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Patterns and Organisation in Evaporation

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5.1 INTRODUCTION

The evaporation of water is a crucial process in hydrology and climate. When the whole planet surface is considered, most of the available radiation energy is consumed in this process. However, a global view alone is insufficient to explain the codependence of surface hydrology and climate. Recent findings indicate that spatial variations in surface water and energy balance at various scales play a large role in the interactions between the surface and atmosphere. Advances in remote sensing have hastened the awareness of the spatial variability of the surface, and also offer some promise to quantify such variability. A point has been reached where the quantification of spatial patterns of evaporation is required in order to address current issues in hydrology and climate.

The evaporation of water at the surface and subsequent exchange with the lower atmosphere is a complex process – even for local scales and simple surfaces. When larger scales and spatial variations are considered, nonlinear processes may become pronounced, and further difficulties arise. Because of its great importance to hydrology and climate, considerable effort has been extended towards understanding and quantifying the evaporation process. Much is known about the process for uniform surfaces at local scales. However, current issues in hydrology and climate involve larger scales and non-uniform surfaces. Here there remains much to be learned. Note that evaporation can follow several avenues, including free water surfaces, soil surfaces, and transpiration by vegetation. Here we use the term evaporation in a generic sense, so that it is inclusive of any of these pathways.

5.2 GOVERNING FACTORS AND MODELS

5.2.1 Governing Factors

Before contending with spatial patterns there must be clear understanding of the processes important to a local surface. Because of the variety of ecosystems

and environmental conditions, the importance of various factors on evaporation differs from case to case. This can lead to confusion and improper generalisations about how to approach the process. We commence with a brief overview of the governing factors and subsequent interactions.

Water Supply

For land surfaces, the upper soil profile or root zone is the storage medium for water. The depth of soil in which water content must be considered must be commensurate with the root zone. Knowledge of surface water content alone is insufficient. Although soil water availability is a necessary condition for evaporation, the rate is not only a function of soil water. However, spatial variations in soil water play a direct role in spatial patterns of evaporation.

Available Energy

When water is sufficiently available, evaporation often proceeds at a rate that is proportional to available energy, usually defined by $R_n - G$, where R_n is net radiation and G is energy flowing into the soil. The large value of latent heat causes a great deal of energy to be consumed when water is available. This has led many models to treat evaporation as proportional to the available energy, and reflects the historical bias of research towards surfaces with relatively large water supplies.

Saturation Deficit

The very large negative values for water potential in the atmosphere require more useful variables such as vapour pressure or specific humidity. The gradient in humidity between the surface and the air has historically been replaced by the saturation deficit of the air, in order to linearise equations and avoid explicit dealings with surface temperature. When surface humidity values are large enough, saturation deficit effectively represents the gradient in water potential.

Turbulence Transport

Supply of water, energy, and a gradient of humidity are not enough to maintain the process, however. The water vapour must be transported away from the surface into the atmosphere, or the humidity gradient would soon decay and reduce the evaporation. So wind and turbulence play a critical role in maintaining values of saturation deficit. Unfortunately, turbulence is a very complex process without an analytical solution. As a result, it is inevitably parameterised in any treatment of evaporation.

Stomatal Conductance

Finally, when plants are considered, the situation becomes much more complex. Plants are *living* things, which limits the use of physical laws and mathematics to describe the processes. The response has been to focus on the behaviour of the stomates, since water vapour must pass through these structures. Indeed,

stomatal conductance is a key mechanism by which we account for the role of the vegetation in this process.

Although stomatal conductance of plants has been studied for many years, predicting the exact behaviour remains somewhat elusive. We know that there are connections between stomatal conductance, transpiration, and several atmospheric variables such as saturation deficit. The connections between the processes are examined at scales from the sub-leaf to canopy by Jarvis and McNaughton (1986). However, the concepts of cause and effect are tenuous. Historically, stomatal conductance was assumed to respond to saturation deficit, and thereby affect transpiration. However, Mott and Parkhurst (1992) showed that transpiration may respond directly to saturation deficit, and stomatal conductance adjusts in response to transpiration. Monteith (1995a) reanalysed 52 data sets, and concluded that they support this hypothesis. Monteith (1995b) discusses the implications of this issue on approaches to model evaporation. Clearly there are complex and nonlinear interactions between plant water status, stomatal conductance, transpiration, and various atmospheric factors. The role of living vegetation in the process is not treated very directly at present.

5.2.2 Problems of Nonlinearity

A major difficulty in modelling evaporation is the strong dependency among the variables. In fact, there are no independent variables as such. Changes in any of the critical factors in principle induce changes in all others, until a new equilibrium can be reached. At small spatial scales the nonlinearities are not always very evident. Hence, many of these have historically been ignored or hidden inside the definitions of various parameters. Indeed, the common consideration of a very shallow layer of atmosphere above the surface does not allow for many of the critical feedbacks. The solution to this problem will be discussed shortly. It involves examination of the entire atmospheric boundary layer.

5.2.3 Models Describing Evaporation

Penman–Monteith Equation

This expression is the most fundamental equation available to examine the evaporation process. It is strictly valid for a leaf, but is generally considered at the scale of a canopy. A uniform surface is implicitly assumed. The equation is developed by linearising the vapour pressure gradient term, to remove any explicit dependence on surface temperature. The final equation is:

$$E = \frac{s \cdot (R_n - G) + \rho \cdot c_p \cdot D/r_a}{s + \gamma \cdot (1 + r_c/r_a)} \tag{5.1}$$

Here s is the slope of the saturation specific humidity versus temperature relation, ρ is density of air, c_p is specific heat of air, γ is c_p/L where L is latent heat of vapourisation, D is saturation deficit or saturation minus actual specific humidity, r_a is aerodynamic resistance, and r_c is stomatal resistance.

