Description of the CAS-Hydro (version 1.2) Catchment Hydrological Model

Model Overview

CAS-Hydro is a dynamic, distributed, object orientated, process based, hydrological simulation model which operates at the landscape scale within surface water dominated catchments. There are two facets to the design philosophy of the model: (1) the model should use a minimal parameter set, with parameters that may be determined for any catchment in the U.K. from published national datasets; and (2) the spatial representation of process description should be explicit, so that the model's hydrological component can be used to address questions related to the impacts of both climate change and distributed land management upon flow extremes and, eventually, water quality. The spatially explicit nature of the hydrological routing allows for the natural development of soil moisture patterns and flow connectivity through the interaction between processes as a function of the system dynamics. Hence, CAS-Hydro uses a fully distributed spatial structure and a continuous temporal representation. The model structure is divided into four key sections: weather; point hydrological processes; landscape; and river channel network.

Weather

The weather module assimilates spatially-distributed and time-dependent information on rainfall, temperature and solar radiation. Rainfall forms the main hydrological input into the catchment. The temperature and solar radiation are used to calculate evapotranspiration rates. Temperature is also used for the timing of the growth of vegetation. The base information can be provided from measured data or output from other models such as Global Climate Models. This information is transformed to the temporal scale required for the simulation either through the use of interpolation functions or using stochastic downscaling techniques.

Processes related to the generation and transmission of runoff occurs at the per-minute timescale. Therefore, to capture the dynamics of the runoff process, rainfall information for a temporal scale of minutes is required. The weather module is able to read in this time series from a file or generate the per-minute rainfall from daily totals by applying a stochastic rainfall generator. The generation of the per-minute rainfall time series is achieved through the use of a stochastic rainfall generator. The rainfall generator uses information on the characteristics of the storms in the region to create the daily time series. The approach is based upon the generator developed by Mulligan (1996). The generator goes through the following sequential steps: (1) total daily rainfall depth is read from a file; (2) individual storm rainfall depth totals are generated from a Monte Carlo model parameterised from observed data until the correct amount of daily rainfall has been assigned; (3) the per-minute rainfall intensities within each storms are generated from a Monte Carlo model parameterised from observed data which assumes that there is no temporal structure to the storm rainfall; and (iv) the storms are placed at random times during the day. This algorithm is able to generate a rainfall time series at the required temporal resolution. The data requirements are the daily rainfall time series for the period of interest and a detailed rainfall time series from a tipping bucket type rain gauge. The tipping-bucket series is used to calculate the probability distribution functions for the Monte Carlo generators. Across a large catchment, there will be significant differences in the amount of rainfall received. The model uses a spatially distributed scaling factor [1] to scale the rainfall depths:

$$rf_s = rf.s \qquad [1]$$

where rf_s is the scaled rainfall depth, rf is the original rainfall depth and s is the scaling factor. The spatial pattern of s can be determined either from the interpolation of observed rainfall trends, correlated with environmental parameters such as altitude or from the published rainfall spatial pattern maps in sources like the Flood Estimation Handbook (Reed et al., 2002).

The weather module uses daily maximum and minimum air temperature which are available from observed records. These values are then interpolated to per-second temperatures using:

$$T_{a(s)} = \frac{\sin\left(\frac{d_s + td + (12*60*60)}{4*60*60}\right) + 1}{2} * (t_{\max} - t_{\min}) + t_{\min}$$

[2]

where $T_{a(s)}$ is the air temperature at the current second, d_s is the current second of the day, td is the time between midday and the maximum temperature occurring, t_{max} is the daily maximum temperature and t_{min} in the daily minimum temperature. This equation is able to fit the observed hourly temperature values with an R^2 value of 0.94. Soil temperature is related to air temperature using:

$$T_s = a.T_a + b$$

[3]

where T_s is the soil temperature, T_a is the air temperature and a and b are coefficients paramertisted from observed data.

The amount of solar radiation received at the surface is calculated from solar geometry and cloud cover. The solar geometry gives information on the amount of solar radiation received at the top of the atmosphere for a given latitude, slope and aspect. The cloud cover information simulates the effects of scattering within the atmosphere. The probability of cloud over is determined by the rainfall amount with a fixed cloud cover on rain days and a stochastic function on non-rain days. The amount of net radiation at the surface is determined from the surface albedo and emissivity.

