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Satellite Based Energy Balance Models for Estimation of Crop Evapotranspiration - A Review of Emerging Technologies

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Abstract

Satellite-based energy balance models represent a significant advancement in the estimation of crop evapotranspiration (ET) by leveraging remote sensing data. These models integrate information from various satellite sensors to calculate the energy fluxes at the Earth's surface, essential for understanding the water dynamics in agricultural landscapes. This abstract provides an overview of the satellite-based energy balance models and their application in estimating crop ET. The models include inputs from satellite sensors such as thermal infrared, visible, and near-infrared bands, which are used to derive surface temperature, vegetation indices, and albedo. These inputs are crucial for calculating the energy available for evapotranspiration processes. Secondly, the energy balance equation, incorporating these inputs along with meteorological data, facilitates the estimation of key parameters such as net radiation, soil heat flux, and sensible heat flux, which together determine ET. Moreover, the integration of satellite-derived data enables spatially explicit and temporally frequent estimation of ET over large agricultural areas, overcoming the limitations of ground-based measurements. This capability is particularly advantageous for monitoring water use efficiency, irrigation scheduling, and drought assessment at regional and global scales. Additionally, the models can accommodate different crop types and land cover conditions, providing flexibility in application across diverse agricultural landscapes. Furthermore, validation studies have demonstrated the accuracy and reliability of satellite-based ET estimates when compared with ground-based measurements, highlighting their utility for decision-making in agriculture and water resource management. Ongoing advancements in satellite technology and data processing techniques continue to enhance the capabilities of these models, promising further improvements in accuracy and spatial resolution.

Key words : Remote sensing; evapotranspiration; energy balance models, SEBI, SEBS, METRIC.

Estimation of evapotranspiration (ET) is an important for the computation of irrigation water requirements, water resources management and determination of water budget especially under dry conditions where water resources are scarce and fresh water source is limited. It is important to develop irrigation water management plans, improve irrigation efficiency, and evaluate water productivity. Estimation of water consumption based on ET models using remotely sensed data has become one of the most widely used method in water resources planning and management over watersheds due to the competition for water between trans-boundary water users (Bastiaanssen et. al. 2005). In climate dynamics, continuous progress has been made to describe the general circulation of the atmosphere and it has proved that the general circulation models appeared to be quite sensitive to the land surface ET information (Brutsaert 1982). For vegetated land surfaces, ET rates are closely related to the assimilation rates of plants and can be used as an indicator of plant water stress (Jackson 1981). Therefore, accurate estimates of regional ET in the land surface water and energy budget modeling at different temporal and spatial scales are essential in hydrology, climatology and agriculture. In various practical applications, there are still no specific ways to directly

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measure the actual ET over a watershed. Conventional ET estimation techniques (i.e., pan-measurement, Bowen ratio, eddy correlation system, and weighing lysimeter, scintillometer, sap flow) are mainly based on site (field)-measurements and many of those techniques are dependent on a variety of model complexities. Though they can provide relatively accurate estimates of ET over a homogeneous area, conventional techniques are of rather limited use because they need a variety of surface accessory measurements and land parameters such as air temperature, wind speed, vapour pressure at a reference height, surface roughness, etc., which are difficult to obtain over large-scale terrain areas and have to be extrapolated/interpolated to various temporal and spatial scales with limited accuracy in order to initialize/force those models (Idso et. al 1975). Remote sensing technology is recognized as the only viable means to map regional- and meso-scale patterns of ET on the Earth's surface in a globally consistent and economically feasible manner and surface temperature helps to establish the direct link between surface radiances and the components of surface energy balance. Remote sensing technology has several marked advantages over conventional "point" measurements: 1) it can provide large and continuous spatial coverage within a few minutes, 2) it costs less when the same spatial information is required, 3) it is particularly practical for ungauged areas where man-made measurements are difficult to conduct or unavailable (McCabe et. al. 2006). Remotely sensed surface temperature can provide a measure of surface from a resolution of a few cm2 from a hand-held thermometer to about several km2 from certain satellites (Hatfield 1983.) Combining surface parameters derived from remote sensing data with surface meteorological variables and vegetation characteristics allows the evaluation of ET on local, regional and global-scales. Remote sensing information can provide spatial distribution and temporal evolution of NDVI (Normalized Difference Vegetation Index), LAI (Leaf Area Index), surface albedo from visible and near-infrared bands and surface emissivity and radiometric surface temperature from mid and thermal infrared bands, many of which are indispensable to most of the methods and models that partition the available energy into sensible and latent fluxes components(Mauser et. al. 1998).

This paper provides an overview of various methods and models that have been developed to estimate land surface ET on a field, regional and large scales, based mainly on remotely sensed data.

2. Overviews of Remote Sensing-Based Evapotranspiration Models in the Past Decades : The ET models using remote sensing data can be broadly categorized into two types: (semi-)empirical methods and analytical methods, with (semi-)empirical methods relying on statistical correlations between remotely sensed variables and ET rates, and analytical methods integrating detailed representations of energy and water fluxes at the land surface.

2.1 Surface Energy Balance method for estimating Evapotranspiration : Surface energy balance governs the water exchange and partition of the surface turbulent fluxes into sensible and latent heat in the soil-vegetation-atmosphere continuum. The residual method of surface energy balance is one of the most widely applied approaches to mapping ET at different temporal and spatial scales. When heat storage of photosynthetic vegetation and surface residuals and horizontal adjective heat flow are not taken into account, the one-dimensional form of surface energy balance equation at instantaneous time scale can be expressed numerically as:

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

Each of the three components of the energy balance equation, including Rn, G and H, can be estimated by combining remote sensing based parameters of surface radiometric temperature and shortwave albedo from visible, near infrared and thermal infrared wavebands with a set of ground based meteorological variables of air temperature, wind speed and humidity and other auxiliary surface measurements. These calculations based on surface energy balance models will be addressed in the following sections.

