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"THE USE OF SPACE BASED TECHNIQUES IN MONITORING EVAPOTRANSPIRATION, CONSTRAINTS TO POSSIBLE APPLICATIONS IN AFRICA" by B Chipindu, Agricultural Meteorology Programme, Department of Physics, University of Zimbabwe.

ABSTACT

There has been many research programmes investigating the potential use of data derived from space based sensors in monitoring land surface processes such as evapotranspiration. The programmes have demonstrated that evapotranspiration can be estimated using remotely sensed data in combination with routinely measured meteorological data. Good estimates of evapotranspiration will result in reliable estimates of water available for agriculture, industrial and domestic use. There are many constraints faced by practitioners in developing countries resulting in underutilization of spaced based techniques in monitoring evapotranspiration.

This paper presents the space based techniques that have been applied in monitoring evapotranspiration. It then highlights the constraints faced by practitioners when applying space based methods of estimating evapotranspiration. The growing population in developing countries and increased agricultural production have increased the demand for freshwater. Recurrent droughts have also reduced the precipitation. The assessment and management of water resources requires accurate measurements or estimates of precipitation and evapotranspiration. Space based techniques have the advantage of a high spatial coverage in assessing evapotranspiration.

1. INTRODUCTION

Evapotranspiration (ET) and precipitation are the major components of the hydrological cycle. The ground water resources strongly depend upon the balance between ET and precipitation. They both govern the amount of runoff that is available from a watershed or a river basin. ET also determines, to a large extent, the response characteristics of a watershed, to produce storm runoff and flooding as a result of heavy precipitation.

The large-scale distribution of soil moisture content and the evaporative process over land are among the most important boundary conditions of the earth-atmosphere system. The diagnosis of these fields from observed data is necessary for the initialization and verification of models such as hydrological, numerical weather prediction and global climatic models.

Monitoring ET has important implications in modelling regional and global climate and the hydrological cycle, as well as assessing environmental stress on natural and agricultural ecosystems (Kustas and Norman, 1996). Results from climate models indicate that changes in the available moisture released to the atmosphere can have significant feedback effects on cloud formation, which in turn greatly impacts the radiation budget and precipitation fields at global and continental scales (Kustas and Norman, 1996).

In many parts of the world, the available water resources are presently being tapped close to the limit. Most of the African countries have agriculturally based economies in which the provision of water and the sustainable management of the water resources are of paramount importance. Most of the commercial farming relies on irrigation rather than rainfall as a source of water for plant growth. The erratic nature of the rainfall in recent years has necessitated the construction of many water storage reservoirs and the communal farmers are being encouraged to shift to irrigation instead of relying on rainfall. The farmers compete with many urban centres for the available water resources. The assessment and management of water resources requires accurate measurements or estimates of precipitation and evapotranspiration.

Precipitation is well mapped because there are many rainfall stations manned by trained observers as well as volunteer observers. Many parts of Africa do not have synoptic stations, making it difficult to map evapotranspiration from measurements. Space based techniques, which provide a global and almost continuous set of data sometimes with good spatial and temporal resolution, have a potential for use in monitoring regional evapotranspiration.

In recent years, many models/algorithms for monitoring evapotranspiration using data derived from space–based sensors have been developed and evaluated with a few selected datasets, mostly in Europe and North America. The methods vary in complexity from empirical/statistical approaches to more physically-based analytical approaches and ultimately to numerical models that simulate the flow of heat and water through the soil, vegetation and atmosphere.

This paper describes a few methods of mapping evapotranspiration using data derived from space. It also highlights the constraints faced by researchers and practitioners from Africa in applying the methods.

