EVAPOTRANSPIRATION and REMOTE SENSING

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ABSTRACT

Evapotranspiration is the process in which water in the liquid phase is extracted from the soil and transferred to the atmosphere in the vapor phase. This transfer can occur as evaporation directly from the soil surface or transpiration from plant leaves. There are three things required for the process to occur: 1) energy (580 cal/gm) for the change of phase of the water; 2) a source of the water, i.e. adequate soil moisture in the surface layer or the in the root zone of the plant; and 3) a sink for the water, i.e. a moisture deficit in the air above the ground.

Remote sensing can contribute information to the first two of these conditions by providing estimates of solar insolation, surface albedo, surface temperature, vegetation cover, and soil moisture content. In addition there have been attempts to estimate precipitation and shelter air temperature from remotely sensed data. The problem remains to develop methods for effectively using these sources of information to make large area estimates of evapotranspiration.

INTRODUCTION

Evaporation and transpiration (evapotranspiration or ET) estimates are required for several purposes, including irrigation scheduling, water balance calculations, run-off prediction and meteorological and climatological studies. Long-term estimates of evapotranspiration may be made using water balance methods, as in lysimetry or evaporation pans, or at a larger scale in river basins. However, there is considerable uncertainty in these measurements, particularly for short time periods, and it is often difficult to generalize the measurements to large areas. Many areas do not have any hydrologic measurements and for these areas Thornthwaite (1947) suggested an empirical approach to estimate long-term evaporation from routine meteorological observations, principally monthly mean temperature. Penman (1948) derived a more physically-based expression which uses standard meteorological data to estimate potential evapotranspiration. This was for a short grass cover with an adequate water supply. This model has served as the theoretical basis for most of the currently used models. However their application is still limited to locations where standard meteorological data are available. In principle, remotely sensed measurements, offer methods for extending these models to much larger areas including those where data may be sparse. Although no method is as yet operational the utility of such measurements is sufficiently great to encourage further work.

To understand better how remote sensing may contribute it is

instructive to examine first the basic energy and moisture balance equations.

Energy and Moisture Balance

The energy balance of a soil or vegetated surface is governed by a conservation equation. In the absence of precipitation or advection, we may write

$$R_n + G + H + LE = 0 \qquad (1)$$

where R_n is the net radiation flux, G the soil heat flux, and H the sensible heat flux and LE the latent heat flux in the atmospheric boundary layer. The latter is the product of the water vapor flux, E, and the latent heat of vaporization of water per unit mass, L.

The net radiation flux is the sum of the incoming and outgoing short and long wave radiation fluxes:

$$R_n = (1 - \alpha_s)R_s + e_L (R_L - \sigma T_c^4) \qquad (2)$$

where R_s is the incoming short-wave radiation and R_L is incoming long-wave radiation, α is the short-wave soil/crop reflectivity or albedo, e_L is the long-wave emissivity, σ is the Stefan-Boltzman constant and T_c is the soil surface temperature in Kelvins.

The soil heat flux is related to the temperature gradient in the soil dT/dz and to the thermal conductivity λ

$$G = \lambda \; dT/dz \qquad (3)$$

This relationship leads to a form of the diffusion equation

$$d(\lambda \; dT/dz)/dz = C_V \; dT/dt \qquad (4)$$

where C_V is the volumetric heat capacity of the soil. This equation may be solve numerically to give a temperature profile in the soil, taking either T_c or G as the upper boundary condition and some known temperature at a given depth in the soil as the lower boundary condition.

The sensible and latent heat fluxes may both be estimated from the transport equations. The sensible heat flux H may be expressed as

$$H = \rho \; C_p \; (T_a - T_c)/r_a \qquad (5)$$

where ρ is the air density, C_p is the specific heat of the air and r_a is the aerodynamic resistance. The latent heat flux may be expressed as

$$LE = \frac{\rho \; C_p}{\gamma^{\prime}} \; \frac{(e_a - e_s)}{r_a - r_s} \qquad (6)$$

where $\gamma^{\prime}$ is the psychrometric constant, e_a is the atmospheric vapor pressure in the boundary layer, e_s is the saturated vapor pressure at the temperature T_c, and r_s is the stomatal diffusion resistance to water vapor transport.

