Review to estimate Evapotranspiration from remote sensing data: some examples from the simplified relationship to the use of mesoscale atmospheric models.

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Abstract

Different methods have been developed to estimate evapotranspiration from remote sensing data. Among them direct methods based upon the energy balance equation and using thermal infrared (TIR) data like the simplified relationship. This method has been applied for various situations: from small spatial scale using airborne TIR images to continental scale with NOAA data. More recently indirect estimates using assimilation procedures and Soil-Vegetation-Atmosphere Transfer (SVAT) models have been developed. In this last case, the combination of different wavelength domains is often required so as to get input parameters of these models to characterize the different surfaces like albedo, emissivity or Leaf Area Index. A brief review of these different approaches is presented. Some examples are shown on the site of the Alpilles Reseda project, where various types of models (Sebal, Meso-NH...) were used to estimate surface fluxes from remote sensing data. The main physical bases and assumptions of these models are also discussed in this paper.

Résumé

La connaissance de l'évapotranspiration est importante pour mieux gérer les besoins en eau des cultures, mais son estimation est parfois délicate à large échelle. La télédétection en fournissant des informations spatialisées sur les principales caractéristiques des surfaces permet d'accéder aux différents termes du bilan d'énergie. Ce papier rappelle les méthodes mises en oeuvre ces dernières années pour estimer l'évapotranspiration à partir de données satellitales : de la relation simplifiée utilisant des images thermiques jusqu'aux modèles de transferts Sol-Végétation-Atmosphère (TSVA) plus complexes à une ou trois dimensions ayant la possibilité d'assimiler des données de télédétection. La combinaison de différentes bandes spectrales est de plus en plus utilisée pour élaborer des cartes des principaux paramètres d'entrée de ces modèles comme l'albedo, l'émissivité, l'indice foliaire (LAI)...Une brève description des méthodes est présentée avec des exemples pris sur le projet Alpilles-Reseda où différents modèles ont été utilisés. Les principales hypothèses et bases physiques de ces approches sont discutées.

Introduction

Detailed knowledge of land surface fluxes of latent and sensible heat is important for monitoring the climate of land surface, for evaluating parameterization schemes in weather and climate models used to predict fluxes exchanges between the surface and the lower atmosphere, and for agricultural applications such as irrigation scheduling. The main methods classically used to measure evapotranspiration (*ET*) are available at the field scale (like the Bowen ratio, eddy correlation system, soil water balance), but do not allow the flux estimation over large geographical areas. For operational applications, water managers and irrigation engineers need to have accurate estimations of *ET*. Nowadays, in numerous countries, the method recommended by FAO (FAO 56 method) is used. It consists in estimating the crop evapotranspiration (Etc) for a crop canopy using a reference evapotranspiration (Etr) and a crop coefficient (Kc). The Penman – Monteith allows to compute Etr

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over a grass under optimum soil moisture conditions with a constant value of the surface canopy resistance considering then the grass as a single big leaf (Allen *et al*, 1998, FAO 56 method). However, the surface resistance can vary according to the day, the weather conditions, particularly the available radiation and the vapor pressure deficit (Ortega *et al*, 2003). The determination of crop coefficients is also debatable because a lot of factors occur. The *ET* crop surfaces under non-standard conditions is adjusted by a water stress coefficient or by modifying *Kc*. Actual evapotranspiration (*Etact*) corresponds to the real water consumption according to weather parameters, crops factors, management and environmental conditions. The crop type, variety and development stage should be considered when assessing evapotranspiration from crops. Differences in resistance to transpiration, crop height, ground cover, roots...result in different *ET* levels.

Remote sensing data with the increasing imagery resolution is a useful tool to provide such information over various scales. Different methods have been developed to use this information. It is always difficult to classify these methods, because there are often intermediate approaches which combine physical and empirical relationships. Nevertheless, we proposed in this paper three model categories which are based on :

- Empirical direct methods where remote sensing data was introduced directly in semi-empirical models to estimate *ET* (for example, the simplified relationship using thermal infrared (TIR) data). We will present the main assumptions of this model in the first section of this paper. It allows to characterize crop water use both at the local scale from ground measurements and at the scale of large irrigated areas from satellite data using the cumulative temperature difference *(Ts-Ta)*, also known as a stress degree day (*SDD*).
- Residual methods of the energy budget combining some empirical relationships and physical components. Most current operational models (such as Sebal, S-Sebi described further) use remote sensing directly to estimate input parameters and *ET*.
- Indirect methods generally use more complex models simulating the different terms of the energy budget (ISBA, Meso-NH). Remote sensing data can occur at different levels, in the input parameters to characterize the different surfaces, and/or using assimilation procedure to get more adequate parameters to compute *ET*. Some examples of this approach will be shown in the third section

Before presenting these approaches, it is necessary to make a brief review about the energy budget, in order to understand better the relationship between *ET* and surface temperature (*Ts*). Then we will describe some models using remote sensing to estimate *ET* (let us mention that it is not an exhaustive review, we have chosen to illustrate some models widely used. For more details see other references about overviews on the use of remote sensing for evapotranspiration monitoring (Kustas and Norman, 1996), Agricultural and water management, 2003, 58,and look at the site :http://www.cgiar.org/iwmi); At least, in conclusion, we will discuss about the application of these models for crop monitoring and water management, present potentialities and limits, and on future remote sensing tools.

