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Land-surface modelling in hydrological perspective

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Abstract

A comprehensive review of energy-based land-surface modelling, as seen from a hydrological perspective, is provided. We choose to focus on energy-based approaches, because in comparison to the traditional potential evapotranspiration models, these approaches allow for a stronger link to remote sensing and atmospheric modelling. New opportunities for evaluation of distributed land-surface models through application of remote sensing are discussed in detail, and the difficulties inherent in various evaluation procedures are presented. Remote sensing is the only source of distributed data at scales that correspond to hydrological modelling scales. Finally, the dynamic coupling of hydrological and atmospheric models is explored, and the future perspectives of such efforts are discussed.

1. Introduction

With the growing population of the Earth and predicted changes in the global climate, the pressure on the already scarce water resource is likely to increase in the coming years. This has created a need for integrated models that can assess the available water resource as well as predict the impact of future changes in management and climate. Making such accurate predictions is an immense task that can only be achieved through joint co-operation between scientists across multiple disciplines. Being located at the borderline between the atmosphere and hydrology, the land-surface provides the link between several scientific disciplines, and land-surface modelling has been subject to intense research in the hydrological, atmospheric, and remote sensing communities in the last decades. Combining these efforts is of vital importance for the successful predictions of future changes.

Especially land-surface models (LSMs) that are based on a solution of the energy balance at the land surface have been subject to intense research. Since the late 1980s, a large number of advanced energy-based LSMs containing sophisticated parameterisations of vegetation and root zone have been developed, and currently, the energy-based models are probably the most frequently applied LSMs in the scientific community. Three factors are mainly driving the interest in energy-based LSMs.

First of all there is the desire to gain better physical understanding of the land surface-vegetation system through development of more and more sophisticated and advanced models. In this regard, the physical basis of the energy-based LSMs makes them an attractive alternative to the more conceptual types of evapotranspiration models that have traditionally been applied in hydrological modelling. Most energy-based LSMs are one-dimensional column models that describe the root zone and vegetation in great detail. Applied at the plot scale and evaluated against measured land-surface fluxes, these models have helped to gain important information on the fluxes of heat and water between the land surface and atmosphere for many different vegetation types and under many different climatic conditions. In some cases energy-based LSMs have been applied in spatially distributed frameworks, but these rarely include the lateral surface and subsurface flows between the columns.

Secondly, the atmospheric scientific community has contributed to the development of advanced energy-based LSMs. Recognizing the close connection between the atmosphere and land surface, great effort has, since the early 1990s, been put into developing advanced LSMs for atmospheric models working at all scales ranging from storm scale to global scale. Providing information on all fluxes and state variables required at the land-surface boundary in atmospheric models, the energy-based LSMs have a great advantage over the conceptual evapotranspiration models that tend to provide only actual evapotranspiration. This has made energy-based LSMs the preferred choice of atmospheric modellers. In atmospheric models, these LSMs are inherently distributed, but also here, the lateral surface and subsurface flows between cells are rarely considered.

Finally, the remote sensing community has an important role in the rapid development of energy-based LSMs. The physical basis of the energy-based LSMs makes them well suited for utilizing the growing amount of land-surface data available from remote sensing, and energy-based models have therefore become the preferred LSMs in the remote sensing community.

The LSMs have, beyond any doubt, benefited from the combined efforts across multiple scientific disciplines, and in many studies state-of-the-art models have proven to perform well at the plot scale, and in some cases, also when applied in spatially distributed frameworks.

However, when moving from the one-dimensional column models to fully distributed models, it becomes increasingly important to describe the spatial variations in soil moisture to ensure an accurate simulation of the land-surface fluxes. Variations in soil moisture may be induced by factors such as precipitation, soil texture, drainage, irrigation, flooding and shallow groundwater. While most of the energy-based LSMs include a detailed description of the vegetation and root zone, the interactions between groundwater, root zone and surface water, as well as the lateral surface and subsurface flows, are normally neglected, and consequently these models will fail to produce accurate results in areas where such interactions are important.

Contrary to this, hydrologists have a long tradition for developing and applying distributed models that take these interactions into consideration, but so far the energy-based LSMs are rarely applied in integrated hydrological modelling, and ironically, most of the integrated hydrological models that claim to be physically based do, in fact, contain rather conceptual evapotranspiration components. There are several reasons for this. The most important, probably, is that the conceptual evapotranspiration models seem to serve their purpose well, and hence there has been no obvious need to replace them with more advanced alternatives. Moreover, the applicability of the energy-based LSMs has, at least until recently, mainly been limited to intensely monitored experimental areas due to the lack of detailed land-surface and climate data required by these models. To ensure a general applicability of the integrated hydrological models, the conceptual evapotranspiration models requiring less data have been the preferred alternative so far.

The increasing quantity as well as improved quality and resolution of land-surface and near-surface climate data obtained from remote sensing has, however, significantly improved the perspectives of using energy-based LSMs in distributed hydrological modelling. Assisted by the rapid increase in computing power that allows more advanced models to be applied for larger areas and longer periods, energy-based LSMs may now begin to provide an attractive alternative to the conceptual models currently implemented in most hydrological models.