The role of turbulence and stomatal behaviour are both collapsed into resistance terms. Also note that for scales larger than a single leaf, the stomatal resistance term represents some bulk or effective value for the surface. The value of D is generally specified near the surface. Thus, there is no explicit allowance for connections and exchanges with a deeper layer of atmosphere. This equation is a diagnostic equation describing the relationships between key factors of the system. It represents a tool to examine interactions between evaporation and critical factors in the soil, vegetation, and atmosphere.

Simplifications for Special Cases

For extensive surfaces covered with vegetation, the evaporation is large and convection is small. This leads to poor coupling between the surface and atmosphere, and evaporation becomes energy limited. The evaporation flux by definition must approach the value of available energy. This value is called *equilibrium evaporation* (E_{eq}). For extensive vegetated surfaces the actual evaporation is strongly proportional to E_{eq} . This led Priestley and Taylor (1972) to propose that:

$$E = \alpha \cdot E_{eq} \quad (5.2)$$

where α is a parameter, originally defined as 1.26, although McNaughton and Spriggs (1989) demonstrate that α is not constant and depends on dynamic interactions between the surface and atmospheric boundary layer. Nevertheless, this equation is a useful tool for the special case of large and uniform regions with complete vegetation.

Use of Surface Temperature to Estimate Evaporation by Residual

If the entire energy balance equation is considered, E can be estimated by the residual if the other terms are calculated and measured. This involves determination of sensible heat flux (H). Remote sensing methods allow estimation of the surface temperature, which can be used with air temperature to estimate H using similarity theory, as described later. Since remote sensing techniques can sometimes retrieve spatial fields of surface temperature, such an approach can estimate spatial distribution of evaporation. Examples of this approach will be discussed in Section 5.6.

Coupling of Surface Energy Balance to the Atmospheric Boundary Layer

Most of the historical study of evaporation has been conducted at local scales, and considered a layer of atmosphere only a few metres above the surface. This ignores the role of large-scale atmospheric properties and the feedback between the surface and the atmosphere.

Recently, several studies have demonstrated the need to consider a continuous and interactive system that often includes the atmospheric boundary layer (ABL) as well as the air above it. McNaughton and Jarvis (1983) and McNaughton and Spriggs (1986) demonstrate how a growing ABL can entrain warm, dry air from aloft which mixes down to the surface. This can raise the value of saturation

deficit, and enhance evaporation rates. The system is coupled, in that changes in the surface heat and evaporation rates affect the growth of the ABL, which in turn can feed back to alter the surface fluxes. These processes were combined into an elegant model posed by McNaughton and Spriggs (1986). These connections between the surface energy balance and the ABL must be considered in the process. They become especially important for regional scales, or to consider spatial variations in surface fluxes.

5.3 ESTIMATION OF EVAPORATION RATES USING MEASUREMENTS

There are several approaches either to measure evaporation directly, or to estimate it from other measurements. We will cover the most common and reliable approaches.

5.3.1 Local Scales

Eddy Covariance

This is the most direct approach, and attempts to actually measure the flux. The flux of water vapour can be described as:

$$E = \bar{w} \cdot \bar{\rho}_v + \overline{w' \rho'_v} \tag{5.3}$$

where ρ_v is water vapour density, and w is the vertical wind velocity. The primes indicate instantaneous deviations from the temporal mean. The first term represents flux due to the mean vertical wind, while the second term is the turbulence flux. In many conditions over flat surfaces with a suitable averaging period, the mean vertical velocity should be zero. The first term then vanishes, leaving:

$$E = \overline{w' \rho'_v} \tag{5.4}$$

The turbulence flux is equal to the covariance of the vertical wind velocity and a scalar such as water vapour density. In practice, it is not as simple as it appears. Determination of the “suitable” averaging period, presence of non-stationary conditions, non-zero mean vertical velocities, and other issues, pose challenges to making quality flux measurements. These problems are discussed in Mahrt (1998) and Vickers and Mahrt (1997). Some of these issues are also denoted in Baldocchi et al. (1988).

Bowen Ratio

If the evaporation and sensible heat fluxes are expressed in terms of turbulence diffusivities and gradients, then the ratio of sensible to latent heat flux, or Bowen Ratio, can be approximated as:

$$B = \frac{c_p \cdot \Delta \bar{T}}{L \cdot \Delta \bar{q}} \tag{5.5}$$

Critical assumptions made here include equality of turbulence diffusivities for heat and water vapour, and replacing finite differences for differential values of gradients. The energy balance equation can be used with (5.5) to obtain:

$$E = \frac{R_n - G}{1 + B} \quad (5.6)$$

If measurements of available energy and vertical changes in temperature and humidity are made, E can be calculated. This assumes that available energy can be measured *without error*. For uniform surfaces with large values of vertical gradients, the Bowen Ratio technique works well. However, for heterogeneous surfaces, the assumption of equality in heat and water vapour diffusivities is likely to be violated.

