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Scale issues in soil moisture modelling: problems and prospects

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Abstract: Soil moisture storage is an important component of the hydrological cycle and plays a key role in land–surface–atmosphere interaction. The soil-moisture storage equation in this study considers precipitation as an input and soil moisture as a residual term for runoff and evapotranspiration. A number of models have been developed to estimate soil moisture storage and the components of the soil-moisture storage equation. A detailed discussion of the implication of the scale of application of these models reports that it is not possible to extrapolate processes and their estimates from the small to the large scale. It is also noted that physically based models for small-scale applications are sufficiently detailed to reproduce land–surface–atmosphere interactions. On the other hand, models for large-scale applications oversimplify the processes. Recently developed physically based models for large-scale applications can only be applied to limited uses because of data restrictions and the problems associated with land surface characterization. It is reported that remote sensing can play an important role in overcoming the problems related to the unavailability of data and the land surface characterization of large-scale applications of these physically based models when estimating soil moisture storage.

Key words: scale, soil moisture, hydrological cycle.

I Introduction

Soil moisture is an important component in the hydrological cycle, specifically in the planetary water balance. Soil moisture storage plays a key role in land–surface–atmosphere feedback processes. Thus, it is necessary to obtain a better estimate of soil moisture storage to understand the processes involved within the hydrological cycle. Soil moisture models can be used to estimate moisture storage. The fundamental equation for soil moisture modelling can be expressed as follows:

$$\Delta SM = P - ET - R$$

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where SM is soil moisture storage, P is precipitation, ET is evapotranspiration and R is runoff. The runoff term includes surface runoff, subsurface lateral flow and seepage to groundwater. The soil moisture budget can be modelled totally from a hydrological standpoint to calculate runoff (cf. Loague and Freeze, 1985) or it can be modelled as a component of land–surface–atmosphere interaction processes (cf. Milly, 1992); it is also possible to combine both (cf. Mather, 1978). An important issue that needs to be addressed before taking any of these routes is to determine the model's scale. The scale of application determines the complexity of the physical–chemical–biological processes within the models. This article will discuss various issues related to the scale of soil moisture budget modelling. These include the determination of the physical processes and controlling factors of runoff and evapotranspiration at different scales, their modelling and the problems associated with measurement.

It is known that the role of the various controlling factors and associated processes varies at different scales. Klemes (1983) noted that the scale of physical processes is not arbitrary, and their range is not unrealistically continuous. Problems associated with interpolating small-scale controlling factors and dominating processes to the large scale frequently demonstrate this. Models that deal with the estimation of runoff and evapotranspiration also commonly inherit this problem. As a result, modelling at different scales has become an important issue in recent years. The scale of application determines how much complexity should be allowed within the models.

Over the last few decades, our understanding of biophysical-chemical processes has expanded at an astonishing rate. Owing to the increase in computing power during this period, it became possible to integrate these processes into the model to estimate soil moisture. Although this leads to a significant improvement in replicating the real-world processes within the model, model validation and application at different scales became a significant problem. The former is dependent on high-quality and high-resolution data while the latter is dependent on understanding the various dominant processes at different scales. The following sections will focus on the processes that are important for the two components (runoff and evapotranspiration) of soil moisture modelling at different scales, and related issues. Although runoff and evapotranspiration are closely connected and interdependent, they will be treated separately for a clearer understanding of the processes involved. Detailed model equations are provided in the Appendix.

II Runoff

Runoff is largely dependent on the intensity of rainfall, the infiltration capacity of the soil, antecedent soil water condition, soil type, land use and physiography. If rainfall intensity exceeds infiltration capacity, surface runoff occurs (Dunne and Leopold, 1978). Gravity flow and capillary forces significantly influence infiltration capacity. Both these processes are largely dependent on the soil type. Gravity flow is a dominant controlling factor where soil is coarse or pores are large, while capillary forces are important controlling factors where soil is fine or pores are very small. Antecedent soil moisture also determines how much water is going to pass through the soil and thus controls saturation of the soil (which leads to surface runoff).

Vegetation cover and land use influence infiltration capacity and thus the surface and subsurface flow of water. Types of vegetation determine the depth of the root zone

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and the moisture-holding capacity of the root zone. Dense canopy cover protects soils from rain-packing and thus increases the infiltration capacity of the soil. Dense canopy cover also supplies abundant humus which does not allow rainfall intensity to exceed infiltration capacity. It is also found that replacing forests with agriculture (which does not cover the land fully and does not contain a higher amount of organic matter) drastically reduces infiltration capacity and thus increases runoff. Urban land use sharply increases surface runoff. A sloped surface partly controls surface and subsurface runoff, and enhances runoff through gravitational pull. Furthermore, groundwater storage and subsurface runoff influence surface runoff. Local geology, soil type and climate determine the depth of the groundwater table. Groundwater storage influences the soil moisture content of overlying soil layers through capillary flow, which is dependent on the soil texture. This process eventually modifies surface runoff and infiltration capacity by changing soil moisture content. Furthermore, surface runoff determines how much water is left for evapotranspiration and soil moisture storage. Thus, it is clear that various soil-water-related processes are interconnected and interdependent.