Evapotranspiration and energy budget

Evapotranspiration estimation (corresponding to the latent heat flux *LE*) from remote sensing is based on the assessment of energy balance through surface temperature. For instantaneous conditions, the energy balance equation can be written as :

$$Rn = LE + H + G \qquad (1)$$

The available net radiant energy *Rn* is shared between the soil heat flux *G* and the atmospheric convective fluxes (sensible heat flux *H* and latent energy exchanges *LE*).

Given the aerodynamical resistance r_a between the surface and the reference height z_a in the lower atmosphere (generally 2m) above the surface, *H* is expressed as :

$$H = \rho\, cp\, (Ts - Ta)/ra \qquad (2)$$

r_a is a function of wind speed u_a, atmospheric stability and roughness (z_o, $z0t$, depending on vegetation height and geometry).

Rn depends on solar radiation (*Rg*), incident atmospheric radiation (*Ra*), surface albedo (α_s), surface emissivity (ε_s) and surface temperature (*Ts*) :

$$Rn = (1-\alpha_s)Rg + \varepsilon_s\ Ra - \varepsilon_s\ \sigma\ T_s^4 \qquad (3)$$

This means that *LE* is linearly related to the surface air temperature difference at the time of *Ts* measurement, if the second order dependence of r_a on this gradient is ignored.

$$LE=R_n - G - \rho\, cp\ (Ts - Ta)/ra \qquad (4)$$

This equation is widely used for the estimation of instantaneous *LE* (residual method). At midday it is a good indicator of plant water status for irrigation scheduling. For estimation of *LE* over longer periods (seasonal, monthly, daily estimations), the use of ground-based *ET* from weather data is necessary to make temporal interpolation. Several papers have used the tendency for the evaporative fraction (*EF*, the ratio of latent heat flux to available energy) to be nearly constant during the daytime, that allows to estimate daytime evaporation from only one or two estimates of *EF* during the middle of the day (at the satellite acquisition time) (Crago, 2000).

$$EF=LE/(Rn-G)$$

$$ET_{24}=EF*Rn_{24}$$

Another way to estimate *ET* is to compute this term according to the following equation from air vapor pressure e_a and a water vapor exchange coefficient (h_s) : This last method is generally used in models simulating Soil-Vegetation-Atmosphere Transfers (SVAT) and defined in this paper as indirect approaches.

$$LE = \rho\, c_p\ h_s\ (e_s^*(T_s) - e_a) \qquad (5)$$

$e_s^*(T_s)$ is the saturated vapor pressure at the surface temperature T_s. h_s depends on the aerodynamic exchange coefficient ($1/r_a$), soil surface and stomatal resistances of the different leaves in the canopy. For its calculation, information on plant structure is required : leaf area index (*LAI*) and fraction of vegetation cover (*veg*), the minimum stomatal resistance (rs_{smin}). Different parameterisations for the stomatal resistance can be found in the literature linked to climatic and soil moisture characteristics.

The determination of the aerodynamical resistance can be also very variable according to the models taking into account or not the ratio z0t/z0 (often expressed as kB^{-1}=log(z0/z0t). Differences between thin or medium surfaces (grass, soybean, wheat) and tall surfaces appear in this coefficient estimation. Thus the corresponding “aerodynamical surface temperature” defined by extrapolation of air temperature profile down to the level z0t may differ from the radiative surface temperature measured with satellites. Different models generally with 2 layers (described further) have integrated this difference to estimate ET by taking into account kB^{-1}.

Thus, from this basis elements, it appears that the surface temperature (*Ts*) or more exactly (*Ts-Ta*) is related to *ET*, and that *Ts* can be estimated using thermal infrared measurements (either at local scale using ground radiothermometer, either at regional or global scale using satellite data).

In the next paragraph, we will present the main steps and assumptions of these methods using remote sensing data to estimate *LE*.

1. Direct simplified methods

The simplified relationship, firstly derived at field scale by Jackson *et al* (1977) and later analyzed by Seguin and Itier (1983), has widely been used for mapping daily evapotranspiration over large areas from surface temperature measurements (Lagouarde and Brunet,1991, Courault *et al*, 1994). This method assumes that it is possible to directly relate daily (ET_d) to the instantaneous $(Ts - Ta)_i$ measurements as follow :

$$ET_d=Rn+A-B(Ts-Ta)_i \qquad (6)$$

A and B being constant depending on the local situation. Many papers have dealt with the analysis of this relationship and their assumptions (Lagouarde, 1991, Seguin and Itier, 1983, Riou *et al*, 1988). The main hypothesis considers that the ratio H/Rn is constant all along the day, and $G_d=0$. *Ts* can be extracted from measurements acquired in the thermal infrared range with airborne or satellite sensors, if they are corrected of the atmospheric effects. Seguin *et al* (1982) and Steinmetz *et al* (1989) have shown that the accuracy could reach 10-15% at a local scale, but also that *A* and *B* coefficients varied according to the experiment (figure1). Other studies have introduced different parameterizations for these coefficients as function of windspeed, roughness, criterions of atmospheric stability (Vidal and Perrier, 1989, Lagouarde and McAneney, 1992).