Flux Gradient Approach – Monin–Obukhov Similarity Theory

Monin–Obukhov Similarity theory (MOS) can be used to estimate the vertical profiles of wind speed as well as momentum, heat and water vapour fluxes with only a few parameters. It is based on an assumption that the turbulent transport of a quantity is proportional to the product of the turbulence diffusivity, K , and the vertical gradient in mean concentration C . The height-dependent eddy diffusivity is assumed to be a function of the momentum transport and atmospheric stability. For momentum, heat and water vapour, the gradients are related to the fluxes using similarity parameters. Integrated forms of the resulting expressions have been derived (Brutsaert, 1982).

The fact that stability functions continue to be modified, raises concern about the reliability of using gradient type approaches for estimating fluxes. Large Eddy Simulation (LES) suggests that boundary layer depth has an indirect influence on MOS scaling for wind (Khanna and Brasseur, 1997). Williams and Hacker (1993) show that mixed-layer convective processes influence MOS and support the refinements made by Kader and Yaglom (1990). Clearly there are still considerable uncertainties as to the exact forms of the mean profiles as both surface heterogeneity as well as mixed-layer convective processes affect the idealised MOS profiles.

When surface values of temperature and humidity are determined, only values at one height in the surface layer are needed, along with an estimate of the surface roughness for momentum, z_{Om} , and heat, z_{Oh} , and water z_{Ow} , and surface humidity. For heterogeneous surfaces, z_{Oh} has little physical meaning, but there has been more progress in relating z_{Om} to physical properties of the surface (e.g., Brutsaert, 1982).

5.3.2 Regional Scales

Aircraft-based Eddy Covariance

Aircraft-based flux systems can in theory provide large-area flux estimation both in the surface layer and throughout the ABL. However, in a number of field

programs, the latent and sensible heat fluxes measured by aircraft tend to be smaller than those measured by towers several metres above the surface (Shuttleworth, 1991). Sampling errors for both tower and aircraft-based systems are discussed by Mahrt (1998). Under nonstationary conditions, procedures for estimating sampling errors are invalid. Moreover the flux estimate is sensitive to the choice of averaging length. Vickers and Mahrt (1997) and Mahrt (1998) describe the use of a quantity called the nonstationarity ratio, to define when significant errors may exist in the measurements. Processing of aircraft measurements is considerably more involved than tower data, and collection of the data is quite expensive. However, it is the only method to directly estimate fluxes and their spatial variations at regional scales.

Regional Fluxes and Properties of the ABL

Since the atmospheric boundary layer is connected to surface processes at a regional scale, there must be a relationship between the regional surface fluxes and properties of the ABL. One approach to this issue has been to use a similarity theory for the ABL to estimate fluxes from vertical profiles of wind, temperature, and humidity in the ABL (Sugita and Brutsaert, 1991) measured using soundings from radiosondes.

A different approach presented by Munley and Hippias (1991), Swiatek (1992), and Hippias et al. (1994), related temporal changes in ABL properties to surface fluxes using fundamental governing equations for temperature and humidity. The latter two studies suggested that horizontal advection in the ABL was an important process affecting the ability to recover reasonable surface flux values. When a crude estimate of this process was made, agreement of ABL estimates with measured surface fluxes was reasonably good for two semi-arid ecosystems. However, in the application of this approach over other semi-arid landscapes containing significant variability in surface fluxes, greater discrepancies with flux observations, especially in evaporation, have been found (Kustas et al., 1995; Lhomme et al., 1997). One of the reasons for this scatter is footprint issues.

5.3.3 Footprint Issues

In order to interpret an estimate of a surface flux of mass or energy, one must know from where the flux originated. A source area or region upwind of the surface contributes to a measured flux at a given height. This source area is called the “footprint” and is the area over which measurements are being influenced (see Chapter 2, p. 19) The contribution from each surface element varies according to upwind distance from the location of the measurement, and atmospheric diffusion properties. In order to determine the region associated with a flux value or the footprint, some type of model must be used.

There are two main approaches in footprint models: analytical solutions to the diffusion equation, and Lagrangian models. The analytical approaches derive solutions to the diffusion equation using parameterisations such as similarity theory for turbulence diffusion. There are also other critical assumptions made, such as no spatial variation in the surface flux. This results in equations that require only a few inputs, and are relatively easy to implement. Lagrangian models are more complex and numerically simulate the trajectories of many thousands of individual particles. Knowledge of the turbulence field is needed to allow the trajectories to be computed. When the results of many particle journeys are compiled, the relative contribution of various upwind distances to the flux can be determined. Examples of the analytical category are Schuepp et al. (1990), Horst and Weil (1992), and Schmid (1994). Lagrangian approaches are presented in Leclerc and Thurtell (1990) and Finn et al. (1996).

For heterogeneous surfaces, knowledge of the footprint of any flux measurement is absolutely necessary, in order to interpret spatial variations in fluxes. A current limitation is that present footprint models generally assume a spatially constant flux at the surface. In reality, fluxes will vary in space. The effects of spatial variations in surface properties and fluxes on the resulting footprints remain to be determined, i.e. the measurements represent the bulk effects but we cannot use them to easily define detail of the spatial patterns.

5.4 SPATIAL VARIATIONS OF EVAPORATION

It is of great importance in hydrology to be able to quantify the spatial distribution of evaporation. It certainly has some connections to the traditional hydrologic outputs at the catchment scale, such as streamflow. However, the spatial distribution of water balance, especially at larger scales, has strong connections with the atmospheric conditions and hydroclimatology of a region. Qualitatively, the important surface properties that relate to spatial variations in evaporation are understood rather well. Spatial changes in water balance are connected to those of the root zone soil moisture, vegetation density, stomatal conductance, net radiation, saturation deficit, and turbulence intensity.

