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**Paper:**
Ballard, CE, McIntyre, N, Wheater, HS, Holden, J and Wallage, ZE (2011) *Hydrological modelling of drained blanket peatland.* Journal of Hydrology, 407 (1-4). 81 - 93. ISSN 0022-1694

http://dx.doi.org/10.1016/j.jhydrol.2011.07.005

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# Hydrological modelling of drained blanket peatland

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## Abstract

Open ditch drainage is a commonly implemented land management practice in upland blanket peatlands, particularly in the UK, where policy decisions between the 1940s and 1970s lead to widespread drainage of the uplands. The change in the hydrological regime associated with the drainage of blanket peat is poorly understood, yet has perceived importance for flooding, low flows and water quality. We propose a new simplified physics-based model that allows the exploration of the associated hydrological processes and flow responses. The model couples four one-dimensional models to represent a three-dimensional hillslope, allowing for the exploration of flow and water table response throughout the model domain for a range of drainage configurations and peat properties. The model is tested against a data set collected from Oughtershaw Beck, UK, with results showing good model performance for wet periods although less compatibility with borehole observations during rewetting periods. A wider exploration of the model behaviour indicates that the model is consistent with the drained blanket peat hydrological response reported in the literature for a number of sites, and therefore has potential to provide guidance to decision makers concerning the effects of management practices. Through a global sensitivity analysis, we conclude that further field investigations to assist in the surface and drain roughness parameterisation would help reduce the uncertainty in the model predictions.

Keywords: peatlands, runoff, water table, ditches, drainage, modelling

## 1. Introduction

Peatlands are located across the globe, from the tropics to the high latitudes, covering approximately 3% of the Earth's surface. These environments are of particular nature conservation value due to their unique and diverse biodiversity. Moreover, they store soil carbon and water; it is estimated that 10% of the world's freshwater resources and up to one-third of global soil carbon are stored in peatlands (Rubec, 2005).

In the UK, 87 % of peat covered areas take the form of blanket peat. The UK uplands include approximately 2.9 M ha of blanket peatland (Holden et al., 2004), constituting approximately 15% of the global amount of blanket peatland (Milne and Brown, 1997). These regions have traditionally been heavily managed for low density farming, energy, forestry and game rearing. In recent times, in recognition of the significant ecosystem services provided by peatlands (including biodiversity, carbon sequestration, water supply, and recreation (Bonn et al., 2009)), many areas of upland blanket peatland have been designated as 'Sites of Special Scientific Interest', 'Areas of Outstanding Natural Beauty', 'Special Protection Areas', 'Environmentally Sensitive Areas', 'Special Areas of Conservation' and National Nature Reserves (Condliffe, 2009). Peatlands also generate a large proportion of the UK water supply; therefore water quality and colour are also significant considerations (Armstrong et al., 2010). The management of these areas is thus of interest to a range of stakeholders, including physical and social scientists, and land owners and managers. Given the inherent significance of the upper areas of catchments for downstream flooding, with their higher rainfall rates and generally flashier response (Wheater et al., 2008), the management of upland blanket peatland also has the potential to affect flood risk.

In the UK, approximately 50% of upland blanket peatland has been drained (Milne and Brown, 1997), which in England alone amounts to 75,000 ha. Drainage of peats is typically via a series of open ditches, with the aim of improving vegetation and therefore the production of livestock and game (Stewart and Lance, 1983). The rationale is that drainage will remove excess surface runoff and lower the water table, thereby creating more conducive environments for plant species suitable for stock grazing. However, these predicted benefits have rarely been realised since blanket peat cannot sustain anything above very small sheep populations without undergoing severe degradation (Stewart and Lance, 1983); and peatland drainage is also generally considered to have adverse effects on the natural environment.

The implementation of an open ditch drainage scheme causes two key processes to occur: (1) water is drained from the soil matrix directly into the open drains, lowering water tables and creating more soil storage capacity, and (2) surface runoff and direct rainfall are captured in the drains and transmitted to the catchment outlet more rapidly (Holden, 2009b). How peatland drainage affects flooding in the catchment depends on the interaction of these two counteracting processes, where process (1) tends to reduce storm peaks and increase base flows, while process (2) tends to increase the flashiness of the response and increase peak flows. The dominance of each process is likely to depend on a number of factors including: drainage density and geometry, hydraulic conductivity, drain and peat surface roughnesses, topography, event size, and antecedent conditions.

Observations have shown that drained peatlands typically have a shorter time to peak, higher peak flow and a quicker recession than undrained areas, and are associated with increased water table fluctuations (Ahti, 1980; Conway and Millar, 1960; Holden et al., 2006; Robinson, 1986; Stewart and Lance, 1991). The zone of influence of the water table drawdown (i.e. process (1) above) due to the drains is quite limited in blanket peats, due to very low hydraulic conductivities, particularly in the deeper layers (Robinson, 1986; Stewart and Lance, 1983), therefore the spacing of the drains plays a significant role in both the short and long term effects of drainage. The effect is also not uniform in space, particularly in blanket peat on sloping terrains, since the reduction in upslope contributing area is most significant immediately downslope of a drain (Coulson et al., 1990; Holden et al., 2006). In a minority of field studies, drainage of peatlands has been observed to decrease flood peaks (e.g. Burke, 1967; Coulson et al., 1990; Newson and Robinson, 1983). However, in all of these cases some aspects of the site conditions are at the extreme of the range of conditions that are typically encountered in drained peatlands. For example, Burke (1967) studied a site with especially dense (3.5m) drain spacing and drains running along contours, thus minimising drain flow velocity; Newson and Robinson (1983) examined a peaty soil with a hydraulic conductivity higher than typical blanket peats; and Coulson (1990) studied a site with lower altitude and lower annual rainfall than typical for UK peatlands.

Given the uncertainty about best management practices for peatlands, due largely to the complexity of process interactions, there is a need for suitable process-based models to aid understanding of impacts of management interventions. Due to the difficulties in observing and quantifying land management effects at the catchment scale, the simplest case to consider is the response at field or hillslope scales. Some modelling has been performed to examine drainage in the uplands, notably hillslope simulations using SHETRAN (Dunn and Mackay, 1996) and a modified TOPMODEL (Lane, 2002; Lane et al., 2004; Lane et al., 2003). However, while the TOPMODEL simulations can explicitly represent drainage networks, the conceptual nature of the model does not provide detail about the subsurface behaviour, particularly at the sub-drain spacing scale. The conceptual stores in the model that represent saturated subsurface storage are independent of each other, hence, although the topographic index concept partly accommodates the way a cell may be affected by upslope areas, downslope effects, such as the presence of drains and the level of water in them, cannot be simulated. The studies have also not been compared against any observed flow or water table time series. The SHETRAN simulations of Dunn and Mackay (1996) are physically based, but the drainage configurations were limited to alignments with the grid boundaries (due to their representation as 'channel elements'), thereby providing a limited range of potential drain configurations. Also, the inter-drain regions were represented by a single grid cell, not allowing examination of the local changes in water table heights. Hence, although both approaches had some success at the large scale, in order to explore and examine the hydrological processes associated with the management of drained peatlands there is a need for a physics-based model that provides flexibility in the representation of drainage configurations and can provide information about the spatial variability of the internal model states.