The cumulative value of *(Ts-Ta)* named stress degree day (*SDD*) appeared as a significant tool for assessing the global water use of a given crop.

The application of this relationship requires two variables: the maximum air temperature and the daily net radiation. If the last one (R_n) can be obtained by remote sensing (for example incident solar and atmospheric radiations can be computed from the visible and thermal channels of Meteosat, see EARS[2] and EUMETSAT[3]), the problem of the spatial representativity of the air temperature is more arguable and particularly acute for regional studies. Geostatistical models can be used to interpolate local measurements (Courault *et al*, 1994). Accuracy is then around 20 to 30%.

Carlson and Buffum (1989) have proposed to take air temperature at 50m above the surface making the assumption that at this level, atmospheric conditions are more homogeneous. They considered the difference : $(Ts\text{-}Ta)^n$ and expressed *n* and *B* coefficients as function of *NDVI*.

Other authors have used the relationship between *Ts* and a temperature of a well irrigated area (Nieuwenhuis *et al*, 1985, Thunissen et Nieuwenhuis,1990).

Carlson *et al*, (1995), Moran *et al*, (1994) have explored the relationship between *Ts* and *NDVI*, because the amount of vegetative cover affects transpiration. Vegetation indices (like *NDVI*) are also related to surface temperature, *i.e.* more evapotranspiration tends to be associated with lower temperatures. A trapezoid scheme appears in which the different soil moisture conditions can be classified (figure 2). Carlson *et al* (1990) have proposed a method of estimating root-zone moisture availability, soil surface moisture and vegetation fraction using *NDVI* and directional *Ts* combined with a transfer model. Water stress indices have been computed from this scheme and applied at large spatial scale for crop monitoring and water management.

2. Other residual methods of the energy budget

Sebal

Sebal is a model with an intermediate approach using both empirical relationships and physical parameterizations (Bastiaanssen *et al*, 1998_{ab}). This model has been designed to calculate the energy partitioning at the regional scale with minimum ground data. Atmospheric variables (air temperature and windspeed) are estimated from remote sensing data by considering the spatial variability induced by hydrological and energetic contrasts (figures 3-4). The determination of wet and dry surfaces on the studied area is necessary to extract threshold values. The model requires incoming radiation, *Ts*, *NDVI* and albedo maps. Semi-empirical relationships are used to estimate emissivity, roughness length and *G* from *NDVI*. The sensible heat flux is computed from flux inversion at dry non evaporating land units and at wet surfaces types. Latent heat flux is computed as the residual of energy balance.

This model has been used for different applications to estimate monthly and seasonal ET by linearly interpolations the ET values for periods in between two adjacent images (Bastiaanssen, 2000) and applied under several irrigation conditions in different countries (Droogers and Bastiaanssen, 2002).

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[2] EARS: www.ears.nl/EWBMS

[3] EUMETSAT: www.eumetsat.de/fr

SEBI,-S-SEBI, SEBS

Also based on the contrast between wet and dry areas, Menenti and Choudhury (1993) proposed a method to derive the evapotranspiration from the evaporative fraction. The concept was included by Su (2002a) in a more complex framework called SEBS that allows the determination of the evaporative fraction by computing the energy balance in limiting cases. A simplified method derived from SEBI (S-SEBI) has been developed to estimate of surface flux from remote sensing data (Roerink et al, 2000). It determines a reflectance dependant maximum temperature for dry condiions and reflectance dependant minimum temperature for wet conditions, the major advantages being that no additional meteorological data is needed if the surface extremes are present on the images studied.

Other models

Other approaches have been presented in the literature, such as the excess resistance (or kB^{-1}) (Su, 2002), the two sources (Norman *et al*, 1995, Chebhouni *et al*, 2001) and the β approaches (Chebhouni *et al*, 1997). Some of them have given satisfactory results even on sparse vegetation (Zhan *et al*, 1996, Chehbouni et al, 1997, French *et al*, 2000). All these models presented in table 1 can be used for operational applications for water management. The main problems for routine monitoring of surface energy fluxes is to get satellite observations with high spatial and temporal resolutions.