As mentioned previously, the question is, how much process detail should be incorporated into the model for soil moisture estimation? The details of the model will depend on the scale of application. For a small-scale study, it is sometimes possible to record most of the small variations of the different controlling factors. For example, Loague (1992a; 1992b) developed a soil-water content data set composed of nearly 25 000 measurements made at a 100 m^2 basin of Chikasha, Oklahoma. He estimated the impact of soil water on runoff by constructing a 'quasi-physically based rainfall-runoff model'. This model provided improved estimates of peak flow (however, the timing of the peak flow was not satisfactory). It would be very difficult to use this model satisfactorily for a large basin because of the impossible task of gathering such high-resolution data. In such a case, simplifying the model's assumptions and its underlying physics is the answer.

Numerous extensive smaller-scale studies have been performed using sophisticated physically based models. For example, Peck *et al.* (1977) developed a scaling factor to estimate the effects of spatial variability of soils in water balance modelling. They developed this method to interpolate soil properties from fewer observations. The authors estimated the soil properties of the Branch watershed of Tennessee by using the scaling factor. Finally, they used these results in a water balance model that calculates, among other things, runoff. They claimed that the water budget components were in close agreement with the simulated spatial soil-water variability. In subsequent years, scaling theory has been used frequently in various water balance and rainfall-runoff models for small basins to estimate the effects of soil heterogeneity on soil-water content (e.g., Sharma and Luxmoore, 1979; Luxmoore and Sharma, 1980; Clapp *et al.*, 1983). Similar modelling studies by Milly and Eagleson (1987), Loague (1990) and Loague and Gander (1990) in small basins have estimated the effects of porosity and the infiltration rate on runoff and water balance. Hughes (1994) applied four 'deterministic' models (namely, VT1, RAFLES, P-Export and Pitman) to estimate runoff from a 0.18 km^2 grassland catchment in the Eastern Cape Province of South Africa. Model parameters were determined from the physical characteristics of the catchment. Model predictions were satisfactory.

In the light of the above discussion, the following questions can be asked: is it possible to collect data on the worldwide infiltration rate for model calibration, or is it possible to include characteristics of capillary flow under various soil types for the

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whole planet, or is it possible to model the effects of all the different types of vegetation cover on runoff for the whole globe? The answer is no. The land surface is very heterogeneous at the large scale, which makes it impossible to capture its complexity within models for large-area studies. It has also been questioned whether we can extrapolate the results from small-scale studies to the larger regional or global scale. Again, the answer is no. Pilgrim *et al.* (1982) and Pilgrim (1983) identified the problem of transferring results from small basins to large basins. Pilgrim (1983) noted that even if infiltration characteristics are the same in two basins, different infiltration responses owing to the varying sizes of the basins require separate parameterization. For example, the water-storing capacity of the larger basins will be greater than the smaller basins. As a result, runoff response will vary at basins of different sizes under similar precipitation conditions. Pilgrim (1983) concluded that transferring results from smaller-scale studies to the regional or the global is dangerous and counterproductive. He also noted that this may create a major impediment to the understanding of the relationships among processes. Thus, we need to take a different modelling approach in estimating runoff for soil moisture calculation. This includes the need to identify dominating processes and their responses under different conditions at the regional and global scale to develop proper parameters. Gleick's (1987) approach can be adopted as a first step in this direction. He applied a modified Thornthwaite (1948) method to the Sacramento basin to estimate water balance. One of the model outputs was runoff. The size of the Sacramento basin is 41 000 km^2. As a result of the distinct characteristics of the climate and vegetation in the upper and lower basin due to elevation change, he introduced different assumptions into the model for these two subbasins. The two-basin model run provided a better, physically plausible estimate of runoff.

For global-scale runoff estimates, Thornthwaite's (1948) water budget model has been widely accepted as a standard method. This method assumes that runoff only occurs if excess soil moisture is available after evapotranspiration demand has been met and soil moisture recharge has been completed (Willmott *et al.*, 1985). Legates and Willmott (1995) noted that the problem with the Thornthwaite method lies in the assumption that runoff occurs as a residual of water budget. As a result, the inaccuracies associated with the measurement of components of the water budget (such as evapotranspiration and soil moisture) or poor quality of the input data set and/or insufficient parameterization can be compounded in runoff estimates. Regardless of these difficulties, the Thornthwaite model is probably the most satisfactory runoff estimation method available for the large, regional scale to global-scale application. This method is relatively simple and the data requirement is not complex. The integration of a relatively detailed (but not too complex) and realistic description of biophysical-chemical processes into this method would make it an ideal tool for large-scale runoff estimation.