In this paper, a new fine resolution simplified physics-based model is proposed to test hypotheses about hydrological processes and to investigate the effects of peatland land management. The new model aims to allow the impacts of management scenarios to be explored, as an extension to the limited experimental data currently available, and as a complement to any future extensive experimental programmes. The model is tested against flow and water table data from a drained peatland site in the UK. The results from the analysis are used to explore the model performance and to identify processes that require refinement and the data that would reduce the uncertainties in the model predictions. Finally, the wider applicability of the model is assessed.

# 2. Model description

The modelling approach used in this study was to identify the key hydrological processes for intact and drained peatlands from the literature and include them in a model that has an appropriate level of complexity relative to the level of information available on the system hydrological processes. To avoid over-parameterisation, minor processes have been excluded or treated in a simplified manner. In particular, the development has focused primarily on ombrotrophic (rain water fed) blanket peatlands in the UK, where deep groundwater flows are expected to be negligible, and on representing processes known to influence flood generation.

## 2.1. Conceptual model

Blanket peat deposits are typically found draped over gently rolling terrain in areas with a cool climate, high rainfall and impeded substrate drainage. These conditions allow peat formation, which occurs when organic material decomposes slowly due to anaerobic conditions associated with waterlogging (Allaby, 2008). Typically, peats exhibit two major zones: the upper layer (acrotelm), which is composed of live and decaying plant material and can range from 5 to 50 cm thick, and a lower zone (catotelm), which is denser, usually saturated and anoxic (Evans et al., 1999; Holden and Burt, 2003b; Ingram 1978, 1983).

Water tables in blanket peat catchments are generally observed to fluctuate between the top of the catotelm and the ground surface and are highly responsive to changes in the soil water balance (Evans et al., 1999). Due to these typically high water tables, soil water storage does not contribute significantly to the attenuation of winter floods and surface runoff in peat catchments is generally observed to be due to saturation excess (Holden and Burt, 2002a). These local scale responses lead to very flashy responses at the catchment scale (Bragg, 2002; Evans et al., 1999; Holden and Burt, 2003c; Holden et al., 2001) and generally low base flows.

Although soil storage may not be a major factor in attenuating flows, micro-relief elements (Kellner and Halldin, 2002) and land cover (Weiss, 1998) both significantly affect runoff as they can provide local storage and increase the effective roughness of the surface. Additionally, pipes are also often observed in blanket peats and can couple the shallow acrotelm with the deeper catotelm, contributing between 10-30% of the total flow (Holden and Burt, 2002b), although their relative contribution to runoff is lower under saturated conditions, due to the dominance of overland flow processes (Holden, 2005).

The saturated hydraulic conductivity ($K_S$) of peat soil is observed to reduce with depth (Clymo, 2004; Holden et al., 2001; Surridge et al., 2005; Van Seters and Price, 2002), with decreases of as much as five orders of magnitude by a depth of 0.4 to 0.8m (Bradley, 1996). In the catotelm, high compaction and greater humification of the material leads to a greater bulk density (Holden and Burt, 2002a) and a reduction of the voids in the substrate, thereby reducing the saturated hydraulic conductivity. In contrast, macropores due to voids created from the decaying plant material are particularly important in the acrotelm and contribute significantly to the higher hydraulic conductivity of this layer (Holden, 2009a). Shallow throughflow along the boundary of the acrotelm and catotelm is a significant flow mechanism due to the discontinuity of hydraulic conductivities between the two layers (Holden and Burt, 2003a).

Based on literature, a simplified conceptualisation of the hydrological functioning of drained blanket peatlands has been developed (Figure 1), consisting of three main hydrological components: soil blocks, drains between soil blocks, and a collector drain. This is used as the basis of a mathematical model to represent a three-dimensional drained blanket peat hillslope. The parameters in Figure 1, as well as others that are introduced in the following sections, are defined in the Appendix.

## 2.2. Mathematical model

Given that the water table in undrained blanket peats is always observed to be close to the surface, and the high degree of uncertainty in the parameterisation of a full Richards' equation-based model with limited or no unsaturated data, a physics-based unsaturated zone representation has been excluded from the model. This has the added benefit of significantly reducing computational time (Pancioni et al, 2003). By removing the unsaturated zone, it is assumed that exchanges between the subsurface and the surface (i.e. evaporation and infiltration) occur instantaneously; and it is assumed that subsurface lateral fluxes can be described using Darcy's law:

$$V = -\overline{K}_S(h_S)\frac{\partial h_T}{\partial x} \tag{1}$$

where $V$ is velocity in the downslope direction ($LT^{-1}$), $\overline{K}_S$ ($LT^{-1}$) is the depth averaged saturated hydraulic conductivity, $h_S$ is the depth of the water table above the impermeable bed (L), $h_T$ is the total hydraulic head (L) and $x$ is the downslope subsurface ordinate (L). The continuity equation is defined as:

$$\overline{\varepsilon}(h_S)\frac{\partial h_T}{\partial t} = -\frac{\partial q_S}{\partial x} + i - ET_p \tag{2}$$

where $\overline{\varepsilon}$ ($L^3L^{-3}$) is the depth averaged effective porosity, $q_S$ is the unit width subsurface flux ($L^2T^{-1}$) and $i$ and $ET_p$ ($LT^{-1}$) are exchange terms representing the fluxes across the peat surface due to infiltration and evaporation respectively. Decomposing the total hydraulic head ($h_T$) into a fixed component related to the slope and a variable component, $h_S$, and making the Dupuit-Forcheimer approximation that the flow lines are always parallel to the slope, then:

$$\frac{\partial h_T}{\partial x} = \frac{\partial h_S}{\partial x} - \tan\theta_S \tag{3}$$

$$q_S(x,t,x_d) = -\overline{K}_S(h_S)h_S\left(\frac{\partial h_s}{\partial x} - \tan\theta_S\right) \tag{4}$$

where $\theta_S$ is the site slope. Substituting equation 4 into equation 2 gives the Boussinesq equation:

$$\overline{\varepsilon}(h_S)\frac{\partial h_S}{\partial t} = \frac{\partial}{\partial x}\left(\overline{K}_S(h_S)h_S\frac{\partial h_S}{\partial x}\right) - \tan\theta_S\frac{\partial\left(\overline{K}_S(h_S)h_S\right)}{\partial x} + i - ET_p \tag{5}$$

The more commonly used version of the Boussinesq equation (e.g. Beven, 1981; Childs, 1971; Henderson and Wooding, 1964; Verhoest et al., 2002) can be derived from equation (5) by assuming constant $K_S$ and using a transformation of $h_s' = h_s/\cos(\theta_S)$ and $x' = x/\cos(\theta_S)$ (where $x'$ and $h_s'$ are the rotated distance and water table height measures). The gravitational frame of reference was chosen here to assist in the coupling between the subsurface, overland and drain flows and also allows the drain walls to remain vertical. Importantly, the Boussinesq equation still retains dependence on the downslope boundary condition, which will be significant once the model is adapted to represent blocked drains. In their comparison study of the performance of the Boussinesq equations compared with a full Richards' equation representation, Pancioni et al. (2003) concluded that the Boussinesq equation was able to successfully capture the shapes of the storage and outflow profiles, particularly for low air-entry pressure soils under draining conditions. Given the typically low air entry pressure of the acrotelm (Letts et al, 2000), the benefits of reduced parameterisation for the Boussinesq equation are likely to outweigh the performance benefits of a Richards' equation representation.