The purpose of this paper is to provide a review of the different types of energy-based LSMs and discuss some of the new possibilities that will arise when energy-based LSMs are combined with distributed hydrological modelling. Besides the general review, focus will be on utilization of remotely sensed data for evaluation purposes and the possibility of coupling hydrological and atmospheric models dynamically through a shared energy-based LSM.

2. Land-surface modelling

The total radiation absorbed at the land-surface is balanced by emission of thermal, infrared radiation to the atmosphere, latent heat loss associated with evaporation and transpiration, and sensible heat losses and diffusion of energy into the soil. The basic task of any land-surface model is to accurately simulate the partitioning of net radiation at the land surface into these component fluxes, when provided with the relevant information on land-surface and climate data.

A widely used approach to land-surface modelling is to consider the land surface as an electrical analogue. Basically, the electrical analogue expresses that the rate of exchange, F, of a quantity between two points A and B (e.g. the land surface and the surrounding atmosphere) are driven by a difference in potential of the quantity, X, (e.g. vapour pressure, temperature or carbon-dioxide), and controlled by a number of resistances, r, that depend on the local climate as well as the internal properties of the land surface and vegetation:

$$F = C\frac{X_B - X_A}{r_{AB}} \quad (1)$$

C is a constant, depending on the quantity being considered. One of the earliest, and most well-known, models that estimate evaporation using the electrical analogue was developed by Penman (1948), who described evaporation from a wet surface by linking the resistance analogue to the surface energy balance, and furthermore eliminated the need for surface potentials by assuming that the saturation vapour pressure is a linear function of temperature. This was, and continues to be, very useful since surface potentials are rarely measured. In more recent computer models it is, however, common practice to use an iterative solution of the surface energy balance, whereby the surface temperature appears as a by-product of the calculation of surface fluxes.

Penman assumed that the only resistance working between the wet surface and the surrounding atmosphere was the so-called atmospheric resistance. The atmospheric resistance expresses the ability of the air to transport a given quantity away from the surface. In unstable conditions, occurring when the surface is strongly heated, buoyancy will strongly enhance the vertical motions allowing a faster transport, and hence lower the resistance. Under stable conditions, for example on a clear night with light winds, vertical motion is dampened by the stable stratification of air near the surface, leading to higher resistances. Penman's model is illustrated in Fig. 1a.

One of the earliest land-surface models that used the Penman approach, and hence assumed that the land-surface evaporated water at the same rate as a wet surface, was the "bucket" model by Manabe (1969). The bucket model holds a maximum of water, which evaporates at the same rate as a wet surface.

However, the land surface acts as a wet plane only during and immediately after a precipitation event, when the foliage is wet. At all other times, evapotranspiration has two important controls. First, evaporation from bare soils are greatly reduced when the uppermost soil layer dries out, and equally important, the plant retards the transpiration rate because of resistance of the stomata to molecular diffusion of moisture.

Recognizing the influence of surface controls on the total evapotranspiration, Monteith (1965) further developed the Penman equation to take the land-surface controls into consideration by introducing an additional surface resistance. This resistance depend on the vegetation type, soil moisture conditions and local climate. This model structure is illustrated in Fig. 1b.

Models of the Penman-Monteith type is often referred to as “one-layer”, or “big-leaf” models because they do not distinguish between soil evaporation and transpiration, but treat the land surface as one homogeneous surface. Their simplicity and yet physically sound basis has made the one-layer models widely used. In densely vegetated canopies such “big-leaf” models have proven sufficient to describe evapotranspiration (Monteith and Unsworth, 1990). Examples of “big-leaf” models are the TOPUP model by Schultz et al. (1998) and PROMET by Mauser and Schlädich (1998).

In numerical models the land surface is often divided into grids inside which the land-surface model is applied. In those cases where the area covered by a grid consists of clusters of vegetation surrounded by bare soils, the “big-leaf” assumption tends to break down. This initiated the development of the so-called patch-, tile- or mosaic type of models, where the area covered by a grid is divided into fractions of bare soil and vegetation, and the one-layer model is applied separately for each with parameters corresponding to soil and vegetation. The patch type of model was first introduced by Avissar and Pielke (1989). Examples of models that implement the tile approach to distinguish between soil evaporation and transpiration are the ISBA land-surface model by Noilhan and Mahfouf (1996), and SEWAB by Mengelkamp et al. (1999). Patch models are considered one-layer models as the fluxes from each patch are not allowed to interact.

In cases where homogeneous, but sparse vegetation is covering the land, fluxes from the soil surface and vegetation are known to interact. The interaction between soil surface and canopy fluxes has been verified experimentally as well as numerically. Ham and Heilmann (1991) observed that for a cotton crop with a leaf area index (LAI) of 1.6 the sensible heat flux generated at the soil surface accounted for one third of the energy available for transpiration, and a similar result was found for a Texas vineyard by Heilman et al. (1994). In such cases one-layer models will fail to accurately describe the land-surface fluxes, and this has led to the development of two-layer models.