2. Mapping of Evapotranspiration Using Remote Sensing Observations

2.1 The Available Energy for Evapotranspiration

Evapotranspiration and sensible heat flux into the atmosphere require the availability of energy at the earth-atmosphere interface. In computing evapotranspiration, the surface energy balance is the primary boundary condition that must be satisfied, that is,

$$R_n - H - L_v E - G = 0 \qquad (1)$$

where R_n is the net radiation at the upper surface of the layer [W m^{-2}], L_v is the latent heat of vaporization or the latent heat of sublimation [J kg^{-1}], G is the specific energy flux leaving the layer at the lower boundary or the ground heat flux [W m^{-2}], H is the sensible heat flux [W m^{-2}], and $L_v E$ is the latent heat flux [W m^{-2}]. In equation (1) the energy fluxes toward the surface are taken as positive and those away from it as negative. In many cases, R_n is the largest component in equation (1) while G can be from 5 to 50% of R_n depending on vegetation cover and soil moisture conditions (Brutsaert, 1982).

A number of approaches using remote sensing data from satellites have been developed for estimating the four components of R_n, namely:

$$R_n = (1 - \alpha_s) R_S^{\downarrow} + R_L^{\downarrow} - R_L^{\uparrow} \tag{2}$$

where $R_S^{\downarrow}$ is the downward short wave radiation transmitted through the air [W m^{-2}], $\alpha_s R_S^{\downarrow}$ is the reflected short wave radiation [W m^{-2}], $R_L^{\uparrow}$ is the upward long wave (infrared, ir) radiation emitted by the surface [W m^{-2}], $R_L^{\downarrow}$ is the downward long wave diffusive (ir) radiation [W m^{-2}], α_s is the albedo of the surface or the fraction of downward short wave radiation that is reflected by the earth's surface. The fluxes into the surface are positive by definition while fluxes away from the surface are negative. The long wave radiation consists of two components: the downward radiation from the earth's atmosphere, $R_L^{\downarrow}$ and the upward radiation from the surface, $R_L^{\uparrow}$. The upward long wave radiation is usually obtained by assuming that the ground, the canopy or the water surface under consideration is equivalent to an infinitely deep grey body of uniform radiometric temperature, T_s, and emissivity, ε, ranging from 0.9 to 0.99. From Stefan-Boltzmann law the upwelling long wave radiation is given by:

$$R_L^{\uparrow} = \varepsilon \sigma T_s^4 \tag{3}$$

where $\sigma = 4.67 \times 10^{-8}$ W m^{-2} K^{-4} is the Stefan-Boltzmann constant.

The radiometric temperature measured by an infrared radiometer from a space borne platform, T_{rad}, is assumed to approximate T_s (Kustas and Norman, 1996). Both R_S and α have been estimated from satellites using empirical/statistical and physically-based models (Kustas and Norman, 1996). Satellite estimates of the contribution of the net longwave flux at the surface have been developed using sounding data. Other approaches have utilized meteorological data collected at screen level with semi-empirical relationships for estimating $R_L^{\downarrow}$, then use T_{rad} for calculating the upwelling longwave component.

2.2 Determination of Evapotranspiration

One of the most widely applied approaches of mapping daily evapotranspiration using remote sensing observations of surface radiant temperature T_{rad} near midday was pioneered by Jackson *et al.,* (1977). The so-called "Simplified Method" was later adapted and applied by several investigators, such as Seguin and Itier (1983), Nieuwenhuis et al, (1985), Carlson and Baffum (1989) and Carlson et al., (1995).

In the Simplified Method, the net integrated daily evapotranspiration at the surface (ET_{24}) is estimated from a surface radiant temperature measured near the time of local maximum (about 1300-hr local time, T_{013}), a corresponding air temperature (usually measured at screen height, T_{a13}), and the net radiation expressed as an integrated value over a 24-hr period (R_{n24}). They observed that daily differences between net radiation and evapotranspiration could be approximated by the following linear expression (Seguin and Itier, 1983):

$$R_{n24} - L_v E_{24} = B\,(T_{013} - T_{a13})^n \tag{4}$$

where R_{n24} is the net radiation [W m^{-2}] expressed as an integrated value over a 24-hour period, L_vE_{24} is the latent heat flux or evapotranspiration [W m^{-2}] over a 24-hr period, T_{013} and T_{a13} are, respectively the surface radiant temperatures [K] measured near the time of local maximum (about 1300 local time) and air temperature [K] (usually measured at screen height). B and n are pseudo constants given as functions of the normalized difference vegetation index (NDVI), expressed as a scaled index N*. Both N* and T_{a13} are obtained with the aid of remotely determined measurements, which are viewed on scatter plots of T_{a13} versus NDVI.

