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Putting the ‘vap’ into evaporation

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The scientific progress reviewed in this paper took place over a period which corresponds exactly to the full lifetime of my beloved son, Matthew James Shuttleworth, to whom this paper is dedicated.

Abstract

In the spirit of the Special Issue of HESS to which it contributes, this paper documents the origin and development of the science of natural evaporation from land surfaces over the last 30–35 years, since the symposium *A View from the Watershed* was held to commemorate the opening of the new Institute of Hydrology (IH) building in 1973. Important subsequent technical progress includes the ability to measure routinely the diurnal cycle of near-surface meteorological variables using automatic weather stations, and of surface energy and momentum exchanges using automated implementations of the Bowen Ratio/Energy Budget technique and the Eddy Correlation technique, along with the capability to estimate the ‘fetch’ for which these measurements apply. These improvements have been complemented by new methods to measure the separate components of evaporation, including: the interception process using randomly relocated below-canopy gauges, transpiration fluxes from individual leaves/shoots using porometers and from plants/plant components using stem-flow gauges and soil evaporation using micro-lysimeters and soil moisture depletion methods. In recent years progress has been made in making theory-based area-average estimates of evaporation using scintillometers, and model-based area-average estimates by assembling many streams of relevant data into Land Data Assimilation Systems. Theoretical progress has been made in extending near-surface turbulence theory to accommodate the effect of the ‘excess’ boundary layer resistance to leaf-to-air transfer of energy and mass fluxes relative to that for momentum, and to allow for observed shortcoming in stability factors in the transition layer immediately above vegetation. Controversy regarding the relative merits of multi-layer model and ‘big leaf’ representations of whole-canopy exchanges has been resolved in favour of the latter approach. Important gaps in the theory of canopy-atmosphere interactions have been filled, including recognising the need, separately, to represent dry-canopy and wet-canopy evaporation in models and the capability to describe wet-to-dry canopy transitions as well as the ability to describe sparse vegetation canopies which only partly cover the underlying soil. There is progress in methods of estimating crop water requirements, but an important recommendation of this paper is that this progress should continue by introducing use of an effective stomatal resistance rather than crop factors. The paper draws attention to relevant theoretical insight on this issue. Progress in theoretical understanding of evaporation processes has been used in the creation of numerous Land Surface Parameterisations (LSPs), the models used to represent land-surface interaction in climate and weather forecast models, and there have been important advances in describing the behaviour of plant stomata in LSPs. A major investment over the last 25 years in conducting Large-Scale Field Experiments, the better to measure, understand and model coupled land-surface/atmosphere interactions, has resulted in improvements in the capabilities of global climate models and the ability of mesoscale meteorological models to describe the enhanced circulation resulting from different forms of land-surface heterogeneity. Progress in understanding why early equations for potential evapotranspiration can be adequate in certain conditions is reviewed. The paper concludes with recommendations for future research.

Keywords: evaporation, evapotranspiration, surface energy balance, land-atmosphere interactions, land surface parameterisations

Introduction

Hydrological understanding has advanced substantially in the decades since the symposium *A View from the Watershed* commemorated the opening of the new Institute of Hydrology (IH) building at Wallingford, UK, in June 1973 but, arguably, the ‘view from the watershed’ of the process

of natural evaporation has improved more than any other facet of hydrology. In many ways, this change reflects the growth and interdisciplinary diversification of the institution itself, from IH to the Centre for Ecology and Hydrology (CEH). This paper documents and describes the development of the science of natural evaporation from land

surfaces over the last 30–35 years and highlights the important steps along the way in terms of new measurement technologies and important new process understanding and modelling methods. It is the author's opinion that much of the hydrological community's now-increased ability to measure, understand and model the natural evaporation process had its origin in technologies and ideas that first arose in the 1970s and 1980s, when Jim McCulloch held court as the dour Scottish baron reigning over the UK's hydrological enterprise. Documenting the influence of IH and CEH on advances in hydrological science is, therefore, a secondary objective of this paper but it does emerge naturally as the inevitable consequence of describing progress in this field.

Where were we then?

Back in the early 1970s, where did the science of natural evaporation stand? In terms of instrumentation and measurement techniques, past information on climate was available largely in the form of paper records of measurements made with comparatively simple instruments read manually, typically only once a day. Often, measurements comprised maximum and minimum temperature and daily total precipitation, together with sample measurements of atmospheric humidity made with wet and dry bulb thermometers or a hygrometer utilising a human hair, and perhaps wind speed, generally measured as the daily total 'wind run'. Sometimes an indirect measure of solar radiation was added, perhaps as an estimate of cloud cover or the number of sunshine hours recorded, for instance, by a Campbell-Stokes radiometer or a Gunn-Bellani distilometer.

Within the hydrological community, measurements of evaporation were still deduced indirectly from a water balance, most commonly from an evaporation pan, but, in research situations, a lysimeter and, occasionally, a well-instrumented catchment. However, the science of micrometeorology was by then recognised and grew rapidly, driven primarily by the agricultural science community, and rapid development of associated instrumentation included devices to measure fluxes of radiant energy and the evaporation from individual leaves (porometers). Profiles of meteorological sensors on short towers over short agricultural crops were often deployed to quantify aerodynamic exchanges and define components of the surface energy balance using the *Bowen Ratio/Energy Budget* technique.

In terms of a theoretical understanding and modelling of the evaporation process, the scene was dominated largely by the long-established concept of a hypothetical 'potential evaporation rate', determined from meteorological

measurements and variously described but basically thought of as an assumed maximum rate of evaporation from natural surfaces where water was not limiting. The common use of the term 'evapotranspiration' (instead of evaporation) among hydrologists encouraged the belief that natural evaporation was a comparatively simple physically-determined surface-atmosphere exchange, with only a limited relationship to the vegetation covering the land surface. For several decades previously, the simultaneous availability of climate data and estimates of evaporation from evaporation pans, lysimeters, and sometimes catchments, had allowed hydrological researchers to seek relationships between the two in pursuit of a preferred means for estimating the potential evaporation rate. Several of the proposed relationships were essentially empirical (e.g. Wilm *et al.*, 1944; Thornthwaite, 1948; Blaney and Criddle, 1950; Hargreaves, 1975) but the equation of Penman (1948, 1963), which is still in several respects empirical but has a more robust physical basis, was most popular in the UK and in countries exposed to the influence of UK scientists during the colonial era. Re-written in metric units (Shuttleworth, 1993a), the Penman Equation takes the form:

$$E_{Penman} = \frac{\Delta A' + \gamma [6.43(1 + 0.536 U_m) D]}{\Delta + \gamma} / \lambda \quad (\text{mm day}^{-1}) \quad (1)$$

where λ ($= 2.501 - 0.002361.T$) is the latent heat of vaporisation of water (in MJ kg⁻¹), with T the temperature in °C; Δ is the rate of change saturated vapour pressure (in kPa °C⁻¹) at air temperature; A' is the measured or estimated energy available for evaporation from the free water surface expressed as an evaporated water equivalent (in mm day⁻¹); U_m is the wind speed (in m s⁻¹); D is the vapour pressure deficit (in kPa) measured at 2 m; and γ is the psychrometric 'constant' ($= 0.0016286 P/\lambda$, in kPa °C⁻¹), where P is the atmospheric pressure (in kPa). (Note: it is convenient that when A' is in MJ m⁻² day⁻¹ the numerical value of (A'/λ) gives the equivalent water depth in mm day⁻¹ because the density of water is approximately equal to 1000 kg m⁻³). In the early 1970s a simpler form of this equation introduced by Priestley and Taylor (1972) was gaining recognition. In the Priestley and Taylor equation, the second *advection term* in the Penman equation was replaced by a fixed fraction of the first *radiation term*, thus:

$$\lambda E_{PT} = 1.26 \frac{\Delta A'}{\Delta + \gamma} \quad (\text{mm day}^{-1}) \quad (2)$$

However, there were already several clues that the natural evaporation process may not be quite as simple as was then thought, and that the nature of the land surface may have

influence. The Penman-Monteith equation (Monteith, 1965) which explicitly defines a dependence of transpiration on vegetation was increasingly recognised and accepted, and was already being used in *diagnostic* mode to quantify the influence of plants through a so-called surface resistance, but was not yet used in *predictive* mode to estimate transpiration. At least as important, especially in the context of the initial research thrust at IH, previous observations by Law (1956, 1958) suggested that extreme changes in vegetation cover (from forest to pasture, or *vice versa*) could alter the evaporation rate at a particular location with a particular climate; because of this, the water resources available as runoff might be altered dramatically. Concern that the estimation of available water resources might be flawed by imperfect understanding of the natural evaporation process, set the scene for a major investment in research, in developing instrumentation and establishing field methods and experimental sites. Prime Minister Harold Wilson's belief in the 'white hot heat of technological revolution' was in vogue in the UK and, with McCulloch directing the show, IH took centre stage.

Steps along the way

Progress in science occurs as interplay between new measurements, which reveal weaknesses in existing ideas or allow the testing of new ideas, and innovative thought, which results in the creation of new hypotheses that can be tested as the means for developing understanding. Over the last 30–35 years, progress in land-surface evaporation science has involved both the creation of new measurement methods and the emergence of new ideas and understanding of near-surface processes; together these led to a recognition that evaporation is not just a land-surface process but a coupled process that involves feedback between the surface and the overlying atmosphere. The following sections overview the most important developments that have led to present-day understanding of the science of natural evaporation since the original *A View from the Watershed* symposium.

MEASUREMENT METHODS AND INSTRUMENTS

Routine measurement of weather

The established relevance and value of routine measurements of weather variables for hydrology, in general, and for understanding and estimating evaporation in particular, ensured that investments in seeking their routine, automated measurement using robust instruments was worthwhile. IH was a leader in automatic weather station development (McCulloch and Strangeways, 1966;

Strangeways, 1972) and new opportunities for field studies were the direct result. Creating the capability to make such measurements without the need for a human observer other than to provide occasional service visits meant that it became possible to monitor climate in remote locations where regular manual data capture was impossible (e.g. Curran *et al.*, 1977). Just as important, routine data logging meant that samples could be taken regularly and frequently to monitor the daily cycle of measured variables. The ability to record the daily cycle and to identify periods with precipitation subsequently proved crucial when the component processes involved in natural evaporation were studied in separate detail.

Stimulated by the still-developing science of micrometeorology, the development of robust sensors to measure weather variables was rapid in the late 1960s and early 1970s. In particular, the inclusion of sensors of the radiant energy reaching the earth's surface (i.e. solarimeters and net radiometers) among the suite of instruments routinely monitored subsequently proved extremely important. It enabled and stimulated an increasing acceptance among hydrologists that the process of natural evaporation is perhaps best considered as the partition of the energy that is available at the land surface between evaporation and warming the overlying atmosphere, rather than as merely a residual water loss in the catchment water balance. Although progress was arguably most rapid at that time, there has been steady and worthwhile incremental progress in sensor development since, particularly with respect to humidity sensors. Meanwhile, building on the early breakthroughs using magnetic tape systems at IH, progress in data capture technology has advanced dramatically over the last two decades with the advent of high-capacity memory chips, and IH's pioneering efforts in remote data capture (e.g. Strangeways and Lisoni, 1973) have blossomed through the use of wireless telephony and satellite data capture.

Above-canopy profile measurements

The (now largely superseded) *Aerodynamic Method* and (still current) *Bowen Ratio/Energy Budget Methods* hitherto used by micrometeorologists to measure evaporation over short agricultural crops (e.g. Tanner, 1960; Denmead and McIlroy, 1970) had required only modest investment in infrastructure. It required an act of faith and a step change in the investment of finance, time and effort to extend their use to forest vegetation. Among the first pioneering studies of forest micrometeorology, the 'Thetford Project' (Stewart, 1971; Stewart and Thom, 1973; Oliver, 1975) was particularly important, because it took measurements over a very extensive level fetch of uniform mature forest that

had been planted in a comparatively dry part of the UK where rainfall is much less than in the more mountainous western UK, where most forests grow. The use of this particular site was significant because the combination of a near-perfect micrometeorological site together with the fact that, relative to the aerodynamic roughness length of the vegetation, measurements were being made very near to the underlying canopy, allowed stringent testing of the then-current theory of near-surface turbulent transfer, as discussed below. Obtaining evaporation data at the Thetford site was also important because evaporation measurements were thus available for similar forest stands in both dry and wet climates to help distinguish the separate contributions to overall evaporation (transpiration and interception) and demonstrate the need for separate modelling of these components.

Implementing the Thetford Project required technical innovations, including new methods for mounting and levelling instruments on tall towers, and coping with the challenge of measuring the very small gradients of temperature and humidity that occur over forests as a result of the efficient turbulent transfer. Ultimately this last challenge resulted in adopting a new approach to measuring gradients (Black and McNaughton, 1971; McNeil and Shuttleworth, 1975), which involved the regular mechanical interchange of sensors between different heights to reduce the effect of systematic offset errors in the sensor output (Fig. 1). Subsequently this approach became a standard method for measuring the Bowen ratio (e.g. Tanner *et al.*, 1987; Kanemasu *et al.*, 1992), along with a modification of it that involves ducting samples of air from different heights alternately to a common humidity sensor (e.g. Smith *et al.*, 1992; see also <http://www.campbellsci.com/bowen-ratio>). The technical difficulties involved in applying the Bowen Ratio/Energy Budget to measure forest evaporation also stimulated an early evaluation of alternative methods such as the eddy correlation technique (McNeil and Shuttleworth, 1975), and this subsequently led to further development of the eddy correlation hardware which allowed its routine application, as described below.

Measurement of canopy processes

Motivated by the observations of Law (1956, 1958) and Rutter (1967), new emphasis was given to measuring the separate components of evaporation for forest canopies at IH in the 1970s and, in particular, the measurement of that portion of precipitation that is intercepted by the forest canopy and quickly re-evaporated. A whole new branch of hydrology concerned with techniques to measure and model this 'interception loss' has grown out of these pioneering studies in the decades since then. Interception loss is

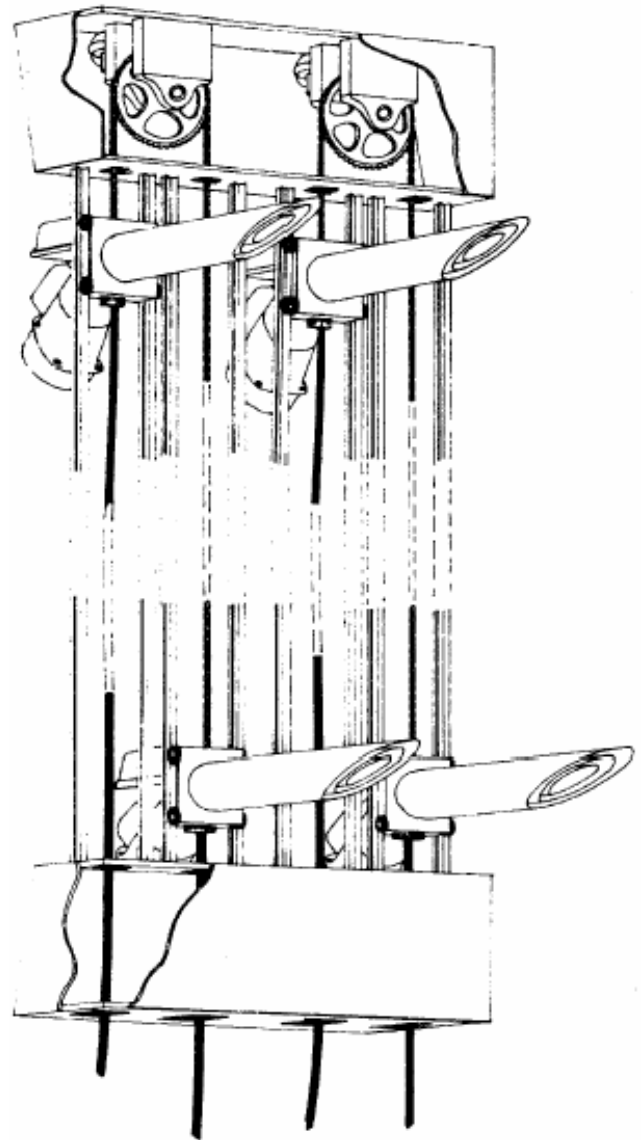


Fig. 1. Schematic diagram of the Thermometer Interchange System, which determines the Bowen ratio by measuring the difference in temperature and humidity at two heights more accurately than a static system because systematic offsets errors in the pairs of sensors are interchanged. [Taken from McNeil and Shuttleworth, (1975)].