III Evapotranspiration

Modelling evapotranspiration (ET) for soil moisture estimates at different scales can be as difficult as calculating runoff. ET is primarily controlled by solar radiation, the supply of water and atmospheric humidity (or vapour pressure deficit). Other important controlling factors are stomatal resistance, aerodynamic resistance and surface resistance. Soil characteristics also play an important role by determining the

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water-holding capacity at the surface level and at the root zone, which in turn influences the water available for plants to transpire. As in modelling runoff, the scale of study may determine the degree of complexity that should be introduced into the ET model without weakening its theoretical foundation or simplifying its biophysical-chemical processes.

A large number of highly sophisticated models that replicate detailed biophysical processes have been developed over the past few decades. In most cases these are referred to as 'physically based' models. In this discussion, these models are grouped into two categories, namely, combination and eddy-diffusion-type ET models. Models in the combination category integrate energy balance and aerodynamic terms. These models estimate ET by calculating the supply of energy and the 'turbulent transport of water vapor from an evaporating surface' (Rosenberg *et al.*, 1983: 248). Penman (1948) was the first to develop this type of model, and Monteith (1965) proposed some major improvements to Penman's method by adding resistance terms. This is why the combination-type equations developed during subsequent years are known as Penman and Penman–Monteith equations. Owing to the wide variety of combination-type ET models, they can be further categorized into four subgroups, namely, energy balance, interception, single layer and multilayer models.

Energy balance equations were the first step in developing a physical basis for ET. Penman's (1948) method of ET estimation is the classic example. His method successfully integrates physical terms into ET estimation. The data requirements for his method are simple and, as a result, it is the most widely used physically based model. However, it has been criticized for underemphasizing the importance of ventilation relative to radiation in maintaining regional evaporation, and also for underestimating ET under strong sensible heat convection. The latter criticism is probably linked to the fact that this model was initially based on cool and moist climatic data.

Van Bavel (1966) and Priestley and Taylor (1972) developed similar energy balance methods. Surface roughness length plays an important role in the ET estimation system constructed by van Bavel. Since this model is sensitive to windiness, it underestimates ET rate under calm conditions and overestimates ET rate under strong winds. The distinct advantage of the Priestley and Taylor (1972) method is that it can be used with satellite data to calculate reference crop ET for large remote areas where data are not available. It also provides good ET estimation in humid areas.

Interception models are characterized by the introduction of the concept of interception loss during a storm. Thom and Oliver (1977) modified the Penman (1948) method to develop a model for estimating actual ET. They used atmospheric stability terms extensively in their method. However, this model has been criticized because the surface resistance term is not totally related to stomatal resistance. Gash (1978) modified the Thom and Oliver (1977) method to provide separate estimates of interception loss. Unfortunately, this model is only applicable to tall vegetation. Later, Gash (1979) proposed a model for interception loss from tall crops that omitted evaporation from tree trunks. In this model, real rainfall is represented by a series of discrete storms. These storms are separated by a time interval that is sufficient to dry up the canopy and tree trunks.

Rutter *et al.* (1971) developed a model to estimate rainfall interception in forests. This model can calculate a running water balance during a period of rainfall, throughfall and ET. It can also estimate changes in canopy storage. This method made an important contribution to calculating interception loss from wet and partially wet surfaces.

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Despite the model's success, it has been criticized for its failure to formulate and adopt an algorithm that could deal with the changing depth of vegetation and the resulting variation in interception and evaporation loss. Sellers and Lockwood's (1981) multilayer model attempted to overcome some of these weaknesses.

Single-layer combination-type models assume that 'all the components of element of vegetation are exposed to the same microclimate' (Shuttleworth, 1991: 112). In other words, these models treat the whole boundary layer as a single layer. In these models, aerodynamic and stomatal resistance control ET. Monteith (1965) made the first successful breakthrough in developing a single-layer model which combines energy and aerodynamic terms. Monteith used resistance terms extensively in his model. His method estimates relative evaporation rates from dry and wet surfaces. It assumes that the relationship between transpiration and leaf area is influenced by the closure of stomata when they are shaded from sunlight. Unlike many established models, Monteith raised the important question of advection. Although this method made remarkable progress by incorporating several very important plant physiological phenomena that control ET, it failed to address the influence of soil surface resistance, horizontal fluxes, evaporation from tree trunks, interception loss and the three-dimensional nature of various ET-related processes.