2.1.1 Net Radiation Equation (Rn) : Surface net radiation (Rn) represents the total heat energy that is partitioned into G, H and LE. It can be estimated from the sum of the difference between the incoming (Rs) and the reflected outgoing shortwave solar radiation (0.15 to 5 µm), and the difference between the down welling atmospheric and the surface emitted and reflected longwave radiation (3 to 100 µm), which can be expressed as:

$$R_n = (1 - \alpha_s)\, R_s + \varepsilon_s\, \varepsilon_a\, \sigma\, T_a^4 - \varepsilon_s\, \sigma\, T_s^4 \qquad (2)$$

where s is surface shortwave albedo, usually calculated as a combination of narrow band spectral reflectance values in the bands used in the remote sensing, Rs is determined by a combined factors of solar constant, solar inclination angle, geographical location and time of year, atmospheric transmissivity, ground elevation, etc., s is surface emissivity evaluated either as a weightedaverage between bare soil and vegetation or as a function of NDVI(Bastiaanssen et al. 1998), a is atmospheric emissivity estimated as a function of vapor pressure.

The uncertainties of various methods of estimating the net shortwave and longwave radiation fluxes was reviewedby Kustas and Norman1996 and found that a variety of remote sensing methods of surface net radiation estimation had an uncertainty of 5-10% compared with ground-based observations on meteorologically temporal scales. Bisht et al. 2005 proposed a simple method to calculate the instantaneous net radiation over large heterogeneous surfaces for clear sky days using only land and atmospheric products obtained using remote sensing data from MODIS-Terra satellite over Southern Great Plain (SGP). Similarly, Allen et al. 2007proposed an internalized calibration model for calculating ET as a residual of the surface energy balance from remotely sensed data when surface slope and aspect information derived from a digital elevation model are taken into account.

2.1.2 Soil Heat Flux (G) : Soil heat flux (G) is the heat energy used for warming or cooling substrate soil volume. It is traditionally measured with sensors buried beneath the surface soil and is directly proportional to the thermal conductivity and the temperature gradient with depth of the topsoil. The one used in SEBAL (Surface Energy Balance Algorithm for Land) to estimate the regional-scale G is expressed as follows:

$$G = 0.30(1-0.98\ NDVI^4)\ R_n \qquad (3)$$

As G varies considerably from dry bare soil to highly well-watered vegetated areas, it is inappropriate to extrapolate ground-based measurements to values of areal areas. Under current circumstances, it is still impossible to directly measure G from remote sensing satellite platforms. Fortunately, the magnitude of G is relatively small compared to Rn at the daytime overpass time of satellites. Estimation error of G will thus have a small effect on the calculated latent heat flux. Many papers have found that the ratio of G to Rn ranges from 0.05 for full vegetation cover or wet bare soil to 0.5 for dry bare soil and this ratio is simply related in an exponential form to LAI, NDVI, Ts and solar zenith angle based on field observations. The value of G has been shown to be variable in both diurnal and yearly cycle over diverse surface conditions. However, the assumption that daily value of G is equal to 0 and can be negligible in the daily energy balance is generally regarded as a good approximation. Comparisons of G between results from these simplified techniques and observations at micrometeorological scales showed an uncertainty of 20-30%.

2.1.3 Sensible Heat Flux (H) : The sensible heat flux (H) is the heat transfer between ground and atmosphere and is the driving force to warm/cool the air above the surface. In the single-source energy balance model, it can be calculated by combining the difference of aerodynamic and air temperatures with the aerodynamic resistance (ra) from:

$$H = \rho c_p \, (T_{aero} - T_a) / r_a \tag{4}$$

Aerodynamic resistance ra is affected by the combined factors of surface roughness (vegetationheight, vegetation structure), wind speed and atmospheric stability, etc. Therefore, aerodynamicresistance to heat transfer must be adjusted according to different surface characteristics except whenthe water is freely available. Hatfield et al. 1983 have shown that ra decreased as the wind speedincreased, regardless of whether the surface was warmer or cooler than air, and ra decreased if thesurface become rougher. Various methods for calculating ra have been developed ranging fromextremely elementary (a function of wind speed only) to quite rigorous ones (accounting foratmospheric stability, wind speed, surface "aerodynamic" roughness, etc.), with thecommonly applied one being:

$$r_a = \frac{\ln\left[\dfrac{Z_a - d}{Z_{om}} - \psi_1\right] \ln\left[\dfrac{Z_a - d}{Z_{oh}} - \psi_2\right]}{k^2 u} \tag{5}$$

Jackson et al. 1983 found that Ts-Ta varied from -10 to + 5°C under medium to low atmosphericmidity, which shows that neutral stability cannot prevail under a wide range of vegetation cover andsoil moisture conditions. Under stable and unstable atmospheric stability conditions, theMonin-Obukhov length was introduced to measure the stability and it needs to be solved with Hiteratively:

$$A = \frac{u^3 \rho c_p T_a}{k_g H} \tag{6}$$

where if A < 0, unstable stability, A >0, stable stability.

For unstable conditions (usually prevailing at daytime) with no predominant free convection, ψ_1 and ψ_2 can be expressed as:

$$\psi_1 = 2\ln\left(\frac{1+x}{2}\right) + \ln\left(\frac{1+x^2}{2}\right) - 2\arctan(x) + \frac{\pi}{2} \tag{7}$$

$$\psi_2 = 2\ln\left(\frac{1+x^2}{2}\right) \tag{8}$$

$$\text{With } x = \left(1 - 16\frac{Z_a - d}{A}\right)^{0.25} \tag{9}$$

For stable conditions (usually prevailing at night-time), the formula proposed by Webb andBusinger et al. 1971 was adopted to account for the effects of atmospheric stability on ra :

$$\psi_2 = \psi_2 = -5\left(\frac{Z_a + d}{A}\right) \tag{10}$$

Hatfield et al. 1983 have shown that ET rates could be over-estimated when the canopy-airtemperature difference is greater than about ±2°C, if the aerodynamic resistance is not corrected foratmospheric stability.