measured as the difference between the incident gross rainfall measured above the forest canopy or in a nearby clearing, and the sum of throughfall and stemflow (sometimes called net rainfall) measured at ground level. In the early years, a few troughs placed beneath the canopy (Rutter 1967; Rutter *et al.*, 1971) were used to measure throughfall but poor sampling and systematic under-measurement associated with wetting the troughs stimulated investigation of alternative approaches. It was recognised that intensive sampling using an array of funnel or trough gauges placed randomly beneath the forest canopy was needed to overcome the great spatial variability of

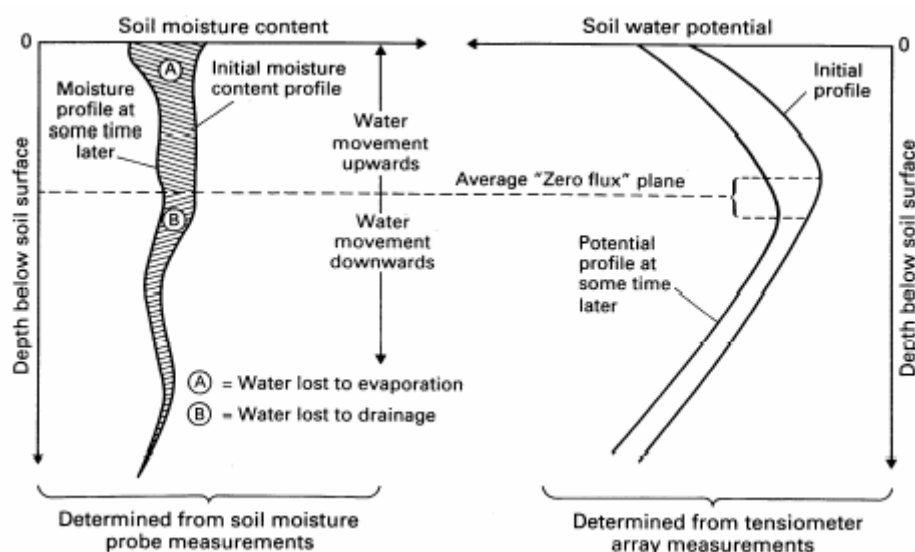


Fig. 2. Illustration of the measurement of evaporation using soil moisture depletion supplemented with the determination of an average 'zero flux' plane to discriminate between (upward) evaporation and (downward) drainage. [Taken from Shuttleworth (1993)].

throughfall, as well as stemflow gauges designed and used to estimate the (usually small) amount of rain running down the trees (Gash *et al.*, 1980). This general approach has now largely become accepted practice and continues to be widely used in many studies worldwide (e.g. Lloyd and Marques, 1988; Valente *et al.*, 1997). An alternative approach (e.g. Calder and Rosier, 1976) involving the use of a large plastic sheet to capture all the net rainfall was also developed in the early years of IH and has sometimes been used since (Calder *et al.*, 1986; 1996), although its application can sometimes be problematic (Ward and Robinson, 2000; Roberts *et al.*, 2004).

Just as techniques were developed to study the rainfall interception process in greater detail, new instruments and methods have increasingly been used over the last three decades to study the transpiration process in plants. Porometers, instruments that measure water vapour loss by individual leaves and shoots directly, have been used in hydrology since the late 1960s to study within-canopy variations in transpiration and to investigate the factors that influence the stomatal control of plants on transpiration flux; see Roberts (2006). Over the last few decades, sap-flow sensors (Dugas *et al.*, 1992; Allen and Grime, 1995; Hall *et al.*, 1998) have had a major impact on plant level studies by allowing routine measurements of whole-plant transpiration and for diagnosing differences in sap flow in individual stems and roots (e.g. Hultine *et al.*, 2004).

The development and use of sensors to measure soil moisture, which was aggressively pursued at IH starting in the late 1960s (Bell, 1969, 1987; Vachaud *et al.*, 1977; Dean, 1994), also bore fruit by allowing the indirect measurements

of evaporation and, in dry periods, transpiration, for individual plots of vegetation, especially agricultural crops (Hodnett, 1986). The use of such sensors in combination with tensiometers to measure soil moisture tension (Bell, 1987) allowed estimates of total crop evaporation using the 'zero plane displacement/soil moisture depletion' method (Fig. 2). More detailed studies of the separate contributions to evaporation from vegetation and soil for sparse crops and row crops also proved important, and these were facilitated by the use of micro-lysimeters (Wallace *et al.*, 1993; Daamen *et al.*, 1993; Wallace *et al.*, 1999; Jackson and Wallace, 1999) and capacitance probes (Robinson and Dean, 1993) to estimate evaporation from the soil surfaces.

Eddy-correlation measurements

For evaporation science, the instrumental development at IH in the late 1970s and early 1980s that has subsequently had the most dramatic and widespread impact was the development of the 'Hydra' system (Fig. 3), an integrated micrometeorological system to measure surface-energy fluxes based on the eddy correlation principle (Shuttleworth *et al.*, 1988). The Hydra system improved on the earlier 'Fluxatron' system (Dyer *et al.*, 1967) that used traditional mechanical sensors and analogue electronics, by using then state-of-the-art fast-response sensors (Shuttleworth *et al.*, 1982; Moore, 1983), solid state electronics and in-device software for real-time flux calculation and data storage (Lloyd *et al.*, 1984) and, not least important, retrospective flux correction procedures (Shuttleworth, 1988a; Moore, 1983). In some respects, the development of the Hydra system parallels the development of the automatic weather

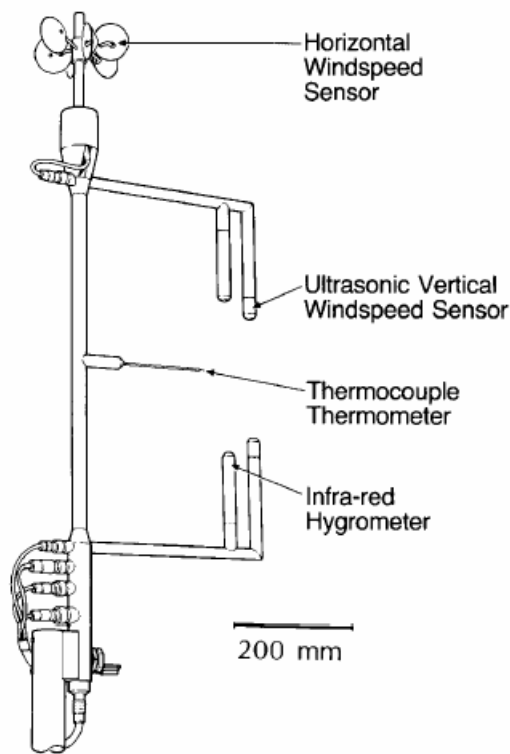


Fig. 3. The Mark 2 Hydra sensor head comprising vertical sonic anemometer transducers, an infrared hygrometer detector (top) and chopped light source (bottom), a fine-wire thermocouple thermometer, and, vertically above the rear frame support, a fast-response cup anemometer. [Taken from Shuttleworth *et al.* (1988)].

station a decade earlier in that it was capable of unsupervised, remote application and it had the ability to resolve the daily cycle, in this case of surface energy fluxes. This enabled evaporation and sensible heat and momentum fluxes to be measured over extended periods with limited supervision for a range of land covers (e.g. Shuttleworth *et al.*, 1984a; Gash *et al.*, 1989; Shuttleworth *et al.*, 1989; Wallace *et al.*, 1993). However, the more profound impact of the development of the Hydra system was that it clearly demonstrated, for the first time, that it was actually possible to make *routine* eddy correlation measurements of surface exchanges with limited supervision. This encouraged a new confidence in the use of the eddy correlation technique, which many other researchers then adopted and applied, particularly over forests where alternative surface flux measuring methods are least effective.

The basic principles used in early fast-response, infra-red absorption hygrometers (Hyson and Hicks, 1975; Raupach, 1978; Moore, 1983) were adopted and developed to allow similar fast response measurements of CO₂ (e.g. Ohtaki and Matsui, 1982; Ohtaki, 1984; Auble and Meyers, 1992; Leuning and King, 1992). Fuelled by international interest in climate change and the global carbon cycle, there

has been, over the last 15 years, a huge explosion in the use of the eddy correlation method to measure surface energy and CO₂ fluxes, to the extent that many major continental and regional networks of eddy correlation instruments are now deployed worldwide, including EUROFLUX, AMERIFLUX, OZFLUX and ASIAFLUX. These numerous regional networks are themselves coordinated through the international programme FLUXNET (<http://www.fluxnet.ornl.gov/fluxnet/>). However, recognising some responsibility for having stimulated this (in the words of Alan Greenspan, the Chairman of the US Federal Reserve) “irrational exuberance” to apply the eddy correlation technique, the author feels obliged to sound a note of caution at this time. Systematic underestimation of surface fluxes almost always occurs when using the eddy correlation technique, especially at night. When measuring evaporation, such underestimation is troublesome but not as important as when measuring CO₂ flux. Under-measurement of evaporation is less worrisome because the extent of loss during the day can be estimated (and perhaps corrected for) by calculating the *recovery ratio* for surface energy fluxes relative to a measured energy budget. Moreover, evaporation is usually small at night, so its under-measurement is less important. However, in the case of CO₂, the extent of underestimation cannot easily be quantified by subsidiary measurements, and night time (respiration) fluxes are as large and important as daytime fluxes. Great care is needed when measuring CO₂ using the eddy correlation technique because the long-term average net carbon exchange between the atmosphere and the ground is measured as the small difference between large daytime and night time fluxes, and the likely under-measurement using the eddy correlation technique is very difficult to estimate at these times. Proper recognition should always be given to the effects of advection and coordinate rotation, especially when measurements are attempted in other than flat terrain, and to the effect of angle of attack on sonic anemometer sensors (Gash and Dolman, 2003).

Area average micrometeorological measurements

Over the last two decades, there has been a substantial improvement in understanding of the degree to which measurements of evaporation made with micrometeorological sensors (such as Bowen ratio instruments, or eddy correlation devices) are representative of the vegetation canopy over which they are mounted. Gash (1986), starting from the approach of Pasquill (1972), deduced a simple analytical approach for estimating the *fetch* for which above-canopy measurements are representative. This stimulated a substantial subsequent literature (e.g. Schuepp *et al.*, 1990; Schmid and Oke, 1990; Leclerc and

Thurtell, 1990; Wilson and Swaters, 1991; Horst and Weil, 1992) which has resulted in much better understanding in this important field of micrometeorological measurement theory.

In the last decade, there has also been growing interest in a turbulence theory-based approach to deducing area surface energy exchanges indirectly from fluctuations in temperature and humidity in the near-surface atmospheric boundary layer (Hill, 1992; Ochs and Wilson, 1993). The technique involves transmitting a beam of electromagnetic radiation and measuring the intensity variations of the received signal, these being related to the movement of heat and moisture in the path between transmitter and receiver, with the relative contributions from heat and moisture depending on the wavelength of the radiation. In the visible and near-infrared waveband, the signal is much more sensitive to heat movement, while in the microwave waveband it is more sensitive to moisture movement (Hill, 1997). The approach is popular because it is relatively cheap and the resulting instruments — called scintillometers — are robust and can be used to estimate line-average fluxes over distances of up to several kilometres (e.g. De Bruin *et al.*, 1995; McAneney *et al.*, 1995). Because it is based on Monin-Obukov theory, in principle the method is strictly reliable only when applied near (but not too near) a rough surface in the well-developed boundary layers over uniform vegetation. However, field tests (Lagouarde *et al.*, 1996; Chehbouni *et al.*, 1999; Meijninger *et al.*, 2002a,b; Beyrich, 2002) suggest that, in practice, it may well be usable over surfaces with heterogeneous vegetation cover.

In principle, measurements of evaporation based on the water balance of lysimeters, evaporation pans and catchments and on the direct measurement of vapour flow (i.e. eddy correlation) do not involve theoretical assumptions, but theoretical assumptions *are* involved in several micrometeorological methods, including the Bowen Ratio/Energy Budget method and the scintillometer approach. Over the last decade attempts have been made to provide estimates of evaporation over large areas which have a higher level of empiricism or which involve more theoretical assumptions or model calculations. Estimates of sensible heat have been based on remotely sensed measurement of surface temperature, estimated or measured air temperature, and on different assumptions regarding the exchange coefficient between these two (Kustas *et al.*, 1989; Shuttleworth and Gurney, 1990; Hall *et al.*, 1992; Moran *et al.* 1994; Stewart *et al.*, 1994; Chehbouni *et al.*, 1996, 1997; Boegh *et al.*, 2000, 2002). If a separate estimate is made of the radiation exchange and soil heat flux, evaporation can then be estimated from the energy balance. The alternative approach of calculating surface exchanges over large areas

using models (Cosgrove *et al.*, 2003a, b; Sheffield *et al.*, 2003; Mitchell *et al.*, 2004; Rodell *et al.*, 2004) has come to be called the Land Data Assimilation System (LDAS) approach because the models provide the framework into which several types of data including remotely sensed surface temperature can be incorporated using modern data assimilation techniques. It is feasible to create and apply such LDAS systems now because of the very substantial progress that has been made in understanding and modelling the natural evaporation process for land surfaces over the last three decades.

PROCESS UNDERSTANDING AND MODELLING

By the early 1970s, micrometeorological research had been active for several decades and there had been at least one productive decade of agricultural meteorology and plant physiology studies. As a result, most of the basic physical and physiological building blocks were in place for a theory to describe evaporation for the two limiting cases of bare soil and vegetation with complete canopy cover and, in the case of vegetation cover, for the limiting cases of totally dry or totally wet vegetation canopies. However, hydrologists and civil and agricultural engineers were largely unaware of this understanding and it was, at that time, not yet applied by practising hydrologists nor recognised in hydrological or meteorological models.

Process understanding prior to 1970

The basic physics of heat flow in soil, which is a fairly simple application of the diffusion equation and energy conservation (subject to the boundary condition of soil surface temperature), was already understood and had found its way into textbooks (e.g. Monteith, 1973). The outstanding limitation then, as now, is the requirement for detailed knowledge of the spatial distribution of thermal properties in the soil and their relation to (perhaps changing) moisture content. There had also been substantial research relating to the capture, reflection, and transmission of radiation by bare soil, leaves and canopies (e.g. Bowers and Hanks, 1965; Tagueva and Brandt, 1961; McCree, 1972; Stanhill, 1970), and typical values for the albedo for water and many soils and crops were known (e.g. Monteith, 1973). Attenuation of incoming radiation within crops with full canopy cover was sometimes represented by a simple attenuation (Beer's law) approach. Thus, the basic radiative interaction between the earth's surface and the atmosphere was reasonably well understood and quantified, although later work stimulated by remote sensing and climate modelling interests resulted in more detailed measurements and more complex representation and modelling of this interaction (e.g. Idso

et al., 1975; Goudriaan, 1977; Graser and van Bavel, 1982; Ross, 1975; Meador and Weaver, 1980; Dickinson, 1983; Sellers, 1985).