A much improved version of the single-layer model for ET-interception was developed for urban areas by Grimmond and Oke (1991). This model is based on the methods developed by Penman (1948), Monteith (1965), Rutter *et al.* (1971) and Shuttleworth (1978; 1979). The Grimmond and Oke (1991) method has been satisfactorily applied to urban areas, where it has successfully integrated an anthropogenic heat-flux term and a turbulent source-area concept, and has estimated ET from wet, partially wet and dry surfaces.

The multilayer ET models divide the atmospheric boundary layer and soil surface into several horizontal layers, estimate the interception of solar and thermal radiation, and calculate sensible and latent energy flux for each layer. Shuttleworth (1991) noted that these models are the best available methods for estimating ET provided that complex data requirements are fulfilled and submodels are available. Multilayer models can be divided into two groups. One includes the models that divide the atmospheric boundary layer and soil surface into several layers, and integrate these explicitly into the model (e.g., Sellers and Lockwood, 1981; Shuttleworth and Wallace, 1985; Choudhury and Monteith, 1988). The second group includes models that subdivide only the canopy layer into several horizontal layers, and do not integrate the layered soil surface as explicitly as the previous group (cf. Waggoner and Reifsnyder, 1968; Sinclair *et al.*, 1976).

Shuttleworth and Wallace's (1985) multilayer model estimates evaporation from sparse crops using a Penman–Monteith-type combination equation. This model divides the boundary layer into several layers: the soil surface; the layer between soil surface and mean canopy layer; the mean canopy layer; and the layer between the mean canopy layer and the reference height above the canopy. It also includes the concepts of aerodynamic resistance and canopy resistance, and the less well-known concept of bare soil resistance. These resistance terms can be applied at different horizontal layers (e.g., substrate surface resistance; bulk boundary layer resistance between the vegetation surface and the canopy air stream; transfer resistance between the mean canopy layer and a reference height; and transfer resistance between the substrate surface and the adjacent boundary layer). Two Penman–Monteith-type equations have been devised, one for

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the latent heat flux from the substrate, and one for the plant canopy. Summation of these two provides values for the whole boundary layer. The weakness of this model is its failure to address the three-dimensional nature of the various boundary processes. Also, Shuttleworth and Wallace (1985) arbitrarily divide the boundary layer into 'the crop' and 'the soil' layer. The model considers only the soil beneath the vegetation. For row crops, it does not allow any room to deal with the soil between the rows. A similar model has been developed by Choudhury and Monteith (1988), where the canopy layer and soil layer are divided into two more layers. Energy flux is defined by resistances. This model is only applicable to homogeneous surfaces. Choudhury and Monteith (1988) assume a small gradient for lateral fluxes, overlooking horizontal fluxes – one of the major weaknesses of their model. Compared with integrated crop and soil multilayer models, crop-only multilayer models are less realistic in their assumptions. For example, Waggoner and Reifsnyder (1968) assume a soil surface that does not exchange water or sensible heat with the adjacent atmospheric layer. This assumption affects the model's estimation of latent energy flux by influencing vapour pressure and resistance terms.

Eddy diffusion models incorporate the effects on evapotranspiration of turbulent transfer of water vapour. The theoretical basis of these models and their assumptions offered a new direction for understanding ET-related processes. Some of the important contributions in eddy diffusion modelling were made by Garratt and Hicks (1973), Rosenberg *et al.* (1983), Butler (1986), Myers and Paw (1987), Massman and Dijken (1989) and Wilson (1989).

Myers and Paw (1987) included the latent heat of vaporization, the density of the air, the saturation-specific humidity of the leaf temperature and the specific humidity of the air – which all play important roles in eddy diffusion. Their model validated the flux–gradient relationship above the canopy and also turbulent transport within the canopy. Additionally, Myers and Paw (1987) included resistance terms in their model. However, although the resistance terms are not a primary determinant of ET in Myers and Paw's model, the model is an important development in reconciling combination-type and eddy-type diffusion models. Similarly, Rosenberg *et al.* (1983), Butler (1986) and Massman and Dijken (1989) integrate resistance terms. As expected, diffusion terms are explicitly added to these models. Compared with these, Wilson (1989) does not address resistance terms at all. The theoretical basis of this model is dependent on the assumption of the turbulent transport of water vapour. Wilson criticized combination methods because of their 'failure' to address the process of turbulent diffusion.

Legg and Monteith (1975) criticized eddy diffusion models for their inability to comprehend the role of physiological resistance in the ET process. Since eddy diffusion models provide a good description of the turbulent transport of water vapour, and since combination-type methods offer a reasonable description of the physiological control of the ET process, a combination of both approaches would be an ideal basis for ET modelling.

