

A framework for coupling atmospheric and hydrological models

ALAIN PIETRONIRO

National Water Research Institute, Environment Canada, Saskatoon, Saskatchewan S7N 3H5, Canada

e-mail: al.pietroniro@ec.gc.ca

ERIC D. SOULIS, KEN SNELGROVE & NICK KOUWEN

Department of Civil Engineering, University of Waterloo, Waterloo, Ontario N2L 3G1, Canada

Abstract There is a strong global research effort in coupling atmospheric and hydrological models for improved hydrological flow modelling and improved atmospheric simulation. The land surface is an important hydrological control as it is the primary influence in the surface water budget and it is almost always a requirement in the implementation of either hydrological or atmospheric models. Sophisticated soil–vegetation atmospheric transfer (SVAT) schemes are currently being implemented in global climate models (GCMs), regional climate models (RCMs) and day-to-day operational forecasting numerical weather prediction models (NWP). Rarely have these been incorporated into hydrological models. This paper focuses on describing the coupling of an atmospheric model (both NWP and RCM) with a hydrological model. A conceptual framework for model development was initiated using different levels of coupling in order to work towards a complete two-way coupled model.

Key words numerical weather prediction model; hydrological model; land surface–atmosphere coupling; SVAT model; surface water budgets

INTRODUCTION

The linking of hydrological and atmospheric models is often based on uncoupled (linked) model approaches where hydrological models are linked to atmospheric models through simple forcing of the hydrological models, using the atmospheric model output. Such linking however can be misleading since the land surface is treated independently in each model, resulting in inconsistent basin state variables. This can be avoided if the two models share the same land surface scheme. It is often suggested that combining climate models with a hydrological streamflow models through a common SVAT scheme can potentially improve the land surface representation, benefiting the streamflow prediction capabilities of the hydrological models as well as providing improved estimates of water and energy fluxes into the atmosphere. This paper focuses on describing the coupling of an atmospheric model—both numerical weather prediction (NWP) and regional climate models (RCM)—with a hydrological model.

THE WATFLOOD–GROUPED RESPONSE UNIT APPROACH

Physically-based simulation models often require that the watershed is broken down into smaller units to more closely represent the observed hydrological and hydraulic

phenomena. Distributed models are defined by their ability to incorporate the distributed nature of watershed parameters and inputs into a modelling framework. Fully-distributed models apply detailed physics in differential form but are too complex and data intensive to solve for large basins. Lumped hydrological models often lack the detail of physics and distributed inputs. Two widely accepted distributed model procedures are the hydrological response units (HRUs) approach and the grouped response units (GRUs) approach. An HRU is an areal element within a basin where the hydrological properties are definable and would not be significantly different if a smaller scale of discretization were used. This approach is appropriate for small basins and grids, since it is data and CPU intensive. A more sensible approach for large basins is the GRU, which is a grouping of all areas with a similar land cover (or other attribute), such that a grid square will contain a number of distinct GRUs. Runoff generated from the different groups of GRUs are then summed together and routed to the stream and river system (see Fig. 1). Two GRUs with the same percentages of land-cover types, rainfall and initial conditions will produce the same amount of runoff regardless of how these land cover classes are distributed. The major advantage of the GRU approach is that it can incorporate the necessary physics while retaining simplicity of operation.

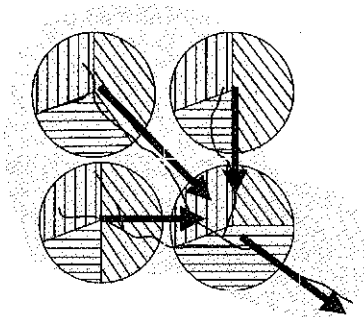


Fig. 1 The GRU approach to basin discretization used in WATFLOOD and WATCLASS. The shading with each circle represents the land-cover based GRU within a grid square. Water is routed along the stream in the downslope direction as indicated by the arrows.