The acrotelm-catotelm layering is represented in the model through depth-averaged saturated hydraulic conductivity $\overline{K}_S$ and depth-averaged porosity $\overline{\varepsilon}$ defined by

$$\overline{K}_S(h_S) = \frac{\int_{z=0}^{z=h_S} K_s(z)}{h_s} dz \qquad (6)$$

and

$$\overline{\varepsilon}(h_S) = \frac{\int_{z=0}^{z=h_S} \varepsilon(z)}{h_s} dz \qquad (7)$$

This depth-averaging provides an approximation of the dual layer system for application in the one-dimensional Boussinesq equation. As the model solution is sensitive to discontinuities in the hydraulic conductivity a smoothing function is used to describe the variation of hydraulic conductivity with depth:

$$K_S(z) = K_{Sc} + (K_{Sa} - K_{Sc})(1 + \tanh[(z - d_c)100])/2 \qquad (8)$$

where $z$ is the coordinate measured vertically from the impermeable lower boundary, $K_{Sc}$ is the saturated hydraulic conductivity of the catotelm, $K_{Sa}$ is the saturated hydraulic conductivity of the acrotelm, and $d_c$

is the thickness of the catotelm. A step function is used to describe the variation of effective porosity with depth:

$$\begin{aligned}\varepsilon(z(z \le d_c)) &= \varepsilon_c \\ \varepsilon(z(z > d_c)) &= \varepsilon_a\end{aligned} \qquad (9)$$

where $\varepsilon_c$ is the effective porosity of the catotelm and $\varepsilon_a$ is the effective porosity of the acrotelm. Examples of these relationships are shown in Figure 2.

Natural soil pipes have not been explicitly represented in the model, as the data required to parameterise a pipe model are unlikely to be available for typical model applications; pipe flow contributions are assumed to be accounted for in the acrotelm hydraulic conductivity. As hydraulic conductivities are known to be very low at depth (Letts et al., 2000), a zero-flux boundary is imposed at the depth of the drains. Fluxes into the peat, represented by $i$ in Equation 5, are firstly from any surface water (reinfiltration at a maximum rate equal to the saturated hydraulic conductivity), and then directly from rainfall (snow is not explicitly represented). The rainfall infiltration rate is set at the smaller of the rainfall rate or the saturated hydraulic conductivity of the upper layer. When the soil is saturated no infiltration is allowed, due to the no-flux condition at the lower boundary. Infiltration and saturation excesses are added to the overland flow. For the peat blocks, water is firstly evaporated at the potential rate from any surface water ($ET_{OF}$), and then from the acrotelm ($ET_p$). Soil evaporation is assumed to cease when the water table is below the acrotelm. Evaporation from the drains and collector drain ($ET_d$ and $ET_c$) occurs at the potential rate while water is present.

Overland and channel flows are represented by the kinematic wave equation, an approximation of the Saint Venant equations of gradually varied unsteady flow commonly used for representing surface flow dynamics (Singh, 1996). The approximation neglects the acceleration and pressure terms in the full equations, replacing the momentum equation with a steady state depth-discharge relationship. The general form of the kinematic wave equation is:

$$\frac{\partial H}{\partial t} = -\frac{\partial Q}{\partial y} - Sink \qquad (10)$$

where H is the flow depth (L), Q is the unit width flux ($L^2T^{-1}$), *Sink* represents sink and source terms for the channel ($LT^{-1}$) and $y$ is a distance ordinate (L). When applied to the drain, collector drain and overland flow, the following three equations are generated:

Drain:

$$\frac{\partial h_d}{\partial t} = -\frac{\partial q_d}{\partial x_d} + \left(Rain - ET_d\right) + \frac{\partial}{\partial x_d}\left(\left(q_S\left(end,t,x_d\right) - q_S\left(1,t,x_d\right) + q_{OF}\left(end,t,x_d\right)\right)\frac{\Delta x_d}{W_d}\cos\beta_d\right) \quad (11a)$$

Collector drain:

$$\frac{\partial h_c}{\partial t} = -\frac{\partial q_c}{\partial x_c} + \left(Rain - ET_d\right) + \frac{\partial\left(q_d\left(end,t\right)\right)}{\partial x_c} \quad (11b)$$

Overland flow:

$$\frac{\partial h_{OF}}{\partial t} = -\frac{\partial q_{OF}}{\partial x} + \left(Rain - ET_{OF} - i\right) \quad (11c)$$

where $h_d$ is the drain flow depth (L), $h_c$ is the collector drain flow depth (L), $h_{OF}$ is the overland flow depth (L), *Rain* is the unit area rainfall ($LT^{-1}$), $q_{OF}$ is the unit width overland flow ($L^2T^{-1}$), $q_d$ is the unit width drain flow ($L^2T^{-1}$), $W_d$ is the width of the drain (L), $x_d$ is the drain ordinate (L) and $x_c$ is the collector drain ordinate (L).

The depth-discharge relationship for the drains was represented by the Manning equation, as friction factors quoted in the literature are more commonly Manning's roughness coefficient values. The depth-discharge relationship for the drains is:

$$q_d\left(x_d,t\right) = \frac{h_d\left(\frac{W_d h_d}{\left(W_d + 2h_d\right)}\right)^{2/3}\sqrt{\tan\left(\theta_d\right)}}{n} \quad (12)$$

where $n$ is the Manning's roughness coefficient and $\theta_d$ is the slope of the drain, where $\theta_d = \sin^{-1}\left(\sin\theta_S \sin\beta_d\right)$. For the collector drain:

$$q_c\left(x_c,t\right) = \frac{h_c\left(\frac{W_d h_c}{\left(W_d + 2h_c\right)}\right)^{2/3}\sqrt{\tan\left(\theta_S\right)}}{n} \quad (13)$$