Two-layer models consist of a single, semi-transparent, canopy layer located above the soil surface such that the only way for heat and moisture to enter or leave the surface layer is through the canopy layer, whereby the component fluxes are allowed to interact. A widely used structure for two-layer models is the structure proposed by Shuttleworth and Wallace (1985). This structure incorporates a bulk stomata resistance for the vegetation similar to that used in "big-leaf" models, but also introduces a resistance at the substrate surface to control soil evaporation. By assuming that the aerodynamic mixing within the canopy is sufficiently good to allow the hypothetical existence of a "mean canopy air-stream", this formulation allows the fluxes of heat and water from the substrate and canopy to interact before they are exchanged with the atmosphere. Two-layer models of this type include three aerodynamic resistances, which control the transfer between the leaf-surface and mean canopy air-stream, soil-surface and mean canopy air-stream and mean canopy air-stream to a reference height located above the crop. The Simple Biosphere (SiB2) model by Sellers et al. (1996) represents this type of model. The structure of two-layer models is shown in Fig. 1c.

Some studies have questioned the use of a soil resistance to describe soil evaporation. Daamen and Simmons (1996) found that the use of a single soil resistance depending solely on soil moisture content will provide a reasonable estimate of cumulative evaporation from soils over a period of several days if well calibrated, but that this method is considerably less accurate on a daily or hourly basis. This, and results from similar studies, have led to the development of a modified type of two-layer models, where soil evaporation is accounted for by alternative methods.

Examples of such modified models are the DAISY SVAT by van der Keur et al. (2001) and SWEAT by Daamen and Simmonds (1994, 1996) and Daamen (1997). In both models the use of soil resistance formulations was circumvented by calculating soil evaporation as a function of the evaporative demand limited by the hydraulic properties of the upper soil layer.

From a structural point of view the two-layer approach is easily extended to include more than one vegetation layer. In such multi-layered models the effects of the vertical canopy structure are considered, which may be required to describe the fluxes, e.g., from a forest/understory system or other vegetation types with a complex vertical structure. The exchange rates are calculated for each layer and the canopy-scale fluxes are obtained by integration of these fluxes over the canopy depth. Examples of such models are given by Gu et al. (1999) and Baldocchi and Harley (1995). Gu et al. (1999) distinguish between incomplete and complete multi-layer models. They define incomplete multi-layer models as models that describe the vertical differentiation in the solar radiation environment and wind speed, whereas other factors, such as air temperature and air humidity, are assumed to be constant over the canopy depth. In contrast, complete multi-layer models predict vertical changes of these variables in an attempt to achieve better representation of the physical and biological reality. Application of these models requires detailed information on crop canopy architecture, crop physiology, turbulence etc. on a layer-by-layer basis, and they are generally computationally very demanding. The structure of the multi-layer models is shown in Fig. 1d, here exemplified by a four-layer model (three canopy layers and one surface layer).

Finally, a new generation of land-surface parameterisations has emerged in recent years. In these models, the exchanges of water and heat at the vegetated land surface are linked to exchanges of CO_2. This linkage between fluxes of heat and water and CO_2 emerged from the fact that the physiological control of evapotranspiration by plants seems to act as an optimisation mechanism that seeks to maximize carbon fluxes by photosynthesis, and reduce the water loss from the plant by closing stomata. These mechanisms have been implemented in both two-layer models (e.g. SiB2 by Sellers et al., 1996) and multi-layer models (e.g. Gu et al., 1999).

2.1. Model selection

As pointed out by Raupach and Finnigan (1988), the choice of model in any given situation is a trade-off between the desirable but incompatible traits of realism and simplicity.

Realism is desirable from the point of view that realistically structured, and consequently very detailed models, will provide realistic results. The more of the physical processes involved in the generation of land-surface fluxes we can parameterise, the more likely it is that we can predict the behaviour of the land surface under different conditions. This is, of course, only true if we can measure or realistically estimate all parameters in such detailed models, preferably at the scale of application.

Simplicity is desirable from the point of view that simple models will require less parameters, which will extend the applicability of the model outside the intensely monitored experimental areas. Moreover, simple models tend to be less computationally demanding. However, simplicity comes with a cost: by replacing the physically-based model structure with the simpler, conceptual models that require less data, the resulting conceptual parameters become increasingly difficult to infer from observations, and the dependence on calibration data increases.

Realizing that no single model will be the "best" choice for all possible applications, the background for selection of an "appropriate" complexity of land-surface models will be discussed in the next sub-sections.

2.1.1. Multi-layer models

As noted in the previous section, the resistance networks are easily extended into multi-layered models, assuming that the resistance analogy holds true for such models. However, as discussed by Finnigan and Raupach (1987) and Raupach (1988), the diffusion theory on which the resistance networks are based, frequently fails inside and just above the canopy where pronounced counter-gradient fluxes of heat and water have been observed (Denmead and Bradley, 1985, 1987). The basic reason for this failure is that the eddy motions responsible for the vertical turbulent transfer within the canopy have vertical length scales comparable to those of the canopy height. Hence the assumption of "fine-grained" mixing required by the diffusion theory breaks down. van den Hurk and McNaughton (1995) showed that the non-diffusive part of the transport processes in two-layer resistance models could be accounted for by adding a "near-field" resistance in series with the leaf-boundary layer resistance, but found the effect to be small, and later concluded that diffusion theory seems to be an adequate basis for two-layer models (McNaughton and van den Hurk, 1995). This was further supported by results of studies by Sauer and Norman (1995), Sauer et al. (1995) and Wilson et al. (2003).