The atmospheric boundary layer (ABL) is the layer of air directly above the earth's surface in which the effects of the surface have influence on time scales of less than a day, and in which fluxes of momentum, heat, water vapour and other atmospheric constituents are carried largely by turbulent motion. It has two main sub-layers, a well-mixed *outer region* and a distinctive *inner layer* or *surface layer*, which is typically about 10% of the total depth of the ABL, where flow is dependent mainly on surface characteristics. Any theory of evaporation that describes the surface-atmosphere processes must include adequate representation of turbulent transfer in the surface layer, usually in the form of so-called *aerodynamic resistance*. There can be no complete and universal theory of near-surface turbulence because of the essentially chaotic nature of turbulence and the fact that aerodynamically perfect surfaces of infinite extent are difficult to find in nature. However, to the extent that a theory of near-surface turbulence can be defined from the conservation equations for momentum, mass and energy, with empirically-derived scaling functions, such a theory was believed to be in place by the early 1970s and, to a large extent, it was. Seminal work by distinguished researchers including Taylor, Karman, Prandtl, Kolmogorov, Monin, Obukhov and Kazanski over several decades (see Garratt, 1992, for a review of these contributions) and major field experiments in the 1950s, late 1960s and early 1970s (see Dyer, 1974, for a review of the results of these experiments) had resulted in an accepted description. This description included parameterising surface layer exchange in terms of parameters that characterise the underlying surface (such as vegetation), specifically the aerodynamic roughness length and the zero-plane displacement. The description also included empirical *stability functions*, $\Phi_{M,V,H}$ (ξ), for momentum, water vapour, and sensible heat, respectively, which described the effect of atmospheric buoyancy and which are parameterised in terms of $\xi = (z-d)/L$, where z is the height above the ground and L is the Monin-Obukhov length. These stability functions were assumed to be universally applicable and independent of the underlying surface. Within this description, the aerodynamic resistance, $r_a^{M,V,H}(z_m)$ to the level z_m for momentum, water vapour, and sensible heat, respectively, is given by:

$$r_a^{M,V,H}(z_m) = \int_{d+z_0}^{z_m} \frac{dz}{K_{M,V,H}} = \int_{d+z_0}^{z_m} \frac{\Phi_{M,V,H}(\xi) \cdot dz}{k u_* (z-d)} \quad (\text{s m}^{-1}) \quad (3)$$

where u_* is the friction velocity and k is von Karman's constant (~ 0.41). Following Dyer's (1974) review, the

generally accepted forms for $\Phi_{M,V,H}(\xi)$ became:

$$\Phi_V = \Phi_H = (\Phi_M)^2 = (1 - 16\xi)^{-0.5} \quad \text{for } \xi \leq 0 \quad (4)$$

$$\Phi_V = \Phi_H = \Phi_M = (1 + 5\xi) \quad \text{for } \xi > 0$$

According to this theory, in conditions of neutral atmospheric stability when $\xi = 1$, $\Phi_M = \Phi_V = \Phi_H = 1$, and the aerodynamic resistances for momentum, water vapour, and sensible heat transfer have the same value, r_a . Substituting for u_* from:

$$u_* = \frac{k \cdot u_m}{\ln[(z_m - d)/z_0]} \quad (\text{m s}^{-1}) \quad (5)$$

where u_m is the wind speed at height z_m and k is the *von Karman* constant (~ 0.41), Eqn. (3) simplifies to:

$$r_a(z_m) = \frac{\ln[(z_m - d)/z_0]^2}{k^2 \cdot u_m} \quad (\text{s m}^{-1}) \quad (6)$$

Immediately adjacent to the surface of evaporating water, wet soil and vegetation with leaves covered with water, the air is very close to saturation. However, the air in transpiring canopies is not saturated because the true source of water is then inside the leaves and water loss is controlled by diffusion through the plants' stomata. Monteith (1965) had proposed diagnosing the vegetation's control on transpiration in terms of an additional resistance, the *surface resistance*, r_s . This resistance was assumed to be, at least approximately, an indirect measure of the overall stomatal control in the canopy, and Monteith pointed out that it could be deduced from meteorological measurements made above the canopy by allowing for the effect of aerodynamic resistance using the equation:

$$r_s = \left(\frac{\Delta}{\gamma} \cdot \beta - 1 \right) r_a + (1 + \beta) \frac{\rho c_p D}{\gamma A} \quad (\text{s m}^{-1}) \quad (7)$$

where $\beta = [(A - \lambda E) / \lambda E]$ is the *Bowen ratio* (dimensionless), ρ is the density of air (in kg m^{-3}), c_p ($= 1.013 \text{ kJ kg}^{-1} \text{ }^\circ\text{C}^{-1}$) is the specific heat of air at constant pressure, and A is the measured or estimated energy available for evaporation now expressed in energy terms (W m^{-2}). Alternatively, if the value of r_s is already available, Eqn. (5) can be rearranged into a predictive form to provide an estimate of the transpiration from the equation which came to be known as Penman-Monteith equation, i.e.

$$\lambda E = \frac{\Delta A + (\rho c_p D) / r_a}{\Delta + \gamma [1 + r_s / r_a]} \quad (\text{W m}^{-2}) \quad (8)$$

If the canopy is totally wet, as it is during and immediately after rain, the within-canopy air is then close to saturation and evaporation rate is no longer controlled by stomata, only

by aerodynamic resistance. In such conditions, the latent heat flux can be calculated by substituting $r_s = 0$ into Eqn. (8) to yield:

$$\lambda E_p = \frac{\Delta A + (\rho c_p D) / r_a}{\Delta + \gamma} \quad (\text{W m}^{-2}) \quad (9)$$

Rutter *et al.*, (1971) had used Eqn. (9) as the basis of a model of the evaporation of the water intercepted by forest vegetation and re-evaporated back to the atmosphere. The model Rutter proposed, which he and co-workers later tested in the field and developed further (Rutter *et al.*, 1975; Rutter and Morton, 1977), was essentially a water balance model (Fig. 4) that describes the evolution of water stored on the forest canopy and trunks of the trees during and subsequent to a rain storm. The canopy and trunks are assumed to be capable of storing a maximum amount of water, S and S_t , respectively, and the state variables in the model are the current amount being stored on the canopy and the trunks, C and C_t , respectively. The units of S , S_t , C , and C_t , are depth of water, usually mm. A proportion, p , of the precipitation is assumed to fall freely through the canopy and a proportion p_t is diverted to the trunks of the trees. Water caught on the canopy is assumed not to drain if $C < S$ but, when $C \geq S$, to drain rapidly at rate $D = D_s \exp [b(C - S)]$ where D_s ($\sim 0.002 \text{ mm min}^{-1}$) and b (typically 3.7 mm^{-1}) are empirical parameters selected to represent measured drainage rates. A running water balance is kept of the amount of water stored

on the canopy from the equations:

$$\begin{aligned} \frac{dC}{dt} &= \left(\frac{C}{S}\right) \lambda E_p - D_r \quad (\text{for } C \leq S); \\ \frac{dC}{dt} &= \lambda E_p - D_r \quad (\text{for } C > S) \end{aligned} \quad (10)$$

Equivalent equations are used to make a running water balance for the trunks, and the net interception loss is the time integral of the calculated evaporation loss from the canopy and trunk stores.

Aerodynamic resistance

Notwithstanding the great progress that had been made in understanding near-surface turbulent transport over several decades, troubling aspects of the theory remained. In the late 1960s and early 1970s, wind tunnel and field studies (e.g. Chamberlain, 1968; Thom, 1971, 1972) suggested that there were differences in the efficiency with which momentum and other entities such as water vapour and heat were transferred between the underlying surface and the atmosphere above. Transfers other than momentum are subject to a so-called *excess resistance* (Chamberlain, 1968), which is additional to the aerodynamic resistance for momentum. Thom (1972) associated this excess resistance with differences in the efficiency with which fluxes are either created or destroyed by the elements that make up the structure of the underlying surface. Specifically, the excess

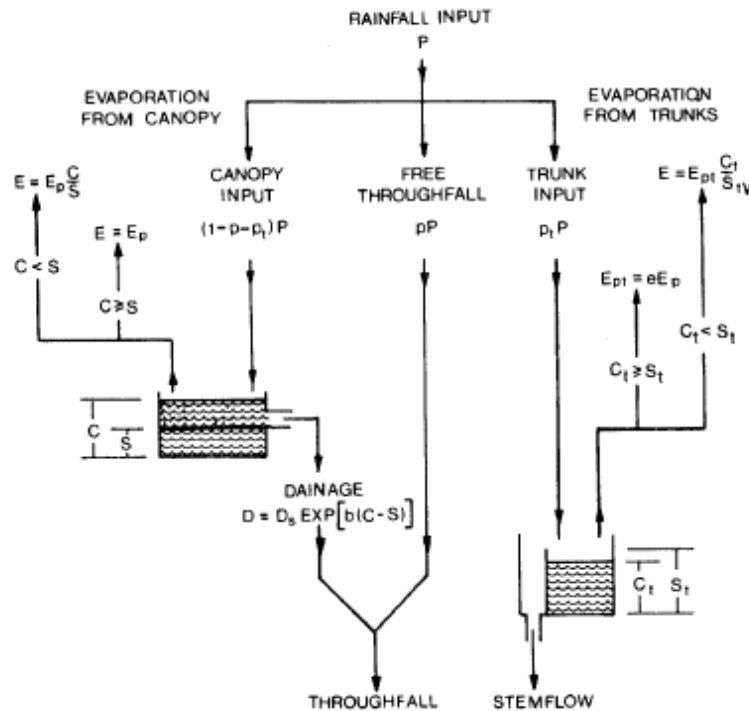


Fig. 4. Conceptual framework of the Rutter model (Rutter *et al.* 1971) of rainfall interception from forest vegetation. [Taken from Gash and Morton (1978)].

resistance was considered to be associated with the fact that it is easier to transfer momentum to the elements that make up the surface (leaves, twigs and branches in the case of vegetation) than to other entities. Momentum can be transferred by both pressure forces and molecular diffusion, but other entities can be transferred only by molecular diffusion. Expressed in terms of resistances, the difference is, therefore, due primarily to the difference between r_{bM} , the boundary layer resistance for momentum transfer to the vegetation elements making up the canopy, and r_b , the (higher) boundary layer resistance for the transfer of water vapour and sensible heat, assuming here, for simplicity, that these are equal. There is also a contribution associated with the effective average sink for momentum being higher than the effective average source/sink height for other transferred entities because the lower boundary layer resistance for momentum means the divergence for momentum is rapid near the top of the canopy.

Although the basic difference in boundary layer resistance is the main factor contributing to excess resistance, for simplicity and consistency with existing theory, Monteith (1973) and later researchers chose to represent this difference by postulating that momentum *appears* to originate at height $(d+z_0)$ and other entities *appear* to originate at a different (lower) level $(d+z_p)$. Several empirical formulae have been proposed for excess resistance (and, therefore, implicitly z_p) under different conditions (e.g. Thom 1972; Stewart and Thom, 1973; Brutsaert, 1982; Chamberlain, 1984). Fortunately, investigation of the effective value of z_p (e.g. Garratt and Francey, 1978) for a range of surfaces covered in permeable (fibrous) vegetation has suggested that $\ln(z_0/z_p) \sim 2$ (Fig. 5). Consequently, Shuttleworth (1993a) proposed setting $(z_0/z_p) = 10$ to provide an order of magnitude estimate of the effect of excess resistance when estimating evaporation in practical applications. In neutral conditions, if the aerodynamic resistance for water vapour and sensible heat transfer are assumed equal, then allowing for excess resistance, Eqn. (5) is modified to become:

$$r_a(z_m) = \frac{\ln[(z_e - d)/z_0] \cdot \ln[(z_e - d)/(z_0/10)]}{k^2 \cdot u_m} \quad (\text{s m}^{-1}) \quad (11)$$

where z_e is the height at which humidity is measured. Thom (1972) suggested that setting the difference in boundary layer resistances for heat fluxes and momentum to $(5.6/u_*^{2/3})$ "may well provide an accurate enough first approximation". However, if $\ln(z_0/z_p)$ is considered to be constant as Garratt and Francey (1978) suggest (see Fig. 5), then setting $(z_0/z_p) = 10$ as per Shuttleworth (1993a), the difference in boundary

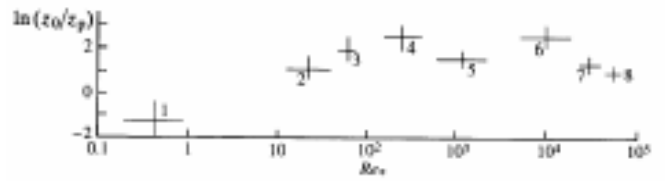


Fig. 5. Variation of $\ln(z_0/z_p)$ with roughness Reynolds Number for surface types as follows: 1, sea; 2, vineyard; 3, short grass; 4, medium grass; 5, bean crop; 6, savannah scrub; 7 and 8, pine forest. [Taken from Garratt (1992)].

layer resistances becomes $(5.6/u_*)$. Assuming this form, and recognising that the boundary layer resistance for momentum is believed to be typically five to ten times less than for energy fluxes, it is here proposed that, as a rough approximation, r_b is estimated adequately by:

$$r_b = 6/u_* \quad (\text{s m}^{-1}) \quad (12)$$

A further troubling aspect of near-surface turbulent transport theory was first recognised when analysing data gathered by IH during the Thetford Project. At this near-perfect micrometeorological site, data were taken over forest and, relative to the aerodynamic roughness length of the vegetation, these data were therefore gathered near the underlying canopy. This allowed stringent testing of turbulent transfer theory very near to the surface. The then-current theory was found to be imperfect (Thom *et al.*, 1975) in that the empirical stability functions previously derived over shorter vegetation covers (and assumed to be universally applicable) were inconsistent with observations made a few metres above forest, in the *roughness sublayer* or *transition* layer. Investigation of this phenomenon in wind tunnels and in field studies (Raupach, 1979; Raupach *et al.*, 1980; Garratt, 1980; Raupach and Thom, 1981) suggested this increase is associated with wakes generated by individual plants. Subsequently, theoretical accommodation for this phenomenon was made by the introduction of an additional height-dependent, empirical function (Garratt, 1992); this multiplies the stability factors and, thus, represents the increased efficiency of turbulent transfer near the canopy. This additional factor reduces asymptotically and approaches unity at a height z_* , with (z_*/z_0) typically in the range 10–150. With this assumed empirical form, the definition of the aerodynamic resistance, $r_a^{M,V,H}(z_m)$, to the level z_m , for momentum, water vapour and sensible heat, previously given as Eqn. (3) becomes:

$$r_a^{M,V,H}(z_m) = \int_{d+z_0}^{z_m} \frac{dz}{K_{M,V,H}} = \int_{d+z_0}^{z_m} \frac{\Phi_{M,V,H}(\xi) \cdot \varphi(z/z_0) \cdot dz}{k u_* (z-d)} \quad (\text{s m}^{-1}) \quad (13)$$

where:

$$\phi = \exp[-0.7(1-z/z_*)] \quad \text{for } z < z_*; \quad \text{and } \phi = 1 \quad \text{for } z \geq z_*$$

A theoretical shortcoming of this empirical approach is that it includes a physically unrealistic discontinuity when $z = z_*$.

Representing full canopy cover

Just as it is important to describe turbulent transport above and within the canopy using aerodynamic resistance, and the transfer between canopy air and vegetation using boundary layer resistances, so it is important in the case of dry vegetation to include description of the plants' control through their stomata. This control is recognised through another resistance, the *stomatal resistance*, which for an individual leaf is defined as:

$$r_{st} = \frac{\rho c_p [e_{sat}(T_s) - e_s]}{\gamma \lambda E_l} \quad (\text{s m}^{-1}) \quad (14)$$

where r_{st} is the stomatal resistance of the leaf for unit surface area, λE_l is the latent heat loss from the leaf, $e_{sat}(T_s)$ is the saturated vapour pressure at the area-average surface temperature, T_s , of the leaf, and e_s is the average vapour pressure immediately adjacent to the leaf surface. Multilayer computer models of the interaction between vegetation canopies and the atmosphere were already being made by the mid-1970s (e.g. Waggoner and Reifsnnyder, 1968, 1969; Goudrian, 1977). Such models (Fig. 6a) represent the capture of radiant energy at several levels in the canopy and the heat exchanges between leaves and air at these levels in terms of the average stomatal resistance to water vapour flow and the leaf boundary layer resistance to water vapour and sensible heat flow. They also describe the aerodynamic

resistance to energy flow between each level.