In order to utilise data in Holden et al. (2008) for depth varying overland flow friction factors, the overland flow depth-discharge relationship is calculated using a Darcy-Weisbach equation, given by:

$$q_{OF}(x,t,x_d) = h_{OF}\sqrt{\frac{8g\tan(\theta)h_{OF}}{f}} \tag{14}$$

where g is the acceleration due to gravity (9.81 ms$^{-2}$) and *f* is the Darcy-Weisbach friction factor. Holden et al. (2008) investigated values of *f* for four different land cover types, which in order of increasing roughness were: bare (Ba), *Eriophorum* (E), *Eriophorum/Sphagnum* mix (E/S) and *Sphagnum* and *Juncus* (S/J). *f* was also found to vary with overland flow depth. The mathematical relationship between *f* and depth proposed by Holden et al. (2008) has a discontinuity in the relationship at approximately 1cm and *f* tends to infinity as the overland flow depth tends towards zero. Both of these properties cause numerical difficulties when introduced into the continuous simulations. Therefore a continuous polynomial that passes though the origin was identified for each land cover type with the general form $f(h_{OF}) = ah_{OF} + bh_{OF}^2$, where parameters *a* and *b* were optimised to closely recreate the original data. Parameter *b* could be described as a function of *a*; the final relationship used is $f(h_{OF}) = ah_{OF} - (2.21a+3.82)h_{OF}^2$, with values of *a*: (Ba) 20.79, (E) 5.05, (E/S) 3.48 and (S/J) 1.90. Parameter *a* was used as a proxy for *f* for the purpose of model calibration and sensitivity analysis.

The resulting model couples four one-dimensional models that represent respectively subsurface, overland flow, drain flow and collector drain flow (Figure 1). The one-dimensional models are run simultaneously with feedbacks between the subsurface and surface through the infiltration and evaporation terms, and between the subsurface and the drains through the drain depth and seepage face water level. By limiting the model to four one-dimensional models rather than a fully integrated three-dimensional model, some of the computational demands of the modelling procedure are reduced (assuming that fewer nodes and fewer equations will lead to reduced computational time) and the parameterisation of the model can be limited to those parameters for which information can be taken from the literature (such as the saturated hydraulic conductivity and surface roughness). The model uses inputs of rainfall and potential evaporation, and outputs flow and water depths throughout the model domain.

The model space is discretised into a number of soil blocks (Figure 1), which are bounded upstream and downstream by drains. The block lengths tested ranged from 5m (close drain spacing) to 500m (to simulate intact peatland). The model space domain may include a large number of blocks, depending on the application. Although the drains may be at any angle relative to the contours, the soil blocks are always aligned downslope, meaning that surface and subsurface flow in the block are always perpendicular to the contours and parallel to the edge of the block. In this way, flow paths on and in the soil block may be represented by the single dimension and there is no exchange flow between the soil blocks. This representation neglects any cross-slope flow paths that may be present.

The partial differential equations to describe the variation of flow depths with time for each of the one-dimensional models were discretised in space using finite differences. The resulting ordinary differential equations were then integrated in time using Matlab's ODE15s solver (Shampine and Reichelt, 1997; Shampine et al., 1999). The solver uses an adaptive time grid, which limits the numerical error associated with each time step to within a user defined tolerance. For the soil block and overland flow calculations, nodes in the x-direction are in a $\log_{10}$ space, allowing nodes to be more closely spaced toward the boundaries. By using a varying x-spacing, computational efficiency can be increased, by focusing nodes in the regions of rapidly varying flows.

# 3. Model calibration and testing

## 3.1. Case study application

Oughtershaw Beck, a tributary of the River Wharfe, is a catchment of approximately 13.8 $km^2$ (Lane et al., 2004) located at 54°13'54" N, 2°15'09"W, in the Yorkshire Dales, Northern England (Figure 3). The average annual rainfall is 1850mm (Wallage et al., 2006). The catchment ranges in elevation from 353m at the outlet to 640m, and is primarily blanket peat with an average thickness of 2m. The catchment is underlain by carboniferous limestone and millstone grit that is covered with a glacial boulder clay deposit (Wallage et al., 2006). Open cut drainage was installed over a large portion of the catchment in the 1960s. There was no maintenance of the drains in the intervening period, but Holden et al. (2007) surveyed the drains in the area, finding that most had either remained the same dimensions as when cut or had eroded; there were very few that had naturally infilled or become vegetated.

A monitoring programme ran from December 2002 until August 2004, consisting of 6 boreholes in a transect across a drained peatland site within the catchment (Figure 4a), with water table depths below the surface recorded at approximately 10 minute intervals. The boreholes were monitored continuously over a 419 day period starting on 17th February 2003. A 25° V-notch weir was located in a drain approximately 32 m downstream of the borehole transect (see Figure 4); the notch was 28cm above the base of the drain and water level measurements were taken at approximately 5 minute intervals. The weir equation, calibrated in-situ, is $Q_W=0.21\ h_W^{2.3}$, where $Q_W$ ($m^3s^{-1}$) is weir flow and $h_W$ (m) is water level over the crest. 304 days of reliable observations are available during the 21 month period. Because the field experiment was not originally designed to support a physics-based model, a detailed survey of the site was not completed prior to the removal of field equipment; therefore information regarding exact ground surface heights is unknown. A transect survey across the site in the approximate location of the boreholes indicates that the ground surface level fluctuates by up to 10cm around the average slope of the site. The schematic map of the site, shown in Figure 4a, is based on information from aerial photographs and topographic maps. A rain gauge was located approximately 300m from the site, with measurements made at 15 minute resolution. As evaporation data were not available for the location, an approximate time series of daily potential evaporation was synthesised using the EARWIG weather generator (Kilsby et al., 2007). The dataset is unique in its simultaneous high resolution measurements of rainfall, drain flow and water table in blanket peat and therefore provides an important opportunity to calibrate and test the model performance.

The model boundaries were defined by the drains at the top and bottom of the transect, the weir at the outlet and the upstream end of the central drain (Figure 4b). 10 soil blocks were used in the simulation (5 upslope and 5 downslope from the central drain), each with 20 nodes, with spacing of these nodes ranging from 40cm near the drain to 3.5m in the centre of the soil block. Borehole locations were explicitly added as nodes, to avoid interpolation errors when comparing the model output against observations. The original model configuration was altered slightly in order to incorporate the routing effect of the weir: the drain upstream of the weir was modelled as a reservoir with outflows set by the weir equation measured in the field. This assumes that the residence time in this drain is dominated by the storage effect of the weir, and that travel time of the wave is negligible compared to the simulation output time step, which is considered reasonable because of the short drain length.

The model was calibrated by performing a Monte Carlo analysis. 2000 random samples were taken from the *a priori* parameter ranges shown in Table 1. The calibrated model parameters were: the acrotelm and catotelm hydraulic conductivities, the thickness of the acrotelm, the angle of the drain, the surface slope and the type of land cover. Some of the *a priori* ranges were more easily constrained (i.e. the slope and drain angle) as there was some knowledge about these parameters from information such as maps and aerial photographs. However, we chose not to fix these parameters, in order to investigate the parameter sensitivity and also to reflect the uncertainty in this information. The drain length was fixed at 46m based on the results of a long term mass balance, and the acrotelm and catotelm porosities were set as functions of their respective hydraulic conductivities following the relationship presented by Letts et al. (2000). Simulations were then performed for each of the *a priori* parameter sets for a 50 day calibration period from 24 September 2003 with a preceding 50 day model warm up period (not used for comparison against the observations) to allow sufficient time for the model behaviour to become independent of the user-defined initial conditions. The simulations took 6-10s per simulation day using an Intel Core 2 Duo Processor (E6850, 3.00 GHz).