The WATFLOOD model (Kouwen *et al.*, 1993) divides a watershed into a number of GRUs and discretizes the basins into a series of a square grids. The objective of using the GRU method of discretization is to model hydrologically-consistent sub-areas of the watershed, each with known properties. The surface water budget is computed for each GRU within a grid square (see Kouwen *et al.*, 1993 for details) and infiltrated using the well-known Green-Ampt approach. The model then computes overland flow from the Manning equation, when the infiltration capacity is exceeded by the water supply, and the depression storage has been satisfied. Infiltrated water is stored in a soil reservoir referred to as the upper zone storage (UZS). Water within this layer percolates downward or is exfiltrated to nearby water courses using simple storage–discharge relationships. All GRUs within a grid contribute to shared lower zone storage (LZS). An initial LZS is determined through trial and error by matching stream baseflow at the outlet to the observed hydrograph. Groundwater or LZS is replenished

by recharge from the UZS according to a power function. A groundwater depletion function is used to gradually diminish the baseflow. There is only one LZS for each grid. Flow rates through soil depend upon the hydraulic conductivity that is optimized on the basis of land cover. The total inflow to the river system is found by adding the surface runoff and interflow from all GRUs in a grid and the baseflow. These flows are added to the channel flow from upstream grids and routed through the grid to the next downstream grid. The routing of water through the surrogate channel system is accomplished using a storage routing technique. More sophisticated routing models are available, but the application of such models does not appear to be warranted. In fact, for large watersheds, differences between the routing methods may well be smaller than the noise in the data (M. Ponce, personal communication, 1990). Input from a digital elevation model (DEM) and land cover are the spatial data sources needed to generate the required physiographic files.

THE MODELLING FRAMEWORK

The GRU approach is appropriate for atmospheric–hydrological coupling for a number of reasons. Firstly, the landscape-based approach for model parameterization requires a minimum amount of ground-based data for model calibration. Although not an optimal scenario for modelling, this situation is especially true for large or remote basins where it is not possible to determine detailed physiographic information. The GRU approach also allows for detailed physics in the vertical water budget on a landscape basis. This is entirely consistent with the mosaic approach currently being implemented in atmospheric models. The WATFLOOD model theoretically can incorporate any surface vertical water budget (SVAT) model and is currently coupled to the Canadian Land Surface Scheme (CLASS; Verseghy, 1991). The resulting model, WATCLASS forms the basis for coupling with both weather and climate atmospheric models.

The WATFLOOD hydrological model was linked to the CLASS-SVAT scheme in order to add an energy component to the existing WATFLOOD-SVAT (referred to as SIMPLE—Kouwen *et al.*, 1993) and improve the vertical water balance calculated by CLASS alone, while allowing for proper streamflow routing between elements. This coupling has occurred at different levels. Level I involves coupling of an atmospheric model with the land surface scheme (CLASS), with limited or no runoff calculations. In level II coupling, there is full interaction between land surface and runoff models (WATCLASS), driven by prescribed measured, forecast, and/or analysed fields of near-surface data, but no feedback to the atmosphere. In level III, there is full interaction between both atmosphere and land surface models, and between land surface and runoff models. We refer to level 0 coupling as model linking where the hydrological model uses the atmospheric model output as the meteorological forcing field to drive the hydrological model. In this case there is truly no feedback to the atmosphere. These linked model systems are often misrepresented as coupled systems.

COUPLING THE MODELS

The method developed by Soulis *et al.* (2000), which is a physical representation of the current soil drainage scheme in WATFLOOD, was adopted for use within the

WATCLASS model. The method is based on topographic index approaches that focus on the near-surface water balance of valley slopes (Beven & Kirkby, 1979). The current CLASS scheme has no horizontal drainage (see Fig. 2(a)), resulting in the topsoil remaining wet too long after a rainfall event. This results in evapotranspiration that is overestimated between runoff events, while infiltration will be too low during these events. To address this, a physically-based transfer functions was included in the WATCLASS model to allow movement of water from the soil column into the micro-drainage system in a physically realistic manner. This is depicted in Fig. 2(b).