In general, both combination and eddy diffusion type models are based on theoretically sound assumptions. But the question is, again, are these models suitable for large-scale applications? The answer, again, is no: all these models were tested and applied to very small areas, and are thus largely suitable for small-scale applications. Moreover, the assumptions and parameterization schemes of these models are only suitable for such applications. It is, for example, impossible to collect data for stomatal resistance, aerodynamic resistance and soil surface resistance of the various types of soil under

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variable climatic and soil-water content conditions; for the leaf area index of the various types of vegetation; for the interception loss from the different types of vegetation under storms of differing intensity and character; and for the surface roughness length for large basins/regions or for the globe to calibrate and validate the models. Despite their sophistication, these models adopt a 'big leaf' approach for stomatal resistance estimation which assumes identical stomatal resistance for all types of vegetation. Such an assumption clearly implies the unavailability of data not only for macroscale but also for microscale study. Thus, Shuttleworth (1991) is correct when he suggests that, despite the superiority of multilayer and single-layer models for ET calculation, the lack of short-term meteorological data and the unavailability of a stomatal resistance submodel is a major impediment to the application of such models (even for the small scale). He noted that the calculation of a standard evpoaration rate and its subsequent modification by a crop factor would provide a useful solution to this problem.

Thornthwaite's (1948) ET estimation method is one of two means towards solving this problem. Several large-scale studies have been performed using this method or a modified version of it (e.g., Willmott *et al.*, 1985; Serafini and Sud, 1987; Mintz and Walker, 1993). Willmott *et al.* (1985) used Thornthwaite's ET method to estimate global potential ET in the process of calculating the planetary water budget. Legates and Willmott (1995) noted that this method's performance is relatively satisfactory when compared with the Jensen–Haise, Priestley–Taylor and Penman methods. However, Willmott *et al.* (1985) claim that Thornthwaite's method systematically underestimates potential ET when compared with lysimeter records. Legates and Mather (1992) also point out that, owing to the model's biased precipitation estimates, and in order to calculate correct streamflow and runoff estimates, the method must underestimate potential ET. Willmott (1984) suggests that, by using a simple linear regression analysis, it is possible to remove the model's systematic bias and hence attain a relatively satisfactory estimate. Another drawback in the method lies in its calculation of potential ET under nonstressed conditions. To overcome this, Dooge (1992) introduced a correction factor, and Mintz and Walker (1993) further refined the model by introducing an 'equivalent temperature' component to estimate potential ET. Mintz and Walker also found that the true potential ET decreased from 6.2 mm day^{-1} to 5.1 mm day^{-1} (the former using a shelter-high temperature).

Serafini and Sud (1987) used a modified version of the Thornthwaite method to estimate global ET. They modified the moisture availability function of Nappo's (1975) ET equation (see the Appendix). Serafini and Sud achieved satisfactory results in estimating planetary ET. Manabe (1969) used a modified version of Budyko's (1956) method of ET calculation, the Geophysical Fluid Dynamics Laboratory (GFDL) general circulation model (GCM), to estimate the ET and soil moisture of the globe. Delworth and Manabe (1988) report that there are notable differences between potential ET estimated by the GFDL GCM and the very rough approximations calculated by Budyko's method for the summer months (June–September). They also have found that the GFDL GCM application resulted in an ET estimate of 1–3 m for the southern USA. Rind *et al.* (1990) calculated an ET estimate of more than 3 m for the same season and region. The application of Budyko's (1956) rough estimation method has produced much more realistic assessments of 1–2 m. Further, UNESCO (1974) found that ET for the southern USA ranges between 0.8 and 1.8 m during the summer months.

From the above discussion it is apparent that the theoretical basis of these models

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is weak and largely assumes a linear relationship. Therefore, incorporating a solid theoretical basis into these methods would help to improve our predictive power.

An alternative approach to the above methods is the use of the BATS (Dickinson and Sellers, 1988) or SiB (Sellers *et al.*, 1986) models for global ET estimation. These models are theoretically sound, physically based methods that have been developed for large-scale applications. However, the problem with these models lies in the fact that, currently, data are not available to calibrate or validate the models' differing components. A promising solution to this problem of calibration and verification can, nevertheless, be found in the use of remotely sensed data. Several studies have recently been conducted successfully using satellite data to estimate fluxes (cf. O'Kane, 1991; Running, 1991; Schmugge and Becker, 1991).