At the same time, a separate school of thought prevalent in the UK preferred the much simpler so-called *big leaf* approach to describe plant canopy exchange with the overlying atmosphere (Fig. 6b). The big leaf approach had its origin in the Penman-Monteith equation (Monteith, 1965); essentially it assumes that the exchange of the whole canopy can be adequately represented by assuming that all the radiation capture and the partitioning of energy into latent and sensible heat can be described as if it occurred at a single level, the effective source-sink height. At this level, the whole-canopy parallel-average values of stomatal resistance and boundary layer resistance control the exchange between the hypothetical big leaf and the surrounding air, these resistances being appropriately scaled down from the resistances for individual leaves by dividing by the leaf area index (LAI). The aerodynamic resistance for latent and sensible heat is then used to represent the turbulent transfer of energy fluxes upward into the atmosphere. In the simpler big-leaf model, the canopy-average boundary layer resistance and the aerodynamic resistance act in series for both the latent and the sensible heat transfer, and are often combined as a single aerodynamic resistance which is estimated as being that for momentum appropriately corrected for the excess resistance, as discussed above.

The relative merits of multi-layer computer modelling of whole-canopy exchanges versus the simpler big leaf approach were debated throughout the late 1960s and early 1970s. However, drawing on earlier recognition (Shuttleworth, 1975) that water surfaces have a small but finite *intrinsic resistance* to vapour exchange similar to stomatal resistance, Shuttleworth (1976) demonstrated that it was possible to write a multi-layer model in analytically-

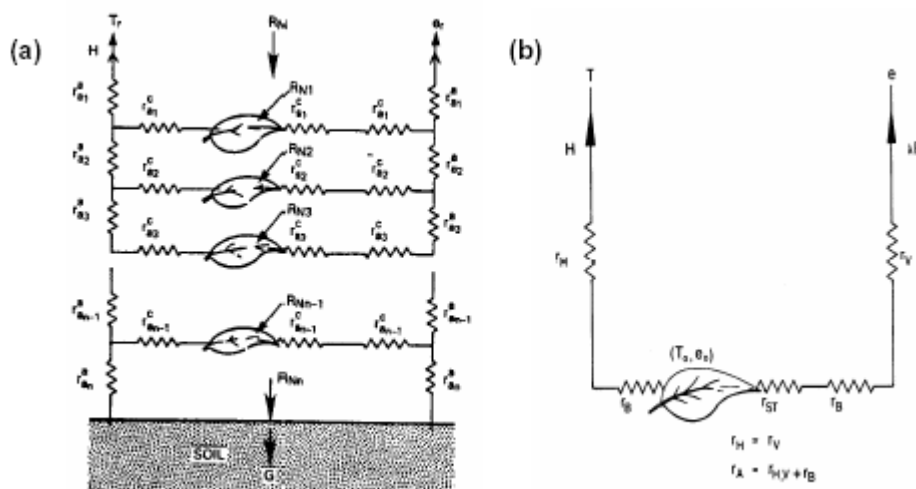


Fig. 6. Schematic diagrams of the resistance networks used to represent the controls on sensible heat and latent heat exchanges between plant canopies and the overlying atmosphere in (a) a typical multilayer model and (b) the "big leaf" model.

continuous form and create a one-dimensional description of general applicability (including during wet and partly wet conditions) in which evaporation is described by a combination equation similar to the Penman-Monteith equation. Subsequently Shuttleworth (1977a) applied similar physical principles to derive a theory describing the deposition of fog and cloud droplets on canopies; and developed the analytical multi-layer approach to provide a simpler representation (Shuttleworth, 1978) and to include the effect of below-canopy energy fluxes (Shuttleworth, 1979a). During the 1970s, the big leaf approach began to gain preference over multi-layer computer modelling of evaporation partly because the two had been shown to be formally equivalent (Shuttleworth, 1976a), but mainly because it was increasingly realised that multi-layer canopy models require a level of detail in the specification of canopy properties and canopy structure that limit their use in research to particular locations and times where such detailed knowledge is available.

A second significant step forward in describing the evaporation process also occurred at about this time. Using either the multi-layer or big leaf model, it was already possible to describe totally dry canopies and totally wet canopies, and the Rutter model (Rutter *et al.*, 1971, 1975) was available to calculate the net loss of intercepted water as the canopy dried after a storm. However, there was no physically-based way to describe the transition between dry and wet canopies. Indeed, because the Penman-Monteith or big leaf approach assumes a parallel average of all leaf-level resistances in the canopy, in principle if just one leaf is covered in water, the effective canopy-average resistance is very small. Clearly this is unrealistic and data from the Thetford Project were used to demonstrate this fact (Shuttleworth, 1976b). A debate ensued (Monteith, 1977, Shuttleworth, 1977b). The resolution of this issue was already implicit in Shuttleworth's (1976a) paper, but Shuttleworth (1978) also demonstrated it was feasible to describe the transition between wet and dry canopies using the simpler big leaf approach. Simplifying into the nomenclature of this paper, assuming equality of the boundary layer and aerodynamic resistances for water vapour and sensible heat and, for conciseness, defining:

$$r_b' = r_b (\Delta/\gamma + 1) \quad \text{and} \quad r_a' = r_a (\Delta/\gamma + 1) \quad (15)$$

Shuttleworth (1978) showed that, r_s^{ALL} , the surface resistance in dry, wet, and partly wet conditions has the form:

$$r_s^{ALL} = \frac{r_b' (r_b' + r_s)}{W (r_b' + r_s) + (1 - W) r_b'} - r_b' \quad (16)$$

where W is the fractional surface wetness of the canopy. Shuttleworth also showed that W is related to (C/S) , the

fractional fill of the canopy store in Rutter's (1971) model, through the relations:

$$W = 1 \quad \text{for } (C/S) > 1 \quad \text{and} \quad (17)$$

$$W = \frac{(C/S)[1 - Z]}{[1 - Z(C/S)]} \quad \text{for } (C/S) \leq 1$$

where $Z = [r_s (r_a' - r_b')] / [r_b' (r_s + r_b')]$

Figure 7 is an example of how Eqns. (16) and (17) successfully describe the transition from fully wet to totally dry canopies, using data from the Thetford project.

Breaking apart the evapotranspiration concept

As previously stated, the long-established concept of a hypothetical potential evaporation rate still dominated hydrological practice in the early 1970s, and was the normal approach when representing evaporation in hydrological and meteorological models. However, the brisk pace of research into the separate processes involved in natural evaporation (transpiration and the evaporation of intercepted water) and IH's focus on understanding evaporation from forest stands (where these two processes are most distinct) had rapid impact. Experimental evidence soon showed that the energy used to evaporate water from wet canopies commonly exceeded that which was locally available, so that advection of energy from elsewhere must have occurred even over extensive forests such as Thetford Forest (Stewart, 1977). Similarly, abundant data from the Thetford Project (Stewart and Thom, 1973; Gash and Stewart, 1977; Gash *et al.*, 1978) showed that forest transpiration is usually significantly less than potential evaporation. Meanwhile, the use of potential evaporation persisted and the Priestley and Taylor (1972) equation (Eqn. (2)) was gaining increasing recognition as a means of estimating its value. Ultimately, by combining data and understanding from both the Thetford Project and studies in IH's Plynlimon catchments, Shuttleworth and Calder (1979) were able to demonstrate the categorical failure of using potential evapotranspiration estimates to describe forest evaporation (Fig. 8). They suggest the possibility that a much simpler model, which recognised, separately, forest transpiration and the evaporation of intercepted water, was likely to be a more productive approach. In this way, evaporation from a forest was shown to depend strongly on the forest's rainfall climatology. In the case of the UK, for instance, forests in the west, where rainfall is frequent, evaporate a great deal of intercepted rainfall and take energy from the atmosphere as needed to support this evaporation. The relative importance of transpiration is greater for forests in the eastern UK and on average they evaporate less, perhaps at less than the potential rate. Thus, by the end of the 1970s, research at IH had

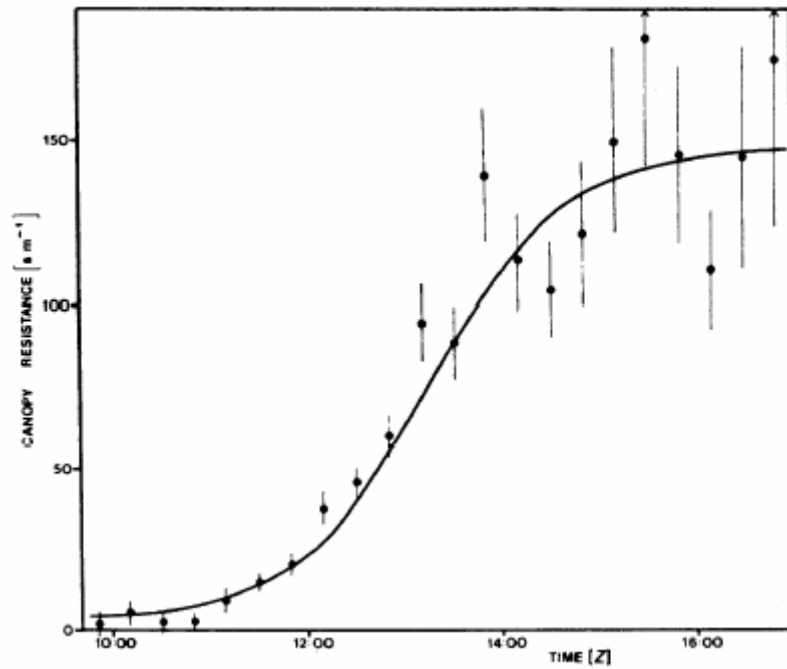


Fig. 7. The measured and modelled variation in the surface resistance for Thetford Forest between a totally wet and a totally dry canopy. [Taken from Shuttleworth (1978)].

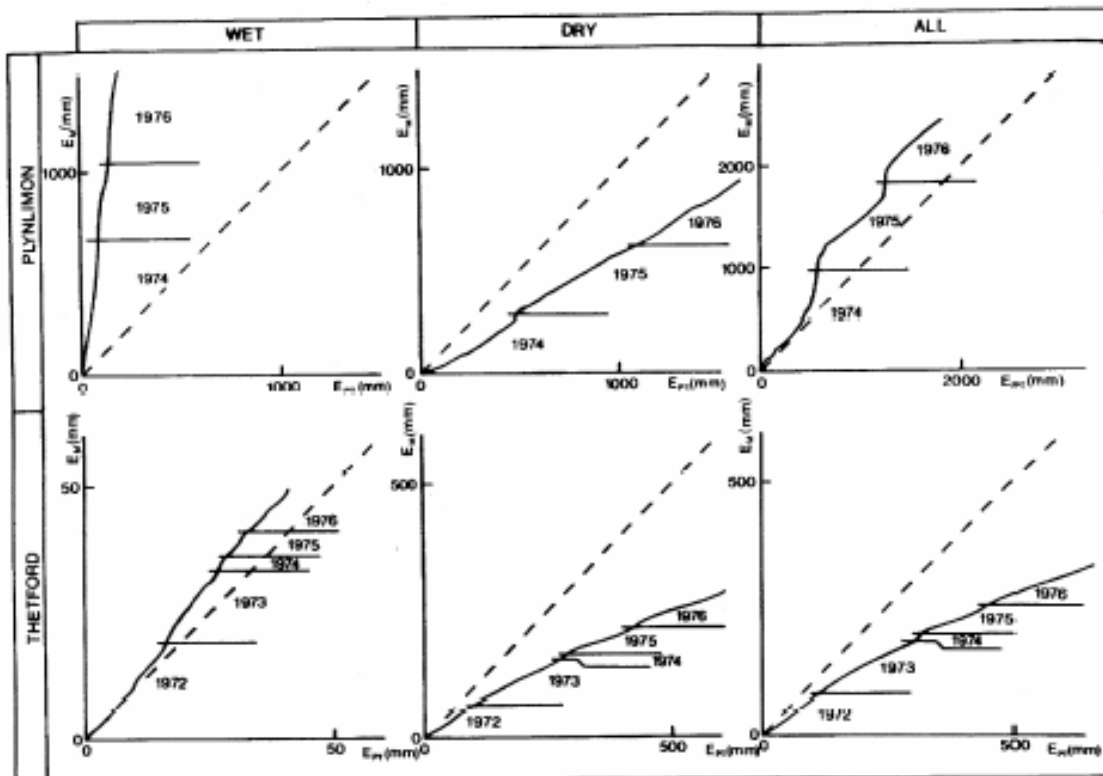


Fig. 8. Comparison between, E_M , the evaporation measured at the Plynlimon and Thetford sites, and E_{PT} , the estimate given by the Priestley-Taylor equation, expressed as a cumulative sum over several years of data collection. A separate comparison is made for both wet and dry periods, and for the total loss in all conditions. (Note: the results for Thetford are not the cumulative evaporation loss for complete years.) [Taken from Shuttleworth and Calder (1979)].

demonstrated the absolute necessity of modelling the transpiration and interception processes independently, at least in the case of tall, aerodynamically rough vegetation.

Other significant theoretical developments at this time complemented the progress described in the last paragraph. Thom and Oliver (1977) made an in-depth re-evaluation of one of the equations most commonly used to estimate potential evaporation, the Penman equation (Penman, 1948). They demonstrated that, in the climate where the equation was derived (south-east England), and for the surfaces for which it was originally derived (open water, bare soil and grass), some aspects of the empiricism used were compensating implicitly for physical and physiological processes that had subsequently been recognised as having a role in the evaporation process. Specifically, they showed that the so-called 'wind function' in Eqn. (1), i.e., $f(u_m) = 6.43(1+0.536)$, is serving two purposes:

- (i) it provides a reasonable average description of the effect of thermal stratification on a more rigorous formula for aerodynamic resistance with an assumed value of sensible heat flux of 50 W m^{-2} , this value being not atypical of the conditions for which the equation was derived; and
- (ii) when compared to a physically rigorous formula of aerodynamic resistance in neutral conditions, the implicit (small) value of z_0 which it contains is such as to compensate for the absence of a surface resistance in the denominator of the Penman-Monteith equation for crops for which $(r_s/r_a) \sim 1.4$, at least in the climate for which the equation was derived.

Equally important, stimulated by recognition that potential evaporation is often a poor estimate of the interception loss from forests and that the detailed above-canopy meteorological data required by the Rutter model (Rutter *et al.*, 1971) are rarely available, at about this time Gash (1979) derived a comparatively simple analytical model of the evaporation of intercepted rainfall from a forest; this acknowledged that, in addition to the amount of water that a forest canopy can hold, the duration of rainfall is a critical controlling factor for this particular evaporation process. Subsequent studies (Gash *et al.*, 1995; Valente *et al.*, 1997) have provided significant improvements in both the Gash analytical model and the Rutter numerical model by including a more rigorous representation of sparse forest canopies that reverts to the classic Rutter and Analytical model for closed canopy conditions.