There have been some advances in determination of spatial fields of some of the above properties using remote sensing information. In particular, net radiation, *surface* soil moisture, and vegetation density can be estimated spatially with remote sensing and auxiliary data (Kustas and Humes, 1996; Carlson et al., 1994).

We can define several issues that pose difficulties in assessing the spatial patterns in water balance, including difficulties associated with the definition and description of heterogeneous surfaces, and the effects of such surfaces on fluxes and the aggregation of fluxes over the landscape. These must be resolved in order to develop the ability to quantify spatial variations in the surface fluxes.

5.5 DIFFICULTIES POSED BY HETEROGENEOUS SURFACES

When surfaces are heterogeneous, several issues arise. First, most models and measurement approaches either explicitly or implicitly assume a uniform surface. Second, the spatial variability in critical properties can cause nonlinear processes to become important.

5.5.1 The Notion of Heterogeneity

Heterogeneity is a rather descriptive term, and is often used somewhat ambiguously. Unfortunately, there is at present no universal approach to quantify the degree of heterogeneity. This is partly because the importance or effects of nonuniformity seem to depend upon the process that is being considered. The difficulty in quantifying what we mean by heterogeneity is indicative of the complexity of the entire issue of water and energy balance of inhomogeneous surfaces. Here we discuss some of the recent approaches to this problem.

Heterogeneity exists at all spatial scales, from variations within individual leaves (Monteith and Unsworth, 1990), to the canopy level where evaporation and sensible heat may originate from significantly different sources (Shuttleworth and Wallace, 1985), to larger scales where nonuniformity can affect atmospheric flow (Giorgi and Avissar, 1997). Besides scale, the type of heterogeneity may also be important. For example, de Bruin et al. (1991) showed that variations in temperature and humidity fields have a different effect on Monin–Obukhov similarity than variations in the wind field.

For purposes of estimating evaporation either directly via measurement of eddies, or indirectly using flux–gradient relationships, heterogeneity at the canopy scale and larger is of primary concern. At smaller scales, physically-based methods which consider both biological and fluid dynamics have been developed for scaling from the leaf to canopy scale (Norman, 1993; Baldocchi, 1993). However, they can be quite complicated and may only be applicable under ideal conditions, such as a canopy that is horizontally homogeneous (Baldocchi, 1993). The issue is how to define when the surface can no longer be treated as homogeneous.

5.5.2 Determining when a Surface is Heterogeneous

No exact methodology or theory exists to determine a priori when a surface can no longer be considered uniform. Measurement of turbulent fluxes and statistics is one indirect method, where deviation of the Monin–Obukhov similarity functions from those determined over uniform surfaces has been shown to be an indicator of heterogeneity (e.g., Chen, 1990a,b; de Bruin et al., 1991; Roth and Oke, 1995; Katul et al., 1995). Similarity theory requires the correlation between temperature and humidity to be near unity. This is not true for non-uniform surfaces (Katul et al., 1995; Roth and Oke, 1995), due to the source and/

or sink of evaporation differing from that of sensible heat flux. Unfortunately, these approaches do not provide a measure of the degree of heterogeneity.

Remote sensing may hold potential as a means of quantifying surface spatial variability by calculating spatial power spectra for surface radiance or reflectance values (Hips et al., 1996). This requires pixel resolution fine enough to discriminate between plant and soil, which is often not available from satellites. Moreover, the shape of spatial power spectra depends upon the spatial resolution of the surface data (Hips et al., 1995). This brings forward a critical issue. The degree of heterogeneity or spatial variability may be dependent upon the spatial resolution at which the surface is observed (see Chapter 2, p. 19).

Another indirect approach suggested by Blyth and Harding (1995) uses remotely sensed surface temperature along with wind and temperature profiles in the surface layer, to derive the roughness lengths of heat and momentum. The relationship between these values is related to heterogeneity of the surface. Both theory and observations indicate that transfer of momentum is more efficient than heat (Brutsaert, 1982). For homogeneous surfaces the ratio of roughness length for momentum, z_{OM} , and heat, z_{OH} , is essentially a constant, usually expressed as the natural logarithm $\ln(z_{OM}/z_{OH}) = kB^{-1}$ where $kB^{-1} \sim 2$. Many studies, especially for partial canopy cover surfaces, have found kB^{-1} significantly larger than 2 with values generally falling between permeable-rough, $kB^{-1} \sim 2$, and bluff-rough, $kB^{-1} \sim 10$ (Verhoef et al., 1997). So the ratio of the roughness lengths is an indirect indicator of the degree of departure from a uniform surface. This result is caused by several factors which include effects of the soil/substrate on the remotely sensed surface temperature observation, canopy architecture and the amount of cover (McNaughton and Van den Hurk, 1995).