The surface roughness plays a significant role in the determination of sensible heat flux and itchanges apparently with leaf size and the flexibility of petioles and plant stems. The effectiveroughness for momentum zom is considered to be some unspecified distance above a zero planedisplacement height where the wind speed is assumed to be zero when log-profile wind speed isextrapolated downward, rather than at true ground surface. Some papers have specified zom isequal to zoh and can be either a function of vegetation height (Soer, G.J.R. 1980), in which zom is typically 5 to 15percent of vegetation height depending on vegetation characteristics, or estimated from windprofiles, using an extrapolation of the standard log-linear wind relationship to zero wind speed.(Brutsaert 1982)showed that the heat transfer was mainly driven by molecular diffusion while themomentum transfer near the surface was controlled by both viscous shear and pressure forces. Becauseof the differences between heat and momentum transfer mechanisms, there is a distinction between zomand zoh, which has caused an additional resistance (often expressed as aerodynamic definition of kB-1:kB-1 = $\ln(z_{om}/z_{oh})$ to heat transfer or an excess (extra) resistance. Stewart et al. 1994 haverelated the excess resistance to the dimensionless bulk parameter kB-1 using the following expression:

$$r_e = (kB)^{-1} / (ku)^* \quad (11)$$

Verhoef et al. 1997 showed that kB-1 was sensitive to measurement errors both in themicrometeorological variables and in the roughness length for momentum and its value over bare soilcould be less than zero. Massman 1999 used a physically based “localized near-field” Lagrangiantheory to evaluate the effects of kB^{-1} on the vegetative components in the two-source energy balancemodels and on the combined effects of soil and vegetation in a single-source model. Su et al. 2001proposed a quadratic weighting based on the fractional coverage of soil and vegetation to calculate thekB-1 in order to take into account any situation from full vegetation to bare soil conditions. Whatshould be noted is that the determination of the surface roughness still remains a challenging issue forlarge scale retrieval of the turbulent heat fluxes in spite of the efforts made in the past.

Klaassen and van den Berg 1985 showed that the measurement (or reference) height should be set at 50 m instead of 2 m at the bottom of the mixed layer and calculation of ET of crops over roughsurfaces could be improved with increasing reference height.

T_{aero}, the temperature at level of d+zoh, which is the average temperature of all the canopy elementsweighted by the relative contribution of each element to the overall aerodynamic conductance,may be estimated from extrapolation of temperature profile down to $z = d + z_{oh}$ and is recognized as thetemperature of the apparent sources or sinks of sensible heat. T_{aero} is found to behigher (lower) than T_s under stable (unstable) atmospheric conditions and they are nearly the sameunder neutral conditions. Kustas and Norman 1996 concluded that the differences between Taeroand Ts could range from 2°C over uniform vegetation cover to 10°C for partially vegetated areas.

The bulk transfer equation (resistance-based model) has beenpredominately applied since 1970s over a local/regional scale with various vegetation covers. The average difference of H estimated by different authors based on the bulk transfer equation is about 15-20%, which is around the magnitude of uncertainty in eddy correlation and Bowen ratiotechniques for determining the surface fluxes in heterogeneous terrain.

In the single-source surface energy balance models, the main distinction of various methods is howto estimate the sensible heat flux. Some of them are based on the spatial context information(emergence of representative dry and wet pixels) of land surface characteristics in the area of interest.

2.2 SEBI (Surface Energy Balance Index) and SEBS (Surface Energy Balance System) : SEBI, first proposed by Menenti and Choudhury 1993, along with its derivatives like

SEBAL, S-SEBI (Simplified-SEBI), SEBS, METRIC (Mapping Evapotranspiration at high Resolution with Internalized Calibration) etc., are typically single-source energy balance models based on the contrast between dry and wet limits to derive pixel by pixel ET and EF from the relative evaporative fraction when combined with surface parameters derived from remote sensing data and a certain amount of ground-based variables measured at local and/or regional scale. The dry (wet) limit, no matter how it was specifically defined, often has the following characteristics: 1) generally maximum (minimum) surface temperature, 2) usually low or no (high or maximum) ET. In the SEBI method, the dry limit is assumed to have a zero surface ET (latent heat flux) for a given set of boundary layer characteristics (potential temperature, wind speed and humidity, etc.). So the sensible heat flux is then equal to the surface available energy, with the Ts,max inverted from the bulk transfer equation being expressed as:

$$T_{s,max} = T_{pbl} + r_{a,max} \frac{H}{\rho c_p} \tag{12}$$

Correspondingly, the minimum surface temperature can be evaluated from the wet limit, where the surface is regarded as to evaporate potentially and the potential ET is calculated from Penman-Monteith equation with a zero internal-resistance. The Ts,min is expressed as:

$$T_{s,min} = T_{pbl} + \frac{r_{a,min} \frac{R_n - G}{\rho c_p} - {}^{VPD}\!/_{\gamma}}{1 + {}^{\Delta}\!/_{\gamma}} \tag{13}$$

The relative evaporation fraction can then be calculated by interpolating the observed surface temperature within the maximum and minimum surface temperature in the following form:

$$\frac{LE}{LE_p} = 1 - \frac{r_a^{1}(T_s - T_{pbl}) - r_{a,min}^{-1}(T_{s,min} - T_{pbl})}{r_{a,min}^{-1}(T_{s,max} - T_{bpl}) - r_{a,min}^{-1}(T_{s,min} - T_{pbl})} \tag{14}$$

where the second part of the right hand side of Equation (14) is the so-called SEBI, which varies between 0 (actual = potential ET) and 1 (no ET).

