which one can designate as: extremely dry (0.2 or less), moderately dry (0.2–0.4), neither dry nor moist (0.4–0.6), moist (0.6–0.8) and very moist (0.8 or greater). In the

final analysis, the purpose of a soil-atmosphere model is to serve as a rather imperfect filter of extraneous atmospheric noise embedded in a rather ambiguous measurement and to allow that measurement to be replaced by a purer representation of the soil moisture.

## 7 PERSPECTIVES

The moisture availability patterns presented in Figures 6, 11 and 14 correlate significantly with antecedent rainfall and represent an effective soil moisture parameter which, when inserted into a boundary layer model, will presumably yield the measured surface temperatures and the correct surface evaporation and sensible heat fluxes. Heilman *et al.* (1976) and Heilman and Moore (1980) have demonstrated not only that there is a strong correlation between soil moisture and the remotely determined surface temperature response over grassland but that those measured surface temperatures can be used to predict crop yields. Recent numerical modeling experiments (McCumber, 1980; Shukla and Mintz, 1982; Benjamin, 1983) underscore the importance of soil moisture in prediction of both large and small-scale atmospheric circulations. Moreover, soil moisture is undeniably a vital parameter for agriculture and hydrology.

The question is not whether soil moisture is a useful parameter but whether the soil moisture values derived from the infrared method can be sufficiently detailed and accurate to be of use and whether the information derived from the satellite can be delivered efficiently to potential users. Much more investigation must be done in order to assess the applicability and practicality of the method. Interpretation of the meaning of the soil moisture parameter must remain tentative until more is understood about it. Specifically, several theoretical problems need to be addressed:

1) How is  $M$  defined in a vegetative canopy or over other complex surfaces, especially where the radiating surface lies a significant distance above the ground? How can the heat transport be parameterized within the canopy and substrate without resorting to in-situ measurements and still achieve a closure?

2) To what extent is the temperature of the effective radiating surface

affected by ventilation when the radiating surface is at some distance above the ground?

3) Does the moisture parameter vary significantly during the day as plant stomata open and close in response to a variety of stresses? What rooting depth does the soil moisture parameter represent?

4) Are the values of evapotranspiration and sensible heat flux calculated with the derived  $M$  fields essentially correct even if the exact nature of the surface cannot be defined or realistically modeled?

5) What resolution of the satellite is necessary to achieve optimum results and what scale of analysis yields the best agreement with ground truth measurements? Is the higher resolution of a satellite such as HCMM desirable (compared to that of GOES), as Polansky (1982) suggested, or will the derived soil moisture parameter prove more useful when analyzed on a large scale with relatively crude resolution, as Wetzel and Atlas (1983) seem to suggest?

At present the institution of an operational remote soil moisture analysis program is technically, if not practically, feasible. The chief impediments to the operational use of the infrared method are (1) lack of rapid acquisition and alignment of satellite sub-images, (2) the presence of cloud, and (3) the necessity of incorporating large-scale meteorological data into the models and, therefore, the requirement that numerous sub-images, each corresponding to a separate weather regime or radiosonde station, be analyzed.

Our experience suggests that even with the problem of clouds, a portion of a designated target area, say the Great Plains, could be mapped each day using either GOES or NOAA satellite data. With the aid of continuity, an updated pattern of  $M$  could be determined every several days and represent a weekly average in much the same way that the Crop Moisture Index currently represents a highly smoothed and time-averaged parameter. (The analyses would be performed only over relatively flat terrain.)

Remote sensing of soil moisture is potentially a very valuable tool, the exact value of which remains to be determined. Highest priority in future work on this problem should be placed on facilitating image acquisition and alignment and evaluating the results of existing models. Further model improvement should await the acquisition of ground truth information since it would be difficult to evaluate any theoretical

innovation in a model without independent measurements and since various technical difficulties are more pressing and produce errors which may greatly exceed any likely model improvement. Such model improvement should concentrate on parameterizing plant canopies in a manner that allows the solutions to be closed with available satellite and large-scale meteorological data.

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