Table 1. Some semi-empirical models for *LE* and *H* fluxes. Symbols : A_1,A_2,A_3,B : empirical coefficients, *cp*: specific heat of air, i: instantaneous, d: daily, ra : aerodynamical resistance (above canopy), rc: aerodynamical resistance at the soil surface, rex: excess resistance, ta : air temperature at some height above canopy (generally 2m), Taer: aerodynamical temperature (mean temperature at some height in the canopy), Tv : vegetation surface temperature, Tg: soil surface temperature, Ts, radiometric surface temperature (from Olioso *et al*, 1999)

Simplified relationship (Seguin et Itier, 1983)	$LE_d=Rn_d-A_1-B_1(Ts_{14h}-Tamax)$
methods based on excess resistance (Kustas, 1990) (Lhomme *et al*, 1992) (Moran *et al*, 1994)	$H_i=\rho cp(Ts-Ta)/(ra+rex)$ $LE_i=(1-A_2)Rn_i-H_i$
Approaches based on a relation between radiometric and a so-called aerodynamic temperature (Troufleau *et al*, 1997) (Chehbouni *et al*, 1997)	$H_i=\rho cp(Taer-Ta)/ra$ $(Taer-Ta)=(1-A_3)(Ts-Ta)$
Two source approach (Norman *et al*, 1993)	$H_i=\rho cp((Tv-Ta)/ra+(Tg-Ta)/(ra+rc))$

Problems linked to the surface temperature obtained from remote sensing

Most methods use TIR data. Atmospheric corrections and surface emissivity affect the retrieval of surface temperature and thus influence the quality of the information extracted from remote measurements. Two categories of corrections may be applied: direct methods using atmospheric sounding combined with radiative transfer model, indirect methods using only satellite observation (Tovs or split window method). Dual angle observation (ATSR) improve the estimation. Typical uncertainties in atmospheric correction are about 1-3°.

The effect of emissivity is important and can lead to significant error. The most promising method for obtaining both surface directional infrared temperature and surface directional emissivity is based on high spectral resolution (Norman *et al*, 1995). The table 2 shows the importance of error of *(Ts-Ta)* on the sensible heat flux *H*.

Table 2 Error in sensible heat flux arising from a 1°C error in *(Ts-Ta)* for several conditions (*in* Norman *et al*, 1995).

Canopy height	Wind speed	Error H
1	1	8
1	5	40
10	1	17
10	5	87

Note that models like Sebal use an automatic internal calibration of ΔTs vs Ts function so that atmospheric correction of Ts is not necessary. Any bias introduced from the lack of correction is cancelled with the working of model. It is a positive point for operational applications because it decreases the processing time.

Spatial and temporal resolution of TIR data

Frequent data acquisitions are needed for proper crop monitoring during the growing season, but only meteorological satellites offer the necessary frequency of measurements, and the spatial resolution remains still too coarse to define each type of crop. On the other hand, data in the visible and near-infrared wavelengths, used for computing vegetation indices, are available at resolutions an order of magnitude smaller than TIR, and hence provide higher resolution information on vegetation cover. Recently Kustas *et al* (2003) have explored the relationship between these two spatial and spectral resolution (*NDVI* and *Ts*) and proposed a disaggregation procedure for estimating the subpixel variation in Ts. They used then a remote sensing based energy balance model (DisALEXI) for estimating the surface fluxes. This disaggregation technique appears as a promising way for evaluating *Ts* at the field scale.

Meteorological variables - models integrating the atmospheric boundary layer

In order to avoid the difficulties of obtaining meteorological variables on large areas, some models integrate the planetary boundary layer (PBL) to simulate the evolution of parameters like air temperature, windspeed…Radiosoundings or outputs from GCM are then necessary to initialise the atmosphere. 1D (Lagouarde and Brunet, 1991), 2D (Hasager *et al*, 2002) or 3D (Anderson *et al*, 1997, Norman *et al*, 1995, Courault *et al*, 2002) approaches estimating surface fluxes have used remote sensing data at different levels. The inclusion of energy balance in a PBL model has also been exploited for deriving the fluxes on the basis of the rate of change of surface temperature during the morning hours (Meciakalski *et al*, 1999).

The microscale aggregation model (2D) described by Hasager et Jensen (1999) uses surface temperature images. A roughness map is obtained from landuse map. A set of equations per land cover type defines the relation between thermal roughness and LAI (Hasager *et al*, 2002). The model solves the linearized atmospheric flow equations by Fast Fourier Transforms (FFT). The maps of friction velocity, u*, and temperature scalar T*, are calculated through iteration including the Monin-Obukhov stability functions. From the u* and T* maps, the effective values of *z0m* and *z0t* are calculated, and then the surface fluxes.

Although these methods have operational applications like drought detection at continental scale, or water reserve estimation for irrigation, the accuracy is always difficult to estimate. It is a reason why these last years, indirect methods based on assimilation procedures have been more developed, because they allow, among other things, to get intermediate variables linked to the crop development (like LAI) or to the soil water status.

3. Indirect methods

These methods can be also defined as “determinist” approaches because the models (generally SVAT models) describe the exchanges between soil plant and atmosphere according to the physical processes occurring in each compartment with generally a fine time step (second, hour). Different complexity levels appear according to the process description: for example, if the vegetation and soil behavior are separated, then evaporation and transpiration are computed with a surface temperature for each part (it is more realistic for comparison with TIR data acquired at different hours and angles). Different schemes can be found to represent the vegetation: one big leaf with one surface resistance to multi-layer models, where radiative and energy budgets are computed for each layer (see Olioso *et al*, 2002, Olioso *et al*, 1999 for more details on these approaches). The finer the surface and the process description is, the more parameters are needed. Some of them can be estimated by remote sensing data. There are 3 ways to use this spectral and spatial-temporal information.
- to force the model input directly with the remote sensing measurements
- to correct the course of state variables in the model at each time remote sensing data are available (sequential assimilation)
- to assimilate remote sensing data which consists in initializing again or changing some parameters, not only for one remote sensing measurement but on a data set acquired
for several days (variational assimilation) (figures 5 and 6).