According to Seguin and Itier, 1983) equation (4) is of limited utility in areas where *in situ* measurements are lacking. Perceiving this limitation, Carlson and Baffum (1989) suggested that air temperatures at 50 m be used instead of those at screen level. This is because the 50-m air temperature is relatively independent of local surface conditions and can be approximated by regional-scale averages of surface air temperatures, such as reported on weather maps, or from the output of regional-scale atmospheric prediction models. In equation (4), T_{a13} refers to the 50-m air temperature.

Equation (4) was verified using data collected during the intensive field operations by Carlson *et al.*, 1995). They concluded that knowledge of one side of equation (4) is not sufficient to obtain the evapotranspiration, because the integrated net radiation must still be determined. However, they suggested that an estimate of net radiation could be made with reasonable accuracy for clear sky conditions using mathematical formulae. They also noted that, although satellite remote sensing of radiant surface temperature is feasible during clear sky conditions only, it should be possible to calculate ET_{24} with some degree of accuracy over a season, given a measurement of R_{n24}, by determining the so-called evaporation fraction (ET_{24} /R_{n24}**)** from satellite images, for a succession of clear days and then interpolating the ratios for cloudy ones.

Bastiaanssen *et al*., (1994) developed the Surface Energy Balance Algorithm for Land (SEBAL). The instantaneous evapotranspiration (ET_a) act flux is calculated as a residual of the surface energy budget equation (Eq 1) as:

$$ET_a = R_n - H - G \qquad (5)$$

The net radiation is computed by subtracting all outgoing radiant fluxes from incoming radiant fluxes. In equation 2, the broad band surface albedo α is derived from the narrow band spectral reflectances measured by the satellite. The incoming shortwave radiation $R_S^{\downarrow}$ is computed using the solar constant, the zenith angle and sun-earth distance in combination with a clear sky transmission factor derived from field observations. The incoming longwave radiation $R_L^{\downarrow}$ is computed using a modified Stefan-Boltzmann equation. Outgoing longwave radiation $R_L^{\uparrow}$ is computed using the Stefan-Boltzmann equation with a calculated surface emissivity and surface temperature. Surface temperatures are computed from the satellite measurements of thermal radiances.

In equation (5), the sensible heat flux (H) and soil heat flux (G) are subtracted from the net radiation flux at the surface (R_n) to compute the residual energy available for evapotranspiration (L_vE). Soil heat flux is empirically calculated as a G/Rn function using vegetation indices, surface temperature and surface albedo. Sensible heat flux is computed using surface wind speed observations, estimated surface roughness and surface to air temperature differences that are obtained through a sophisticated self-calibration between dry and wet pixels (van Dam and Malik, 2003).

SEBAL converts the instantaneous latent heat flux (L_vE) into daily (L_vE_{24}). The daily values of evapotranspiration are converted into monthly values by applying the Penman-Monteith equation.

According to van Dam and Malik, (2003) SEBAL computations can only be executed for cloudless days.

Bastiaanssen *et al*., (1994) tested the SEBAL model in the Castilla la Mancha area in Spain. Space, air and ground-truth data were collected simultaneously over the Castilla la Mancha area (100 x 100 km) by various agencies, as part of the HAPEX (Hydrological and Atmospheric Pilot Experiment) of the European project on Climate and Hydrological Interactions between Vegetation, Atmosphere and Land (ECHIVAL) surfaces. The project was executed during the summer of 1991.