Sparse canopies

Shuttleworth and Wallace (1985) provided the missing

framework for describing evaporation from sparse vegetation canopies through an equation for which the equations describing evaporation for complete canopy cover and bare soil (each appropriate forms of Eqn. (8)) are asymptotic limits. The equation assumes that there is blending of heat fluxes from the leaves and the soil in the mean canopy airflow at the height of the effective canopy source (Fig. 9) and it takes the form:

$$\lambda E = C_c (\lambda E)_c + C_s (\lambda E)_s \quad (18)$$

$$\text{where: } (\lambda E)_c = \frac{\Delta A + (\rho C_p D - \Delta r_{a,c} A_s) / (r_{a,a} + r_{a,c})}{\Delta + \gamma [1 + r_{s,c} / (r_{a,a} + r_{a,c})]} \quad (19a)$$

$$(\lambda E)_s = \frac{\Delta A + (\rho C_p D - \Delta r_{a,s} [A - A_s]) / (r_{a,a} + r_{a,s})}{\Delta + \gamma [1 + r_{s,s} / (r_{a,a} + r_{a,s})]}$$

$$C_c = [1 + (R_c R_a) / \{R_s (R_c + R_a)\}]^{-1};$$

$$\text{and } C_s = [1 + (R_s R_a) / \{R_c (R_s + R_a)\}]^{-1} \quad (19b)$$

$$R_a = (\Delta + \gamma) r_{a,a}; \quad R_s = (\Delta + \gamma) r_{a,s} + \gamma r_{s,c};$$

$$R_c = (\Delta + \gamma) r_{a,c} + \gamma r_{s,c}$$

where A and A_0 are the energy available to the canopy and the soil, respectively; $r_{a,a}$ is the aerodynamic resistance between the effective source height and screen height; $r_{a,c}$ and $r_{a,s}$ are the aerodynamic resistance between the effective source height and the canopy and soil, respectively; and $r_{s,c}$ and $r_{s,s}$ are the surface resistance of the canopy and soil, respectively. Shuttleworth and Wallace (1985) provided working assumptions for the parameters in Eqns. (18), 19(a) and (19b) for the purpose of illustration, including using Beer's Law to estimate A and A_0 and linear interpolation to estimate values for $r_{a,c}$, $r_{a,s}$ between closed canopy and bare soil limits. However, the precise definition of A and A_0 and $r_{a,a}$, $r_{a,c}$, $r_{a,s}$, $r_{s,c}$ and $r_{s,s}$ remained speculative. By fitting simple functions to curves in Shaw and Pereira (1982), Choudhury and Monteith (1988) specified relevant functions to estimate z_0 and d from LAI. These were then adopted and extended by Shuttleworth and Gurney (1990) to provide LAI-dependent formulae for estimating the aerodynamic resistances in Eqns. 19(a) and (19b). Shuttleworth and Gurney also provided a diagnostic form of the Shuttleworth-Wallace equation which, at least in principle, can be used to deduce the canopy resistance of vegetation of varying density from foliage temperature.

There has subsequently been extensive use of Eqns. (18) and (19a), in many cases with the formulae for aerodynamic resistances given by Shuttleworth and Gurney (1990). The equation has proved particularly useful to describe growing

agricultural crops and row crops (e.g. Lafleur and Rose, 1990; Wallace *et al.*, 1990a,b; Stannard, 1993; Farahani and Bausch, 1995; Federer *et al.*, 1996; Farahani and Ahuja, 1996; Gilles *et al.*, 1998; Teh *et al.*, 2001). In some cases, the basic theoretical framework of Shuttleworth and Wallace (1985) has been extended to include multi-species canopies and the exchange of CO₂ (e.g. Wallace, 1996; Wallace and Verhoef, 2000).

Land surface parameterisations

In retrospect, it is remarkable that the prevailing understanding of the evaporation process for vegetation-covered ground as documented in IH Report No. 56 (Shuttleworth, 1979b) has changed so little. Most of the fundamental understanding of the natural evaporation process and how this could be combined in models was known, including much understanding that arose after 1970. Moreover, most of the description of evaporation measuring methods is still current, and there was already recognition that some of the empirical equations previously used to estimate potential evaporation might be viewed as successive approximations of the more rigorous understanding that had, by then, become available. Since that time, there have been numerous attempts to incorporate process understanding of natural evaporation into computer models of surface energy exchange in different combinations, with differing detailed parameterisations of individual processes and with different emphases on component processes.

For many years, the meteorological community led the way in writing such models (e.g. Deardorff, 1978; Dickinson *et al.*, 1986, 1998; Sellers *et al.*, 1986, 1996a, 1996b; Abramopoulos *et al.*, 1988). Perhaps this is because meteorologists have a crucial need to describe the surface energy balance correctly the better to simulate weather and climate because the atmosphere is largely driven from below, while many in the hydrological community still viewed evaporation merely as a loss in the catchment water balance and assigned importance to its modelling commensurate with this perception. Within the meteorological modelling community, models of surface exchange are often termed *Land Surface Parameterisation* (LSP) schemes and, in the late 1980s and through the 1990s, there was an amazing proliferation of such models. Building LSPs almost became a cottage industry: any intelligent graduate student with a computer and access to a library could build their own LSP, and many did. In due course, an attempt was made to bring some order through the creation of the Project for Inter-comparison of Land-surface Schemes (PILPS; Henderson-Sellers *et al.*, 1993). PILPS conducted organised comparisons of models and between models and data, with

a view to recognising different levels of sophistication among the many models available and defining the level of complexity required. It is not within the scope of this paper to describe the detailed evolution in the complexity with which understanding has been combined in LSPs. Readers interested in this topic are, therefore, referred to recent reviews of LSPs (e.g. Pitman, 2003; Yang, 2004). However, two important aspects in the development of LSPs are relevant to the present paper, namely the growth in understanding of the stomatal control on evaporation written into LSPs, and the use of remotely sensed data to parameterise LSPs.

Initially, the sub-models of evaporation included in hydrological and meteorological models that recognised vegetation explicitly contained comparatively simple parameterisations of the whole-canopy effect of stomata, i.e. in the simple nomenclature used in this paper, of the surface resistance, r_s , or its reciprocal the surface conductance, g_s . Initially the r_s (or g_s) was merely assigned a typical value, or a prescribed daily trend, sometimes with some allowance for the effect of soil moisture limitations (e.g. Gash and Stewart, 1977; Shuttleworth and Calder, 1979; Shuttleworth, 1988b; Dolman *et al.*, 1991). However, alternative (essentially empirical) models of stomatal behaviour soon emerged which sought to capture the response of plant stomata to the environment in which the vegetation is growing, i.e. their observed response to the specific humidity deficit of the surrounding air, D , and air temperature, T ; the solar radiation incident on the canopy, S ; and the soil moisture deficit, θ , in the upper soil where the plant roots are found. Typically, environment control is represented as *stress factors*, and the first approach to including stress factors (e.g. Dickinson, 1986; Sellers *et al.*, 1986; Dolman *et al.*, 1991; Wright, 1996a, 1996b) was as a series of multiplicative factors in the so-called Jarvis–Stewart formula for stomatal conductance (Jarvis, 1976; Stewart, 1988), i.e. in a formula similar to:

$$g_s = g_0 f_D(D) f_T(T) f_S(S) f_W(\theta) \quad (\text{m s}^{-1}) \quad (20)$$

where g_0 is a biome-dependent (i.e. vegetation cover dependent) maximum value of stomatal conductance, and f_D , f_T , f_S , and f_W , (all in the range 0–1) are the stress factors associated with D , T , S and θ , respectively. Over the last few decades, considerable effort and resources have been devoted to providing field-calibrated values for g_0 and the expressions f_D , f_T , f_S and f_W , as described later.

Subsequent studies of the behaviour of plants suggested a different approach to modelling stomatal control that is somewhat less empirical and, therefore, hopefully more universal from one plant species to the next and, perhaps, is

only dependent on whether, in terms of their photosynthetic function, species are C_3 or C_4 plants. In such models, the assimilation of carbon is viewed as the controlling factor and stomatal conductance is described by (and sometimes a derivative of) the so-called Ball–Berry equation (Ball *et al.*, 1987), i.e.:

$$g_s = m(A_n/C_s)P_l F_e + g_{\min} \quad (21)$$

where g_{\min} is a prescribed minimum stomatal conductance; m is a slope parameter (~ 9 for C_3 plants); A_n is the net carbon assimilation; C_s is the partial pressure of carbon dioxide and P_l is atmospheric pressure adjacent to the leaf; and F_e is humidity dependent stress factor, which in the original Ball–Berry equation was numerical equal to the relative humidity, but which in some LSPs (e.g. Dickinson *et al.*, 1998) is assumed to be a function of vapour-pressure deficit. The introduction of this alternative means of describing the behaviour of stomata has sometimes been referred to as the 'greening of LSPs'. In this formula, the simplest estimate of A_n , is given (Farquhar and Sharkey, 1982) by:

$$A_n = \min(J_c, J_e, J_s) \quad (22)$$

where J_c, J_e, J_s are functions expressing the assimilation rates when limited by the Rubisco enzyme, light and transport capacity respectively for C_3 and C_4 plants; see Collatz *et al.* (1991, 1992), Sellers *et al.* (1996b) and Cox (2001). In practice, it has been observed that the transition between these three limiting rates is not abrupt but gradual, and some LSPs (e.g. Collatz *et al.*, 1991; Cox, 2001) have devised mathematical ways to simulate this smooth transition.

Thus, a key difference between the Ball–Berry formula for stomatal conductance (resistance) and the Jarvis–Stewart formula is that stress factors, apart from that for humidity, are no longer combined as a product; rather one factor is considered to be the dominant limitation on carbon assimilation and, hence, on stomatal conductance. Although not directly relevant to the present paper, it is worth noting that LSPs continue to make progress in describing vegetation and that LSPs are now seeking to describe the evolution of the biome (i.e. vegetation cover) represented in the GCM at a particular place in response to long-term changes in modelled climate (Cox *et al.*, 2000; Cox, 2001; Huntingford, 2004; Foley *et al.*, 1996, 1998, 2000; Kucharik *et al.*, 2000; Oyama and Nobre, 2003, 2004). Ultimately, this capability may also become relevant in long-term, large-scale hydrological modelling studies.

LSPs, like hydrological models, require specification of the parameters used in the model to describe the soil and overlying vegetation. Because LSPs are designed to be used

in meteorological models that operate at global scale, they require the specification of soil parameters globally, and sometimes represent several (~ 10 – 12) separate biomes. Originally, soil parameters were derived by digitising global soil maps (FAO/UNESCO, 1974) and the location and extent of biomes by digitising vegetation maps (Matthews, 1983; Wilson and Henderson-Sellers, 1985). More recently, several researchers have used satellite-derived vegetation cover and measures of seasonality (Los *et al.*, 2000; Bounoua *et al.*, 2000; Buermann *et al.*, 2001; Bonan *et al.*, 2002a, 2002b). The more complex LSPs (Dickinson *et al.*, 1986, 1998; Sellers *et al.*, 1986, 1996a) in which canopy processes are represented in detail require the specification of many (~ 20 – 30) parameters, including the key parameters associated with turbulent exchange, radiation exchange, stomatal control and the interception of precipitation by the canopy for each biome. Originally, the parameters for a particular biome were selected as "intelligent guesses guided by the literature" (e.g. Dickinson *et al.*, 1986; Sellers *et al.*, 1986), but more recent field studies have provided data to define parameter values for particular biomes (as described later), and indirect estimation of some parameters using remote sensing data has also been explored (Sellers *et al.*, 1996b; Zeng *et al.*, 2000; Schaudt and Dickinson, 2000). It is interesting that, in a similar way, remote sensing data are now also increasingly being used to parameterise and evaluate some of the more advanced hydrological models (e.g. MIKE SHE: Graham and Butts, 2005; Overgaard *et al.*, 2006a, 2006b).

The parallel track

Over the last 30–35 years, as basic understanding of the natural evaporation process and the complexity of models of evaporation grew among hydrologists and meteorologists, many within the irrigation engineering community took a different route. Perhaps this was because this application requires less accuracy in evaporation estimates and the range of conditions for which estimates are needed is narrower (estimates are generally needed for short, artificially well-watered agricultural crops growing in climates with limited rain). It is just as likely to have been because in this area of science it is often necessary to provide estimates of evaporation using limited data, i.e. from traditional climate station data rather than the detailed meteorological data that are now available, for example, from automatic weather stations. Regardless of the cause, the irrigation engineering community preferred to focus attention on refining the traditional potential evapotranspiration rate approach.

Consequently, progress has been made in several related areas. For example, greater emphasis is now given to obtaining better estimates of the so-called *reference crop*

evaporation rate, the usual starting point for estimating crop water requirements using formulae whose origin is more physically-based, such as the Penman equation and Priestley-Taylor equation (e.g. Doorenbos and Pruitt, 1977) and, as discussed in more detail below, the Penman–Monteith equation (Shuttleworth, 1993a; Allen *et al.*, 1998). However, in some situations, limited data precludes use of these two equations, and simpler (but probably less reliable) temperature-based estimates may then be required. Hence, an effort was made to define the most reliable of the several alternative empirical temperature-based estimation formulae and the Hargreaves equation (Hargreaves and Samani, 1982, 1985) is now preferred (Jensen *et al.*, 1990; Allen *et al.*, 1998). In some situations, the only bases for estimating evaporation are pan evaporation data; consequently, efforts were made to specify correction factors to adjust pan measurements so that they estimate reference crop evaporation better. This resulted in correction factors tabulated as a function of the siting of the pan and ambient meteorological conditions (e.g. Doorenbos and Pruitt, 1977; Shuttleworth, 1993a; Allen *et al.*, 1998). Estimating crop water requirements from reference crop evaporation rate is sometimes called the ‘two-step’ approach, because once an estimate of reference crop evaporation has been made, this must then be adjusted to estimate evaporation for the crop using empirical, crop-specific, multiplicative *crop factors*. Substantial effort has been devoted towards defining crop factors and their estimated values now exist in tabulated form for many crops (e.g. Doorenbos and Pruitt, 1977; Shuttleworth, 1993a; Allen *et al.*, 1998).

In estimating crop water requirements, one very significant step forward towards including greater physical realism was that a panel of experts organised by FAO in 1990 recommended that the preferred definition of reference crop evaporation rate should recognise the presence of both aerodynamic and stomatal resistance operating in the Penman-Monteith equation. Consequently, the currently preferred definition (e.g. Shuttleworth, 1993a; Allen *et al.*, 1998; Allen, 2005) is as follows:

Reference crop evaporation is defined as the rate of evaporation from an idealised grass crop which completely covers the ground and with fixed height of 0.12 m, an albedo of 0.23, and a surface resistance of 69 s m⁻¹.

In terms of evaporation rate, such a crop closely resembles previous definitions of a reference crop, namely, an extensive short, green grass cover of uniform height, actively growing, completely shading the ground and not short of water. Nonetheless, the now more precise specification

allows improved definition of the capture of solar radiation in the case of the albedo, the control of transpiration by stomata in the case of the surface resistance, and the effect of aerodynamic transfer in the case of the crop height, h . However, important assumptions are implicit in using the Penman-Monteith equation in this way, the most significant being that the equation is realistic strictly only when applied over periods when meteorological conditions are fairly constant, i.e. over the 20–60 minute time scale, but in this case it is being applied at the daily time scale. Significant changes occur in both ambient weather variables and surface fluxes during the course of a day, including the regular day-to-night reversal of the sensible heat flux. When applying the Penman–Monteith equation to estimate crop water requirements using daily-average values, it is assumed that use of the value of r_a appropriate for neutral conditions adequately captures the average aerodynamic influence of the crop in both unstable daytime conditions, when most evaporation occurs, and stable night time conditions, when little evaporation occurs. It is also assumed that an effective all-day average value of r_s is appropriate and can be applied. Presumably the empirical calibration of the value of r_s for the reference crop helps to compensate for some of the consequences of these assumptions.