5.5.3 Application of Single and Dual-source Approaches to Heterogeneous Surfaces

There is a fundamental problem in representing a heterogeneous surface as a single layer or source, which is implicit in the application of, for example, the Penman–Monteith equation, because of the significant influence of the soil/substrate on the total surface energy balance. Thus, the surface resistance to evaporation has lost physical meaning because it represents an unknown combination of stomatal resistance of the vegetation and resistance to soil evaporation (Blyth and Harding, 1995). This has prompted the development of two-source approaches, whereby the energy exchanges of the soil/substrate and vegetation are evaluated separately (e.g., Shuttleworth and Wallace, 1985). Nevertheless, some studies reported the Penman–Monteith equation to be useful for evaporation estimation over heterogeneous surfaces (e.g., Stewart and Verma, 1992; Huntingford et al., 1995). In fact Huntingford et al. (1995) found little difference in performance of two-source approaches versus the Penman–Monteith for a Sahelian savanna. However, these studies arrive at reliable evaporation estimates only after the stomatal response functions are opti-

mised with the measurements from the particular site. Therefore, as a predictive tool, the Penman–Monteith approach will be tenuous for heterogeneous surfaces without a priori calibration. By performing such a priori calibration, much simpler formulations such as the Priestley–Taylor equation can yield evaporation predictions similar to two-source approaches for heterogeneous surfaces (Stannard, 1993).

5.5.4 Application of Surface-layer Similarity above Heterogeneous Surfaces

For several decades Monin–Obukhov Similarity (MOS) theory has been used to relate mean profiles of scalars and wind to the turbulent fluxes of heat and momentum (Brutsaert, 1982; Stull, 1988). However, serious limitations exist in the application close to the canopy due to roughness sublayer effects (e.g., Garratt, 1978, 1980). For heterogeneous surfaces we are presently unable to resolve the relative influence of all the mechanisms involved, and more importantly have been unable to develop a unified theory to correct MOS for effect of the roughness sublayer on mean profiles and turbulent statistics (Roth and Oke, 1995).

An example of the effect of heterogeneity on MOS profiles is shown in Figure 5.1 for a desert site containing coppice dunes and mesquite vegetation (Kustas et al., 1998). In Figure 5.1 d_0 is the zero plane displacement. This is a length to account for the fact that in tall vegetation, the source and sinks are above the ground surface, so the heights are specified as distances above a new reference value which makes the relationship between fluxes and gradients valid. While the roughness sublayer does not appear to affect the wind profile, the actual temperature profile departs significantly from the idealised MOS predicted profile. This is probably due in part to the complicated source/sink distribution of heat (Coppin et al., 1986). Over this site, the heat sources are the interdune regions and heat sinks are mesquite vegetation randomly distributed over the surface. As a result, significant scatter between predicted and measured heat fluxes has been reported using the above MOS equations (Kustas et al., 1998).

5.5.5 Effects of Heterogeneity on Surface Fluxes and Aggregation

As mentioned, determination of the spatial distribution of the critical surface properties that relate to evaporation is becoming possible at many scales with advances in remote sensing. However, there are issues about how to properly determine and interpret variables of interest from remote sensing data. For example, the interpretation of radiometric temperature in terms of the heat flux process is far from simple (Norman and Becker, 1995). Remote sensing estimates of vegetation are subject to variations in density and geometry. Only

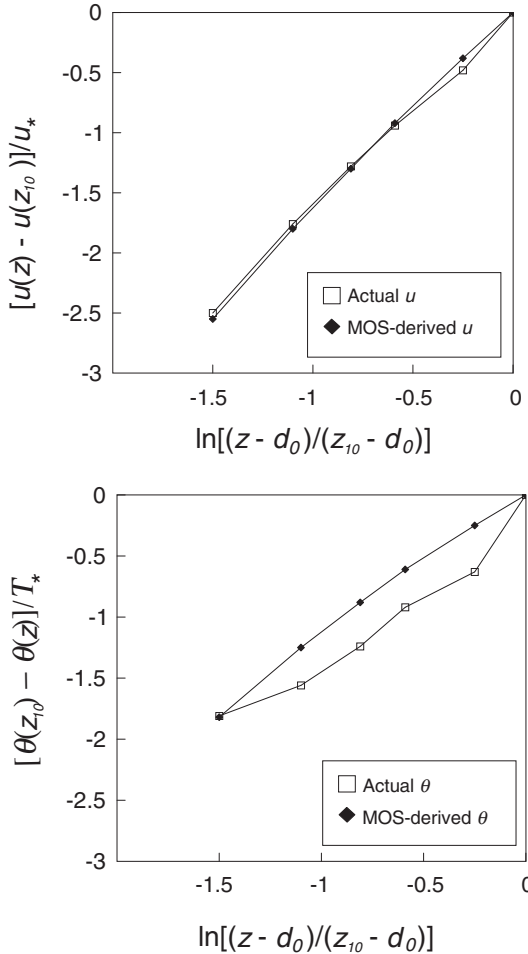


Figure 5.1. Plots of normalised wind $[u(z) - u(z_{10})]/u_*$ and temperature $[\theta(z_{10}) - \theta(z)]/T_*$ versus $\ln[(z - d_0)/(z_{10} - d_0)]$, with u_* and T_* estimates from the eddy covariance measurements. Actual versus MOS-derived normalised profiles of u and θ representing an average of all unstable profiles (see Kustas et al., 1998).

upper soil moisture can be estimated by remote sensing, while plants respond to water in the entire root zone.

In order to model the fluxes, the actual patches of surface types must be delimited. Identifying various patches is not trivial, as it requires determination of the properties that are of hydrological importance, as well as the magnitude of spatial changes which are significant. Also, the scales of heterogeneity must be determined so that the models can be implemented at commensurate spatial scales, i.e. the characteristic scale of the process must match the modelling scale (see Chapter 2, p. 27).