SEBAL was used to map out the land surface energy balance, pixel wise and to determine its associated resistance to total evaporation within an acceptable level of accuracy (Bastiaanssen *et al*., 1994). Considering the various factors, which limited the applicability of remote sensing and the limited data set, which was available, the results of the study were promising (Bastiaanssen *et al*., 1994). Quantitative estimates of surface layer moisture content compared well with field observations.

Zhang *et al*., (1995) developed the one-layer resistance model for estimating regional evapotranspiration (ET) using remote sensing data. They argued that the real ET process from a heterogeneous surface can be conceptualized as a one-layer process from an "averaged" surface. This "averaged" surface represents the characteristics of the real heterogeneous surface in transferring sensible and latent heat fluxes. It can be argued that if model parameters are defined as horizontal averages over area scales in which persistent features exist in large numbers, then a one-layer model may be satisfactory.

The one-layer resistance model uses the energy balance (Eq.1). The transfer of sensible heat is computed as:

$$H = \rho c_p \frac{(T_s - T_a)}{r_{ah}} \tag{6}$$

where ρ is the density of the air [kg m^{-3}] and c_p is the specific heat of the air at constant pressure [J kg^{-1} K^{-1}]; T_s is the remotely sensed temperature of the 'averaged' surface [K]; T_a is the air temperature at the reference height [K]; r_{ah} is the aerodynamic resistance for sensible heat [s m^{-1}].

Equation (6) is a one-layer bulk transfer equation and it is based on the assumption that the radiometric temperature measured by a thermal infrared radiometer is identical to the aerodynamic temperature (Zhang *et al*., 1995). In the case of full canopy cover, there exists a near-equivalence between these two temperatures and it is found that estimates of ET using radiometric temperature are in good agreement with observed values. However, the deviation between the two temperatures is large for partial canopy cover conditions. In this case, there might be significant uncertainties in the sensible heat flux estimated using the radiometric temperature.

The latent heat transfer can be expressed as (Zhang *et al.*, 1995):

$$L_v E = \frac{\rho c_p}{\gamma} \frac{e_s(T_s) - e_a}{r_{av} + r_s} \tag{8}$$

where γ is the psychometric constant; e_s is the saturation vapour pressure [hPa] at the surface temperature,(T_s), e_a is the vapour pressure [hPa] at the reference height; r_{av} is the aerodynamic resistance [s m^{-1}] for water vapour, r_s is the surface resistance [s m^{-1}].

Zhang *et al.*, (1995) tested the one layer resistance model using data obtained during a Special Observing Period (SOP) in the HAPEX-MOBILHY (Hydrologic Atmospheric Pilot Experiment, Modélisation Du Bilan Hydrique) experiment. They made a comparison between the evapotranspiration calculated by the model and the SAMER measurements. A good agreement was revealed between the model estimates and the measurements. Zhang *et al*. (1995) also compared the calculated regional ET from the one-layer resistance model with the regional average values of ET obtained from the SAMER stations located in the agricultural area. There was good correspondence between the calculated regional evapotranspiration and the area averaged SAMER measurements.

Mecikalski *et al*., 1999 developed the Atmosphere-Land Exchange Inversion (ALEXI) model by modifying a version of the Two-Source Time-Integrated Model (TSTIM), which is an extension of an earlier Two-Source Model (TSM) developed by Norman *et al*. (1995). The ALEXI model uses the rate of change of surface radiometric temperature, $\Delta T_{RAD,}$ [K] to partition the net radiation, R_n, at the earth's surface into fluxes of sensible heat, H, latent heat, L_vE, and soil heating, G, (Mecikalski *et al*., 1999). The surface radiometric temperature at view angle φ is given by (Mecikalski *et al*., 1999):

$$T_{RAD}(\varphi) \approx f(\varphi)\, T_c + [1 - f(\varphi)]\, T_s \tag{9}$$

where T_s is the surface temperature. The radiometric temperature is obtained to brightness temperature measured by satellites.