Equations (2), (3), (5) and (6) are a mathematical model of the surface energy balance in terms of the soil surface temperature and a set of meteorological measurements, which at least in principle, are measured or estimated routinely. The energy and moisture fluxes are intimately related because the latent heat flux gives the amount of water evaporated and transpired.

One formulation of potential evapotranspiration that lends itself to remote sensing inputs is that developed by Priestly and Taylor (1972). They obtained for saturated surfaces

$$LE = \alpha \, [s/(s + \gamma^{\prime})](R_n - G) \qquad (8)$$

where s is the slope of vapor pressure versus temperature curve, and α is an empirical evaporation constant which they determined to be 1.26. Barton (1979) and Davies and Allen (1973) have modified the result for unsaturated surface by treating α as a function of the surface layer soil moisture. Barton used airborne microwave radiometers to remotely sense soil moisture in his study of evaporation from bare soils and grasslands.

Remotely Sensed Measurements

Several types of remotely sensed observations may be made by measuring the electromagnetic radiation in a particular waveband either reflected or emitted from the earth's surface. The incoming solar radiation can be estimated from satellite

observations of cloud cover primarily from geosynchronous orbits. For clear sky conditions, the surface albedo may be estimated by measurements covering the entire visible and near-infrared wavelengths, while measurements at narrow spectral bands can be used to determine vegetative cover. The surface temperature can be estimated from measurements at thermal infrared wavelengths, i.e. the 10.5 to 12.5 micron waveband, of the emitted radiant flux and from some estimate of the surface emissivity (for actual surfaces e is usually close to unity). The microwave emission and reflection or backscatter from soil, primarily for wavelengths between 5 and 21 cm, is dependent on the dielectric properties of the soil which are strong functions of the soil moisture content. There are uncertainities in the determination of soil moisture values from microwave measurements introduced by factors such as surface roughness and vegetative cover but it appears that microwave methods can estimate the soil moisture content in the surface 5 cm layer of the soil with 4 or 5 levels of discrimination.

Remotely sensed measurements may be made from a variety of platforms. Operational work is largely conducted using satellites, which offer repetitive coverage but suffer from reduced spatial resolution compared to aircraft. Currently only visible and infrared data are available from satellite. Aircraft platforms are primarily used for experimental studies because of the greater control possible.

Applications of Remotely Sensed Measurements

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In recent years there has been much progress made in the remote sensing of a number of parameters which can contribute to the estimation of evapotranspiration. These include: surface temperature, surface soil moisture, surface albedo, vegetative cover, and incoming solar radiation. There has been little progress made in the direct remote sensing of the atmospheric parameters which affect evapotranspiration such as: surface air temperature, water vapor gradients, and surface winds. Thus approaches for evapotranspiration estimation using remotely sensed data will have to work around these missing data.

The evaporation from the soil surface is directly related to the vapor pressure difference between the surface and lower atmospheric boundary layer, and the water supply in the soil. Assuming that there is a suitable supply of water in the soil, the most important controlling factor is the difference $[e_a - e_s]$ in equation (6), which is largely determined by the corresponding temperature difference. Thus there is a strong relationship under certain conditions between the canopy and air temperature differences and the evapotranspiration rate. This comparison of canopy and air temperatures has been used to detect crop water stress for irrigated crops in the southwestern part of the United States and these relationships have been described in a series of papers from the group at the US Water Conservation Laboratory in Phoenix (Jackson et al., 1981 and references cited therein).

Estimates of the net radiation from geostationary satellite data are used by Kanemasu (1982) in a Priestly-Taylor type of equation to estimate evapotranspiration. The resulting moisture

flux is then used to drive a water balance model for predicting crop yields. They also used Landsat multispectral data to obtain vegetation indices.

A more complete but less practical approach is that adopted by Soer(1980), Rosema et al. (1978), and Carlson (1981), amongst others, in which the energy balance (equations 1 - 6) is modelled and the moisture balance is inferred from the heat flow in the soil. Equations (2), (4), (5), and (6) can be manipulated to obtain an expression for the surface energy balance in terms of the surface temperature and a set of meteorological measurements which, at least in principle, are measured or estimated routinely The initial estimate of the soil/crop surface temperature T   will almost certainly not satisfy the continuity requirement of equation (1) because of numerical simplifications and so it is necessary to use an iterative numerical optimization technique to ensure that the condition in equation (1) is satisfied. The technique may be adapted to give estimates of the surface temperature T_c at every time step at which routine meteorological observations are available. The Tergra (Soer, 1980) model uses a Businger-Dyer (Businger et al.,1971 and Dyer, 1967) approach to estimate the turbulent diffusion resistance r_a. This gives r_a as a function of wind velocity and the stability of the atmospheric boundary layer just above the surface. A simple empirical parameterization of the stomatal resistance r_s and an explicit finite difference soil heat flux model are used.