Point-scale hydrological processes

The model simulates rainfall interception by vegetation, the storage of water on the soil surface, the infiltration of water into the soil, the recharge of the aquifer from the soil and the generation and acceptance of runoff and through flow. The first process in the vertical cascade is the interception of a fraction of the rainfall by the vegetation canopy. From here, the water may drain to the soil surface or be lost in evaporation. The evaporated fraction is termed the interception loss. The water that arrives at the soil surface may infiltrate, be held in the surface depression store or runoff. The infiltration rate is determined by the soil hydraulic properties and the current soil moisture status. If the water is not able to infiltrate, then it may be held in the surface depression store. This is the water that can be held in the surface roughness without running off. If the surface depression store is full, then runoff will occur. Runoff from an upslope cell is termed run on. The water that has infiltrated will move into the main soil matrix and then either enter the groundwater as recharge or move laterally in the soil as through flow. Water can also be evaporated from the surface depression store or from the soil as well as from the canopy as evapotranspiration.

The process of rainfall interception considers the leaf canopy to be a non-leaking store. The rainfall is divided into direct throughfall and intercepted water. The intercepted fraction fills the canopy store until it overflows. The water is then able to leave the canopy store as evaporation. This approach has been applied in the Pattern^{lite} model (Mulligan and Reaney 2000) and the CASC2D model (Johnson *et al.* 2000). The size of the canopy store and the percentage intercepted relates to the vegetation type and amount of biomass.

The depth of the surface depression store can be determined from the surface gradient and roughness. Kirkby *et al.* (2002) suggest:

$$\frac{dp}{\alpha} = 0.11\exp\left(\frac{-0.02\beta}{\alpha}\right)$$

[4]

where *dp* is the surface depression storage (mm), *α* is the surface roughness and *β* is the slope gradient (degrees). The value for *α* can be related to the random roughness coefficient (RR) (Ilmaras *et al.* 1966) by:

$$RR = 0.657\alpha$$

[5]

In this application, the Priesley-Taylor equation ([7]) is used to calculate evapotranspiration. The Penman-Monteith combination equation is probably the best estimator of potential evapotranspiration for all types of vegetated surfaces (Dingman, 1994). However, the equation has large data demands requiring information on temperature, solar radiation, wind speed, relative humidity and vegetation characteristics. The complete dataset may not be available for all locations. In these situations, the Priestley-Taylor equation is more suitable since it only requires information on net radiation, air temperature and pressure. This equation assumes that air moving over large distances over a homogeneous and well watered surface would become saturated. Therefore the mass-transfer term in the Penman- Monteith equation can be removed. The evapotranspiration under these conditions is termed the equilibrium potential evapotranspiration (PET_{EQ}).

$$PET_{PT} = \frac{\alpha_{pt}\Delta(R_n - G)}{\Delta\gamma}$$

[6]

where: PET_{PT} is the potential daily evapotranspiration according to the Priesley – Taylor equation; R_n is the net radiation; G is the soil heat flux; Δ is the slope of the saturation vapour pressure temperature relationship; γ is the psychrometric constant; and α_{PT} is the Priesley – Taylor constant location parameter (under normal conditions, $\alpha_{PT} = 1.26$).

The amount of solar radiation received at the surface is determined by the amount of energy arriving at the top of the atmosphere, the transmission of the energy through the atmosphere and the reflection of energy by the surface cover. The amount of energy arriving at the top of the atmosphere is determined by a set of equations that use the Earth - Sun geometry, the position on the Earth and the date and time of day. Full details of these equations can be found in Dingman (1994).

The amount of energy that reaches the Earth's surface is less than the amount received at the top of the atmosphere due to scattering. The amount of scattering is dependent on the depth of the atmosphere and the local weather conditions, the most important being cloud cover. Within the model, the reduction in energy receipt though a cloud free atmosphere is determined by [8]:

$$R_{ES} = R_{TA}.0.5$$

[7]

where R_{ES} is the amount of solar radiation reaching the Earth's surface and R_{TA} is the amount of solar radiation at the top of the atmosphere. This value is reduced by 50% on days determined to have cloud cover. All rain days are classified as cloudy and the cloud cover on non-rain days is determined using a Monte Carlo model parameterised from observed data.