The observed data points were interpolated to the same time samples as the model output (10 minutes). The model performance was determined for each parameter set using the Root Mean Square Error (RMSE) for observed discharges (over all the parameter sets RMSE ranges from 0.049 to 0.095 l/s), and for water table depths for all six boreholes (RMSE ranges from 0.021 to 0.161 m). None of the sampled parameter sets could simultaneously optimise the RMSE for all seven sets of observations. In order to accommodate the multi-objective nature of the problem and also recognising the uncertainty in the data and the model, rather than performing verification and predictions with a single "optimal" parameter set, the parameter sets considered most consistent with the observed hydrology of the site are selected. These are referred to here as the "*behavioural*" parameter sets (B). The behavioural parameter sets are selected by firstly taking only the best 5% for the weir flow simulations (100 parameter sets), then further reducing this set by keeping only the 50 parameter sets that had the best average RMSE for all six boreholes. Less emphasis is placed on the borehole observations in this combined criteria because our primary interest is to replicate peak flow hydrographs, and as there is inevitable uncertainty in the borehole simulations associated with heterogeneity as well as the uncertainty related to the inexact datums from which the borehole measurements were made. The selection of the criteria was arbitrary; however it achieved the purpose of constraining the model towards the observed behaviour within an uncertainty analysis framework.

Figure 5 shows the confidence limits of the predictions obtained using the *a priori* parameter sets and those obtained using the *behavioural* parameter sets, plotted with observed weir and borehole data for the largest event during the calibration period. This shows that the behavioural parameter sets give good performance during the main flow peak and demonstrates that non-behavioural parameter sets were typically rejected as they produced soil conditions that were too dry preceding the onset of the rainfall event and therefore tended to underestimate the first peak. Following the flood peak, all simulations reflect saturated conditions with the water table at the ground surface; the observations reflect a similar water table level, fluctuating between -2cm and the surface.

To test the model outside the calibration period, simulations using only the *behavioural* parameter sets were performed for the entire observation period (Figure 6). Regions of missing data in Figure 6 are periods when observations were not made, and periods with missing simulations are periods where rainfall data were not recorded. Figure 7 illustrates a period of relatively poor performance and high uncertainty in the transition from a dry to wet period. In this period, the observed rewetting (the time it took for the water table to increase by approximately 10cm) took approximately 3 hours according to the observations, whereas in the simulation it took approximately 1 day (although the simulations did react earlier). For the second flow peak, the upper bound reflects the observations; however, the lower bound indicates no flow. Once the soil became saturated, the third peak shows that the simulation improved significantly, although the lower bound prediction is only approximately 40% of the observed flow and is also delayed by approximately 45 minutes.

Figure 8 shows good flow performance and reduced uncertainty in a consistently wet period. During this period the water table was always very close to the surface (<5cm) in both the observations and the model outputs. Figure 6 demonstrates that ground water levels are generally well predicted in the winter time, when evaporation is low and the water table is very close to the surface. However, during summer periods, the magnitude of the drawdown tends to be under-predicted and the time for rewetting tends to be over-predicted. Predictions for the borehole located 10cm upstream from the drain were consistently worse than those in the centre of the upslope soil block.

The *behavioural* parameter sets can also be used to examine the sensitivity of the model to each of the parameters, by making comparisons between the frequency distributions of the *behavioural* parameters $\mathrm{F}(\theta_x|B)$ and the frequency distributions of the *a priori* parameter $\mathrm{F}(\theta_x)$ (following the approach of Spear and Hornberger, 1980). Figure 9 shows the cumulative distribution functions (cdfs) of the *a priori* and *behavioural* distributions for each parameter; the greater the deviations of the behavioural cdfs from the *a priori* cdfs, the more sensitive the model prediction is to the parameter. The significance and magnitude of the difference between these distributions (and therefore the sensitivities) is quantified using the Kolmogrov-Smirnov (KS) test (see McIntyre et al. 2003). All behavioural parameter distributions were significantly different from the *a priori* parameter distributions at the 95% confidence interval. For the parameter ranges tested in this example, the parameters ordered from most sensitive to least sensitive, based on their KS test statistic values, are: acrotelm saturated hydraulic conductivity ($K_{Sa}$), acrotelm thickness ($d_a$), drain angle ($\beta_d$), catotelm saturated hydraulic conductivity ($K_{Sc}$), land cover ($a$) and site slope ($\theta_d$). The model sensitivity to the evapotranspiration was also tested by running 10 parameter sets, randomly selected from the original 2000 parameter sets, each for five different EARWIG stochastic realisations for a 150 day period. The mean difference in RMSE between the best performing and worst performing simulations for each parameter set was 0.0023 l/s for the flow simulations and 0.0036 m for the water table simulations. This variation was considered to have little significance on the selection of behavioural parameter sets.

### 3.2. Generalised parameter space response

The model performance in the case study application suggests that the model captures the key processes in drained blanket peatlands under wet conditions. For sites that may be modelled with the same structure but different parameter values, the model was used to explore aspects of hydrological response throughout the potential parameter space. In these simulations, the original model structure shown in Figure 1 was used, rather than the version adapted to accommodate the weir. It is assumed that all possible surface roughnesses for peatland sites can be represented by values of $a$ between the smoothest (Ba) and roughest (S/J) land cover types. The parameter ranges for this broader exploration are shown in Table 1. This allows a qualitative validation of the model results relative to responses reported in the literature for a range of sites as well as providing a more general picture of the sensitivity of the flow peaks to the model parameters.

The model parameter space was quantised and simulations performed for all the possible parameter combinations. The model domain was fixed to a 500m x 500m area, and tested with seven design storms taken from the Flood Estimation Handbook (Robson and Reed, 1999), assuming a winter profile. The seven events were: 10 year return period with 1 hour duration, 10 year 2 hour, 10 year 6 hour, 10 year 12 hour, 10 year 18 hour, 2 year 12 hour and 50 year 12 hour. As only large design storms were examined, evaporation was not included in the model. Initial water table levels were set as the steady state solution for infinite duration rainfall of 0.1 $md^{-1}$ and drains were assumed to be empty. The peak flows were found to be independent of this choice of initial condition. In order to reduce parameterisation, the depth averaged hydraulic conductivity was assumed to be constant within each simulation, therefore removing the acrotelm-catotelm representation. The results from the study are shown in Figure 10. For each sampled value of a parameter $x_i$ (i=1 to 6 representing the six sampled parameters), the mean of the peak flow values over all rainfall events and sampled values of the other parameters is calculated and plotted against $x_i$. The $x_i$ values have been scaled to range between 0 and 1; the hydraulic conductivity is shown on a log scale.