It does, however, seem likely that the dense discretization of the vegetation in multi-layer models will cause the problem to persist, and the gradient diffusion theory has been replaced by the more advanced Lagrangian random-walk theory in several multi-layer models (e.g. Gu et al., 1999; Baldocchi and Harley, 1995; Wilson et al., 2003). This, and the fact that these models require a very large number of parameters that would be very difficult to obtain distributed information on, lead to the conclusion that the complete multi-layer models are inevitably complex, and remain mainly a tool for understanding the physical processes that control the exchange of mass and energy between the land surface and atmosphere. Consequently, these models are not considered to be suitable for distributed hydrological modelling.

Eliminating multi-layer model from the list of candidates, the choice has to be made between the single- and two-layer model types.

2.1.2. Single or two-layer models

Besides the canopy resistance and aerodynamic resistance required in single-layer models, two-layer models require three additional resistances to be parameterised. This represents a step up in complexity, as the parameterisation of each resistance require a number of additional parameters to be specified. Taking the view that all model parameters should be inferred through calibration, the problem of equifinality and over-parameterisation will arise.

Beven (1989) was the first to address the problem of over-parameterisation and equifinality in distributed hydrological modelling and Franks et al. (1997) specifically addresses the question on how much complexity in a land-surface model that can be supported by the available observations. They showed that even when using a relatively simple one-layer model (TOPUP by Schultz et al., 1998), there appears to be too many degrees of freedom in terms of fitting the model predictions to calibration data, and they showed that good fits may be achieved in many areas of the parameter space. They (and others) argue that the complexity of land-surface models needs to be reduced in order to eliminate the equifinality problem.

Demarty et al. (2004) applied a multi-objective approach for retrieving quantitative information about the surface properties from different surface measurements to determine the potential of a land-surface model to be applied with "little" a-priori information, and their results suggested that complex LSMs can be driven with limited a priori information. Future research along these lines may help to gain more insight in the problem of equifinality for advanced land-surface models.

Also, when reducing the number of model parameters, it may become more difficult to infer the parameters from, e.g., remote sensing, so even if less model complexity will reduce the problem of over-parameterisation and equifinality, it may not necessarily make the models more applicable.

The bulk-surface resistance used in one-layer models illustrates this dilemma. It basically represents the four resistances in two-layer models, which reduces the number of parameters that needs to be specified. However, as pointed out by Raupach and Finnigan (1988), the cost of lumping four resistances into one is that the bulk surface resistance tends to be less well-behaved than the individual resistances in the two-layer formulation.

Finally, the scale of application plays an important role in coupled applications. Jarvis and McNaughton (1986) and McNaughton and Jarvis (1991) showed that increasing scale leads to an increasing number of negative feedback paths acting through the planetary boundary layer. These negative feedback paths tend to stabilize the system and diminish the sensitivity of transpiration to changes in surface resistance. They conclude that at the regional scale the use of multi-layer models is never justified.

It is obvious that no single model will outperform all other models in all situations and that model selection should be based on the scale and purpose of application as well as the available data.

2.2. The issue of scale

It has been (rightly) argued that due to the mismatch between the scale for which the theory behind the developed models is assumed valid (typically point scale), and the scale of the typical application, the parameters in physically-based models are in reality conceptual, and that even when mean parameters are available, the highly non-linear behaviour of most land-surface models cannot be expected to simulate the mean response of a particular area.

This has led to a number of land-surface models that take this sub-grid variability into consideration (e.g. Kavvas et al., 1998; Famiglietti and Wood, 1994). As pointed out by Beven (1995), any theory to take this sub-grid variability into consideration must be based on some knowledge about the sub-grid variability and that this would appear to make any scaling theory a very distant prospect, given the data gathering techniques that were available in 1996.

Sellers et al. (1997) analysed the consequences of using simple averages of topographic slopes, vegetation parameters and soil moisture in a two-layer LSM (SiB by Sellers et al., 1986). They found that the relationships describing the effects of moderate topography on the surface radiation budget are near-linear and thus largely scale invariant, and so was the relationship linking the canopy conductance to transpiration. The relationship linking root zone soil moisture content to transpiration was found to become increasingly non-linear as the soil dried out. However, their results showed that soil wetness variability decreases significantly as the soils dries out, which partially cancels out the effect of the non-linear functions. They conclude that, for practical purposes, the two-layer model seems to be relatively robust and scale-invariant with respect to these variables.

Dolman and Blyth (1997) also investigated the influence of small-scale land-surface heterogeneity and concluded that in most cases it is possible to find effective parameters for surface resistances by taking simple geometric or arithmetic averages of the component resistances, but that the use of more sophisticated techniques improves calculations.

3. Application of remote sensing for evaluation of distributed land-surface models

Remote sensing (RS) is simply defined as the observation of a target by a sensor without physical contact. From a hydrological point of view, RS is the process of inferring (near-)surface parameters from measurements of the reflected and emitted electromagnetic radiation from the land surface. Both active sensors that send a pulse and measure the reflected pulse and passive sensors that measure emissions and reflectance of natural sources are used in this context. Conversion of the measured electromagnetic fluxes into physical land-surface parameters that can be utilized both for monitoring and modelling purposes is currently a major research topic.