Once the solar radiation reaches the Earth surface, it can be either directly reflected or reflected as long wave radiation. The amount of reflected radiation is denoted by the albedo of the surface. The amount of reflected radiation is given by.

$$r_{sw} = R_{ES} * a$$

[8]

where r_{sw} is the reflected short wave radiation and a is the surface albedo.

The amount of incoming solar energy that is reflected as long wave radiation is determined by both the surface emissivity and the temperature:

$$r_{lw} = e_{ms}.\left(5.6696 * 10^{-8}\right).\left(T_a + 273.15\right)^4$$

[9]

where r_{lw} is the reflected long wave radiation, e_{ms} is the surface emissivity and T_a is the air temperature (℃). Once these reflected amounts have been subtracted from the total solar radiation reaching the surface, the remainder is capable of powering the evapotranspiration process.

The determination of the amount of actual evapotranspiration from the potential amount predicted by the Priesley-Taylor equation is complex. To approach this problem, we divide the predicted potential evapotranspiration into three key parts: (1) the direct evaporation of standing water from exposed surfaces; (2) the transpiration of water through vegetation; and (3) the evaporation of water from the soil surface that has travelled through the soil matrix. Water is evaporated in the following order: (1) water on the vegetation; (2) transpiration; (3) water on the soil surface; and (4) water in the soil. The evaporation of standing water either from the vegetation canopy or from the soil surface occurs at the potential rate. The amount of potential transpiration (t_p), is related to the leaf area index (LAI) of the vegetation (Don Scott 2000, p. 278):

$$t_p = PET_{PT} * \left(-0.21 + 0.7^{LAI}\right)$$

[10]

The amount of actual transpiration is related to the rooting depth of the vegetation and the availability of the water within the dynamic layer and the main soil store. The rate of water extraction from either store is limited by the moisture retention characteristics of the soil, described below. The depth to which water can be extracted is limited by the rooting depth. When the water table falls below this depth, transpiration ceases.

The amount of water that can be evaporated from the soil by travelling through the soil matrix is limited by the retention characteristics of the soil. This relationship has been linearised and takes into account the increased tension at low soil moisture levels:

$$e_{\theta} = PET_{PT}\theta$$

[11]

where e_{θ} is the soil moisture dependent evaporation rate, e is the potential evaporation rate and θ is the soil moisture (m^3 water / m^3 pore space). To aid the fitting of the predicted evapotranspiration to the observed values, e may be adjusted by an empirical correction factor.

Infiltration is based upon a soil moisture storage based simplification of the Green and Ampt (1911) equation developed by Kirkby (1975; 1985):

$$i_t = a + \frac{b}{\theta}$$

[12]

where a and b are coefficients and θ is the soil moisture. The use of the storage term in [12] gives two key advantages: (1) it enables the equation to model the infiltration rate over areas rather than points, which is of great help in scaling the model since the storage term can relate to any area such as $1cm^2$ or $1km^2$ (Beven, 2000); and (2) the use of a changing soil moisture term, rather than a time based model, allows the equation to be applied to an irregular rainfall time series.

Overland flow can be generated in one of three ways within the model: (1) as infiltration excess (Hortonian); (2) as saturated; or (3) as return overland flow. Infiltration excess overland flow occurs when the rainfall rate is greater than the current infiltration capacity of the soil, as defined by [12]. In this case, the rainfall will infiltrate at the maximum rate and the excess water will become overland flow. Under saturated soil conditions, no more rainfall is able to infiltrate into the soil and hence all of the rainfall is converted to overland flow. Return flow occurs when the through flow into a cell exceeds the storage capacity of the cell and hence the excess water overflows out of the top of the cell.