A fundamental drainage element is conceptualized within the WATFLOOD and WATCLASS framework as a grid element. The element can be viewed as an assembly of sloped blocks, each with three soil layers and each with a connection to the drainage system (Fig. 3). One of the morphological properties we require is L_S , the typical length of a block within the element that supplies, or has the potential to supply, a segment of a receiving stream of length L_V . This can be determined from the drainage

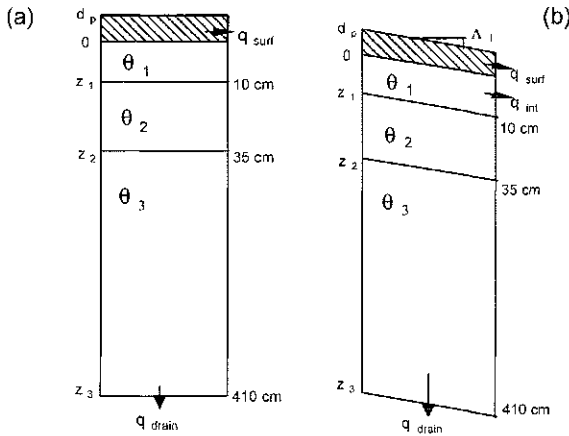


Fig. 2 Soil moisture representation in CLASS and WATCLASS.

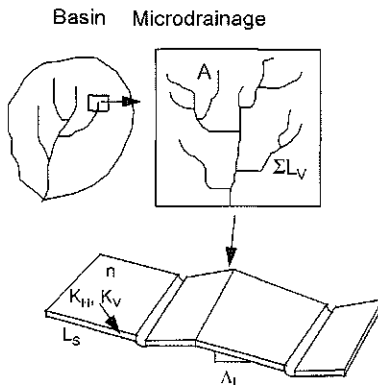


Fig. 3 Conceptualization of micro-drainage within a GRU incorporated in the WATFLOOD model. The square grid depicted above represents a WATFLOOD grid square. The slope facets represent the different GRUs within the grid.

density, D_D , of the element, defined as $\sum L_V/A$ where A is element area (Soulis *et al.*, 2000). Drainage density is landform dependent and typically ranges from 2 to 100 per kilometre (Dingman, 1994); L_S is the average distance from the divides in the micro-drainage system to the stream channels and is equal to $1/2D_D$ (Dingman, 1978). This distance represents the “slow” portion of the water flow system, via overland routes or through the soil matrix. It is important to note that the WATFLOOD and WATCLASS models require that a stream element be present within each grid unit. Also relevant is Λ_I , the typical valley slope, which we call internal slope in order to distinguish it from the overall slope of the element. Λ_I provides the topographic gradient for flow from the soil blocks. An element has a mosaic of sloped tiles, with average dimensions L_S and L_V and average slope Λ_I , drained by a system of microchannels. Elements are large enough that the majority of the inter-element flow is channel flow.

As in WATFLOOD, the excess surface water will drain to the micro-drainage system as overland flow. Horizontal near-surface flow, called interflow, will occur through the soil matrix and the macropore structure, leaving the block through the seepage face. Finally, drainage will occur from the bottom of the soil column into the groundwater system to appear later as base flow in the stream. WATCLASS overland flow is well represented by Manning’s equation, which is the momentum equation applied to open channel flow. The form for a wide channel is:

$$v = \frac{1}{n} d_e^{2/3} \Lambda_I^{1/2} \quad (1)$$

where v is overland flow velocity, d_e is effective depth (depth above ponded water), Λ_I is internal slope, and n is Manning’s roughness coefficient. The depth of flow at the stream edge will depend on how much of the slope is contributing overland flow and how much concentration is occurring. Since the two factors offset each other, we assume the best estimate of depth of flow at the stream bank is the average effective depth. Therefore the flow entering a stream segment is:

$$Q_{\text{over}} = \left(\frac{1}{n} \right) d_e^{5/3} \Lambda_I^{1/2} L_V \quad (2)$$

where Q_{over} is overland flow. In terms of flow per unit horizontal area:

$$\begin{aligned} q_{\text{over}} &= Q_{\text{over}} / L_V L_S \\ &= \left(\frac{2D_d}{n} \right) d_e^{5/3} \Lambda_I^{1/2} \end{aligned} \quad (3)$$

where q_{over} is overland flow runoff. Horizontal flow from within the soil layers, which we define as interflow, is less straightforward but can be considered a major flow mechanism (Dingman, 1994). The primary interflow mechanism is saturated and unsaturated matrix flow, enhanced by macropore flow near the surface. Moisture content typically varies from saturation immediately after a heavy rainfall event, to the wilting point after a sustained dry period. Interflow occurs almost entirely when soil moisture is between saturation and field capacity. It can be derived from the Richards equation, with the added complication that hydraulic conductivity is not constant, often decreasing several orders of magnitude in the top 10–20 cm, with horizontal to vertical ratios of 10–100 (Bear, 1972). A rigorous analysis would require a detailed solution to