The quality and resolution of remote sensing data products have improved much in recent years and have provided a valuable source of distributed information with a promising potential for application in distributed hydrological models (Schultz, 1998; de Troch et al., 1996; Tenhunen at al., 1999; Waring and Running, 1999). Designing models to utilize this distributed information in land-surface modelling is of great importance (Wessmann et al., 1999), since spatial information obtained from RS is the only source for distributed data that can be characterized as realistic at scales above plots and small experimental catchments (Refsgaard, 2001).

The use of energy-based land-surface models in distributed hydrological models opens for the possibility for using remote sensing data for model evaluation purposes, which is addressed in the next section.

3.1. Application of remote sensing for model evaluation

General methodologies to calibration and evaluation of distributed hydrological models have been subject to much discussion during the last decades (Beven, 1989, 1996, 2001, 2002; Konikow and Bredehoeft, 1992; de Marsily et al., 1992; Refsgaard, 1997, 2001; Refsgaard and Henriksen, 2004; Rosbjerg and Madsen, 2005), mainly due to the large number of parameters that are allowed to vary during calibration, and the lack of methods to perform a truly distributed evaluation of the model performance. Still today one of the most commonly used evaluation approaches in distributed hydrological modelling is evaluation against measured river discharge, while the utilization of spatial data is rare (Refsgaard, 2001). While discharge remains an excellent indicator of how well a model is able to reproduce the water balance in a river basin, it does not necessarily give any information on the performance of a model at scales smaller than the area upstream the gauge. This has led some authors to express doubts about the applicability of this calibration/evaluation method in distributed models (Beven, 1989; Bergström, 1991; Refsgaard, 1997, 2001).

Ideally, a distributed evaluation of any land-surface model would require distributed observations of the land-surface fluxes that make up the energy balance. However, since none of these can be measured directly with the current RS technology, models will have to be evaluated against derived measures, such as surface temperatures, or by deriving the relevant surface fluxes of heat and water from the data types that are more easily derived from RS. These topics are addressed in the next sub-sections.

3.1.1. Distributed evaluation against thermal observations

Thermal observations of the Earth's surface have long been recognized as a valuable source of information for evaluation of the surface energy balance over large regions (Price, 1980), and the direct, but complex relationship between surface temperature and evapotranspiration has been illustrated in several studies (Rivas and Caselles, 2004; Jackson et al., 1981; Shuttleworth and Gurney, 1990; Mauser and Schädlich, 1998). Recognizing this close relationship, remotely sensed maps of land-surface temperatures have previously been used to evaluate model performance by comparing maps of simulated surface temperatures to their observed counterpart.

Mauser and Schädlich (1998) utilized the close relationship between latent heat flux and surface temperature to evaluate the one-layer PROMET model in a 100×150 km area in Bavaria, Germany. They compared remotely sensed NOAA-AVHRR surface temperatures to simulated surface temperatures and found the spatial pattern of surface temperatures simulated by PROMET to be very similar to the observed. Through a pixel-by-pixel comparison between simulated evapotranspiration and observed surface temperature they found a clear trend of increasing simulated evapotranspiration with decreasing observed surface temperature.

Silberstein et al. (1999) used a number of Landsat-TM images to evaluate the performance of the COUPLE model at the small catchment scale ($\sim$1 km^2) in two catchments in Australia. They concluded that their model gave excellent results when compared to Landsat-TM data. They found that the temperature difference between pasture and forest can be more than 15°C in summer and 2–3°C in winter.

Biftu and Gan (2001) used NOAA-AVHRR and Landsat-TM surface temperatures to evaluation the semi-distributed DPHM-RS model for the Paddle River Basin, Alberta. They compared average simulated and observed surface temperature for four different land-use classes and found that their model was able to reproduce observed surface temperatures within 2°C on clear days.

The studies described here are all examples of how surface temperatures have been included in the evaluation procedure, whereby a truly independent measure of the distributed performance has been obtained. Such comparison is an important step toward a distributed evaluation, as it allows for an identification of areas where a difference between simulated and observed temperatures indicates that there may be inconsistencies between the model and reality.

It does not, however, provide any direct information on how a difference in surface temperature translates into a difference in latent heat flux. This makes it difficult to determine when a deviation between simulated and observed surface temperature should be considered significant. A logical extension of the pure temperature evaluation seems to be evaluation against flux-maps derived from remote sensing.

3.1.2. Distributed evaluation against latent heat fluxes

The conversion of remote sensing data into surface fluxes for monitoring purposes and to improve evaluation of land-surface models have been subject to intense research.

To eliminate the need for determining plant stress parameters and near-surface humidity, a common approach to estimation of latent heat (LE) from remote sensing is to calculate latent heat as the residual of the land-surface energy balance.

$$LE = R_n - H - G \tag{2}$$

where R_n is net-radiation, H is sensible heat and G is soil heat flux.

Soil heat is normally considered a fixed fraction of the net-radiation (Norman et al., 1995; Anderson et al., 1997; Boegh et al., 2000, 2002, 2004), and since previous studies have shown that net-radiation can be accurately determined from RS data (e.g. Boegh et al., 1999), the main task becomes the determination of sensible heat flux from remote sensing data.