The soil depth exerts a major control on the hydrological operation of a point in the landscape. There is a relationship between geomorphological form / position and the soil depth (Huggett and Cheesman 2002). Following this approach, the surface topography is classified into ridges, slopes, channels and plane areas. For each of these geomorphological forms, a soil depth is assigned within a logical structure:

$$\text{Channels} > \text{Plane} > \text{Ridges} > \text{Slopes} \quad [13]$$

The rate of recharge is determined by the minimum of the hydraulic conductivity at the base of the soil profile and the hydraulic conductivity of the bedrock.

Landscape scale processes

Each model cell generates and receives lateral fluxes in the form of runoff and through flow. The model holds the spatial information within a raster grid structure. This is due to the availability of datasets in this format and the efficiency of the structure both in terms of computational expense and storage.

The amount of through flow is determined by Darcy's Law in the saturated zone. The lateral flow in the unsaturated zone is considered to be insignificant and hence is not included.

$$tf_v = wt.y.K_d \frac{dh}{dx}$$

[14]

where th_v is the through flow volume per second (m^3s^{-1}), *wt* is the height of the water table above the bedrock (m), *y* is the width of the routing facet (m), K_d is the soil conductivity at the water table depth (ms^{-1}) ([15]), *h* is the hydraulic head (m) and *x* is the horizontal distance (m) between model cells. The soil conductivity is defined by:

$$K_d = K_{sat} \exp\left(\frac{-d}{dc}\right)$$

[15]

where K_{sat} is the soil saturated conductivity, *d* is the water table depth and *dc* is the decay factor for the change in conductivity with depth.

Overland flow may consist of a combination of laminar, transitional and turbulent flows (brahams *et al.* 1986). In these situations, the Darcy-Weisbach equation is the most appropriate since it is able to describe all of these flow conditions (Baird 1997):

$$v = \sqrt{\frac{8gRs}{ff}}$$

[16]

where *ff* is the friction factor, *g* is the gravity constant, *R* is the hydraulic radius and *s* is the slope of the energy gradient. Following Abrahams *et al.* (1992), the friction factor, *ff*, is related to the surface cover rather than the flow hydraulics.

The horizontal flux routing is performed by the FD8 algorithm (Quinn *et al.* 1991). This algorithm uses multiple flow directions and hence allows both the dispersion and concentration of flow. On hillslopes, flow is distributed to all of the lower neighbouring cells. The amount of flow assigned to each cell is determined on a slope-weighted basis as proposed by Quinn *et al.* (1991) and Freeman (1991):

$$F_i = \frac{\beta_i^v}{\sum_{i=l}^{8} \beta_i^v}$$

[17]

where β_i is the slope from the central node to neighbour *i* and *v* is a positive constant. The *v* constant is a flow concentration factor, the greater the value of *v*, the greater the flow concentration (Holmgren 1994). Holmgren (1994) recommends values of *v* = 4 - 6 for distributed modelling.

River channel network

Movement of water through the river channel network is modelled using the Muskingham-Cunge model (Ponce and Lugo 2001). Each river reach is associated with a landscape cell and receives inflow from overland flow and soil through flow. A reach may also be connected to an abstraction or discharge to other channels in the network. The outflow from a river reach is determined by:

$$Q = (C_0.U) + (C_1.U_1) + (C_2.Q_1)$$

[18]

where *Q* is the current discharge, Q_1 is the discharge from the previous time step, *U* is the inflow from the upstream reach, U_1 is the inflow from the upstream reach from the previous time step and C_0, C_1 and C_2 are routing coefficients:

$$C_0 = \frac{-1 + C + D}{1 + C + D}$$

$$C_1 = \frac{1 + C - D}{1 + C + D}$$

$$C_2 = \frac{1 - C + D}{1 + C + D}$$

[19]

where *C* is the Courant number and D is the cell Reynolds number. The Courant number is the ratio of the flood wave celerity, *c* ([20]), to the grid celerity ([21]). The cell Reynolds numbers is the ratio of the hydraulic diffusivity to the grid diffusivity (22):

$$c = mQ^k$$

[20]

where *m* and *k* are coefficients.

$$C = c\frac{\Delta t}{\Delta x}$$

[21]

where t is time and x is space.

$$D = \frac{q}{Sc\Delta x}$$

[22]

where q is the reference unit-width discharge and S is the channel slope.