Parameterization of the SEBI approach was first proposed by defining theoretical pixel-wise ranges for LE and T_s to account for spatial variability of actual evaporation due to albedo and aerodynamic roughness. This parameterization was essentially a modification from CWSI (Crop Water Stress Index) proposed by Idso et al. 1981 and Jackson et al. 1981. The theoretical CWSI accounted for the effects of the net radiation and wind speed in addition to the temperature and vapor pressure required by the empirical CWSI.

Taking into account the dependence of external resistance on the atmospheric stratification, Menenti and Choudhury 1993 proposed an approach to calculate the pixel-wise maximum and minimum surface temperature and redefined CWSI as a pixel-wise SEBI at given surface reflectance and roughness to derive the regional ET from the relative evaporative fraction. The CWSI was based on surface meteorological scaling while the SEBI used planetary boundary layer scaling.

Subsequently the SEBAL, SEBS and S-SEBI models have been developed based on this conception of SEBI. The main distinction between each of these models and other commonly applied single source models is the difference in how they calculate the sensible heat flux or precisely how to define the dry (maximum sensible heat and minimum latent heat) and wet (maximum latent heat and minimum sensible heat) limits and how to interpolate between the defined upper and lower limits to calculate the sensible heat flux for a given set of boundary layer parameters of both remotely sensed T_s, albedo, NDVI, LAI, Fr and ground-based air temperature, wind speed, humidity, vegetation height, etc.. Assumptions in SEBI, SEBAL, S-SEBI, SEBS models are that there are few or no changes in atmospheric conditions (mainly the surface available energy) in space and sufficient surface horizontal variations are required to ensure dry and wet limits existed in the study region.

The Surface Energy Balance System (SEBS), detailed by Su 2001 and Su et al. 2003, 2005 with a dynamic model for the thermal roughness and the Bulk Atmospheric Similarity theory for PBL scaling and the Monin-Obukhov Atmospheric Surface Layer (ASL) similarity for surface layer scaling, is an extension from the concept of SEBI for the estimation of land surface energy balance using remotely sensed data in a more complex framework. SEBS consists of: 1) a set of tools for the calculations of land surface physical parameters, 2) calculation of roughness length for heat transfer, 3) estimation of the evaporative fraction based on energy balance at limiting cases. In SEBS, at the dry limit, latent heat flux is assumed to be zero due to the limitation of soil moisture which means sensible heat flux reaches its maximum value (i.e., H_{dry} = R_n-G). At the wet limit, ET takes place at potential rate (LE_{wet}), (i.e. ET is limited only by the energy available under the given surface and atmospheric conditions, which can be calculated by a combination equation similar to the Penman-Monteith combination equation assuming that the bulk internal resistance is zero), the sensible heat flux reaches its minimum value, H_{wet}. The sensible heat flux at dry and wet limits can be expressed as:

$$H_{dry} = R_n - G \tag{15}$$

$$H_{wet} = \frac{(R_n - G) - \frac{\rho C_p}{r_a}\frac{VPD}{\gamma}}{1 + \frac{\Delta}{\gamma}} \tag{16}$$

where r_a is dependent on the Obukhov length, which in turn is a function of the friction velocity and sensible heat flux.

The EF_r and EF then can be expressed as:

$$EF_r = 1 - \frac{H - H_{wet}}{H_{dry} - H_{wet}} \tag{17}$$

$$EF = \frac{EF_r * LE_{wet}}{R_n - G} \tag{18}$$

H can be solved using a combination of a dynamic model for thermal roughness and the Bulk Atmospheric Similarity theory of Brutsaert 1999 for Planetary Boundary Layer scaling and the MoninObukhov Atmospheric Surface Layer similarity for surface layer scaling.

In SEBS, distinction is made between the ABL (Atmospheric Boundary Layer) or PBL (Planetary Boundary Layer) and the ASL similarity. Inputs to the SEBS include remote sensing data-derived land parameters and ground-based meteorological measurements, such as land surface temperature, LAI, fractional vegetation cover, albedo, wind speed, humidity, air temperature. Jia et al. 2003 described a modified version of SEBS using remote sensing data from ATSR and ground data from a Numerical Weather Prediction model and validated the estimated sensible heat flux with large aperture scintillometers located at three sites in Spain. With the surface meteorology derived from the Eta Data Assimilation System, Wood et al. 2003 applied SEBS to the SGP region of the United States where the ARM (Atmospheric Radiation Measurement) program had been carried out by the U.S. Department of Energy. Derived latent heat fluxes were compared with the measurements from the EBBR sites and results indicated that the SEBS approach had promise in estimating surface heat flux from space for data assimilation purposes. SEBS has been used to estimate daily, monthly and annual evaporation in a semi-arid environment. Su 2002 showed that SEBS could be used for both local scaling and regional scaling under all atmospheric stability regimes.

Results from Su et al. 2005 have shown that accuracy of ET value estimated from SEBS could reach 10-15% of that of in-situ measurements collected during the Soil Moisture-Atmosphere Coupling Experiment even when evaporative fraction ranged from 0.5 to 0.9.

Advantages of the SEBS are that: 1) uncertainty from the surface temperature or meteorological variables in SEBS can be limited with consideration of the energy balance at the limiting cases, 2) new formulation of the roughness height for heat transfer is developed in SEBS instead of using fixed values, 3) a priori knowledge of the actual turbulent heat fluxes is not required. However, too many required parameters and relatively complex solution of the turbulent heat fluxes in SEBS can be the source of more or less inconveniences when data are not readily available.

2.2.1 S-SEBI : A new method, called S-SEBI, developed by Roerink et al. 2000 to derive the surface energy balance, has been tested and validated with data from a small field campaign conducted during August 1997. The main theory of S-SEBI is based on the contrast between a reflectance (albedo) dependent maximum surface temperature for dry limit and a reflectance (albedo) dependent minimum surface temperature for wet limit to partition available energy into sensible and latent heat fluxes.