Many works have been done on these assimilation procedures and have shown that the most adequate variables which can be estimated using remote sensing are the surface and stomatal resistances, and soil moisture (Olioso *et al*, 1999, figure 6). Numerous studies have used radiative surface temperature (Soer, 1980, Ottlé et Vidal Madjar, 1994), or microwaves (Wigneron *et al*, 2002). Thus, for example, *Ts* derived from NOAA data has been used to find parameters linked to the irrigation with the SVAT called MAGRET applied over the agricultural region of “la Crau” in the South-East of France (Courault *et al*, 1998). These parameters were the beginning and the end of irrigation, frequency and quantity brought. MAGRET is a simplified SVAT simulating hourly values of *Ts* and the main surface fluxes. This application was based on a global calibration dealing with the comparison between Ts simulated by the model and Ts estimated from NOAA data for 10 days along the cultural cycle.
The main assumptions are to consider the surface as homogeneous with uniform variables, since remote sensing data acquired at large spatial scale result often from the combination of different elements. “Effective” parameters are then defined corresponding to these composite surfaces (Noilhan and Lacarrère, 1995). Other approaches search to disaggregate the pixel content into elementary responses for each landuse class (Courault *et al*, 1998).

The main parameters extracted from remote sensing measurements are vegetation fraction, LAI, albedo, emissivity, (most of them are estimated using information in the solar domain, table 3). Roughness and parameters linked to the stomatal resistance are still difficult to access and often estimated from a knowledge of the type of canopy and its phenological stage.

Table 3. Main biophysical variables derived from remote sensing data classified according to wavelength ranges and models, Fonct: crop model simulating the vegetation development (from Baret, INRA Avignon, personal communication)

Biophysical variables	solar	IRT	Active µwaves	Passive µwaves	Process models
Albedo	++				SVAT
Vegetation cover	++	+			SVAT
FAPAR	++				Fonct
LAI	++	+	+	+	SVAT&Fonct
Water content in veg			++	++	Fonct
Temperature		++		+	SVAT&Fonct
Chlorophyl	++				Fonct
Leaf water content	++				Fonct
Soil water content	++		++	++	SVAT&Fonct
Soil roughness			++	++	SVAT
Vegetation height (roughness)	++	+	+		SVAT&Fonct

MESO-NH[4] is a 3D atmospheric model mainly developed by the Aerology Laboratory and the CNRM[5] from Toulouse. The surface scheme based on the force restore method is ISBA (Noilhan and Planton,1989) which has been widely used in 1D version coupling assimilation methods with remote sensing data (Calvet *et al*, 1998, Olioso *et al*, 2002a). The assimilation procedures are not yet introduced in the 3D atmospheric model, but all input data may be derived from remote sensing. An example of an evapotranspiration map is shown in figure 7 where LAI, vegetation fraction (figure 8) were computed from POLDER images using neural network (Weiss et Baret, 2002). Albedo in the visible and near infrared range were estimated using Liang's coefficients (Jacob *et al*, 2002), roughness and other parameters linked to the stomatal resistance were derived from the landuse map obtained from SPOT images. Surface temperature and fluxes were then estimated by the model and compared with TIR images acquired during the experiment (figure 9). The results were globally satisfactory, even if the main difficulty still remains the determination of the initial soil moisture variability on the whole area(Courault *et al*, 2002).

Comparison between models

During the Alpilles – Reseda[6] project, several models have been used to estimate the surface fluxes with remote sensing data. Figures 7 and 10 show two *LE* maps obtained respectively with MESO-NH and a simplified energy balance model for the same date in April 97 (Olioso *et al*, 2002b). Although the spatial resolution was not exactly similar (20m for fig 10, 50m for MESO-NH), the same pattern appeared on the 2 maps. The fluxes showed a great spatial variability according to the development stage of the different crops as expected: high values for well developed crops (winter wheat in April, alfalfa well supplied in water), and low values for dry soils (the last rain was in January). *(Ts-Ta)* varied from 0 to 15°c for this date. The difference observed between the two maps were due to the variability of input parameters of the two models which were not the same.