The Tellus model (Rosema, et al., 1978) is somewhat similar except that the model is calibrated against both a daytime and a

nightime surface temperature. A Newton-Raphson numerical
integration scheme is used to optimize two parameters: the surface
relative humidity , effectively related to the resistance to water
transport across the surface boundary layer, and the thermal
inertia of the soil. The finite difference scheme uses an
expanding grid with thinner soil layers near the surface (Rosema
et al.,1978). This model has been used to create look-up tables,
which were used with airborne thermal infrared surface temperature
data and albedo estimates to give maps of cumulative daily
exapotranspiration (Dejace et al., 1979). Other studies (Gurney
1978) have shown that the Tergra and Tellus models give results
relatively close to those estimated or measured by other means.
Carlson (1981) developed a similar model, with a simple soil heat
flux component. Using satellite data from the Heat Capacity
Mapping Mission he obtained reasonable results under non-stressed
conditions from vegetated surfaces. Elkington and Hogg(1981) used
an approach based on the Tergra model to estimate
evapotranspiration. However they usd a simple set of inputs ,
estimating net radiation from the Brunt equation and the other
boundary conditions from routine meteorological observations taken
at nearby standard weather stations. Their results are
sufficiently encouraging to suggest that the simplifications which
would be required for an operational approach are indeed possible.

Bernard et al.(1981) in a simulation study have examined the
use soil moisture estimates from a radar to calculate
evapotranspiration rates. The microwave estimates of the surface
soil moisture were used as the upper boundary condition in a water

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balance model. The study provides a good indication of the utility of microwave soil moisture sensing for estimating evapotranspiration.

A more physically based and realistic model has been developed by Camillo and Schmugge (1981). Both energy and moisture fluxes are modelled. The space-time dependence of temperature and moisture content are described by a set of diffusion type partial differential equations, and the model uses a predictor/corrector method to numerically integrate them, yielding both moisture and temperature profiles in the soil as functions of time. The model was used to simulate energy and moisture fluxes under field conditions using data from Phoenix, Arizona. The qualitative agreement between the observations and the model was very good, indicating that the model is a reasonably accurate representation of the physical processes involved. A further example of the use of the model is given in Gurney and Camillo (1982). This model, because it simulates the moisture profile explicitly in addition to the temperature profile, requires an initial moisture profile. At least in principle this may also be derived using remotely sensed microwave measurements. As the model is most sensitive to changes in the surface soil temperature and soil moisture, which is that part of the profile accessible to the microwave thermal infrared measurements, this is a very realistic application of remotely sensed data. If it is possible to take initial conditions from remotely sensed measurements and the boundary conditions either from routine meteorological observations or remotely sensed measurements, the

estimation of actual evapotranspiration from remotely sensed measurements becomes feasible.

There are various other measurements which have either been suggested as inputs to evapotranspiration estimates or in principle could be used as additional model inputs. Multispectral information may be used to estimate the health of vegetation and the biomass (Tucker, 1980, Holben, et al.,1980, and Kanemasu, 1982) and hence to give an indication of the evapotranspiration rate.

The major problems with all remote sensing methods of evapotranspiration are: 1) the process of transpiration is still not well understood and parameterized for structured crops such as cereals or complex vegetation, such as trees; 2) in the presence of vegetation the surface temperature T_c estimated by a thermal infrared sensor is at an unknown level within the vegetation; and 3) the most appropriate use of microwave observations of surface soil moisture in the presence of vegetation needs to be determined. For the future we expect that the most practical method will probably use a multi-spectral approach including repetitive observations at the visible, near and thermal infrared, and microwave wavelengths. This will afford the possibility of estimating solar insolation, surface vegetative cover and/or albedo, surface temperature, and surface soil moisture from remotely sensed data and incorporating them into models of the type described here.

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