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Simplified modeling of catchment-scale evapotranspiration via boundary condition switching

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ABSTRACT

Evapotranspiration is one of the most difficult terms of the water balance to estimate and model, because of its complex links with atmospheric, hydrological, and ecological processes. In this paper we show how a switching procedure for the boundary conditions at the soil surface achieves a good description of evapotranspiration in a catchment-scale process-based hydrological model. The switching algorithm relies on one parameter, a threshold soil water pressure head (ψ_{min}), to distinguish between atmosphere-controlled and soil-limited evapotranspiration. We successfully applied the model to a small water-limited catchment in southwestern Victoria, Australia, mainly used as pasture and where an extensive hydrological data set is available. Our simulation results show that the model is capable of reproducing satisfactorily the hydrological regime of the catchment without the need for a detailed multiparameter calibration. Specifically, the observed daily flow hydrographs, typical of ephemeral streams, are reproduced satisfactorily for both the calibration and validation phases. Water table levels proved to be more difficult to match, even though the overall groundwater dynamics are well captured by the model. The comparison between measured and observed evapotranspiration rates demonstrates the capability of ψ_{min} to describe the conversion of potential evaporative demand into actual evapotranspiration. The effect of ψ_{min} on the components of the catchment water balance and the model numerical performance were investigated for two different soil types through a sensitivity analysis. The modeled reduction of evapotranspiration with decreasing soil water potential is shown to be analogous to the commonly adopted Feddes formulation of water stress. The results show that the boundary condition-switching algorithm, with a proper choice of ψ_{min} and soil retention curves, can represent a simple and effective way to account for the impacts exerted on the catchment hydrological response by shallow rooted vegetation.

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1. Introduction

Evapotranspiration (*ET*) is one of the fundamental terms of the hydrologic cycle at all scales, yet it is also one of the most difficult to model, being influenced by many factors, such as air temperature, soil moisture, vegetation type, and atmospheric advection [1,2]. Therefore, the practical application of hydrological models where *ET* plays a significant role is typically subject to large uncertainties.

Several hydrological catchment-scale models solve the coupled equations for surface and subsurface water flow [3], building on the blueprint paper by Freeze and Harlan [4]. Generally known as physically based integrated surface–subsurface hydrological models (ISSHMs [5,6]), popular codes include Integrated Hydrology Model (InHM [7]), MODHMS [8], tRIBS [9], GEOtop [10], ParFlow [11], HydroGeoSphere (HGS [12]), CATHY [13], and PAWS [14]. Despite the generally high computational effort they require, usually due to the detailed resolution of topography and the fully three-dimensional solution of variably saturated subsurface flow, recent advances in computational resources have made it possible to extend the applications of such models from short-term

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simulations of rainfall-runoff processes for small catchments (usually at scales of less than 0.5 km^2 and durations of a few hours [e.g., 15]) to increasingly larger catchments (i.e., up to about 10^3 km^2 [e.g., 9]) and time scales (i.e., up to several decades [e.g., 16]).

Calibration and validation of ISSHMs are generally difficult to achieve and are mostly done by comparing only the discharge at the catchment outlet. This is usually sufficient for short-term rainfall-runoff predictions, where evapotranspiration typically plays a minor role. However, an accurate estimation of many other surface and subsurface states and fluxes, evapotranspiration in particular, becomes critical for the reliability of long-term model applications. A number of different approaches are used in ISSHMs to estimate actual *ET*, with various degrees of complexity. One of the most commonly used methods expresses actual evapotranspiration, ET_a, as a fraction, β, of the potential evapotranspiration, ET_p; ET_a is then treated as a sink term distributed over a predefined soil depth dependent on root depth and density. In the commonly adopted formulation by Feddes et al. [17], β, with values ranging from 0 to 1, is a function of the soil water pressure head, ψ, and accounts for both water and oxygen stress.

Several models use variations of the Feddes' formulation. In InHM [18], for example, β is a function of soil saturation, while the first version of PAWS [14] uses the formulation proposed by Lai and Katul [19]. Other models, such as MODHMS [20] and HGS [21], rely on the more complex formulation developed by Kristensen and Jensen [22], where actual transpiration and soil evaporation are computed separately as a function of potential evapotranspiration, soil moisture, Leaf Area Index (LAI), and root distribution. A more mechanistic approach was adopted in the first version of tRIBS [9], which is similar to what is used in GEOtop [10]. Both codes implement the hydrology-vegetation model developed by Wigmosta et al. [23], where total evapotranspiration is divided into evaporation from wet canopy, canopy transpiration, and bare soil evaporation. These are computed as functions of the potential evapotranspiration, expressed through a comprehensive energy balance at the surface, via estimation of soil, canopy, and aerodynamic resistances. tRIBS was later coupled to a model of plant physiology and spatial dynamics [24], linking water and energy processes of river basins to plant life regulatory processes, with a particular focus on ecohydrology of semiarid environments. More recently, some physically based hydrological models have been coupled to land surface models (LSMs), allowing the integration of the detailed description of variably saturated flow in soils and surface routing to accurate computation of energy, water, and carbon flux exchanges between the land surface and the atmosphere. Examples include the integration of the Community Land Model (CLM) into ParFlow [25] and PAWS [26], as well as the coupling between CATHY and Noah-MP [27]. A similar flux-based approach is used by the Soil Water Atmosphere Plant (SWAP) model [28].