Richards' equation. However the data and computational requirements for such an approach are substantial. Soulis *et al.* (2000) proposed a conceptualization of the phenomenon. Fundamental to this development is expressing the variation in hydraulic conductivity with depth by a power law as:

$$K_H(\theta) = K_0(\theta) \left(1 - \frac{h}{D}\right)^e \tag{4}$$

where K_0 is surface hydraulic conductivity, D is the depth of layer 1, h is the relative distance below the surface and e is an exponent normally between 1 and 5. Combining with the Richards equations for unsaturated flow in a porous media gives:

$$\frac{\partial S}{\partial t} = \frac{1}{\theta_s} \frac{\partial}{\partial x} \left[K_{s0} d^e S^c \Lambda_f \right] \tag{5}$$

where $S = \theta/\theta_s$, $d = 1 - h/D$ and K_{s0} is the saturated hydraulic conductivity at the surface, equal to $K_0(\theta_s)$. The solution to this differential equation for an initially saturated aquifer is approximated as the power law function and the drainage of this saturated layer is approximated by (Soulis *et al.*, 2000):

$$q_{INT} = 2D_D K_{0H}(\theta_s) \left(\frac{\theta_1 - a\theta_s}{\theta_c - a\theta_s} \right)^f \frac{D}{e+1} \Lambda_f \tag{6}$$

where $f = a \cdot c \cdot g(e)$ and $g(e) = \frac{(e+1)(e+2)}{(5e+2)}$. The model is depicted schematically in

Fig. 4. At $t = 0$, flow from the seepage face is fully saturated. After $t = t_c$, the water table drops below the surface of the face and the interflow becomes a mixture of

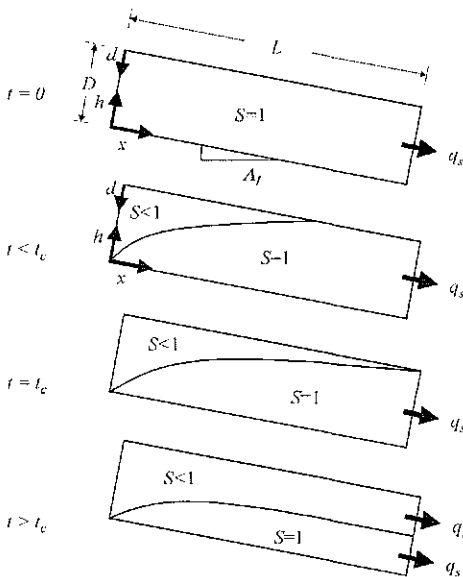


Fig. 4 Schematic of outflow from a shallow aquifer (after Soulis *et al.*, 2000).

saturated and unsaturated flow. Behind and above the water table S declines in both time and space. For typical values of e , which is normally between 0 and 3, $g(e)$ ranges from 0.8 to 1.3. All the parameters for equation (6) can be determined *a priori*. This includes the function $g(e)$, which represents land cover influences on soil structure. However, field studies and/or hydrograph optimization techniques are needed to determine its effective value for a grid element, which may vary from predicted values. The WATFLOOD model currently uses a default interflow procedure that is equivalent to assuming $g = 1$. Our tests with boreal forest data suggests that a value of g between 0.8 and 0.9 is appropriate in that environment (Whidden, 1999). Percolation, the downward movement of water from the unsaturated zone to the groundwater system and ultimately to the river system as base flow, is well represented by Darcy's law and is given as:

$$q_{\text{down}} = K_v (\theta_3) \quad (7)$$

The above formulation for all three runoff mechanisms provides a physically-based approach for consistent application of the WATCLASS model as a hydrologically-based SVAT scheme that can be run at level I, II or III model coupling.