Olioso *et al* (2002b) have compared 3 models on the same dataset of ALPILLES: a direct flux equation using Ts, Ta and the exchange coefficient computed using the Monin-Obukhov theory, the SEBAL model which computed Ta and windspeed, and the 2D aggregation model (Hasager) where Ta and windspeed were taken from radiosounding measurements. The results showed that the differences on flux estimation were mainly due to the way of obtaining the surface parameters and meteorological

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[4] MESO-NH Non –Hydrostatic Mesoscale atmospheric model, http://www.aero.obs-mip.fr/~Meso-NH

[5] CNRM : Centre National de Recherches Météorologiques

[6] Alpilles Reseda : was an CEE project http://www.avignon.inra.fr/reseda

variables, particularly air temperature and roughness. An accurate description of the model inputs (surface parameters and meteorological variables) is therefore a first stage for the estimation of surface fluxes, which are crucial to get realistic *LE* values. The other conclusions on the main results about this experiment can be found in the special issue of *Agronomie* (2002,vol22).

Discussion-Conclusion

An accurate estimation of evapotranspiration is very useful for an appropriate water management both at the farm and the irrigation project level. In numerous countries, the method recommended by FAO is used. However the spatial and temporal variations of the surface characteristics can not be taken into account with accuracy by this method. The use of remote sensing brings a significant contribution for assessment of crop water status either in view irrigation scheduling or in global assessment of crop water use and its spatial variations within an irrigated area (Vidal *et al*, 1987).

Evapotranspiration may be estimated from remote sensing data with different approaches: direct methods using TIR data, indirect estimates using assimilation procedure combining different wavelengths to get various input parameters (in particular related to vegetation water status). Some methods are based on the spatial variability present in remote sensed images (like the Sebal or S-Sebi models) and try to use no additional meteorological data to estimate *ET* for routine applications.

The interest of using SVAT models is not only because they generally describe with more accuracy the crop functioning, but also because they allow to access to intermediate variables like soil moisture or LAI, which are related to physiologic and hydrologic processes which can be linked to other meteorological or hydrological models.

However, the use of remote sensing for operational applications presents still several problems :

The determination of ET for crop monitoring requires the routine processing of images on a near-real-time basis. The relatively long turn-around time for image delivery and the cost involved with the acquisition of high-resolution imagery make their use for operational application often unattractive.

➚*Data accuracy*

Most methods use TIR data. Atmospheric corrections and surface emissivity are necessary to get accurate Ts. Some models like Sebal with their internal calibration avoid this problem and are then more attractive for operational applications. Thus Sebal has been applied on a near-real-time basis to estimate actual evaporation in Sri Lanka on 10-day basis from June 1999 to 2000 using NOAA AVHRR radiances (Bastiaanssen, 2003).

➚ *Spatial and temporal resolution*

The thermal infrared measurements appear as useful tools for water use in irrigated area. For a global monitoring purpose, the availability of advanced very high resolution radiometer (AVHRR) imagery from NOAA series meteorological satellites on daily basis at most of the national meteorological services worldwide at no extra cost, makes them a viable alternative for operational estimation of evaporation. But more detailed observations would be needed for analysing the spatial distribution of water use in the irrigation network. The NOAA resolution (1km) is too coarse for that purpose. A higher resolution can be achieved by Landsat (90m in TIR), but both the frequency (every 16 days) and time acquisition (for example 10:00 over France) are limiting factors. Moreover the future of Landsat is uncertain, because of the cooling techniques are too heavy and it makes the payload too expensive. There is currently no operational solutions for this problem. So, we have to find methods combining informations at different wavelengths and resolutions.

The arrival of new satellites like ASTER (15m in 3 visible near infrared bands and 90m in 5 TIR band from 8.1 to 11.6µm) allows to combine high spatial resolution with other sensors with high temporal resolution (like MODIS or GOES). The method proposed by Kustas *et al* (2003) to disaggregate the pixel to estimate subpixel *Ts* is promising and allows to estimate ET combining Ts with an energy balance model (DISALEXI).

➚*meteorological forcing*

It is important to take into account the spatial variability of climatic data, particularly air temperature, which is a key variable in the exchanges. Meteorological variables may be directly measured but often the station density is poor. They can however be estimated by models simulating the evolution of the planetary boundary layer (MESONH). Some models like SEBAL use spatial information in images to derive air temperature, but their estimations depend on the spatial variability of the studied area. These simplified methods worked correctly when the atmospheric conditions are constant over the image and sufficient wet and dry pixels are present throughout the scene. When different windspeed occur and change values of extreme Ts (min and max), or if wet and dry pixels cannot be found on the same images (eg England no dry area, Sahara not wet pixels, Europe, variable atmospheric conditions) external meteorological data (radiosoundings or weather prediction model output) are necessary (Roerink *et al*, 2003). Other models use air temperature at 50m making the assumption that atmospheric conditions are more homogeneous at this level.

There is a critical need to understand the feedback between the land surface and atmosphere at various scales. The role of land surface modifying the climate is not yet adequately considered in climate models, however its effect like irrigation is significant for temperature (De Ridder et Gallée, 1998). The current parameterizations of land processes are still too coarse and currently the trend is to describe the different surfaces with more accuracy. The derivation of accurate surface parameters from remote sensing is key for determining the main terms of the energy balance depending on the type of vegetation. It is also important for having an exhaustive view of the vegetation cover types in order to analyze in details model results and evapotranspiration estimations. For that specific purpose thermal infrared wavelengths appears as the best suited, shortwave channels allow to quantify the effect of water stress on biomass by the use of vegetation index. With the increasing spatial resolution and the sensor profusion, we can expect that remote sensing will continue to play an essential role in partitioning the surface energy budget into sensible heat and evapotranspiration, and to provide information at a low cost for improving the use of scare water. resources.