IV Soil moisture

It is evident from the above discussion of runoff and ET that, depending on the scale of application, a separate modelling approach is needed to estimate soil moisture. For example, Sharma and Luxmoore (1979) used scaling theory and Monteith's (1965) method to determine successfully the soil moisture budget when estimating the water balance of a small watershed in Oklahoma. Similar work has been carried out by Peck *et al.* (1977), Federer (1979) and Clapp *et al.* (1983). The data requirements for calibration and validation are so extensive for all these models that they are applicable only to small-scale studies. On the other hand, Thornthwaite's water balance method has been applied extensively to regional or global-scale soil moisture calculation. Willmott *et al.* (1985), Serafini and Sud (1987) and Mintz and Walker (1993) successfully applied this method to calculate the global soil moisture budget. Gleick (1987) used a modified version of the Thornthwaite method to estimate the water budget of the Sacramento basin. Delworth and Manabe (1988) modified Budyko's (1956) method in their soil-moisture budget calculation for the globe under $2 \times CO_2$ conditions. In his water balance model, Thornthwaite (1948) introduced a soil-moisture resistance function which is a ratio of the actual and maximum possible soil moisture conditions at the root zone. This function was later modified by Nappo (1975). Nappo assumed that soil moisture is readily available to plants until soil-moisture storage capacity is reduced to 30%. Under storage conditions lower than this, the ability of plants to extract moisture from the soil decreases rapidly. Legates and Willmott (1995) note that owing to the assumption that precipitation would fall every day of the month, ET would be overestimated and runoff underestimated. Further, the assumption of a uniform 15 cm root zone for worldwide vegetation cover is unrealistic: the depth of grasslands' root zones can be shallower than 15 cm, while tropical forests' root zone depths are usually much depeer than 15 cm. As discussed earlier, the depth of the root zone influences the moisture-holding capacity of the soil which, in turn, affects the soil-moisture function term in the model. Thornthwaite's water budget model assumes a single layer of soil (Legates and Willmott, 1995), which is not realistic. Moreover, it is difficult to acquire detailed information about soil for global-scale studies. However, regardless of these weaknesses, the Thornthwaite method is one of the few models that can be applied to large-scale regional studies.

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V Conclusions

Owing to their successful inclusion of biophysical-chemical processes, 'physically based' models should be an ideal tool for soil moisture estimation. However, data restrictions make it difficult to apply these sophisticated models to larger-scale studies. Despite this drawback, these models can be used for research and educational purposes. In the mean time, Thornthwaite-type models can be modified, and a more solid theoretical basis can be provided for large-scale applications. Furthermore, as a result of data restrictions and owing to the extensive parameterization scheme, the SiB or BATS-type models are not ready for regional and global-scale studies. The use of remotely sensed data for land surface characterization, model calibration and model validation is quite promising, and more emphasis should be placed on developing proper methods to utilize these data. Additionally, some of the model-building exercises should engage themselves in improving the data archives.

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Appendix

Energy balance models

Penman (1948)

$$\lambda E = \frac{sQ_N + \gamma E_a}{s + \gamma}$$

and

$$E_a = f(U)\ (e_s - e_a)$$

where Q_N is net radiation, γ is the psychrometric constant, s is the slope of the saturation vapour pressure curve, E_a is actual evapotranspiration, e_s and e_a are the saturation and actual vapour pressure and U is the wind speed in km day^{-1} at 2 m height. Several methods of $f(U)$ estimation were proposed by Doorenbos and Pruitt (1975), Thom and Oliver (1977) and Stigter (1980).

Doorenbos and Pruitt (1975)

$$f(U) = 0.27(1 + U/100)$$

Thom and Oliver (1977)

$$f(U) = 0.26(1 + U/100)$$

Stigter (1980)

$$f(U) = 0.37(1 + U/160)$$

Van Bavel (1966)

$$\lambda E = -\frac{s(R_n + S) + \gamma L B_v(e_s - e_a)}{s + \gamma}$$

where s is the slope of the saturation vapour pressure curve, S is soil heat, R_n is net radiation, L is the latent heat of vaporization and B_v can be expressed as follows:

$$B_v = \frac{\rho_a \epsilon k^2}{P} \frac{U}{[\ln(z/z_0)]^2}$$

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where k is von Karman's constant, P is atmospheric pressure, ρ_a is the density of moist air and ϵ is the ratio of the molecular weight of air and water, z is the height above surface and z_0 is roughness length.

Priestley–Taylor (1972)

$$\lambda E = \alpha \frac{s}{s + \gamma} (R_n + S)$$

where α is an empirically derived constant, 1.26.

Interception models

Thom and Oliver (1977)

$$\lambda E = \frac{\Delta Q_N + \gamma E_{\dot{a}}}{\Delta + \gamma(1 + n)}$$

where Q_N is net radiation, γ is the psychrometric constant, Δ is the slope of saturation vapour pressure versus the temperature curve for water at air temperature, n is the ratio between aerodynamic and surface resistance and $E_{\dot{a}}$ is a modified version of the equivalent term in the Penman equation:

$$E_{\dot{a}} = 13.8(e_s - e_a)(1 + U/100)/\ln^2(z/z_0)$$

where e_a and e_s are the actual and the saturation vapour pressure measured at a height z, z_0 is the estimate of the aerodynamic roughness parameter for the vegetation and U is wind speed.