A theoretical explanation to S-SEBI, when a wide range of surface characteristics changing from dry/dark soil to wet/bright pixels exist, can be given: 1) at low reflectance (albedo), surface temperature keeps almost unchangeable because of the sufficient water available under these conditions, such as over open water or irrigated lands, 2) at higher reflectance (albedo), surface temperature increases to a certain point with the increases of reflectance due to the decrease of ET resulting from the less water availability, which is termed as "evaporation controlled", 3) after the inflexion, the surface temperature will decrease with the increases of surface reflectance (albedo), which is called the "radiation controlled"

In S-SEBI, the evaporative fraction is bounded by the dry and wet limits and formulated by interpolating the reflectance (albedo) dependent surface temperature between the reflectance (albedo) dependent maximum surface temperature and the reflectance (albedo) dependent minimum surface temperature, which can be expressed as:

$$EF = \frac{T_{s,max} - T_s}{T_{s,max} - T_{s,min}} \tag{19}$$

Where $T_{s,max}$ corresponds to the minimum latent heat flux (LE_{dry} = 0) and maximum sensible heat flux (H_{dry} = Rn-G) [the upper decreasing envelope when T_s is plotted against surface reflectance (albedo)], $T_{s,min}$ is indicative of the maximum latent heat flux (LE_{wet} = Rn-G) and minimum sensible heat flux (H_{wet} = 0) (the lower increasing envelope when T_s is plotted against surface reflectance). $T_{s,max}$ and $T_{s,min}$ are regressed to the surface reflectance (albedo):

$$T_{s,max} = a_{max} + b_{max}\,\alpha_s \tag{20}$$

$$T_{s,min} = a_{min} + b_{min}\,\alpha_s \tag{21}$$

If the atmospheric conditions over the study area can be regarded as constant and sufficient variations in surface hydrological conditions are present, the turbulent fluxes then can be calculated with S-SEBI without any further information than the remote sensing image itself. Results from Roerink et al. 2000 have shown that measured and estimated evaporative fraction values had a maximum relative difference of 8% when measurements obtained from a small field campaign during 1997 in Italy were compared with the S-SEBI derived outputs. Accuracy for the daily evapotranspiration using the S-SEBI method was found to be lower than 1 mm/d over a barrax test site in the framework of the DAISEX (Digital Airborne Imaging Spectrometer Experiment) campaigns. Sobrino et al. 2007 used the S-SEBI model with AVHRR data acquired from 1997 to 2002 over the Iberian Peninsula to analyze the seasonal evolution of daily ET and a RMSE of 1.4 mm/d has been shown when results derived from S-SEBI were checked against with high resolution ET values. Good results inferred from S-SEBI have been also reported by several other authors in different parts of the world.

The major advantages of this S-SEBI are that: 1) besides the parameters of the surface temperature and reflectance (albedo) derived from remote sensing data no additional ground-based measurement is needed to derive the EF if the surface extremes are present in the remotely sensed imagery, 2) the extreme temperatures in the S-SEBI for the wet and dry conditions vary with changing reflectance (albedo) values, whereas other methods like SEBAL try to determine a fixed temperature for wet and dry conditions. However, it should be noted that atmospheric corrections to retrieve T_s and α_s from satellite data and determination of the extreme temperatures for the wet and dry conditions are location-specific when atmospheric conditions over larger areas are not constant any more.

2.3 SEBAL and METRIC : SEBAL, developed by Bastiaanssen 1995 and Bastiaanssen et al. 1998 to evaluate ET with minimum ground-based measurements, has been tested at both field and catchment scales under several climatic conditions in more than 30 countries worldwide, with the typical accuracy at field scale being 85% and 95% at daily and seasonal scales, respectively.

One of the main considerations in SEBAL, when evaluating pixel by pixel sensible and latent heat fluxes, is to establish the linear relationships between T_s and the surface-air temperature difference dT on each pixel with the coefficients of the linear expressions determined from the extremely dry (hot) and wet (cold) points. The dT can be approximated as a relatively simple linear relation of T_s expressed as:

$$dT = a + bT_s \tag{22}$$

where a and b are empirical coefficients derived from two anchor points (dry and wet points).

At the dry (hot) pixel, latent heat flux is assumed to be zero and the surface-air temperature difference at this pixel is obtained by inverting the single-source bulk aerodynamic transfer equation:

$$dT_{dry} = \frac{H_{dry} \times r_a}{\rho C_p} \tag{23}$$

where H_{dry} is equal to Rn-G.

At the wet (cold) pixel, latent heat flux is assigned a value of Rn-G (or a reference ET), which means sensible heat flux under this condition is equal to zero (when reference ET is applied, both H and dT at this pixel will not equal zero any more). Obviously, the surface-air temperature difference at this point is also zero (dT_{wet} = 0).

SEBAL has been applied for ET estimation, calculation of crop coefficients and evaluation of basin wide irrigation performance under various agro-climatic conditions in several countries including Spain, Sri Lanka, China, and the United States. Timmermans et al. 2007 compared the spatially distributed surface energy fluxes derived from SEBAL with a dual-source energy balance model using data from two large scale field experiments covering sub-humid grassland (Southern Great Plains '97) and semi-arid rangeland (Monsoon '90). Norman et al.

2006 showed that the assumption of linearity between surface temperature and the air temperature gradient used in defining the sensible heat fluxes did not generally hold true for strongly heterogeneous landscape. Teixeira et al. 2009 reviwed the inputs to SEBAL model and assessed ET and water productivity with SEBAL using ground measurements observed over the semi-arid region of the Low-Middle São Francisco River basin, Brazil. Opoku-Duah et al. 2008 employed the SEBAL model with remote sensing data derived respectively from MODIS and AATSR sensors to estimate ET over large heterogeneous landscapes and found that both sensors underestimated daily ET when compared with eddy correlation observations. The selection of dry pixel and wet pixel can have a significant impact on the heat flux distribution from SEBAL.