Assuming the Penman–Monteith equation can be applied with all-day average values and that the density of air is estimated adequately (Shuttleworth, 1993a; Allen *et al.*, 1998) by:

$$\rho_a = (3.486P)/(273 + T) \quad (\text{kg m}^{-3}) \quad (23)$$

where T is in °C, Eqn. (8) can be re-written with energy fluxes expressed in terms of equivalent mass of evaporated water and the daily evapotranspiration (in mm day⁻¹) for a crop is estimated by:

$$E_{crop} = \frac{\Delta(R'_n - G') + \frac{\gamma[187342.D_m/(273 + T)]}{r_{a,crop}}}{\Delta + \gamma \left[1 + \frac{r_{s,crop}}{r_{a,crop}} \right]} \quad (\text{mm day}^{-1}) \quad (24)$$

where

R'_n = daily total net radiation exchange for the crop, expressed in mm day⁻¹

G' = measured or estimated daily total soil heat flux, expressed in mm day⁻¹

T = daily average temperature, in °C

U_m = daily total wind speed measured at height z_m , in m s⁻¹

D_m = daily average vapour pressure deficit measured at height z_m , in kPa

$r_{a,crop}$ = effective all-day average aerodynamic resistance to height z_m , in s m⁻¹

$r_{s,crop}$ = effective all-day average surface resistance, in $s\ m^{-1}$

The estimated daily evaporation for a reference crop is then calculated by substituting the appropriate expression for the aerodynamic resistance of a reference crop and the appropriate value for the stomatal resistance of a reference crop, i.e. $r_{a,crop} \rightarrow r_{a,rc}$ and $r_{s,crop} \rightarrow r_{s,rc}$ in Eqn. (24). The effective aerodynamic resistance for the reference crop, $r_{a,rc}$, is estimated by Eqn. (11) with z_e and z_d both set to 2m, and with:

$$z_0 = 0.123h \quad \text{and} \quad d = 0.67h \quad (\text{m}) \quad (25)$$

and the crop height $h = 0.12$ m. This gives:

$$r_{a,rc}^2 = 208 / U_2 \quad (\text{s}\ \text{m}^{-1}) \quad (26)$$

in which U_2 is the daily-average wind speed measured at 2m. Setting the effective daily-average value of surface resistance for the reference crop to:

$$r_{s,rc} = 70 \quad (\text{s}\ \text{m}^{-1}) \quad (27)$$

it can be shown (e.g., Shuttleworth, 1993a; Allen *et al.*, 1998; Allen, 2005) that Eqn. (23) becomes:

$$E_{rc} = \frac{\Delta}{\Delta + \gamma^*} (R'_n - G') + \frac{\gamma}{\Delta + \gamma^*} \frac{900}{(T + 275)} U_2 D_2 \quad (\text{mm}\ \text{day}^{-1}) \quad (28)$$

where:

$$\gamma^* = \gamma (1 + 0.337 U_2) \quad (29)$$

In the concluding paragraph of the chapter on evaporation in the *Handbook of Hydrology*, Shuttleworth (1993a) called for an extension of the use of the Penman–Monteith equation to allow a one-step approach for estimating crop water requirements in which the effective stomatal resistance for different crops would be used to substitute for the use of crop factors in the traditional two-step approach. However, at the time of writing, the two-step approach remains locked in place, in part due to limited data availability. Because there are as yet no data records to substitute for traditional climate station data in many places around the world, estimated potential evapotranspiration rates based on these simple climate station data are often still preferred. Nonetheless, the author foresees progress in the coming decades through the calculation and distribution of globally available, high-resolution, model-calculated data derived by assimilating all available data — both station data and remotely sensed data — into coupled land-surface/atmosphere models. Such model-calculated data may result

from extensions of ongoing re-analysis exercises (e.g. Gibson *et al.*, 1997; Kalnay *et al.*, 1996; Kistler *et al.*, 2001) or through global LDAS initiatives (e.g. Rodell *et al.*, 2004). There is an obvious opportunity for the WCRP's Global Energy and Water Experiment (GEWEX) and IAHS's Prediction in Ungaged Basins (PUB) programme to work together to meet this important need. Adoption of this worthy and achievable goal by these two international programmes is strongly recommended.

In part, the two-step approach when estimating crop water requirements also remains locked in place because no literature yet defines the effective all-day average stomatal and aerodynamic resistance of irrigated crops equivalent to that available for crop factors. Echoing Shuttleworth (1993a), research in this area is again called for here. Meanwhile, using assumptions equivalent to those involved in the current definition of reference crop evaporation, Shuttleworth (2006) provides the theoretical framework to translate currently available crop factors into estimates of the effective surface resistance, $r_{s,crop}$, required in Eqn. (24). In practice, several crops have a crop height greater than 2m and it is therefore preferable to deduce the value of $r_{a,crop}$ and subsequently to apply this value to make calculations of crop evaporation using the values of wind speed and vapour pressure deficit applicable at a 'blending height' which Shuttleworth (2006) selects to be at 50 m. However, the values of wind speed and vapour pressure deficit measured at a reference height of 2 m can still be used (see Shuttleworth (2006) for greater detail) providing changes are made in the definition of $r_{a,crop}$ and $r_{a,rc}$, and providing the vapour pressure deficit at 50 m is estimated and then used in Eqn. (24), as follows:

$$r_{a,crop}^{50} = \frac{\ln[(50 - 0.67h)/(0.123h)] \cdot \ln[(50 - 0.67h)/(0.0123h)]}{1.67 \cdot (0.41)^2 U_2} = \frac{R_{crop}}{U_2} \quad (\text{s}\ \text{m}^{-1}) \quad (30)$$

$$r_{a,rc}^{50} = \frac{R_{rc}}{U_2} = \frac{302}{U_2} \quad (\text{s}\ \text{m}^{-1}) \quad (31)$$

and

$$E_{crop} = \frac{\Delta (R'_n - G')_{crop} + \gamma \frac{230431}{(275 + T)} \frac{U_2 D_{50}}{R_{crop}}}{\Delta + \gamma \left[1 + \frac{r_{s,crop} U_2}{R_{crop}} \right]} \quad (\text{mm}\ \text{day}^{-1}) \quad (32)$$

In general, the value of D_{10} required in Eqn. (32), is calculated from D_2 using the equation:

$$\frac{D_{50}}{D_2} = f_D + \frac{(f_D r_{a,rc}^2 - r_{a,rc}^{50})}{r_{clim}}$$

with:

$$r_{\text{clim}} = \frac{187342 \cdot \gamma \cdot D_2}{\Delta(R'_n - G')(275 + T)} \quad (34)$$

and:

$$f_D = \frac{(\Delta + \gamma)r_{a,rc}^{50} + \gamma r_{s,rc}}{(\Delta + \gamma)r_{a,rc}^2 + \gamma r_{s,rc}} \quad (35)$$

Thus, Equation (32) becomes the equation to be used to make a one-step estimate of crop-water requirement from crop-specific values of $r_{a,crop}$ and R_{crop} . Equation (30) calculates R_{crop} from crop height using Equation (30) while the required value of $r_{s,crop}$ can be estimated (Shuttleworth, 2006) from the crop coefficient, K_{crop} , using the equation:

$$r_{s,crop} = \frac{Y}{K_{crop}} \cdot \frac{\gamma r_{s,rc} + (\Delta + \gamma)r_{a,rc}^{50}}{\gamma} - \frac{(\Delta + \gamma)r_{a,crop}^{50}}{\gamma} \quad (36)$$

with

$$Y = \left[\frac{f_A r_{a,crop}^{50} + (D_{50}/D_2)_{\text{specific}} r_{\text{clim}}^{\text{specific}}}{r_{a,rc}^{50} + (D_{50}/D_2)_{\text{specific}} r_{\text{clim}}^{\text{specific}}} \right] \quad (37)$$

with:

$$f_a = \frac{\Delta(R'_n - G')_{rc}}{\Delta(R'_n - G')_{crop}} \quad (38)$$

where the factor f_A allows for the possibility that the energy available to support evaporation may differ between a crop and the reference crop if, for instance, they have a different albedo or if there is energy advected in the atmosphere from elsewhere, and:

$$r_{\text{clim}}^{\text{specific}} = 60 \text{ (s m}^{-1}\text{)} \quad (39)$$

When defining this value of $r_{\text{clim}}^{\text{specific}}$, Shuttleworth (2006) assumes $r_{s,crop}$ should be calculated from K_{crop} in conditions when the radiative and advective contributions to reference crop evaporation are such that the FAO (Allen et al., 1998) and Priestley-Taylor estimates of reference crop evaporation are equal. Consequently, *in this specific case* Equation (33) becomes:

$$\left(\frac{D_{50}}{D_2} \right)_{\text{specific}} = f_D + \frac{(f_D r_{a,rc}^2 - r_{a,rc}^{50})}{60} \quad (40)$$

For the purposes of illustration, Table 1 shows the results of calculating $r_{a,crop}$ and R_{crop} for sample crops for a range of wind speeds and crop factors assuming $T = 15 \text{ }^\circ\text{C}$, $P = 100 \text{ kPa}$, and $U_2 = 2 \text{ m s}^{-1}$.

RECOGNISING LAND-SURFACE/ATMOSPHERE COUPLING

Continental-scale representation of land-surfaces

In the 1960s and 1970s, the meteorological community was preoccupied primarily with weather forecasting a few days ahead; in pursuit of this goal, primary importance was assigned, correctly, to defining an adequately realistic description of the initial state of the atmosphere and secondary importance to the influence of surface-atmosphere exchanges. Later, as weather forecasting skill increased, recognition that surface influences also merit attention in weather forecasts grew, as discussed below. Meanwhile, a portion of the meteorological community recognised that if the simulation of weather phenomena in the General Circulation Models (GCMs) used for weather forecasts is reasonably realistic, running GCMs for many years ahead should allow simulation of the *statistics* of weather, i.e. climate, even if such models cannot correctly predict the actual weather more than a few days ahead. As a result of this realisation, a new branch of meteorological science was born (e.g. Charney *et al.*, 1977; Walker and Rowntree, 1977; Shukla and Mintz, 1982; Sud *et al.*, 1985). In the case of GCMs that are being used for climate prediction, the relative importance of initial conditions and surface exchanges is reversed relative to their priority for weather forecasting. The initial state of the modelled atmosphere soon becomes irrelevant but the correct representation of the several inputs to the atmosphere becomes critical because this is what determines future climate. As a global average, about half of the energy input to the atmosphere comes from below (MacCracken, 1985), i.e. half of the radiation from the sun permeates the atmosphere and reaches the earth's surface whence it is returned to the atmosphere as short-wave and long-wave radiation, or as sensible heat to warm the overlying air, or as latent heat in the evaporate water that moistens the atmosphere. Also, the momentum balance of the atmosphere is achieved as the equilibrium between atmospheric pressure forces, the local Coriolis force, and the frictional force associated with the exchange of momentum with the underlying surface. Consequently, if the efficiency of the exchange of momentum with the ground is altered as it might be if, for instance, an extensive rough (forest vegetation) cover is changed to a smoother covering of pasture or agricultural crops, the strength of convergence in the atmosphere may alter (decrease) and the precipitation mechanisms supported by convergence, and precipitation itself, may also alter (decrease). Recognition that the description of surface exchanges of energy and momentum (and later, CO_2) between continental surfaces and the modelled atmosphere must be improved in GCMs for

Table 1. Values of R_{crop} (dimension) and $r_{s,crop}$ (in m s^{-1} given in the main body of the table) calculated from Eqns. (30) and (37), respectively, for a range of wind speeds and crop factors assuming $T = 15^\circ\text{C}$, $P = 100\text{ kPa}$, and $U_2 = 2\text{ m s}^{-1}$.

Crop height (m)	R_{crop} (m s^{-1})	K_{crop} (dimensionless)															
		0.30	0.40	0.50	0.60	0.70	0.75	0.80	0.85	0.90	0.95	1.00	1.05	1.10	1.15	1.20	1.25
0.12	255	1170	777	541	384	272	227	188	153	122	95	70	47	27	8	0	0
0.3	195	1197	793	551	389	274	228	188	152	120	92	67	44	23	3	0	0
0.4	178	1197	734	516	370	266	225	188	156	127	102	79	58	39	22	6	0
0.5	165	1197	701	496	359	262	223	188	158	131	107	86	66	48	32	17	4
0.6	155	1197	679	483	352	259	221	189	160	134	111	90	72	55	39	25	12
0.7	146	1197	663	473	347	256	220	189	161	136	114	94	76	59	44	31	18
0.8	139	1197	649	465	342	255	219	189	162	138	116	97	79	63	49	35	23
0.9	133	1197	638	458	339	253	219	189	162	139	118	99	82	66	52	39	27
1.0	127	1197	629	453	336	252	218	189	163	140	119	101	84	69	55	42	30
1.2	118	1197	621	448	333	251	218	189	164	141	121	103	86	71	58	45	33
1.4	110	1197	614	444	331	250	217	189	164	142	122	104	88	73	60	47	36
1.6	104	1197	601	436	327	248	217	189	165	143	124	107	91	77	64	52	41
1.8	98	1197	591	430	323	247	216	189	165	144	126	109	93	79	67	55	44
2.0	93	1197	583	425	320	245	215	189	166	145	127	110	96	82	69	58	48
2.5	82	1197	575	421	318	244	215	189	167	146	128	112	97	84	72	61	50
3.0	73	1197	569	417	316	244	215	189	167	147	129	113	99	86	74	63	53
3.5	66	1197	555	409	311	242	214	189	168	149	132	116	102	90	78	67	58
4.0	60	1197	544	402	308	240	213	189	169	150	133	119	105	93	81	71	62
4.5	54	1197	535	397	305	239	213	190	169	151	135	120	107	95	84	74	65
5.0	49	1197	528	393	302	238	212	190	170	152	136	122	109	97	87	77	68
6.0	41	1197	521	389	300	237	212	190	170	153	137	123	111	99	89	79	70
7.0	33	1197	515	385	298	236	211	190	171	153	138	125	112	101	91	81	72
8.0	27	1197	505	379	295	235	211	190	171	155	140	127	115	104	94	85	76
9.0	20	1197	497	374	292	234	210	190	172	156	141	128	117	106	96	87	79
10.0	14	1197	490	370	290	233	210	190	172	156	142	130	118	108	98	90	82

climate simulations spawned the interest in new LSPS described above.

LSPs were becoming increasingly sophisticated by the 1980s, due to improved understanding of the evaporation process. However, the adequacy of the estimated parameters that were included in this new generation of LSPs remained an issue. Field studies were clearly required to investigate the reliability of LSPs when describing large and comparatively uniform natural biomes, as well as when seeking to describe simply and efficiently the (usually heterogeneous) land cover that often results when natural vegetation is substantially modified by human land management practices. By the early 1980s, the potential impact on regional and perhaps global climate of large scale deforestation in the tropics, where much of the sun's energy is received, was well recognised. Perhaps inevitably, therefore, the first field study specifically mounted to test the capability of LSPs to represent vegetation took place in the Brazilian Amazon (Shuttleworth *et al.*, 1984a,b,c;

Henderson-Sellers and Shuttleworth, 1987). Facilitated by the new capability to make routine measurements of surface energy and momentum fluxes using the eddy-correlation technique, multi-year measurements were made of the net exchanges and the component processes involved in the surface-atmosphere exchanges for tropical rainforest (Shuttleworth, 1988b, 1989; Shuttleworth *et al.*, 1991). These data were subsequently used to recalibrate and/or validate the representation of this biome by LSPs (e.g. Sellers, *et al.*, 1989; Dickinson, 1989; Shuttleworth and Dickinson, 1989) as the basis for predicting the effect on climate of Amazonian deforestation (e.g. Lean and Warrillow, 1989; Lean and Rowntree, 1993; Shukla, *et al.*, 1990; Nobre *et al.*, 1991).

An important feature of GCMs is that they normally represent the land surface using a grid that has a length scale of hundreds of kilometres, and often seek to do this with a single LSP to describe the vertical exchanges as though they arose at a single point at the centre of each rectangle in the

grid (e.g. Shuttleworth, 1991a). This posed a new challenge to hydrologists used to thinking at the plot or field scale. How can land–surface interactions be adequately represented at the scale of a large GCM grid square, biome, or large river basin? This new challenge motivated a series of large-scale studies of land–atmosphere exchanges which involved the simultaneous deployment of surface measurements and airborne and satellite sensors along with modelling studies to synthesise and study the area-average representation surface exchanges.