Generally, the sensible heat flux is modelled using the electric analogue. Using the one-layer approach, the sensible heat flux is written

$$H = \rho c_p \frac{T_c - T_a}{r_a} \tag{3}$$

Here, ρ [kg/m^3] denotes the density of the air and c_p [J/kg/K] is the specific heat of air at constant pressure. T_c [K] is the aerodynamic temperature at the mean canopy air-stream, which is different from the radiometric surface temperature, T_r [K], that is obtained from remote sensing. It has been shown that T_r and T_c may differ by several degrees (Stewart et al., 1994). A way to compensate for this difference between T_r and T_c is to add an extra resistance to r_a in Eq. (3), whereby the aerodynamic surface temperature can be substituted by the radiometric surface temperature. The excess resistance r_{ex} is calculated using the kB^{-1} factor:

$$r_{ex} = \frac{kB^{-1}}{ku_*} \tag{4}$$

k is the von Karman constant, u_* is the friction velocity, and $kB^{-1}=\ln(z_0/z_{0h})$, where z_0 is the roughness length for momentum and z_{0h} is the roughness length for heat. The need for this excess resistance arises from the fact that heat transfer near a surface is controlled primarily by molecular diffusion, whereas momentum exchange takes place as a result of both viscous shear and local pressure gradients (Chehbouni et al., 1997).

Usually kB^{-1} has to be determined empirically by calibration. Especially for sparse vegetations the value of kB^{-1} is seen to vary widely, and a dependence on vegetation type and conditions as well as climate has been observed in several studies (Troufleau et al., 1997; Massman, 1999; Lhomme et al., 2000). This had led some to question the usefulness of the kB^{-1} approach for sparse canopies (Lhomme et al., 1997; Verhoef et al., 1997).

Moran et al. (1996) proposed a one-layer approach that did not require estimation of the excess resistance. They combined the Penman-Monteith equation with the NDVI-T_r (NDVI: Normalized Difference Vegetation Index) relationship, and by computing the theoretical boundaries on the NDVI-T_r relationship (the boundaries represent zero and potential evapotranspiration) they derived the actual evapotranspiration based on the distance from the observed (T_r, NDVI) to its upper and lower boundary.

Boegh et al. (2002) also proposed a method to derive spatial estimates of atmospheric resistance, surface resistance, and evapotranspiration, while eliminating the need for the excess resistance by relating the vapour pressure at the surface to the air-humidity through the decoupling coefficient by Jarvis and McNaughton (1986). The decoupling coefficient expresses the degree of coupling between the atmosphere and land surface as a function of the relative importance of surface and atmospheric resistance. When the land surface is poorly coupled to the atmosphere, water vapour will accumulate near the surface and get close to saturation. Under strongly coupled conditions, the vapour content near the surface will approximately equal the vapour content in the air above the canopy.

By explicitly calculating the aerodynamic temperature at the mean source height, two-layer models eliminate the need for the excess resistance. However, only the effective surface temperature (T_r), which is a combination of foliage (T_l) and soil surface temperature (T_s), is obtained from satellites. Norman et al. (1995) express the relationship as:

$$T_r = \left(f\,(LAI, \varphi)\, T_l^4 + (1 - f\,(LAI, \varphi))\, T_s^4 \right)^{1/4} \quad (5)$$

The function f expresses the vegetation coverage as a function of leaf area index (LAI) and sensor view angle (ϕ). Hence, to derive the component temperatures from the effective temperature, some additional knowledge that ties the two component temperatures to each other is required. This "additional relationship" has been investigated in a number of studies.

Norman et al. (1995) developed a method where transpiration is initially accounted for by the Priestley-Taylor equation, whereby they were able to relate the canopy temperature to air temperature. This allows both canopy and soil surface temperature to be determined from a single effective surface temperature. The initial guesses of component temperatures were used in an iterative procedure to derive soil evaporation and transpiration that satisfied the energy balance. Anderson et al. (1997) coupled this model with a time-integrated component connecting surface sensible heating with planetary boundary layer development, whereby they eliminated the need for specifying air-temperatures. They tested the model on data collected during FIFE (Sellers et al., 1992) and Monsoon '90 (Kustas and Goodrich, 1994) and found that the model yielded uncertainties comparable to those achieved by models that do require air-temperature as input.

Lhomme et al. (1994) attempted to establish an empirical relationship between T_s to T_l. There are, however, strong indications that such relationships may vary with experimental data (Zahn et al., 1996) and that it is probably related to plant characteristics (McNaughton and van den Hurk, 1995).

Recognizing the influence of leaf area index on the relationship between aerodynamic and radiometric surface temperature, Chehbouni et al. (1996) used results from a soil-vegetation-atmosphere-transfer (SVAT) model coupled to a crop growth model to derive a relationship between aerodynamic and radiometric surface temperature for LAI between 0.05 and 1. They later tested this approach at two different sites, and concluded that this approach showed some signs of being generally applicable, but that more applications were needed to test its generality (Chehbouni et al., 1997).

Boegh et al. (1999, 2000) used the relationship between the Landsat-TM surface temperature and the normalized difference vegetation index (NDVI) representing the fractional vegetation cover to derive the vegetation temperature from the surface temperature.