One of the assumptions made in the SEBAL model is that full hydrological contrast (i.e., wet and dry pixels) is presented in the area of interest. The most key aspect in the SEBAL is to identify the dry pixels while wet pixels are often determined over a relatively large calm water surface or at a location of well-watered areas. The advantages of the SEBAL over previous approaches to estimate land surface fluxes from thermal remote sensing data are: 1) it requires minimum auxiliary ground-based data, 2) it does not require a strict correction of atmospheric effects on surface temperature thanks to its automatic internal calibration, and 3) internal calibration can be done within each analyzed image. However, SEBAL has several drawbacks: 1) it requires subjective specifications of representative hot/dry and wet/cool pixels within the scene to determine model parameters a and b, 2) it is often applied over flat surfaces. When SEBAL is applied over mountainous areas, adjustments based on a digital elevation model need to be made to Ts and u to account for the lapse rate, 3) errors in surface temperatures or surface-air temperature differences have great impacts on H estimate, 4) radiometer viewing angle effects, which can cause variation in Ts of several degrees for some scenes, have not been taken into account.

To avoid the limitations of the SEBAL in mapping regional ET over more complicated surfaces, Allen et al. 2005 highlighted a similar SEBAL-based approach, named METRIC, to derive ET from remotely sensed data in the visible, near-infrared and thermal infrared spectral regions along with ground-based wind speed and near surface dew point temperature. In METRIC, an automatic internal calibration method similar to SEBAL (linearly relating Ts to the surface-air temperature difference) is used to calculate the sensible and latent heat fluxes.

Gowda et al. 2008 have evaluated the performance of the METRIC model in the Texas High Plains using Landsat 5 TM data acquired on two different days in 2005 by comparison of resultant daily ET with measured values derived from soil moisture budget. Santos et al. 2008 have found that combing a water balance model with ET estimated from METRIC model could provide significant improvements in the irrigation schedules in Spain. Tasumi et al. 2005 found that SEBAL/METRIC models had high potential for successful ET estimates in the semi-arid US by comparing the derived ET with lysimeter-measured values.

Main distinctions between METRIC and SEBAL are: 1) METRIC does not assume Hwet = 0 or LEwet = Rn-G at the wet pixel, instead a daily surface soil water balance is run to confirm that for the hot pixel, ET is equal to zero, and for the wet pixel, ET is set to 1.05 ETr, where ETr is the hourly (or shorter time interval) tall reference (like alfalfa) ET calculated using the standardized ASCE Penman-Monteith equation, 2) wet pixels in METRIC are selected in an agricultural setting where the cold pixels should have biophysical characteristics similar to the reference crop (alfalfa), 3) the interpolation (extrapolation) of instantaneous ET to daily value is based on the alfalfa ETrF (defined as the ratio of instantaneous ET to the reference ETr that is computed from meteorological station data at satellite overpass time) instead of the actual evaporative fraction, which can better account for the impacts of advection and changing wind

Table 1. Comparison of different satellite based energy balance models for estimation of evapotranspiration

Aspect	**SEB** (Surface Energy Balance)	**SEBI** (Surface Energy Balance Index)	**S-SEBI** (Simplified SEBI)	**SEBAL** (Surface Energy Balance Algorithm for Land)	**METRIC** (Mapping Evapotranspiration at High Resolution with Internalized Calibration)
Description	General approach for calculating energy balance from remote sensing data.	Index-based approach that uses remote sensing data to estimate ET.	Simplified version of SEBI for ease of application.	Comprehensive algorithm to estimate energy balance components and ET using satellite imagery.	Advanced model that combines remote sensing with ground-truth for high-resolution ET mapping.
Inputs	Surface temperature, albedo, vegetation cover, etc.	Surface temperature, vegetation index, albedo.	Surface temperature, vegetation index.	Surface temperature, albedo, net radiation, vegetation index, wind speed, etc.	Surface temperature, albedo, vegetation index, net radiation, etc.
Methodology	Uses various methods to compute energy balance components, often based on physical equations.	Uses a combination of spectral indices and temperature to estimate energy balance.	Simplifies SEBI by reducing complexity in calculations.	Calculates surface energy fluxes including ET by solving the energy balance equation.	Uses a two-source model approach to separate soil and vegetation contributions to ET.
Output	Surface energy fluxes, including ET.	ET index, often expressed as a ratio or percentage.	Simplified ET estimates.	Detailed ET estimates, along with other energy fluxes.	High-resolution ET maps, with detailed energy fluxes and calibration.
Applications	General energy balance studies, ET estimation.	Remote sensing-based ET estimation, especially in vegetation monitoring.	Rapid ET estimation in various applications.	Detailed land surface modeling, agricultural water management, drought monitoring.	High-resolution ET mapping, precision agriculture, detailed water management.
Advantages	Flexible, can be adapted to various datasets and conditions.	Simple to apply, uses easily obtainable remote sensing data.	Less computationally intensive than full SEBI.	Comprehensive, detailed outputs, widely used and validated.	High accuracy, internal calibration reduces reliance on ground data.
Limitations	Can be complex to implement, requires comprehensive data.	May be less accurate in certain conditions, relies on index-based estimates.	Reduced accuracy compared to full SEBI.	Can be data-intensive and computationally demanding.	Requires high-resolution data and accurate calibration, which may not always be available.

and humidity conditions during the day.