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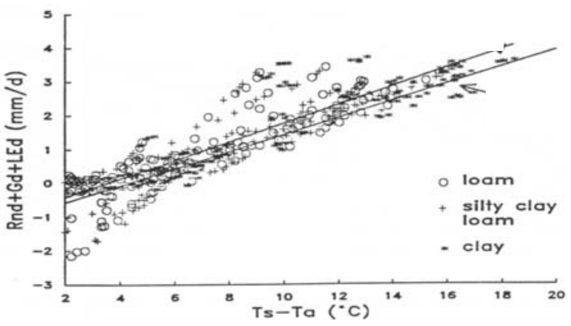

Figure 1. Simplified relationship obtained for different soil types between *(Ts-Ta)* and *H*. (from Chanzy, 1991)

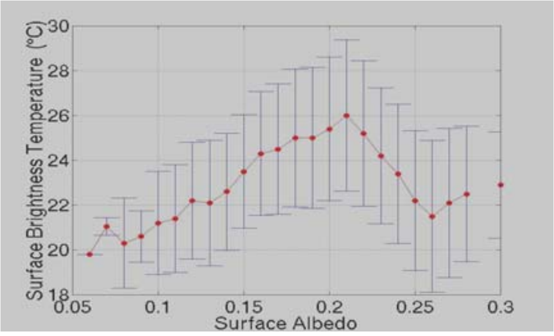

Figure 3. Relation between Albedo and brightness temperature obtained from Polder and TIR measurements over the ALPILLES site in 1997 which allows to derive windspeed (Jacob, 1999)

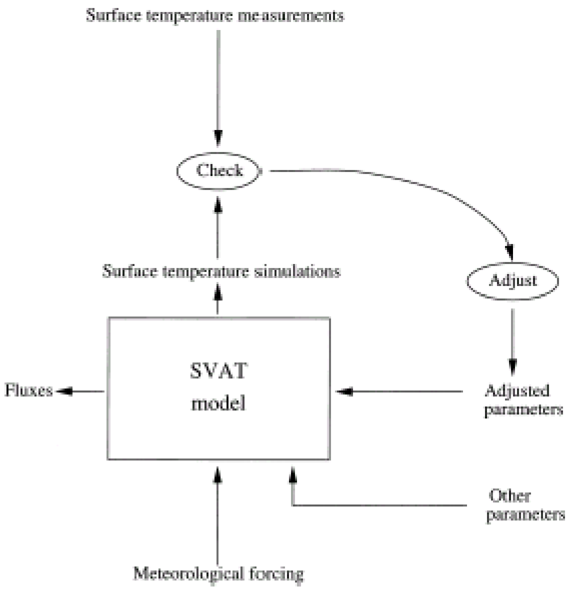

Figure 5. Schematic representation of assimilation method (*in* Olioso *et al*, 1999).

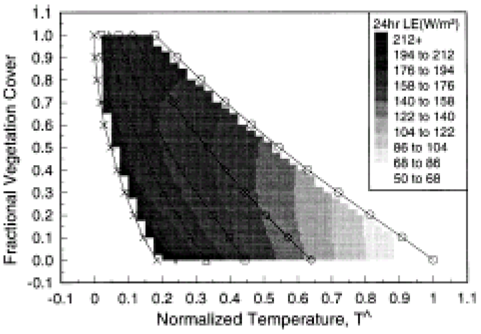

Figure 2. daily average of *LE* simulations for different fractional vegetation cover and soil moistures availabilities (from Capehart, 1996).

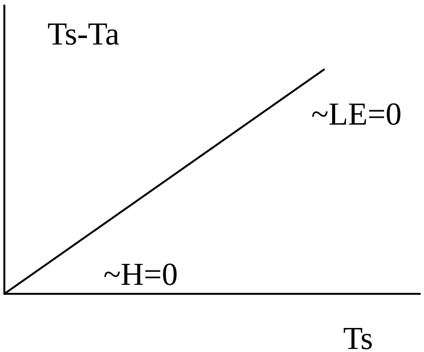

Figure 4. Spatial relation used in SEBAL between Ts and Ta to estimate air temperature.