Gash (1978; 1979)

$$\lambda E = \frac{\Delta Q_N + \gamma E_{\dot{a}}}{\Delta + \gamma(1 + r_{sd}/r_a)} + I(1 - c)$$

where r_a is surface resistance, r_{sd} is stomatal resistance, I is interception loss and c is a correction term. This correction term calculates transpiration under wet conditions. It can be expressed as follows:

$$c = (\Delta + \gamma)/[\Delta + \gamma(1 + r_{sd}/r_a)]$$
$$I = P_s + nS + (E/R)(P - P_s)$$

where P is precipitation input, P_s is the fraction of precipitation in rain storms less than $S/(1 - p)$, S is canopy storage, n is the number of storms with precipitation greater than $S/(1 - p)$, R is the mean rainfall rate in storm conditions, and E is the mean evaporation rate from a totally wet forest canopy in storm conditions.

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Single layer

Monteith (1965)

$$\lambda E = -\frac{s(R_n + S) + \rho_a C_p(e_s - e_a)/r_a}{(s + \gamma)[(r_a + r_c)/r_a]}$$

where s is the slope of the saturation vapour pressure curve at the mean wet-bulb temperature of the air, r_a is atmospheric resistance, r_c is canopy resistance, S is soil heat flux, e_a is the partial pressure of water vapour in the air, e_s is the saturation water vapour pressure, C_p is the specific heat of the air at constant pressure, ρ_a is the density of moist air and γ is the psychrometric constant.

Grimmond and Oke (1991)

$$\lambda E = \frac{s(Q^* + Q_F - \Delta Q_s) + (C_a V)/r_a}{s + \gamma(1 + r_s/r_a)}$$

where Q^* is the net radiation, Q_F is anthropogenic heat flux, ΔQ_s is storage heat flux, C_a is heat capacity, V is vapour pressure deficit and r_a and r_s are aerodynamic and surface resistance.

Multilayer

a Integration of crop and soil

Waggoner and Reifsnyder (1968)

$$\lambda E = (VPD) + \theta \Delta E/\Delta T) C_p(\rho/a)$$

where *VPD* is the vapour pressure deficit of the air at the canopy top, θ is air temperature, a is the psychrometric constant and $\Delta E/\Delta T$ is the change of vapour pressure with change in temperature.

Shuttleworth and Wallace (1985)

$$\lambda E_s = (\Delta A_s + \rho c_p D_0/r_{sa})[\Delta + \gamma(1 + r_{ss}/r_{sa})]^{-1}$$
$$\lambda E_c = [\Delta(A - A_s) + \rho c_p D_0/r_{ca}][\Delta + \gamma(1 + r_{cs}/r_{ca})]^{-1}$$

where λE_s and λE_c are evaporation from the substrate and the plant canopy, respectively; A and A_s are total latent and sensible heat flux from the complete crop and substrate; D_0 is the vapour pressure deficit which integrates the concept of transfer resistance between the mean canopy layer and the reference height; r_{ss} is surface resistance at the substrate surface, while r_{ca} is the bulk boundary-layer resistance which controls the transfer between the vegetation surface and the canopy air stream, and r_{sa} is the transfer resistance between the substrate surface and the adjacent boundary layer.

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Choudhury and Monteith (1988)

$$\lambda E_v = \frac{\Delta_1 R_v + \rho C_p D_b / r_1}{\Delta_1 + \gamma(1 + r_c / r_1)}$$

where R_v is net radiation at the vegetation surface, ρ is the density of moist air, C_p is the specific heat of air, D_b is the saturation deficit, γ is the psychrometric constant, r_1 is aerodynamic resistance between the canopy and air, and r_c is canopy stomatal resistance.

$$\lambda E_s = \frac{\zeta \Delta_2 R_s + \rho C_p [\Delta_2 (T_m - T_b)/r_1 + \eta D_b / r_2}{\Delta_2 + \gamma_2}$$

where R_s is net radiation at the soil surface, T_b is the mean air temperature, T_m is the temperature at the bottom of the wet soil layer, r_1 is the resistance of the wet layers proportional to the depth and inversely proportional to thermal conductivity, ζ and η are the functions of resistances, and r_2 is the resistance between the soil surface and the canopy.

b Crop only

Sinclair et al. *(1976)*

$$\lambda E = L_i \frac{\rho_{v_s}(TL_i) - \rho_{v_i}}{RA_i + RS_i}$$

where L is the leaf area index, RS_i and RA_i are stomatal and aerodynamic resistances and $\rho_{v_s}(TL_i)$ is saturated vapour density.