2.3.1 Constant Reference ET Fraction (ET_rF) : In the METRIC process, Allen et al. 2007 proposed a constant ETrF, which is believed to be able to capture any impacts of advection and changing wind and humidity conditions during the day, to estimate the 24 h total ET. ET_rF is defined as the ratio of the computed ET_i from each pixel to ET_r. ET_r. is the reference ET over the standardized 0.5 m tall alfalfa and computed from meteorological data measured at ground meteorological stations.

$$ET_rF = \frac{ET_i}{ET_r} \quad (24)$$

$$ET_i = 3600\frac{LE}{L \times \rho_w} \quad (25)$$

$$L = [2.501 - 0.00236\,(T_s - 273.15)] \times 10^6 \quad (26)$$

With the assumption of instant ET_rF being same as the average ET_rF over the 24 h average and the consideration of the sloping effects over terrain areas, ETd can be estimated by [36]:

$$ET_d = C_{rad}(ET_rF)(ET_{r,d}) \quad (27)$$

$$C_{rad} = \frac{R_{s,i,Horizontal}}{R_{s,i,pixel}}\frac{R_{s,d,pixel}}{R_{s,d,Hoizontal}} \quad (28)$$

where subscripts i and d indicate instantaneous and daily values respectively, subscripts "pixel" and "horizontal" represent respectively the value for a specific pixel at certain slope and aspect conditions and value calculated for a horizontal surface. For applications to horizontal areas, Crad = 1.0. ETr,d is cumulative daily reference ET.

Conclusion

Despite substantial advancements in remote sensing technology for estimating regional surface turbulent fluxes since the 1970s, several critical challenges persist. These include the need for improved accuracy in retrieving land surface variables, effective scaling from point to regional scales, robust validation against ground-based measurements, and enhanced temporal scaling capabilities. Addressing these challenges through innovative methodologies and interdisciplinary collaborations is crucial for advancing the reliability and applicability of remote sensing techniques in estimating evapotranspiration at global and regional scales. This ongoing effort is essential for supporting various disciplines reliant on accurate water balance assessments and informing sustainable resource management practices worldwide.

References

Allen, R. G., Tasumi, M., Trezza, R. 2007. Satellite-based energy balance for mapping evapotranspiration with internalized calibration (METRIC)-model. J. Irrig. Drain. E. 133, 380-394.

Allen, R. G., Tasumi, M., Morse, 2005. A. Satellite-based evapotranspiration by METRIC and Landsat for western states water management. Presented at the US Bureau of Reclamation Evapotranspiration Workshop, Ft. Collins, CO, USA, Feb.

Businger, J. A., Wyngaard, J. C., Izumi, Y., Bradley, E. F. 1971. Flux-Profile relationships in the atmospheric surface layer. J. Atmos. Sci. 28, 181-189.

Brutsaert, W. H. 1982.Evaporation into the Atmosphere. D. Reidel: London, UK,

Brutsaert, W. 1986, Catchment-scale evaporation and the atmospheric boundary layer. Water Resour. Res. 22, 39-45.

Bastiaanssen, W. G. M. 1995. Regionalization of surface flux densities and moisture indicators in composite terrain. Ph.D. Thesis, Wageningen Agricultural University: Wageningen. The Netherlands,

Bastiaanssen, W. G. M., Menenti, M., Feddes, R. A., Holtslag, A. A. M. 1998. remote sensing surface energy balance algorithm for land (SEBAL): 1. Formulation. J. Hydrol. 212-213, 198-212.

Bastiaanssen, W. G. M., Menenti, M., Feddes, R.A., Holtslag, A.A.M.1998. A remote sensing surface energy balance algorithm for land (SEBAL): 1. Formulation. J. Hydrol. 212-213, 198-212.

Bastiaanssen, W. G. M., Pelgrum, H., Wang, J., Ma, Y., Moreno, J. F., Roerink, G. J., van der, W. T. 1998. A Surface Energy Balance Algorithm for Land (SEBAL): Part 2 validation. J. Hydrol. 212-213, 213-229.

Bastiaanssen, W. G. M., Noordman, E. J. M., Pelgrum, H., Davids, G., Thoreson, B. P., Allen, R. G. 2005. SEBAL model with remotely sensed data to improve water-resources management under actual field conditions. ASCE J. Irrig. Drain. E. 131, 85-93.

Brutsaert, W. 1999. Aspect of bulk atmospheric boundary layer similarity under free-convective conditions. Rev Geophys. 37, 439-451.

Bisht, G., Venturini, V., Islam, S., Jiang, L. 2005. Estimation of the net radiation using MODIS (Moderate Resolution Imaging Spectroradiometer) data for clear sky days. Remote Sens. Environ. 97, 52-67.

Gurney, R. J., Camillo, P. J. 1984. Modelling daily evapotranspiration using remotely sensed data. J. Hydrol. 69, 305-324.

Gowda, P. H., Chávez, J. L., Howell, T. A., Marek, T. H., New, L. L. 2008. Surface energy balance based evapotranspiration mapping in the Texas high plains. Sensors, 8, 5186-5201.

Hatfield, J. L., Perrier, A., Jackson, R. D. 1983. Estimation of evapotranspiration at one time-of-day using remotely sensed surface temperatures. Agri. Water Manage. 7, 341-350.

Hatfield, J. L. 1983. Evapotranspiration Obtained from Remote Sensing Methods. Adv. Irrig. 2, 395-416.

Idso, S. B., Jackson, R. D., Pinter, P. J. Jr., Reginato, R. J., Hatfield, J. L. 1981. Normalizing the stressdegree-day parameter or environmental variability, Agric. Meteorol. 24, 24-45.

Idso, S. B., Jackson R. D., Reginato R. J. 1975. Estimating evaporation: a technique adaptable to remote sensing. Science, 189, 991-992.

Jackson, R. D., Idso, S. B., Reginato, R. J., Pinter, P. J. 1981. Canopy temperature as a crop water stress indicator. Water Resour. Res. 17, 1133-1138.