However, even if there were complete knowledge of the distribution of the critical biophysical properties of the surface, there are other issues to be addressed. At some scales of heterogeneity, nonlinear effects may become important. For example, the properties and processes at one surface may affect those of a nearby surface. Several examples can be posed here. Significant spatial changes in surface water balance, common in semi-arid regions, result

in transport by the mean wind of heat and saturation deficit from drier to wetter surfaces. This can enhance the evaporation and alter the energy and water balance of the latter surfaces. This effect of advection on evaporation is detailed in Zermeño-Gonzalez and Hipps (1997). In addition, Avissar (1998) has shown results with mesoscale models that suggest secondary circulations can form between warm and cool adjacent patches. These may carry significant vertical fluxes of mass and energy, which will not be reflected in local measurements of turbulence transport, nor accounted for in models treating each spatial surface element independently.

Finally, the fluxes and governing properties do not both aggregate linearly. The actual surface fluxes can be added linearly (the flux from each spatial element can be summed, and normalised to yield average flux). However, the spatial averages of the critical properties when input into the flux equation, do not yield the correct value for the average flux (see the discussion on effective parameters in Chapter 3, p. 68). Since, we generally have available, at best, the spatial distribution of the surface properties, the aggregation up to larger regions is a problem.

Ultimately, the above factors create difficulties in properly aggregating the fluxes up to larger regions. This so-called aggregation problem remains unsolved in a general way at present. However, remote sensing may provide spatially distributed hydrologic information critical in addressing scaling issues (Beven and Fisher, 1996). There are several directions which have been posed. These include the determination of effective parameters for surface properties (Lhomme et al., 1994), and treating surface properties as probability density functions, and inputting them into mesoscale atmospheric models (Avissar, 1995). We do not directly address this issue here, but simply note that the spatial distribution of evaporation and the aggregation problem are ultimately connected.

In the meantime there have been attempts to estimate spatial patterns of evaporation using a combination of modelling and remotely sensed information. As a result of the problems discussed above, these methods can be used only under restrictive assumptions and require data that is not commonly available. Nevertheless, they provide a way forward.

5.6 EXAMPLES OF ESTIMATING SPATIAL VARIATIONS OF EVAPORATION

Surface energy balance models using remotely sensed data have been developed and used in generating spatially distributed evaporation maps (Kustas and Norman, 1996). For many of these models, surface temperature serves as a primary boundary condition (e.g., Bastiaanssen et al., 1998). Clearly, the spatial variation of surface temperature is not enough to estimate the variation in evaporation since the amount of vegetative cover, water deficit conditions, and aerodynamic roughness strongly influence the turbulent transport and thus the aerodynamic–radiometric temperature relationship (Norman et al., 1995).

Promising approaches described below, explicitly evaluate flux and temperature contributions from the soil and vegetation using the conceptual modelling

philosophy of Shuttleworth and Wallace (1985). The modelling strategy is to consider the Penman–Monteith type of approach strictly for the vegetated fraction, and a similar resistance type analogue for the soil component (i.e. a two-source approach). In this case, the vapour pressure gradient term is not linearised as in equation (5.1), but is a function of the vegetation and soil temperatures which is derived from remotely sensed observations of canopy cover and surface temperatures and model inversion. Along similar lines, the approach of Norman et al. (1995) uses the Priestley–Taylor approximation for the vegetated component only, but with the extension that the alpha value can approach zero (i.e., no transpiration). This is necessary since the model is constrained by both the energy balance and radiative temperature balance between model-derived component temperatures and the remotely sensed surface temperature observations.

While the above formulations address the issue of aerodynamic-radiometric temperature relationships, determining spatially distributed heat fluxes at regional scales will invariably require incorporating surface–atmospheric feedback processes. Several approaches have made significant progress in this area. Following Price (1990), Carlson et al. (1990, 1994) combined an ABL model with a soil–vegetation–atmosphere–transfer (SVAT) scheme for mapping surface soil moisture, vegetation cover and surface fluxes based on a fundamental relationship between vegetation index (i.e., cover) and surface temperature. Using ancillary data (including a morning sounding, vegetation and soil type information), root-zone and surface soil moisture are varied, respectively, until the modelled and measured surface temperatures are closely matched for both 100% vegetative cover and bare soil conditions. Further refinements to this technique have been developed by Gillies and Carlson (1995), for potential incorporation into climate models. Comparisons between model-derived fluxes and observations have been made by Gillies et al. (1997) using high resolution aircraft-based remote sensing measurements. Approximately 90% of the variance in the fluxes was captured by the model for the conditions of their study.

The Two-Source Time-Integrated model of Anderson et al. (1997) (presently called ALEXI), provides a practical algorithm for using a combination of satellite data, synoptic weather data and ancillary information to map surface flux components on a continental scale (Mecikalski et al., 1999). The ALEXI approach builds on the earlier work with the Two-Source model (Norman et al., 1995) by using remote brightness temperature observations at two times in the morning hours, and considering planetary boundary layer processes. The methodology removes the need for a measurement of near-surface air temperature and is relatively insensitive to uncertainties in surface thermal emissivity and atmospheric corrections on the GOES brightness temperature measurements. Anderson et al. (1997) and Mecikalski et al. (1999) have shown that surface fluxes retrieved from the ALEXI approach compare well with measurements, albeit under some restrictive assumptions. The ALEXI approach is a practical means to operational estimates of surface fluxes over continental scales with 5–10

km pixel resolution. It also connects the surface properties and processes with the development of the atmospheric boundary layer, which is necessary to realistically describe the system.