Eddy diffusion models

Wilson (1989)

$$\langle \overline{E} \rangle = \langle \overline{W \rho_v} \rangle$$

where W is the vertical velocity across the horizontal plane and ρ_v is vapour density.

Massman and Dijken (1989)

$$\frac{d}{dz}\left(K_e(z)\frac{de}{dz}\right) = \frac{-a(z)(e_f - e)}{r_b(z) + r_s(z)}$$

where z is the height above the ground, e is the mean atmospheric vapour pressure within the canopy, and e_f is the vapour pressure within the substomatal cavity of the leaves and is assumed to be constant with height. K_e is the turbulent diffusivity of water

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vapour, $a(z)$ is foliage density, $r_b(z)$ is the individual leaf boundary-layer resistance, and r_s is individual leaf stomatal resistance (all these parameters are a function of height, z).

Butler (1986)
For drops:

$$\partial E_d/\partial\zeta = \alpha_s\rho(q_d - q)/r_d$$

and for leaves:

$$\partial E_l/\partial\zeta = \alpha_l\rho(q_l - q)/(r_s + r_l)$$

where ζ is canopy height, and α_s and α_l are nondimensional area density and nondimensional dry-leaf area density, respectively. r_d, r_s and r_l are resistance terms, q_d and q_l are the specific humidities of saturated air at drop and leaf temperature, respectively, and q is the specific humidity of the air.

Total flux divergence:

$$\partial E/\partial\zeta = \partial E_l/\partial\zeta + \partial E_d/\partial\zeta$$

Myers and Paw (1987)

$$E = 2L_v\rho\,\frac{(\bar{q}_l - \bar{q})}{r_b - r_s}$$

where L_v is the latent heat of vaporization, ρ is air density, r_b and r_s are boundary-layer resistance and stomatal resistance, $\bar{q}_l$ and $\bar{q}$ are the saturation specific humidity at leaf temperature and the specific humidity of the air.

Rosenberg et al. *(1983)*

$$E = \frac{(M_w/M_a)}{P}\,\rho_a C_p K_w\,\frac{\partial e_a}{\partial z}$$

where M_w and M_a are the molecular weight of water vapour and air, K_w is the turbulent exchange coefficient for water vapour, ρ_a is air density, C_p is specific heat at constant pressure, P is atmospheric pressure and $\partial e_a/\partial Z$ is the vertical gradient of vapour pressure.

Temperature-based models

Thornthwaite (1948)

$$ET_p = 16\left(\frac{l_1}{12}\right)\left(\frac{N}{30}\right)\left(10\,\frac{T_a}{I}\right)^{a_1}$$

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where ET_p is mm month^{-1}, l_1 is actual day length (h), N is the number of days in a month, T_a is the mean monthly air temperature and a_1 is defined as

$$a_1 = 6.75_*10^{-7}I^3 - 7.71_*10^{-5}I^2 + 1.79_*10^{-2}I + 0.49$$

where I is a heat index calculated from the sum of 12 monthly values, i, obtained from

$$i = \left(\frac{T_a}{5}\right)^{1.514}$$

Nappo (1975)

$$E_a = E_{p*}M(\eta)$$
$$M(\eta) = 1 - e^{-56.6(\eta)}$$

where $M(\eta)$ is the moisture availability function. Mintz and Serafini (1984) modified the moisture availability function as follows:

$$M(W,W^*) = 1 - e^{\alpha(W/W^*)}$$

where W is the available soil moisture, W^* is the difference between the soil and the moisture in the soil at vegetation wilting point and α is a transformation constant. Serafini and Sud (1987) modified Mintz and Serafini (1984), proposing a soil moisture availability function as follows:

$$M(W,W^*) = 1 - e^{-\alpha(W/W^*)}/(1 - e^{-\alpha})$$

Hargreaves (1974)

$$ET_p = MF(1.8T_a + 32)CH$$

where ET_p is mm month^{-1}, MF is a monthly latitude-dependent factor, T_a is the mean monthly temperature and CH is the correction factor for relative humidity (RH). CH is only used when mean daily relative humidity values exceed 64%.

Solar-radiation-based models

Jensen–Haise (1963)

$$ET_p = R_s(0.025T_a + 0.08)$$

where ET_p is mm day^{-1}, T_a is the mean daily temperature and R_s is daily total solar radiation (mm equivalent of water).

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Solar-radiation and temperature-based models

Hargreaves–Samani (1985)

$$ET_p = 0.00094 * R_a * tF * tD^{\frac{1}{2}}$$

where R_a is the daily extraterrestrial radiation in mm equivalent of water, tF is mean daily temperature and tD is the difference between maximum and minimum daily temperature. R_a is a function of latitude and time of year.

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