Table 2, which draws heavily on Dirmeyer and Hoff (2004), lists most of the large-scale, land-based field experiments undertaken between 1984 and 2004 that gave emphasis to surface–atmosphere energy, momentum, and/or CO₂ exchanges. Some early experiments explored the use of remotely sensed data to upscale land–surface interactions, e.g. FIFE (Sellers and Hall, 1992; Hall and Sellers 1995) and BORES (Sellers *et al.*, 1995; Hall, 1999). Other experiments provided simultaneous data for important vegetation covers across large biomes, e.g. ABRACOS (Shuttleworth *et al.*, 1991; Gash and Nobre, 1997). One important class of experiments, the Mesoscale Field Experiments (MSEs), e.g. HAPEX-MOBILHY (Andre *et al.*, 1988), EFEDA (Bolle *et al.*, 1993), HAPEX-SAHEL (Goutorbe *et al.*, 1994, 1997) emphasised gathering data over an area comparable to that of a GCM grid. Subsequently, the spatial scale of these experiments increased and this resulted in a suite of Continental-Scale Experiments (CSEs), e.g. GCIP (Coughlan and Avissar, 1996; Lawford 1999), BALTEX (Raschke *et al.*, 1998), LBA (LBA, 1996), GAME (Yasunari, 1993) and MAGS (Stewart *et al.*, 1998). One MSE, the Atmospheric Radiation Measurement Southern Great Plains study (ARM-SGP: DOE, 1996) emphasised the parameterisation of radiation processes in the atmosphere.

Over the years there have been attempts, progressively, to evaluate the accumulating understanding that has resulted from this series of large-scale land-surface studies (Shuttleworth 1988c; 1991a,b; 1993b; Shuttleworth and Gash, 2005), including a recent substantial retrospective synthesis carried out as an initiative of the International Geosphere-Biosphere Programme's core project, Biospheric Aspects of the Hydrological Cycle (BAHC; see Gash and Kabat, 2004), to which the reader is referred for greater detail. At a minimum, these many land-surface studies have certainly greatly increased the availability of data for calibrating and evaluating LSP models, and the likely realism of the simulation of surface exchanges in GCMs has benefited, progressively, as these data became available. As an example, Fig. 8 shows the results of a study (Sen *et al.*, 2001) in which data from several land surface studies

were used to calibrate an advanced LSP's representation of some of the most important biomes. The results of this study demonstrate that significant changes in modelled climate result from using recalibrated LSP parameters rather than the parameter values previously assumed.

Just as climate models benefited from large-scale land–atmosphere studies, so did the models used for numerical weather prediction. In practice, the most significant beneficial impact arose from the CSEs, because participation by a regional modelling centre with a weather forecast model was an essential requirement in these experiments. In each CSE, members of the weather forecast modelling community therefore worked directly and regularly with hydrologists and had ready access to relevant observations and land-surface modelling expertise. Weather forecast models with comparatively simple representation of land-surface exchanges were confronted with new experimental data and, in part because the land-surface models were often initially so simple, it was possible to make rapid and dramatic improvements in performance. Thus, progressive, incremental improvement in weather forecast models persisted throughout the CSEs and sometimes more significant steps forward occurred. One notable example of progress was the dramatic improvement in the European Centre for Medium-range Weather Forecasting's (ECMWF) model prediction of heavy flooding in the central Mississippi River basin that resulted from a comparatively small improvement in the model's representation of land-surface exchanges (Beljaars *et al.*, 1996).

Mesoscale land-surface/atmosphere coupling

The simultaneous deployment of multiple plot-scale measurements of surface energy and water exchanges, atmospheric boundary layer sounding systems and aircraft measurements during MFEs, and the associated mesoscale meteorological modelling activity, has generated new understanding of land-surface/atmosphere coupling at the mesoscale. Arguably the most significant early insight was that the nature of this coupling depends profoundly on the length scale of the heterogeneity present in the underlying land-surface cover. This result was first recognised by De Bruin (1989) but, as a result of publication delays in De Bruin's paper, it is sometimes wrongly ascribed to the present author (Shuttleworth, 1988b). If the land-surface is patchy and heterogeneous at length scales of less than about 10 km, then, to a reasonable approximation, each patch of vegetation can be assumed to interface separately at some assumed *blending height* and then to move upwards into the well-mixed portion of the overlying atmospheric boundary layer (ABL) which is typically 0.5–3.0 km deep in daytime conditions. However, if the structure of the

Table 2. List of most of the major international large-scale field experiments that were undertaken between 1984 and 2004 that gave emphasis to surface-atmosphere energy, momentum, and/or CO₂ exchanges [taken from Shuttleworth and Gash, 2005].

<i>Name</i>	<i>Location</i>	<i>Perior</i>	<i>References</i>
Hydrological and Atmospheric Pilot Experiment – Modelization du Bilan Hydrique (HAPEX-MOBILHY)	Southern France	1985–1987	André <i>et al.</i> (1989)
First ISLSCP Experiment Field Experiment (FIFE)	Central Kansas, USA	1987–1989	Sellars and Hall (1992) and Hall and Sellars (1995)
Regio-Klima-Projekt (REKLIP)	Middle and southern upper Rhine Valley	1989	Parlow (1996)
Anglo-Brazilian Climate Observation Study (ABRACOS)	Manaus, Ji-Paraná and Marabá, Brazil	1990–1995	Gash and Nobre (1997)
Hydrological and Atmospheric Pilot Experiment in the Sahel (HAPEX-Sahel)	Western Niger	1991–1993	Goutourbe <i>et al.</i> (1994)
European International project on Climatic and Hydrological Interactions between Vegetation, Atmosphere and Land Surface (ECHIVAL) Field Experiment in Desertification Threatened Areas (EFEDA)	South-eastern Spain	1991–1995	Bolle <i>et al.</i> (1993)
Hei Ho River Basin Field experiment (HEIFE)	Gansu Province, China	1992–1993	Wang <i>et al.</i> (1993)
Atmospheric Radiation Measurement Southern Great Plains site	Kansas & Oklahoma, USA	1992	http://www.arm.gov/docs/sites/sgp/sgp.html
Boreal Ecosystem-Atmosphere Study (BOREAS)	Central Canada	1993–1996	Sellars <i>et al.</i> (1995), Hall (1999)
Mackenzie GEWEX Study (MAGS) Observation at Several Interacting Scales (OASIS)	Mackenzie River basin, Canada Murray-Darling basin, Australia	1994–1995 1994–1995	Stewart <i>et al.</i> (1998) http://www.clw.csiro.au/research/waterway/interactions/oasis
Northern Hemisphere Climate Processes Land Surface Experiment (NOPEX)	Central Sweden	1994–1996	Halldin <i>et al.</i> (1999)
Baltic Sea Experiment (BALTEX)	Baltic Sea basin	1994–2001	Raschke <i>et al.</i> (1998)
Monitoring the Usable Soil Reservoir Experimentall (MUREX)	South-western France	1995–1997	Calvet <i>et al.</i> (1999)
GEWEX Continental-scale International Project (GCIP)	Mississippi River basin, USA	1995–2000	Couglan and Avissar (1996), Lawford (1999)
GEWEX Asian Monsoon Experiment (GAME)	Siberia, Tibet, Thailand, Huaihe River basin, China	1996	Yasunari (1993)
Large-Scale Biosphere-Atmosphere Experiment in amazonia (LBA)	Amazon region of South America	1996	LBA Science Planning Group (1996)
Inner Mongolia Grassland Atmosphere Surface Study (IMGRASS)	Xilnhot, Inner Mongolia	1997–2000	http://iap.ac.cn/english/iap/Divisions/LAGHO.htm
Southern Great Plains (SGP) Semiarid Land-surface-Atmosphere Program (SALSA)	Oklahoma and Kansaa, USA Upper San Pedro River	1997, 1998, 2001 1997–1998	http://hydrolab.arsuada.gov/sgp97/ Goodrich <i>et al.</i> (2000)
Couplage del'Atmosphere Tropicale et du Cycle Hydrologique (CATCH)	Niger, Benin	2000	http://lwww.lthe.hug.inpg.fr/catch
GEWEX Americas Prediction Project (GAPP)	USA	2000–2005 gapp/index.htm	http://www.ogp.naa.gov/mpe/gapp/gapp/index.htm
Cordinated Enhanced Observint Period (CEOP)	Worldwide	2001–2004	http://www.ceop.net/

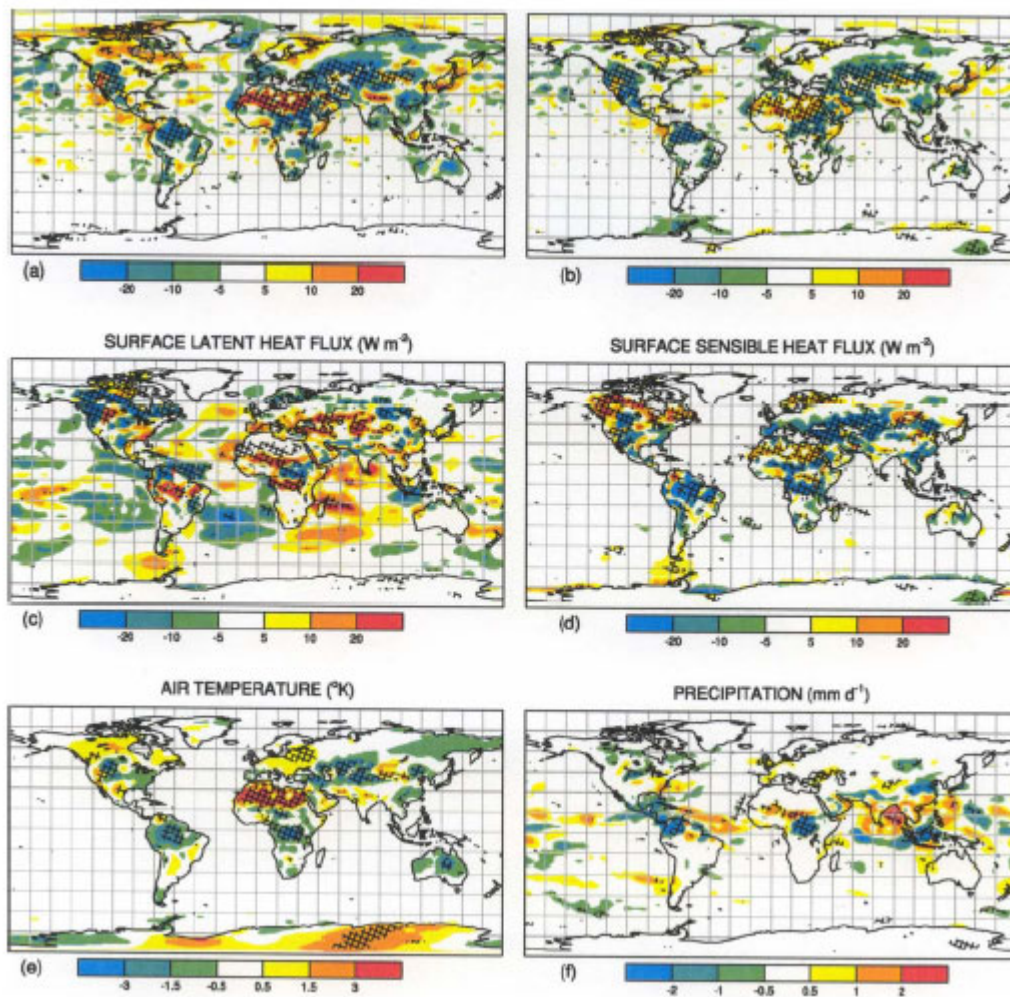


Fig. 8. Eight-year average values for June, July, and August of differences in modelled (a) net solar flux, (b) net radiation, (c) latent-heat flux, (d) sensible heat flux, (e) air temperature, and (f) precipitation between simulations using BATS2 with parameters recalibrated using field data for the tropical forest, boreal forest, semi-desert, mixed crops and farming, and short grass biomes relative to using default parameters. The hatched areas indicate where these differences are significant at the 95% confidence level [Taken from Sen et al. (2001)].

underlying surface is such that there is heterogeneity in surface cover with length scale of tens of kilometres, the ABL is able to recognise and responds to this heterogeneity, and additional mesoscale circulations are generated which can substantially enhance mixing and convection in the atmosphere, perhaps generating additional clouds and/or precipitation.

The result that land surfaces with both small-scale heterogeneity and, paradoxically, with uniform vegetation cover can be assumed to interface with the well-mixed portion of a spatially uniform ABL at an assumed blending height, simplifies representation of the land-surface/atmosphere interaction at the grid scale used in GCMs. In effect, it is assumed as a first approximation that each patch of vegetation (or the uniform vegetation if there is no heterogeneity) responds to the area-average state variables

that characterise the overlying atmosphere at the blending height. Thus, the exchanges for each patch of surface cover can, in principle, be calculated separately using an LSP for each patch, and these can then be assumed to merge with the area-average surface exchange then being merely an average of the exchange for all the patches. Some present day LSPs (e.g. Avissar and Pielke, 1989; Koster and Suarez, 1992) implement this approach directly and seek to include a *mosaic* of LSPs, with each of the most common land covers present in a GCM grid separately described by an LSP. However, for many years, the computational demands associated with implementing this approach fostered the alternative approach of retaining a single LSP and seeking to define effective, *aggregate* parameters for use in it that approximately capture area-average behaviour. In fact, many GCM models still use a single LSP in each grid square and

either assume the parameters of the most common land cover present in that grid square, or use aggregate parameters.

Use of the aggregate LSP parameter approach requires that hypothetical aggregation rules are postulated (e.g. Wieringa, 1986; Mason, 1988, Shuttleworth *et al.*, 1991; Raupach, 1995; Arain *et al.*, 1997; Shuttleworth *et al.*, 1997) and then tested by comparing area-average exchanges calculated using these rules with those calculated by a mesoscale meteorological model in which each patch is represented separately (Arain *et al.*, 1996). Results show that simple aggregation rules can improve simulation of the area-average behaviour of heterogeneous surfaces. One notable result (Burke *et al.*, 2000) of using the aggregate parameter approach to calculate aerodynamic roughness, for example, showed that significant changes in global circulation can result, including a northward shift of the northern hemisphere jet stream from the change in the area-average roughness of the Boreal forest biome (Fig. 10).

Meanwhile, other mesoscale modelling studies have been

used to investigate the consequences of surface cover that is heterogeneous at length scales of tens of kilometres (Giorgi and Avissar, 1997; Pielke, 2001). Studies suggest that such cover can generate enhanced convection and storms (e.g. Avissar and Liu, 1996) and that this effect can occur over extensive areas (Lu and Shuttleworth, 2002). Mesoscale modelling studies have also investigated the potential for defining semi-empirical functions to represent the enhanced mixing and convection generated by the mesoscale circulations associated with surface heterogeneity (Liu *et al.*, 1999). However, this approach has yet to be widely adopted, perhaps because growth in the availability and capability of computers has allowed the grid scale used in mesoscale models to fall to the extent that mesoscale circulations are now beginning to be resolved explicitly in models. It is interesting and not inconsistent that in flat semi-arid areas where relief is limited and where there is much exposed bare soil (and consequently rapid evaporation after rain), results from the HAPEX-Sahel MFE suggest that the

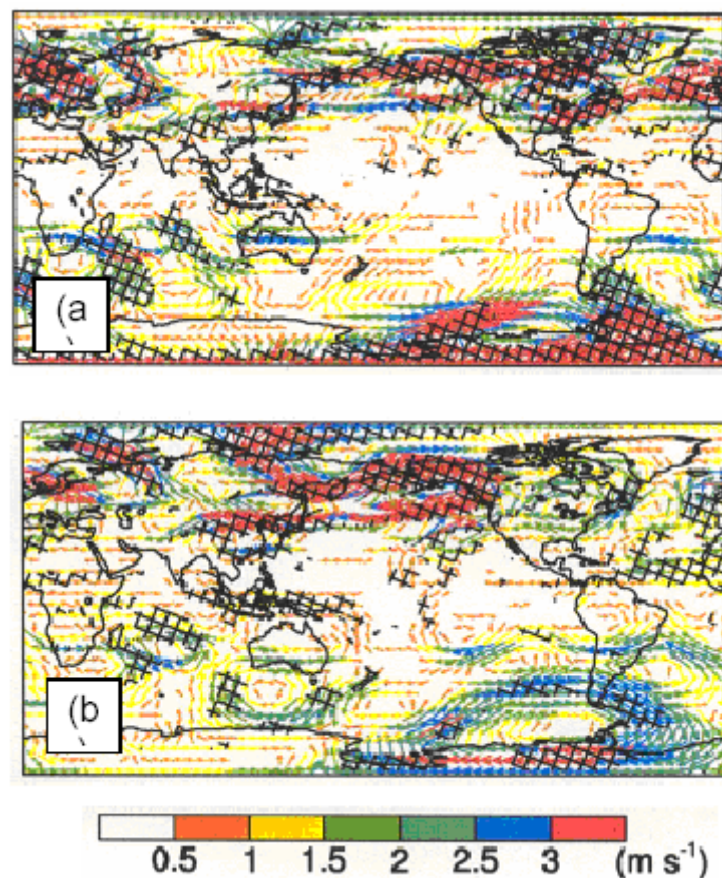


Fig. 10. Modelled 300 mb wind vector in $m s^{-1}$ for a 10 year simulation with the CCM3-BATS model for June, July, and August. (a) shows differences when using aggregate parameters defined following Shuttleworth *et al.* (1997) relative to when using default parameters in BATS; (b) shows the difference when using only the aggregation rule for aerodynamic parameters relative to when using equivalent default parameters in BATS. The hatched areas indicate areas where these differences are significant at the 95% confidence level. [Taken from Burke *et al.* (2000).]

heterogeneity of convective precipitation can itself also stimulate later mesoscale convection and persistent rainfall patterns, with rain falling preferentially where it has fallen before (Taylor and Lebel, 1998; Clark *et al.*, 2003). However, in semi-arid regions where there are significant topographic features, the heterogeneity in surface heating and atmospheric ascent associated with topography is a major, perhaps dominant influence on the initiation of convection and precipitation (e.g. Carbone *et al.*, 2002; Crook and Tucker, 2005; Tucker and Crook, 2005).