A relatively simple two-source model using the framework described by Norman et al. (1995) has been used to generate surface flux maps (Kustas and Humes, 1996; Schmugge et al., 1998). The model was designed to use input data primarily from satellite observations. Several simplifying assumptions about energy partitioning between the soil and vegetation reduce both computational time and input data required to characterise surface properties. The inputs include an estimate of fractional vegetative cover, canopy height, leaf width, surface temperature, solar radiation, wind speed and air temperature. The remote sensing data from the Monsoon '90 experiment (Kustas and Goodrich, 1994), conducted in a semi-arid rangeland catchment in Arizona, have been used to evaluate the model. An example of an evaporation map generated from the two-source model is shown in Figure 5.2. A Landsat-5 TM image was used to generate a fractional vegetative cover and land use map for deriving vegetative height and roughness. A network of surface flux stations (approximate locations displayed as discs in the figure) provided spatially distributed solar radiation, wind and air temperature observations (Kustas and Humes, 1996). Aircraft surface temperature observations for a day with the largest variation in moisture conditions were used. The pixel resolution is 120 m, similar to the resolution of Landsat TM thermal band. The calculated latent heat flux field shows a wide range in values from about 50 to nearly 500 W m⁻². This variation is due in part to a recent precipitation gradient over the study area, with essentially no rainfall occurring in the western quarter of the image and gradually increasing to significant amounts in the north-eastern portion (Humes et al., 1997). In addition, the model computes higher evaporation rates for the areas along the ephemeral channels (the green and blue stripes) which contain more and taller vegetative cover, since there is typically more available water in these areas.

Comparison of model versus observed half-hourly latent heat flux from the flux measurement sites is illustrated in Figure 5.2 (values in W m⁻²). There is qualitative agreement between model and observed fluxes (i.e., higher observed latent heat fluxes are in areas with higher modelled fluxes). However, it is not straightforward to determine how to weight the pixels within the source footprint of the observations. Note that patches with the highest and lowest latent heat fluxes were not within the observation network. This makes it difficult to validate regional flux models with a network of local flux measurements in heterogeneous regions (Kustas et al., 1995). Several pixels surrounding the eight surface flux stations were averaged for three days in which soil moisture conditions were different. The comparison between model and observed latent heat fluxes is illustrated in Figure 5.3. A standard error of approximately 30 W m⁻² and $R^2 = 0.8$ is obtained. These are similar to the results found in the other modelling studies described above.

These examples illustrate that, despite the conceptual problems identified earlier in the chapter, we have made progress towards methods for estimating spatial

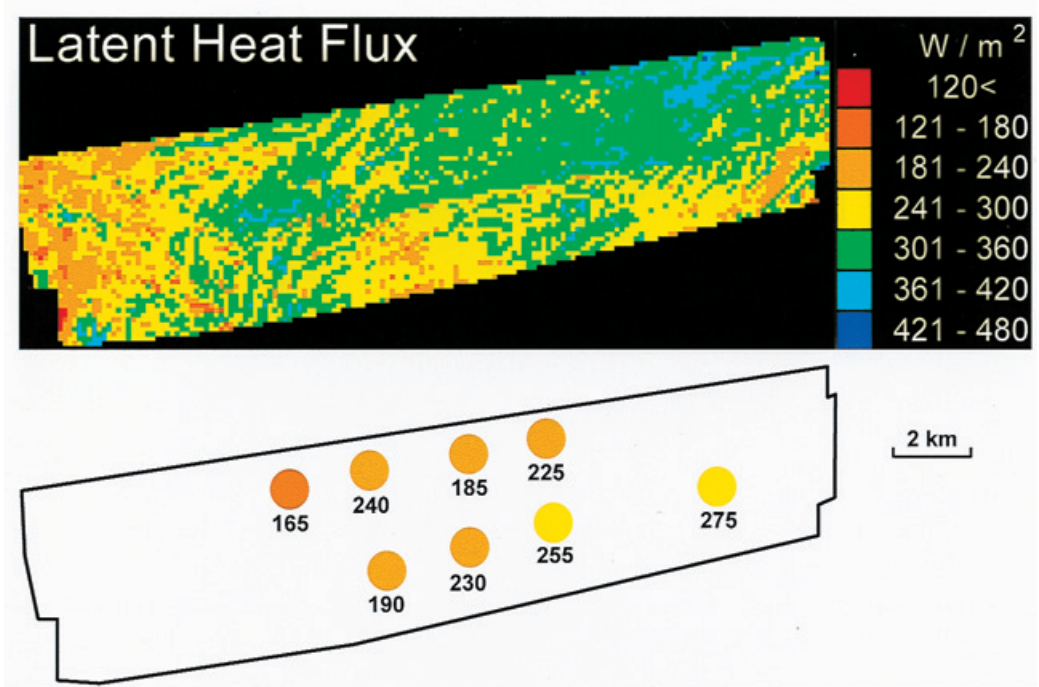


Figure 5.2. Evaporation image created from remote sensing data collected during Monsoon '90 used in a simple two-source model described in Norman et al. (1995) and estimates of evaporation from metflux stations (discs). Note that the size of the discs does not represent the measurement area. See also Kustas and Humes (1996).

variations in evaporation. Presently, these are applicable only under special circumstances, requiring detailed remote sensing data, cloud-free conditions, some limiting assumptions related to the “footprint” problem, and provide only a snapshot view of spatial variations.