The model behaviour was found to be consistent with observations from the literature. For example, at high hydraulic conductivities, drainage is found to be very effective in reducing peak flows; with low hydraulic conductivities (such as in peatlands), drainage is found to increase model peak flows and decrease times to peak, with the effects generally larger in systems with closer drains and lower hydraulic conductivities (e.g. Holden et al., 2006; Robinson, 1986; Stewart and Lance, 1991). At very close drain spacing, the peak model flows begin to reduce, suggesting that spacing contributes to both increased storage and increased conveyance. Examination of the water table profiles also shows that the spatial variation in water table depth observed in the field (Coulson et al., 1990; Holden et al., 2006) is also replicated in the model.

# 4. Discussion

A new hydrological model has been presented for drained blanket peats that can explicitly represent varied drainage networks and the water table response between these drains. The simplified physics-based model allows for the exploration of the internal model behaviour, whilst still being relatively computationally efficient. High quality data from small scale peatland sites are quite limited, and as model complexity increases, there is less likelihood that suitable observational data are available to constrain the model parameters (Freer et al., 2004). Despite the limited complexity of the new model, and the fact that the dataset used for calibration is unique in the UK for the high level of information that it contains about peatlands, there are still challenges in the calibration, in particular, simultaneously optimising the model performance against individual observation time series. There are many possible causes for inconsistency between model outputs and observations, related to the model conceptualisation as well as the quality or suitability of the observations.

Our approach to model calibration in this paper has taken into account responses that it would be reasonable to expect the model to simulate given its relative simplicity. A spatially homogeneous representation of site properties is unlikely to provide consistently accurate representations of multiple point estimates of water table levels. We also note that without a detailed survey of the site, and only the site averaged slope to work from, water table measurements made from a ground surface reference level may have several centimetres of error in them. In the field, it is also difficult to precisely define the surface of a peat, as in reality, the change from peatland vegetation to acrotelm is more of a continuum than a discrete layering. Therefore, the influence of the water table levels was down-weighted in the calibration so that that the simulated response was considered acceptable if it was broadly consistent with the general response of the six boreholes. Despite the reduced weighting of the boreholes in the calibration, they provided important information in the calibration process, particularly in refining the behavioural range of the slope; without the information from the boreholes, the slope would not have been identified as a sensitive parameter.

Even with these challenges, the longer term behaviour of the water table is generally reliably predicted (see Figure 6), with seasonal variability represented well. However, Figure 6 also highlights the relatively poor prediction of the water table near the drain edge. This is unsurprising as the assumptions made in the Boussinesq equation are no longer valid in regions near the drainage ditches, where streamlines begin to converge and the Dupuit-Forcheimer approximation fails (Bear, 1988). Drain edges are also modelled as vertical, whereas in reality they will have some degree of incline. Whether distances to the boreholes are measured from the drain edge at the top or the bottom of the drain will therefore have an impact on the location of the borehole in the model domain. Near the edges of the peat blocks, where the water table level is rapidly varying, water table predictions are very sensitive to movement of a matter of centimetres upslope or downslope of a given location. Should more accurate simulations of the water table within 1m or less of the drain edge be required, it would be necessary to reassess the suitability of the Boussinesq equation.

Despite the inevitable conflict resulting from the desire to accurately represent local scale hydrological processes and the requirement that the model should be computationally efficient, the model performs consistently well during wet periods (e.g. Figure 8). Performance during drying periods and the following recoveries was more poorly represented and had the greatest uncertainty. We assume that the slower recovery of the water table is probably related to the exclusion of an unsaturated zone representation, as in reality water stored in the unsaturated zone would add to the infiltrating water to increase the rate of water table rise. However, we suggest that, in the context of flood response, the loss of precision for these periods is outweighed by the significant gains in computational time (assuming that number of model nodes can be taken as a proxy for computational time, (e.g. Pancioni et al, 2003)), and also note that a poorly constrained complex subsurface representation would be unlikely to provide greater precision in these periods. It is also important to note that our calibration period was during winter; therefore it is possible that if there had been suitable data to use for a calibration period in the summer time, that drying and rewetting of the peat may have been better captured in the behavioural parameter set.

Based on an examination of the response of the modelled flow across the parameter space under large rainfall events, the model parameter to which the peak flow response is most sensitive is the drain spacing followed by the hydraulic conductivity. However, at low hydraulic conductivities (e.g. typical of UK blanket peats) the peak flow is almost independent of the hydraulic conductivity. In that case, apart from drain spacing, the peak flows are most sensitive to the parameters related to the land surface and drain roughnesses. This is unsurprising given that the simulations were for large rainfall events, where any storage in the subsurface could be rapidly filled. The high sensitivity of flow to the roughness parameters also reflects their high uncertainty. Further field investigations of these parameters (e.g. Holden et al., 2008) would greatly enhance any hydrological modelling efforts for blanket peatlands.

# 5. Conclusions

The processes and responses associated with drained peatlands have been captured in a new simplified physics-based model. The model has advantages over previous physics-based and lumped conceptual models, as it provides flexibility in the drainage configurations that can be represented and can provide outputs of the spatial variability of model internal states. The results of the generalised parameter space response indicate that peak flows are sensitive to the geometric properties of the hillslope and drainage configurations, therefore models that are spatially lumped or restricted in their model configuration cannot as accurately distinguish those sites that pose the greatest flood hazard. The model therefore has potential in terms of specifically identifying and prioritising areas for flood hazard mitigation measures in terms of potential reduction of downstream flood risk. The model has been tested against a dataset from the UK and has been shown to perform well in terms of capturing peak flow responses under saturated or near-saturated soil moisture conditions. Poorer performance under drier conditions was explained by lack of an accurate unsaturated zone model, which while not of great concern for flood flow applications, could restrict the model's usefulness for the exploration of other peat management impacts on, for example, low flows and water quality. Although the unknown surface levels at the boreholes created challenges with the simultaneous optimisation of all six boreholes, long term behaviour of water table levels was reasonably well predicted, and the general water table behaviour was consistent with observations from other studies. How far the model can be generalised will need to be explored further through testing against more data sets. The effect of spatial heterogeneity of the model parameters should also be investigated. The modelling process has helped identify the overland and channel flow roughness parameters as being particularly important controls on peak flow response. Further field research towards constraining these parameters is expected to enhance the model performance.

## 6. Acknowledgements

This research was funded by the UK Flood Risk Management Research Consortium Phase 2, EPSRC Grant EP/F020511/1. Time to work on field data used in this paper was gratefully provided by a Philip Leverhulme Prize awarded to Joseph Holden. Thank you also to two anonymous reviewers, whose constructive comments helped us to significantly improve the manuscript.