Appendix References

Abrahams A. D., Parsons A. J., and Hirsch P. J.1992: Field and laboratory studies of resistance to interrill overland flow on semi-arid hillslopes, southern Arizona; in (eds.) Parsons, A. J. and Abrahams, A. D, UCL Press, London, 438

Abrahams A. D., Parsons A. J., and Luk S. H. 1986: Resistance to overland flow on desert hillslopes; Journal of Hydrology, 88, 343 - 363

Allmaras R., Burwell R., Larson W., and Holt R. 1966: Total porosity and random roughness of the interrow zone as influenced by tillage; United States Department of Agriculture Conservation Research Report, 7 - 22

Baird A. J. 1997: Overland flow generation and sediment mobilisation by water; in (eds.) Thomas, D. S. G., Arid zone geomorphology, John Wiley & Sons, Ltd, Chichester, 165 - 184

Beven K. 2000: Rainfall-Runoff Modelling: The Primer; John Wiley and Sons, Chichester, 360

Dingman S. L. 1994: Physical Hydrology; Prentice Hall, New Jersey, 575

Don Scott H., 2000: Soil Physics: Agriculture and Environmental Applications; Iowa State University Press

Freeman T. 1991: Calculating catchment area with divergent flow based on a regular grid; Computers and Geoscience, 17, 413 - 422

Green W. and Ampt G. 1911: Studies in soil physics. Part I. - The flow of air and water through soils; Journal of Agricultural Science, 4, 1 - 24

Holmgren P. 1994: Multiple flow direction algorithms for runoff modelling in grid based elevation models: An empirical evaluation; Hydrological Processes, 8, 327 - 334

Huggett R. and Cheesman J. 2002: Soil and sediments, *Topography and the environment*, Prentice Hall, Harlow, England, 122 - 161

Reed D W, Faulkner D., Robson A, Houghton-Carr H and Bayliss A. 2002: Flood Estimation Handbook; Centre for Ecology & Hydrology

Johnson B., Julien P., Molnar D., and Watson C. 2000: The two-dimensional Upland erosion model CASC2D-SED; Journal of The American Water Resources

Association, 36, 31 - 42

Kirkby M. 1975: Hydrograph modelling strategies; in (eds.) Peel, R, Chisholm, M, and Hagget, P, Progress in Human and Physical Geography, Heinemann, London, 69 - 90

Kirkby M. 1985: Hillslope hydrology; in (eds.) Anderson, MG. and Burt, TP, Hydrological Forcasting, John Wiley and sons, Chichester, 3 - 37

Kirkby M., Bracken L., and Reaney S. 2002: The influence of land use, soils and topography on the delivery of hillslope runoff to channels in SE Spain; Earth Surface Processes and Landforms, 27, 1459 - 1473

Mulligan M. 1996: Modelling the complexity of land surface response to climate variability in Mediterranean environments; in (eds.) Anderson, M. and Brooks, S., John Wiley and Sons, Chichester, 1099 - 1149

Mulligan M. and Reaney S. 2000: PatternLite, A policy relevant version of Pattern for Modulus; in (eds.) Engelen, G, van der Meulen, M, Hahn, B, Uljee, I, Mulligan, M, Reaney, S, Oxley, T, Blatsou, C, Mata-Porras, M, Kahrimanis, S, Giannoulopoulos, P, Mazzoleni, S, Coppola, A, Winder, N, van der Leeuw, S, and McIntosh, B. S., MODULUS: A Spatial Modelling Tool for Integrated Environmental Decision Making, Final Report, The Modulus Project, EU-DGXII Environment (IV) Framework, Climatology and Natural Hazards Programme, Contract ENV4-CT97-0685, July 2000, 145 - 202

Ponce V. M. and Lugo A. 2001: Modeling looped ratings in Muskingum-Cunge routing; Journal of Hydrologic Engineering, 6, 119 – 124

Quinn P., Beven K., Chevallier P., and Planchon O. 1991: The prediction of hillslope flow paths for distributed hydrological modelling using digital terrain models; Hydrological Processes, 5, 59 - 79