Finally, some have attempted to solve this problem by measuring the radiometric surface temperature from two different angles, whereby two equations in the two unknowns T_l and T_s can be established from Eq. (5) (Kustas and Norman, 1997, 1999; Francois et al., 1997; Chehbouni et al., 2001; Merlin and Chehbouni, 2004). View angle effects are, however, most pronounced for sparse canopies, where a change in view angle will cause a large difference in the fraction of soil and vegetation within the footprint of the radiometer, and while the view-angle approach has proven useful when using ground-based measurements, only one viewing angle is usually available for satellite images.

It should be clear from this brief review that converting land-surface temperature into latent heat fluxes requires a number of model assumptions, and knowing that different assumptions lead to deviating fluxes (e.g. Zahn et al., 1996), a comparison between simulated and RS-derived land-surface fluxes will, in reality, be a comparison between two models, usually being based on different structures and assumptions. Hence, it will be difficult to identify the model that is "most correct". This is probably part of the reason why few have attempted such a direct flux comparison (one example is Boegh et al., 2004).

4. Coupling between atmospheric and hydrological models

Understanding the interaction between terrestrial microclimate, hydrology and ecology is a key to determining the effect of land-use and climate change on the hydrological system. The use of energy-based LSMs in hydrological modelling will make a dynamic coupling to an atmospheric model possible. Such a coupled system would provide a unique framework for investigation of land surface – atmosphere interactions at the hydrological scales. A brief review of the couplings between the land surface and the atmosphere at both local and regional scale is given in this section.

4.1. Land surface – atmosphere interactions

Feedback between land surface and atmosphere arises from the fact that if the fluxes of heat and water from the land surface to the atmosphere changes, humidity, temperature and air pressure in the atmosphere will change as a consequence. Since the climate exerts a large influence on the land-surface fluxes, the changed conditions in the atmosphere may significantly feed back to the land surface, working to either dampen or amplify the changes in land-surface fluxes.

A very important factor in determining the significance of feedback is scale. If, e.g., transpiration from a single plant changes, the conditions in the atmosphere will remain unaffected, and hence there will be no feedback effects. However, if transpiration from a large area changes, it will be felt throughout the whole PBL and feedback effects may be pronounced.

It is common practice to perform impact assessment studies of, e.g., land-use changes in hydrological models that are driven by time series of measured climate data. In those cases, a potential change in land-surface fluxes is not allowed to affect the state of the atmosphere, and hence feedback is neglected.

The consequence of neglecting feedback at the regional scale was examined by Jacobs and de Bruin (1992), who coupled a "big-leaf" land-surface model with a one-dimensional PBL model. They found that PBL feedback has a significant influence on the sensitivity of latent heat to changes in land-surface parameters. Similar conclusions were later drawn by Brubaker and Entekhabi (1996) and Kim and Entekhabi (1998), who conclude that it is necessary to examine the impact of any change in land-surface parameters (e.g. land-use change scenarios) in a coupled model system, and that failure to do so may result in model sensitivities that are not only wrong in magnitude, but in sign as well.

Others have investigated the influence of the land-surface properties on the local and regional climate, and the close relationship is now well proven.

At the local scale Segal et al. (1988) investigated the influence of vegetation on thermally induced meso-scale circulations due to non-uniform vegetation cover. They found that when extended areas of dense vegetation not under water stress are adjacent to bare soil areas, thermal circulations comparable in intensity to sea-breezes can be induced. They found that there was no substantial difference in the thermally induced meso-scale circulations generated by a sharp thermal contrast along a flat terrain when compared to that generated by an equivalent, but gradual change from dense vegetation to bare soils along distances less than 30 km.

Similar conclusions were reached by Hong et al. (1995), who investigated the effects of different soil types on the thermally induced circulations between vegetated and bare soil areas and found that the intensity of the "vegetation breeze" is strongly related to soil characteristics.

Such vegetation-induced circulations were experimentally confirmed by micro-meteorological observations from the HAPEX-MOBILHY field experiment (Pinty et al., 1989).

Cuenca et al. (1996) investigated the impact of soil water parameterisation on atmospheric boundary layer formation using an atmospheric model coupled to a soil-plant model. They found little difference in the dry and wet ends of the scale, but significant differences were found for the intermediate soil moisture contents, resulting in differences in planetary boundary layer depths of up to 1000 m.

Song et al. (1997) used a meso-scale atmospheric model coupled to an advanced land-surface model to simulate the local climate over the 15×15 km FIFE site. They found that, even for this relatively homogeneous grassland, local variations in the land-surface fluxes induced up to 2°C spatial variations in the near-surface air temperature.

At the regional scale van den Hurk et al. (2002) compared the results obtained for the Baltic Sea catchment using two different land-surface schemes coupled to the same regional atmospheric model (RACMO). They found that the temporal and spatial distribution of precipitation is sensitive to the choice of land-surface scheme. They conclude that the strong coupling between local evapotranspiration and local precipitation results in clear hydrological feedback mechanisms.

Zeng et al. (2002) investigated the effect of land-surface heterogeneities on the regional climate and found that surface heterogeneities in roughness length and stomata resistance greatly affect the simulation of surface fluxes as well as the wind, temperature and precipitation fields.