## Appendix

List of parameters:

| | |
|---|---|
| $a$ | Proxy overland flow friction factor |
| $\beta_d$ | Drain angle (Degrees) |
| $\varepsilon_a$ | Effective porosity of acrotelm (-) |
| $\varepsilon_c$ | Effective porosity of catotelm (-) |
| $\theta_d$ | Drain slope (Degrees) |
| $\theta_s$ | Site slope (Degrees) |
| $\theta_x$ | *A priori* parameter sets |
| $B$ | *Behavioural* parameter sets |
| $d_a$ | Thickness of the catotelm (L) |
| $d_c$ | Thickness of the catotelm (L) |
| $ET_c$ | Actual evaporation from collector drain ($LT^{-1}$) |
| $ET_d$ | Actual evaporation from drains ($LT^{-1}$) |
| $ET_{OF}$ | Actual evaporation from overland flow ($LT^{-1}$) |
| $ET_p$ | Actual evaporation from peat soil ($LT^{-1}$) |
| $f$ | Darcy Weisbach friction factor (-) |
| $g$ | Acceleration due to gravity (9.81 $ms^{-2}$) |
| $H$ | Generic water depth (L) |
| $h_d$ | Depth of water in drain (L) |
| $h_{OF}$ | Depth of overland flow (L) |
| $h_T$ | Total hydraulic head (L) |
| $h_S$ | Depth of water table above impermeable bed (L) |
| $h_W$ | Height of water above weir crest (L) |
| $i$ | Infiltration ($LT^{-1}$) |
| $K_{Sa}$ | Saturated hydraulic conductivity of acrotelm ($LT^{-1}$) |
| $K_{Sc}$ | Saturated hydraulic conductivity of catotelm ($LT^{-1}$) |
| $L_C$ | Length of the collector drain and site length (L) |
| $L_d$ | Length of the drain (L) |
| $n$ | Manning's n |

$Q$ Generic unit width flux ($L^2T^{-1}$)

$q_c$ Collector drain flow ($L^2T^{-1}$)

$q_d$ Drain flow ($L^2T^{-1}$)

$q_{OF}$ Unit flux of overland flow ($L^2T^{-1}$)

$q_S$ Unit flux of subsurface flow ($L^2T^{-1}$)

$Q_W$ Weir flow ($L^3T^{-1}$)

$t$ Time (T)

$V$ Generic velocity in downslope direction ($LT^{-1}$)

$W$ Drain spacing (L)

$W_d$ Drain width (L)

$x$ Peat block ordinate (L)

$x_C$ Collector drain ordinate (L)

$x_d$ Drain ordinate (L)

$y$ Generic downslope distance ordinate (L)

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*Table 1: Parameter ranges for Oughtershaw Beck Monte Carlo simulations and general sensitivity analysis*

| | Ranges for Oughtershaw Beck Monte Carlo Simulations | | Ranges for Sensitivity Analysis | |
|---|---|---|---|---|
| Parameter | Lower value | Upper value | Lower value | Upper value |
| Acrotelm hydraulic conductivity (m/d) | 0.1 | 4 | (Depth Averaged) 0.001 | (Depth averaged) 10 |
| Catotelm hydraulic conductivity (m/d) | 0.001 | 0.05 | | |
| Acrotelm thickness (m) | 0.05 | 0.2 | FIXED | |
| Drain angle (degrees) | 10 | 20 | 15 | 60 |
| Surface slope (degrees) | 5 | 10 | 2 | 14 |
| Land cover | Sphagnum & Juncus (roughest) | Eriophorum (smoothest) | Sphagnum & Juncus (roughest) | Bare (smoothest) |
| Manning's n | n/a | n/a | 0.2 | 1.4 |
| Drain spacing (m) | FIXED | | 5 | 500 |

Figures:

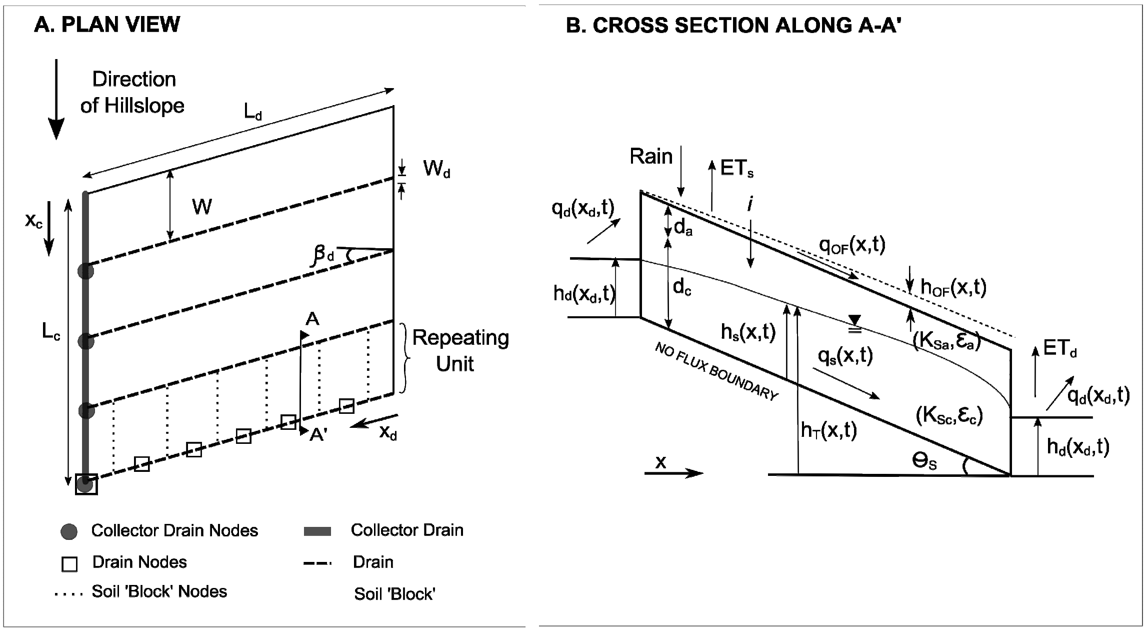


Figure 1: conceptual model of drained peatland

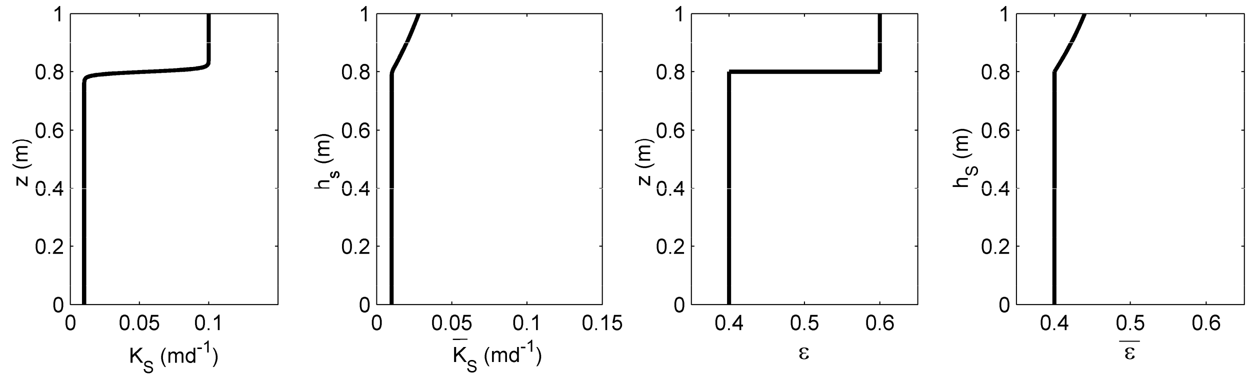


Figure 2. Variation of $\overline{K}_S$ and $\bar{\varepsilon}$ with $h_S$, and variation of $K_S$ and ε with *z,* given $K_{Sa}$=1 md$^{-1}$, $K_{Sc}$=0.01 md$^{-1}$, $\varepsilon_a = 0.6$, $\varepsilon_c = 0.4$, $d_a$ =0.2m and $d_c = 0.8$m