Jackson, R. D., Hatfield, J. L., Reginato, R. J., Idso, S. B., Pinter, P. J. Jr. 1983. Estimation of daily evapotranspiration from one time of day measurements. Agri. Water Manage. 7, 351-362

Jia, L., Su, Z., van den Hurk, B., Menenti, M., Moene, A., De Bruin, H.A.R., Yrisarry, J.J.B., Ibanez, M., Cuesta, A. 2003. Estimation of sensible heat flux using the Surface Energy Balance System (SEBS) and ATSR measurements. Phys. Chem. Earth. 28, 75-88.

Klaassen, W., Van Den Berg, W.1985. Evapotranspiration derived from satellite observed surface temperatures. J. Clim. Appl. Meteorol. 24, 412-424.

Kustas, W. P., Norman, J. M. 1996. Use of remote sensing for evapotranspiration monitoring over land surfaces. Hydrol. Sci. J. 41, 495-516.

Menenti, M., Choudhury, B. 1993. Parameterization of land surface evaporation by means of location dependent potential evaporation and surface temperature range. Proceedings of IAHS conference on Land Surface Processes. IAHS Publ. 212, 561-568.

Mauser, W., Stephan, S. 1998. Modelling the spatial distribution of evapotranspiration on different scales using remote sensing data. J. Hydrol. 212-213, 250-267.

Massman, W. J. A 1999. model study of kBH -1 for vegetated surfaces using 'localized nearfield' Lagrangian theory. J. Hydrol. 223, 27-43.

McCabe, M. F., Wood, E. F. 2006. Scale influences on the remote estimation of evapotranspiration using multiple satellite sensors. Remote Sens. Environ. 105, 271-285.

Norman, J. M., Anderson, M. C., Kustas, W. P. 2006. Are single-source, remote-sensing surface-flux models too simple? In Earth Observation for Vegetation Monitoring and Water Management, D'Urso, G., Osann, Jochum, M.A., Moreno, J., Eds., American Institute of Physics: Melville, New York, USA, Volume 852, pp. 170-177.

Opoku-Duah, S., Donoghue, D. N. M., Burt, T. P. 2008. Intercomparison of evapotranspiration over the Savannah Volta Basin in West Africa using remote sensing data. Sensors 8, 2736-2761.

Roerink, G. J., Su, Z., Menenti, M. 2000. S-SEBI: a simple remote sensing algorithm to estimate the surface energy balance. Phys. Chem. Earth (B). 25, 147-157.

Soer, G. J. R. 1980. Estimation of regional evapotranspiration and soil moisture conditions using remotely sensed crop surface temperatures. Remote Sens. Environ. 1980, 9, 27-45.

Stewart, J. B., Kustas, W. P., Humes, K. S., Nichols, W. D., Moran, M. S., De Bruin, H. A. R. 1994. Sensible heat-flux radiometric surface-temperature relationship for eight semiarid areas. J. Appl. Meteorol. 33, 1110-1117.

Su Z. 2001. A Surface Energy Balance System (SEBS) for estimation of turbulent heat fluxes from point to continental scale. In: Su, Z., Jacobs, J.(Eds.), Advanced Earth Observation-Land Surface Climate. Publications of the National Remote Sensing Board (BCRS), USP-2. 01- 02, 91-108.

Su, Z. 2002. The surface energy balance system (SEBS) for estimation of turbulent heat fluxes. Hydrol. Earth Syst. Sci., 6, 85-99.

Su, Z., Li, X., Zhou, Y., Wan, L., Wen, J.2003. Sintonen, K. Estimating areal evaporation from remote sensing. Proc. IEEE Int. 2, 1166-1168.

Su, Z. 2005.Hydrological applications of remote sensing.

Surface fluxes and other derived variablessurface energy balance. In Encyclopedia of hydrological sciences, Anderson, M., Ed., John Wiley and Sons: Hoboken, NJ, USA,

Su, H., Mccabe, M. F., Wood, E. F., Su, Z., Prueger, J. H. 2005. Modeling evapotranspiration during SMACEX: Comparing two approachs for local- and regional-scale prediction. J. Hydrometeoro. 6, 910-922.

Sobrino, J. A., Gómez, M., Jiménez-Muñoz, J. C., Olioso, A.2007. Application of a simple algorithm to estimate the daily evapotranspiration from NOAA-AVHRR images for the Iberian Peninsula. Remote Sens. Environ. 110, 139-148.

Santos, C., Lorite, I. J., Tasumi, M., Allen, R. G., Fereres, E.2008. Integrating satellite-based evapotranspiration with simulation models for irrigation management at the scheme level. Irrig Sci. 26, 277-288.

Timmermans, W. J., Kustas, W. P., Anderson, M. C., French, A.N. 2007. An intercomparison of the Surface Energy Balance Algorithm for Land (SEBAL) and the Two-Source Energy Balance (TSEB) modeling schemes. Remote Sens. Environ. 108, 369-384.

Teixeira, A. H. D. C., Bastiaanssen, W. G. M., Ahmadd, M. D., Bos, M. G. 2009. Reviewing SEBAL input parameters for assessing evapotranspiration and water productivity for the Low-Middle São Francisco river basin, Brazil Part B: Application to the regional scale. Agric. For. Meteorol. 149, 477-490.

Tasumi, M., Trezza, R., Allen, R. G., Wright, J. L. 2005. Operational aspects of satellite-based energy balance models for irrigated crops in the semi-arid U.S. Irri. Drain. Sys. 19, 355-376.

Verhoef, A., De Bruin, H. A. R., Van Den Hurk, B. J. J. M. 1997. Some practical notes on the parameter kB-1 for sparse vegetation. J. Appl. Meteorol. 36, 560-572.

Webb E. K. 1970. Profile relationships: The log-linear range, and extension to strong stability. Q. J. Roy. Meteor. Soc. 96, 67-90.

Wood, E. F., Su, H., McCabe, M., Su, Z. 2003. Estimating Evaporation from Satellite Remote Sensing. Paper presented at the IGARSS. 7, 21-25.