In the past few decades, greater understanding of the interaction between the overlying ABL and land-surfaces with uniform land cover or cover with small scale heterogeneity (Type ‘A’, see Shuttleworth, 1988b) has substantially improved hydrologists’ understanding of why the concept of potential evapotranspiration served as a comparatively successful early approximation of the rate of evaporation for a range of non-extreme atmospheric and surface conditions. Important papers documenting this growth in understanding include those by McNaughton (1976), De Bruin (1983, 1989) and McNaughton and Spriggs (1986, 1989). The phenomenon is fundamentally

associated with the fact that evaporation occurs primarily during the day into an ABL that is partially, but not totally, contained by a stable inversion aloft, with the ABL underlying a free atmosphere that is usually drier and (in terms of potential temperature) cooler than the ABL. It is helpful to explain the processes involved by considering evaporation into a saturated atmosphere. Figure 11(a) represents evaporation from an open water surface into a well-mixed but well-contained and thermally insulated atmosphere in response to an input of energy. In this situation, the air remains saturated and the incoming energy (shown as radiant energy) is shared in such a way as to retain saturated air by providing latent heat for the evaporation while also warming the air so that it can contain progressively increasing amounts of humidity. The Bowen ratio of the evaporating free water surface is the ratio of the energy used to warm the air relative to that used to evaporate water, and it is easily shown from the energy balance equation that the resulting evaporation rate (when expressed as latent heat flux) is given by:

$$\lambda E = \frac{\Delta}{\Delta + \gamma} R_n \quad (\text{W m}^{-2}) \quad (41)$$

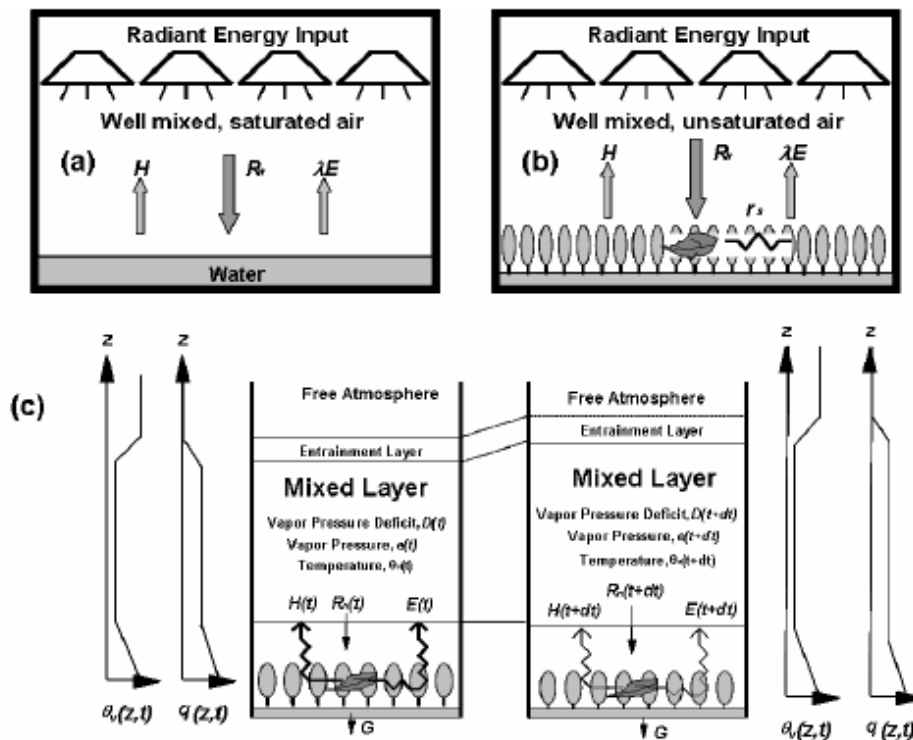


Fig. 11. Schematic diagram illustrating (a) evaporation from an open water surface into a well-mixed but well-contained and thermally insulated atmosphere in response to an input of energy; (b) evapotranspiration from plants into a well-mixed but well-contained and thermally insulated atmosphere in response to an input of energy with the transpiration controlled by a surface resistance (Note: the aerodynamic resistance is assumed to be zero because the air is well-mixed); (c) evapotranspiration into a partly contained growing Atmospheric Boundary Layer underlying a warmer, drier free atmosphere. The process is illustrated in terms of (simplified) atmospheric profiles of virtual potential temperature and specific humidity at two successive times.

where R_n is the radiant energy input to the surface per unit area. Figure 11(b) illustrates a situation similar to Fig. 11(a) except in this case the evaporated water is transpiration from plants and controlled by surface resistance (Note: the aerodynamic resistance is assumed to be zero because the air is well-mixed). The overall energy balance is the same as for the free water case and the evaporation rate is again given by Eqn. (35) but, in this case, the atmosphere maintains a finite *equilibrium vapour pressure deficit*, d_{eq} , so that the transpiration flux can pass through the stomatal resistance, r_s , of the plants. The value of d_{eq} is at all times given by:

$$\delta_{eq} = \left(\frac{\gamma r_s}{\rho C_p} \right) \frac{\Delta}{\Delta + \gamma} R_n \quad (\text{k Pa}) \quad (42)$$

Figure 11(c) illustrates the process of evapotranspiration into a partly contained growing ABL underlying a warmer, drier free atmosphere. The process is illustrated in terms of (simplified) atmospheric profiles of virtual potential temperature and specific humidity at two successive times. The ABL is classified in layers: a fairly shallow (~ 50 – 200 m) *surface layer*, which is strongly influenced by the surface and through which surface fluxes pass by turbulent diffusion; a thicker (~ 500 – 3000 m) *mixed layer* where the air is very well mixed by larger-scale turbulence and where atmospheric gradients of temperature, humidity and wind speed are low; and a fairly shallow (~ 100 – 200 m) *entrainment layer*, which forms the interface between the mixed layer and the *free atmosphere* above. Air is mixed downwards, or entrained, from the drier, warmer free atmosphere into the mixed layer at a rate that depends on the strength of temperature gradient through the inversion but also on the surface sensible heat flux (strictly, buoyancy flux). The mixed layer grows in depth through the day as a result of the entrainment of air from above, and the warmer, drier air entrained at the top of the mixed layer is moistened by the water vapour which enters the mixed layer by surface evaporation and is mixed upwards. Thus, the specific humidity of the air in the mixed layer is fairly constant and changes only slowly through the day. However, the temperature of the mixed layer rises because of the incoming energy both from the warmer free atmosphere above and from the surface sensible heat flux. Consequently, the vapour pressure deficit of the mixed layer rises and is higher than that given by Eqn. (36) and the evaporation rate exceeds the Equilibrium Evaporation rate, typically by about 25%. In practice, simple modelling experiments that describe the daytime evolution of the ABL (e.g. De Bruin, 1983, 1989; McNaughton and Spriggs, 1986; 1989) show that for a range of surface resistances typical of short agricultural crops and for typical start-of-day atmospheric profiles, the daily average evaporation rate can approximate the estimates

given by both Eqns. (1) and (2). Figure 12 illustrates this in the case of Eqn. (2) by showing how the modelled value of α in the general equation:

$$\lambda E = \alpha \cdot \frac{\Delta}{\Delta + \gamma} R_n \quad (43)$$

varies as a function of the surface resistance for a set of initial atmospheric profiles measured at Cabauw in the Netherlands, i.e., in a climate where potential evapotranspiration might traditionally be expected to apply. For surface resistances in the 'agricultural' range (e.g. 20 – 100 s m^{-1}), the modelled value of α is within about 10% of the value 1.26.

Thus, we have come full circle. Over the last 30–35 years, there has been huge growth in detailed understanding of the surface controls on the natural evaporation process associated with the partition of energy and within canopy and near-surface controls on transfers. There has also been huge growth in understanding relevant atmospheric processes and how land-surface and atmospheric processes are coupled at the mesoscale and continental and global scales. In the case of processes that couple the land-surface to the Atmospheric Boundary Layer (ABL), this has enabled hydrologists to recognise why in certain (not uncommon)

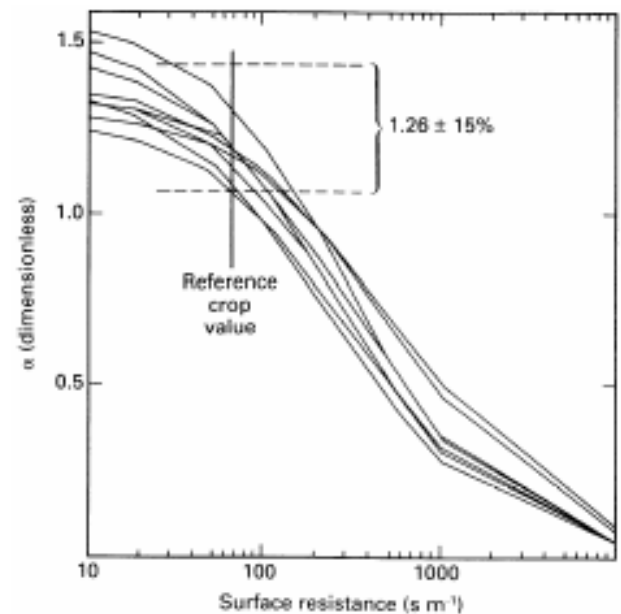


Fig. 12. Values of the constant α in Eqn. (37) calculated using a 'slab' model of the atmospheric boundary layer (McNaughton and Spriggs, 1989) with eight different initial atmospheric profiles measured in a climate where potential evapotranspiration might traditionally be expected to apply at Cabauw in the Netherlands. For surface resistances in the "agricultural" range (e.g., 20 – 100 s m^{-1}), the modeled value of α is within about 10% of the value 1.26, which is the value used in the Priestley-Taylor equation, Eqn. (2).

atmospheric conditions and for certain land covers and surface water availability, some of the more physically-based (but still essentially empirical) relationships that pioneering scientists derived to relate evaporation to measured climate can provide a reasonable first estimate of evaporation rate, although they are not general enough for universal application.

Concluding remarks

This paper describes the origin and development of the science of natural evaporation from land-surfaces over the last 30–35 years since the original *A View from the Watershed* symposium was held to commemorate the opening of the new IH building at Wallingford in 1973. In the course of this review, it becomes apparent that much of the progress that has subsequently occurred had its origin in technology and ideas that first arose at IH in the 1970s and 1980s, when Jim McCulloch was Director of IH.

On the basis of this review, it is clear that very substantial progress was made over the last 30–35 years; but what of the future? Although much is now known, much remains to be known, and not all that is already known is yet being applied. Arguably, the meteorological community is still ahead of the hydrological community in applying currently available near-surface process understanding of natural evaporation in the form of the LSPs in meteorological models. However, it is encouraging that hydrologists are starting to use improved versions of LSPs in hydrological models; these surpass those used by their meteorological colleagues because they include representation of groundwater and surface water flows (e.g. Overgaard, 2006a). Although the procedures used to estimate crop water requirements to schedule irrigation have advanced through the decision to specify reference crop evaporation in terms of more basic physical and physiological understanding, the recommended procedure for estimating actual evaporation still involves the use of traditional crop factors. This paper strongly recommends that attempts be made to make retrospective and near-real-time model-calculated surface climate fields of weather variables available for estimating crop water requirements by assimilating available relevant station and remotely-sensed data into coupled land-surface/atmosphere models. Perhaps this could be achieved through ongoing re-analysis programmes or through global LDAS initiatives. It is also recommended that the community responsible for defining the preferred method for estimating crop water requirements takes the next logical step of proposing that estimates of crop water requirements be made by defining an effective surface resistance, rather than an effective crop factor. Portions of the present paper describe

how existing empirical crop factors can be re-expressed as the required effective surface resistance and it is hoped that providing this capability will stimulate this recommended change.

There are many needs for future research in fields of earth system science that require understanding of the natural evaporation process but, here, three are highlighted. The first relates to the fact that evaporation is linked, fundamentally, to the availability of water and there are features that act to introduce unavoidable heterogeneity into the spatial patterns of available soil moisture that are poorly recognised and understood in models of evaporation. Perhaps the most obvious source of spatial heterogeneity in soil moisture is convective rainfall. A second important source of heterogeneity is micro-topography, i.e. topography that concentrates rainfall into soil moisture distributions that are heterogeneous, with a length scale of tens of centimetres to hundreds of metres. Although modelling of both the patterns of convective precipitation and the influence of micro-topography on evaporation via soil moisture availability might be possible given sufficient computer resources and time, this is not necessarily the only or best way forward. As an alternative, some combination of a statistical representation of spatial patterns of precipitation and topographic forcings and a deterministic representation of the resulting effect on evaporation via a small-scale mosaic of LSPs may be capable of parameterising such surface heterogeneity.

A second area where there remains a very substantial need for new understanding is in measuring, understanding and modelling the evaporative losses associated with frozen precipitation. Currently, canopy interception processes for snow are poorly understood, and the evolution of intercepted snow on canopies is inadequately described. There is also little understanding of the evaporative losses associated with snow-blow, and poor understanding and modelling of evaporation from snow packs and from snow and ice under vegetation canopies. Similarly, the capability to model evaporation from melting snow and ice is poor, especially in cases where the melting snow/ice cover is patchy. This poor understanding of evaporation for frozen precipitation is symptomatic of a more general dearth of knowledge about hydrological and water balance processes in cold environments. Given the profound importance of frozen precipitation as a source of water for human use in many areas, priority should be given in research and funding to acquire knowledge on the process hydrology of snow and ice, especially in mountainous regions.

A third area where research is needed is the effect of direct or indirect human intervention on the evaporation process. Agricultural scientists and foresters continue to seek

improved productivity through selective breeding, and it is plausible to assume that such changes are associated with modified stomatal control or with changes in the phenology of plants, both of which could affect transpiration. Deliberate or unintentional changes in nutrients available to plants may well affect plant vigour and, through this, affect the controls exerted by the plants on transpiration. Some coupling between biogeochemical models and evaporation models will be needed to investigate the importance of this effect. Finally, the inevitable future growth in atmospheric CO₂ will surely alter the behaviour and water requirements of vegetation. Thus, increasing CO₂ will affect hydrology not only through atmospheric physics by changing precipitation and radiation but also through the ecological and physiological changes induced in plant communities on the ground.

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