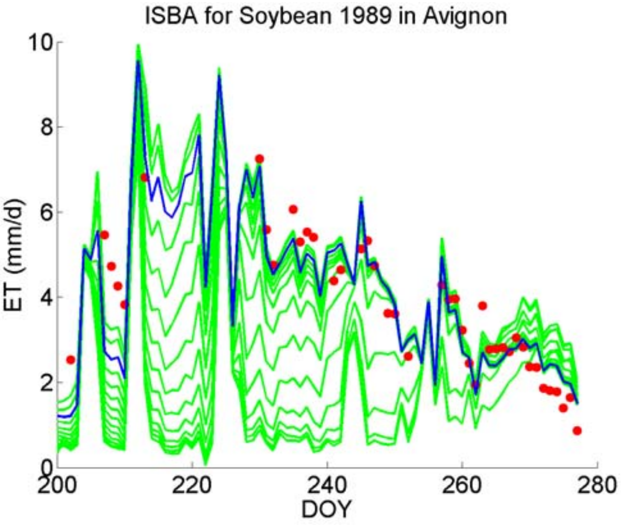

Figure 6. Example of assimilation of remote sensing data in a SVAT model (ISBA) (different simulations have been done (green lines) adjusting the initial soil moisture after comparing Ts estimated and measured (*in* Olioso *et al*, 2002a).

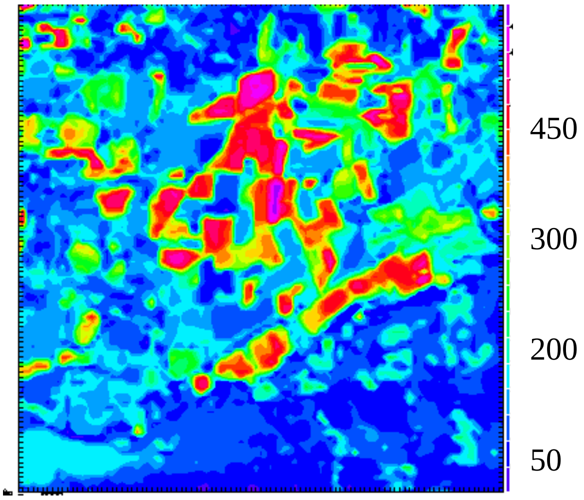

Figure 7. LE map (W/m²) obtained with the MESO-NH model on the Alpilles site for April 18 1997 at midday.

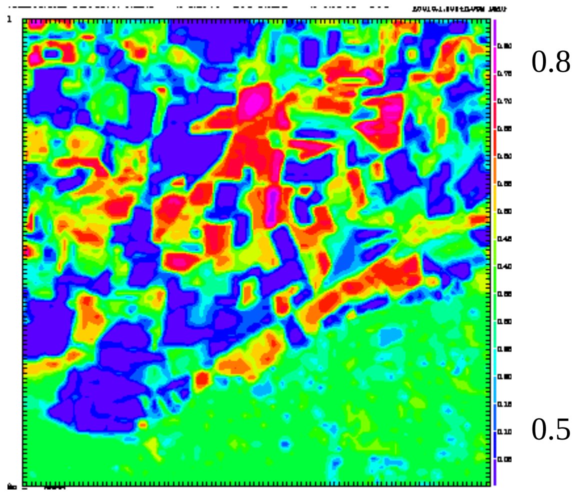

Figure 8. Vegetation fraction computed from a POLDER image acquired on 10 april 1997 over Alpiles area, and used as input data in MESO-NH model.

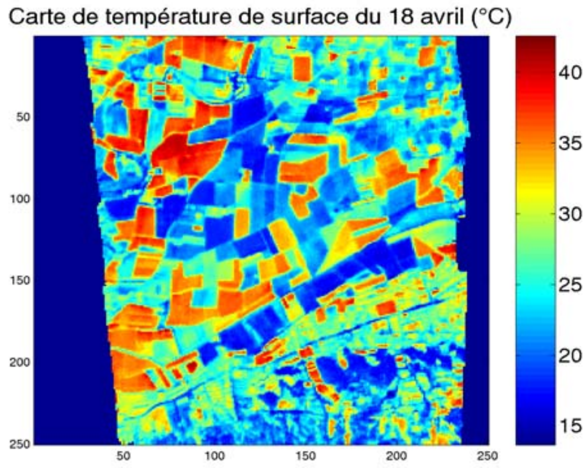

Figure 9. Surface temperature obtained from TIR airborne camera over the Alpilles area for the 18 April 1997 (20m resolution).

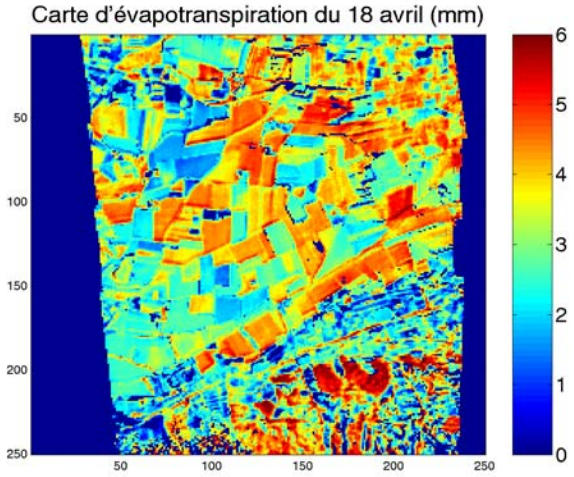

Figure 10. LE map from a simplified model based on the energy balance for April 18 1997 at midday (from Olioso *et al*, 2002b).