Most of the approaches mentioned above were not specifically designed to be integrated into ISSHMs to account for evapotranspiration processes. As a result, they are either complicated to implement, especially with regard to the exchange of fluxes between subsurface solvers and atmosphere, or need calibration/estimation of multiple parameters. This represents a major bottleneck, particularly for larger scale models run over long time scales, where complex representations of evapotranspiration become computationally expensive and vegetation parameters, such as root distribution and leaf area index, are not readily available.

The aim of this study is to explore the possibility offered by the CATchment HYdrology model (CATHY [13]) to use a boundary condition switching procedure controlled by a single parameter to reproduce actual evapotranspiration in catchment numerical modeling. The only parameter needed by CATHY to compute ET_a as a fraction of ET_p is a soil water pressure head, ψ_{min}, which controls the switching between soil-limited and atmosphere-controlled evapotranspiration. The method is here assessed by comparing simulated and measured *ET* fluxes in an experimental catchment located in southwestern Victoria, Australia.

2. Model description

2.1. Main equations

CATHY is an integrated process-based spatially-distributed model for surface–subsurface flow simulations [13]. The model solves the three-dimensional Richards equation for flow in variably saturated porous media coupled to a one-dimensional diffusion wave approximation of the de Saint–Venant equation simulating the overland and channel routing. The two equations can be expressed as

$$S_w S_s \frac{\partial \psi}{\partial t} + \theta_s \frac{\partial S_w}{\partial t} = \nabla \cdot [K_s K_{rw}(S_w)(\nabla \psi + \eta_z)] + q_{ss}(h, \psi), \tag{1}$$

$$\frac{\partial Q}{\partial t} + c_k \frac{\partial Q}{\partial s} = D_h \frac{\partial^2 Q}{\partial s^2} + c_k q_s(h, \psi). \tag{2}$$

The variables in Eq. (1) have the following meaning: $S_w = \theta/\theta_s$ is water saturation, θ and θ_s being volumetric water content and porosity (or saturated moisture content) [m^3 m^{-3}], respectively; S_s is the aquifer specific storage coefficient [m^{-1}]; ψ is pressure head [m]; t is time [s]; ∇ is the gradient operator [m^{-1}]; K_s is the saturated hydraulic conductivity tensor [m s^{-1}]; $K_{rw}(S_w)$ is the relative hydraulic conductivity function; η_z is the unit vector $(0,0,1)^T$, z being the vertical coordinate directed upward; and the source or sink term q_{ss} [m^3 m^{-3} s^{-1}] represents the water flux contribution from the surface to the subsurface, depending on the pressure head and the surface ponding head h [m].

Surface water is routed according to Eq. (2) along a one-dimensional coordinate system, s [m], of hillslope and channel links, defined on the drainage network automatically extracted by a DEM-based pre-processor. In Eq. (2), Q is the discharge along the hillslope or channel link [m^3 s^{-1}], c_k is the kinematic wave celerity [m s^{-1}], D_h is the hydraulic diffusivity [m^2 s^{-1}], and q_s [m^2 m^{-1} s^{-1}] is the inflow or outflow rate from the subsurface to the surface domain. The ponding head h [m] is derived from the discharge Q via mass balance calculations.

The numerical routing scheme derives from the Muskingum–Cunge discretization of the kinematic wave equation based on the matched artificial dispersivity method. It is assumed that hillslope flow concentrates in rills or rivulets. As such, both channel and hillslope flow can be described by the same Eq. (2) defined on the rill or channel network, using different parameter values to distinguish between the two flow regimes [29]. The spatial and temporal variability of the routing parameters c_k and D_h are expressed through the hydraulic geometry scaling relationships described in [13, Section 2.1], depending on the Gauckler–Strickler conductance coefficient, water-surface width, and channel bed slope. Different options based on threshold-type relationships are available for the distinction between overland and channel flow regimes; in this case we used the drainage area criterion. The surface numerical scheme does not support the modeling of flooding and backwater, but lakes and other topographic depressions can be identified and specially treated by a "lake boundary-following" algorithm that is part of the DEM pre-processing procedure [30,31].

The subsurface solver is based on Galerkin finite elements in space, a weighted finite difference scheme in time, and linearization via Newton or Picard iteration [32].

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