WATCLASS APPLICATION

To illustrate the level II linked WATCLASS model, the Boreal Ecosystem–Atmosphere Study (BOREAS) North Study Area (NSA) is examined. This study site, along with the Southern Study Area (SSA), was the focus of intensive observations over the period 1994–1996. Included in these observations were simultaneous measurements of both streamflow and evapotranspiration. This provides a unique opportunity to examine a number of components of the water balance equation:

$$P = E + R + \Delta S \quad (8)$$

where P is the precipitation, E is the evapotranspiration, R is the runoff and ΔS is the change in water storage. Within WATCLASS, P is an independent input variable,

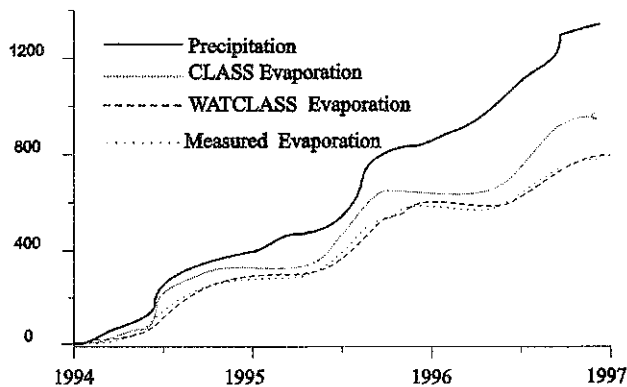


Fig. 5 Measured and simulated evaporation estimates from the NSA Old Black Spruce in BOREAS.

while each of E , R and ΔS are dependent variables calculated within the model. Included in the calculation of the both R and E is a dependence on land surface storage (S) which includes soil moisture. The resulting relationships tightly constrain the water budget so that simulation errors of one component are propagated through the model to the other dependent water balance variables. When both evaporation and runoff are measured simultaneously and are used to constrain the model, partitioning of precipitation into its components becomes much more realistic and parameters controlling the simulation more robust. Applying the WATCLASS model within the BOREAS catchment produces generally drier soil matrix. This in turn reduces the evaporation opportunity and provides a more realistic streamflow hydrograph. Figure 5 shows the impact on model evaporation. The plot shows that the cumulative evaporation over the three-year simulation period is reduced by approximately 200 mm which compares well with the tower measures. In this figure the cumulative rainfall and measured evaporation at the NSA Old Black Spruce are also shown for comparison.

CONCLUSIONS

A robust and consistent method for coupling hydrological and atmospheric models has been presented. Early results show improved runoff and evapotranspiration estimates. Future modelling efforts will focus on the interaction of soil moisture and transpiration from vegetation as well as optimization of land surface parameters using hydrographs.

REFERENCES

- Bear, J. (1972) *Dynamics of Fluids in Porous Media*. Dover Publications, Inc., New York, USA.
- Beven, K. J. & Kirkby, M. J. (1979) A physically based, variable contributing area model of basin hydrology. *Hydrol. Sci. Bull.* **24**(1), 43–69.
- Dingman, S. L. (1978) Drainage density and streamflow: a closer look. *Wat. Resour. Res.* **14**, 1183–1187.
- Dingman, S. L. (1994) *Physical Hydrology*. Macmillan, New York, USA.
- Kouwen, N., Soulis, E. D., Pietroniro, A., Donald, J. & Harrington, R. A. (1993) Grouped Response Units for distributed hydrologic modeling. *J. Wat. Resour. Plan. Manage. ASCE* **119**(3), 289–305.
- Souls, E. D., Snelgrove, K., Kouwen, N., Seglenieks, F. & Verseghy, D. L. (2000) Towards closing the vertical water balance in Canadian atmospheric models: coupling of the land surface scheme CLASS with the distributed hydrological model WATFLOOD. *Atmos. Ocean* **30**(1), 251–269.
- Verseghy, D. L. (1991) CLASS—A Canadian Land Surface Scheme for GCMs. I. Soil model. *Int. J. Clim.* **11**, 111–133.
- Whidden, E. A. (1999) Distributed hydrological modelling in the Canadian boreal forest: subsurface model development, land surface model improvement and modelling results. MASC Thesis, Dept of Civil Engng, Univ. of Waterloo, Waterloo, Ontario, Canada.