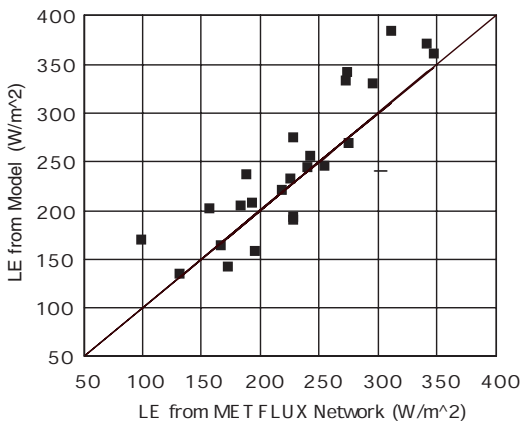


Figure 5.3. Comparison of two-source model-derived LE versus LE observations from the METFLUX network for three days of aircraft remote sensing observations during the Monsoon '90 experiment. See Kustas and Humes (1996) and Schmugge et al. (1998) for details.

5.7 CURRENT FRONTIERS IN EVAPORATION RESEARCH

There are several problems that presently limit our abilities to examine and model spatial variations in evaporation. These include capabilities of making accurate measurements of critical processes over appropriate scales, as well as missing theoretical knowledge about processes and scaling issues.

5.7.1 Measurement Issues

Available Energy

Ultimately, the energy and water balances are inextricably connected. When we consider spatial distribution of fluxes, it is necessary to measure or estimate available energy at various spatial scales. This remains a serious difficulty. Remote sensing information offers promise to allow estimates of spatially distributed net radiation (Diak et al., 1998). However, soil heat flux remains a more serious difficulty, especially for heterogeneous surfaces. In such cases, measurements of spatial averages are nearly impossible, as the number of sites required is likely prohibitive. There are some studies that have related the ratio of G/R_n to remotely sensed radiance indices (Kustas and Daughtry, 1990) and some analytical treatment of this issue (Kustas et al., 1993). However, there is as yet no general solution to this problem.

Longer Timescale Estimates Covering Seasonal and Yearly Trends

There are relatively few studies that have produced a good set of spatially distributed flux measurements to validate models. In addition, these have been generally conducted over rather short time periods, for a variety of reasons. We need to examine the seasonal changes in the fluxes themselves, as well as properties and processes that connect to evaporation and water balance at catchment scales. Little such information is presently available. Some attention is needed to acquiring more data at sites over a number of seasons.

5.7.2 Modelling Issues

Aggregation

Earlier, we briefly addressed the complex issue of aggregation, or how to scale processes and fluxes over a range of spatial scales. Because of the depth and complexity of the subject, we did not cover it in detail. Ultimately specifying spatial variations in evaporation and water balance and their implications to climate will be predicated upon reaching an adequate solution to the scaling or aggregation problem. Currently we appear to be missing fundamental ideas to allow a general theoretical solution to the problem. The atmospheric modelling community involved in Soil–Vegetation–Atmosphere Transfer (SVAT) schemes is starting to recognise the potential of remote sensing information in addressing scaling and aggregation issues in hydrology and meteorology (Avisar, 1998). Preliminary studies using remote sensing data with SVAT schemes indicate the

effects of using aggregated information on large-scale evaporation estimates is relatively minor (e.g., Sellers et al., 1995; Kustas and Humes, 1996; Friedl, 1997). This result, however, depends on the scale of heterogeneity (Giorgi and Avissar, 1997) and on the sensitivity of the model parameterisations to surface properties affecting evaporation (Famiglietti and Wood, 1995). We still lack the knowledge to make any general conclusions about these issues.

Combining Surface–Atmospheric Interaction with Remote Sensing Approaches

Earlier, we pointed out current research efforts attempting to merge ABL models with SVAT schemes. The reason for doing this is that wind, temperature and humidity profiles within the fully turbulent region of ABL (i.e., mixed layer) relate to surface fluxes integrated upwind having length scales several orders of magnitude larger than the ABL depth. With ABL depth, typically on the order of 1 km during daytime convective conditions, the wind and scalar quantities should reflect integrated values of surface heterogeneities roughly 10 km upwind. Therefore, by combining spatially variable information on vegetation cover and type and surface temperature from remote sensing with ABL processes, there is the potential of creating the appropriate links between spatially variable surface fluxes and atmospheric feedbacks. The three examples discussed in Section 5.6 demonstrate possibilities of such an approach. They also indicate the issues involved in linking the ABL, SVAT models, and remote sensing data to represent heterogeneous surfaces. There are still processes not yet expressed in these approaches, such as local or mesoscale advection effects.

5.7.3 Conclusions

As our understanding of hydrology and climate has advanced, the importance of evaporation and its spatial distribution has become more evident. Although there is a wealth of theoretical and measurement information available about evaporation, most of it is confined to rather uniform surfaces, and small spatial scales. Even in these cases, all is not yet known.

The current issues in surface hydrology and climate demand attention to spatial and temporal distributions of evaporation at a range of scales. The feedbacks between the evaporation at the surface and atmospheric processes and circulations are often intricate, and cannot be generally ignored. Inevitably this involves dealing with heterogeneous surfaces, which at best stretch the limits of many of our current approaches. However, the advent of remote sensing information offers to make available the spatial variations of several critical surface properties. The key is how to properly connect this information to the actual fluxes. At this stage we have relatively few cases available where these issues can be carefully examined on the landscape, but clearly some real progress has been made in this issue.