Due to this well-proven relationship between the land surface and the atmosphere, advanced land-surface schemes have now been implemented in many atmospheric models. However, while most of these describe the canopy and root zone in great detail, the interactions between groundwater, root zone, and surface water are normally neglected, which may lead to inaccurate model predictions in areas where groundwater and surface water are closely connected (Chen and Hu, 2004; York et al., 2002).

In contrast, the most advanced integrated hydrological models (e.g. MODHMS, Panday and Huyakorn, 2004; MIKE SHE, Graham and Butts, 2006) describe the subsurface and surface flows and the interactions between surface water and groundwater in great detail, but treat the climate in a very simplistic way, usually being driven by prescribed atmospheric conditions. This implies that any changes in land-surface or hydrological properties are not allowed to feed back to the atmosphere. As earlier described, previous studies have shown that this lack of feedback may potentially lead to errors in scenario simulations dealing with changes in hydrological and land-surface properties. Recent work by Overgaard (2005), using MIKE SHE with an atmospheric model, supports this conclusion.

The simplifications described above are required to limit the computational requirements and to ensure the practical applicability of both hydrological and atmospheric models. It does, however, imply that there is still a gap between the simplified hydrological model components implemented in atmospheric models and the state-of-the-art integrated hydrological models, as well as a gap between the very simplistic treatment of climate data in hydrological models and state-of-the-art atmospheric modelling.

5. Future perspectives

Implementation of the energy-based land-surface models will strengthen the link to remote sensing and make it possible to utilize some of the RS data currently available more efficiently. The link between remote sensing and distributed hydrological modelling will be vital for most future applications and possibly improve the possibilities for making a more spatially detailed evaluation.

The energy-based LSMs also provide the opportunity of coupling dynamically to meso-scale atmospheric models through a shared land-surface model. Such a dynamic coupling would provide a unique framework for investigating issues related to land surface – atmosphere interactions at the hydrological scales.

While such a coupled model system probably will be too computationally expensive to be used for long simulations of large domains, it has the potential to provide important insight on the interactions between the land surface and atmosphere in areas where the parameterisations of the land-surface models implemented in atmospheric models do not suffice.

Moreover, a coupled modelling system could provide hydrologists with information on how the atmosphere will respond to changes in the hydrological and land-surface properties at the hydrological scale, and equally important, provide information about the consequences of neglecting atmospheric feedback in hydrological scenario simulations.

It seems likely that impact assessment of changes in the hydrological system, such as drainage or restoration of wetlands, changes in irrigation practices, or construction of reservoirs, in uncoupled models may lead to errors similar in magnitude as for land-use changes. Overall, this implies that in some cases, depending on the scale and significance of the imposed changes, impact assessment studies in uncoupled models will fail to produce reliable predictions. A coupled model system will provide the means to identify the situations where feedback will significantly affect the simulated impact and in those cases help to produce more accurate predictions.

Predicting the impact of future climate changes will continue to be a major focus area for both hydrologists and atmospheric scientists in the coming years. To the authors' knowledge, no meso-scale atmospheric models are currently capable of running in climate mode, and considering the scale of application of the current regional climate models (RCM), a coupling between a complex hydrological model and a RCM is not feasible at the current time. However, the development of climate models working at the meso-scale is currently being subject to much research, and when these models become operational they will be ideal for climate change impact assessment studies.

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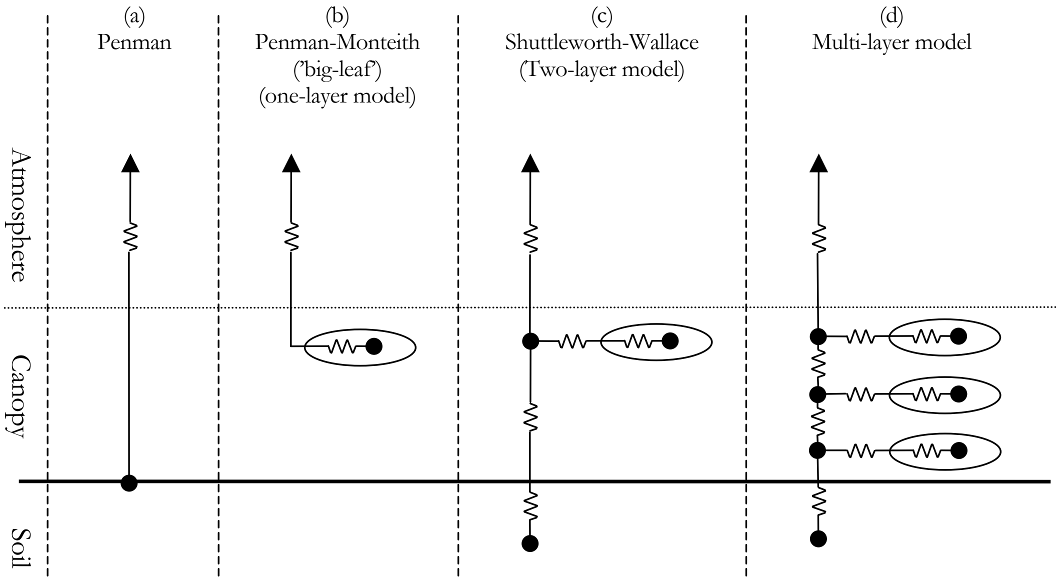

Fig. 1. Illustration of the different resistance models discussed in Sect. 2.

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