The latent heat flux is partitioned as:

$$\lambda E = \lambda E_s + \lambda E_c$$

$$\lambda E_c = \alpha_{PT} f_g \frac{\Delta}{\Delta + \gamma} R_{n,c} \tag{10}$$

where the subscripts s and c represent soil and canopy flux components, respectively, α_{PT} is the Priestly-Taylor coefficient, equal to 1.3, f_g is the fraction of green vegetation, Δ is the slope of the saturation vapour pressure versus temperature curve, γ is the psychrometric constant and $R_{n,c}$ is the component of net radiation absorbed by the canopy.

The ALEXI model requires brightness temperature measurements at two times during morning hours to obtain a temperature-change signal (Mecikalski *et al*., 1999). Based on sensitivity tests reported by Anderson *et al*. (1997), these two observation times have been fixed at 1.5 and 5.5 hours past local sunrise. The ALEXI model generates flux estimates of sensible heat, latent heat, soil heat flux and net radiation at 1.5 and 5.5 hours after local sunrise.

Mecikalski *et al*., (1999) applied the ALEXI model over a region in the Central United States. The model was executed at each 10-km grid scale across the domain, subject to data availability and quality restrictions. All the input quantities that are allowed to vary over the ALEXI model domain were assembled from a variety of satellite imagery and surface weather station reports. The data fields were then interpolated to a grid at the resolution of the model application.

Mecikalski *et al*., (1999) compared the ALEXI model flux estimates with climatological moisture and vegetation patterns as well as with surface-based flux measurements acquired during the joint USDA-NASA Southern Great Plains (SPG-97) Hydrology Experiment. The comparisons were quite promising.

1.5 Possibilities for Applications in Africa

Mapping of evapotranspiration using space-based techniques or remotely sensed data is possible in Africa. However, application of physically-based algorithms or numerical methods may not be suitable for routine (daily- weekly) mapping of evapotranspiration (ET) over many parts of Africa because of various reasons. The first one is the low frequency in satellite coverage over most countries (that is every two weeks). The second one is the course resolution of satellite data. Radiometric surface temperature observations with 1-5 km resolution are available several times per day but this resolution is too course for estimating ET from agricultural fields or for defining variations in ET due to land cover changes (Kustas and Norman, 1997). According to Bastiaanssen et al., (2000), researchers have an influence on the sensor design and flight characteristics (e.g. revisit period, special resolution). These decisions do not necessarily respond to the needs of practitioners in the field of agrometeorology. The third is cloud contamination. Most algorithms for mapping evapotranspiration require cloud free conditions, a requirement rarely met by the atmosphere. The fourth one is the surface heterogeneity. Multiple features of the landscape affect all the wavebands currently being used by satellite-based instruments. For example, the satellite clear sky radiances include radiation emitted by the surface and atmosphere as well as radiation reflected by the surface towards the sensor, (Olioso, 1995). Thus a remotely sensed signal is not uniquely related to a single surface property, and semi-empirical algorithms are required to convert the observed radiance into physical quantities useful ET modelling. This non-uniqueness between the observed radiances and landscape features makes it difficult to apply remote sensing techniques in estimating ET without ground truth data. The fifth reason is the inadequate technology. There are very few satellite receiving stations in Africa and it takes a long time to for one to source the satellite data. The software packages for analysing the satellite data are not readily available in Africa, making it difficult to use the data. Lastly, the practitioners in monitoring evapotranspiration are often unaware of the new technical possibilities, partly because the information about remote sensing remains a preserve of the remote sensing community.

1.6 Conclusion

The mapping of evapotranspiration using space based techniques in Africa has a lot of potential if the developers of the models supply user friendly software to the end users. Research should be carried out in the use of microwave sounders which are not affected by clouds. Personnel from the National Meteorological and Hydrological Services need to be trained in using the different methods of mapping evapotranspiration. They can then provide services to governmental and other agencies involved in managing water resources. Demonstrations to potential end users showing the power of surveying from space are required showing the power of surveying from space, to allow remote sensing to move from being a specialized research tool, to become an important assert for mapping evapotranspiration.

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