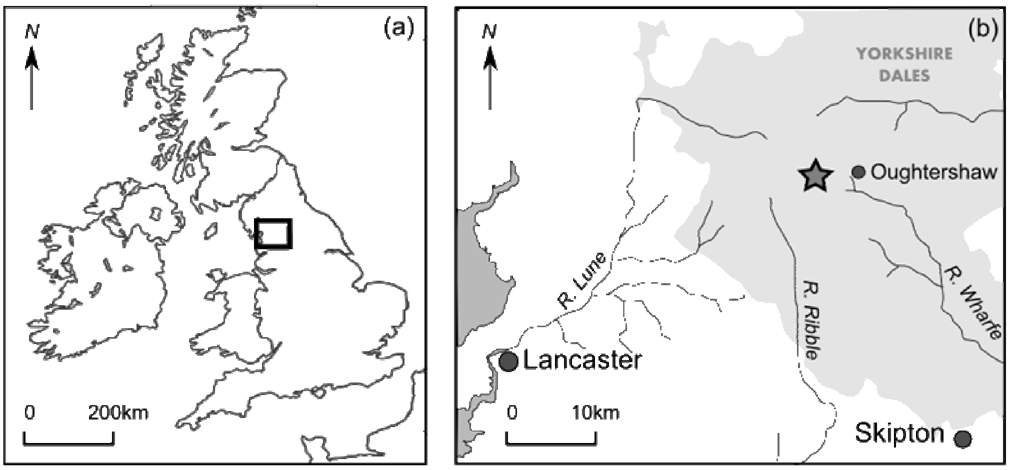


Figure 3: Location map of Oughtershaw Beck; (a) Location within the British Isles (b) Site location within the Yorkshire dales, marked by the star. Major towns in the area are marked with large circles; Oughtershaw is a small hamlet and marked with a small circle.

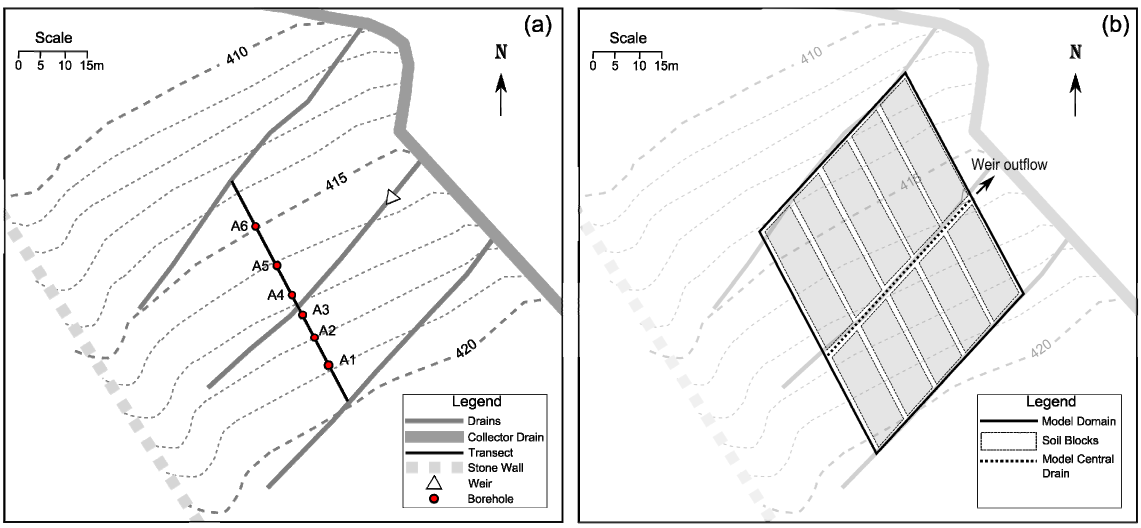


Figure 4: (a) Field site schematic diagram, (b) Model domain and soil blocks

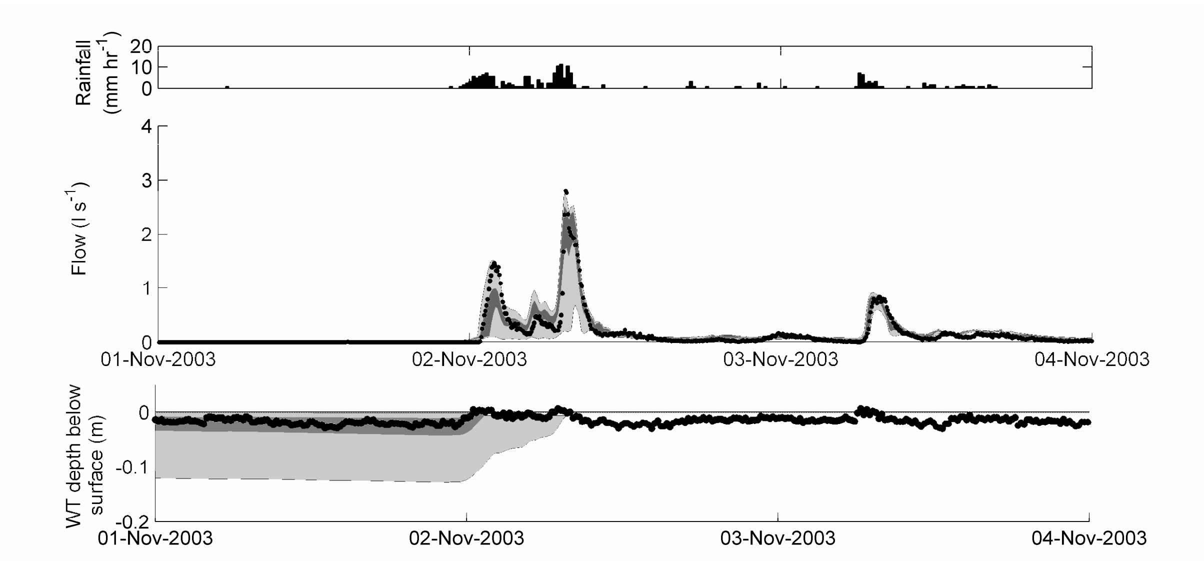


Figure 5: Four day sample from the calibration period, showing the largest peak and water table (WT) depth at borehole A2. Light grey: 90% confidence interval for all simulations; dark grey: 90% confidence interval for behavioural simulations; black dots: observations

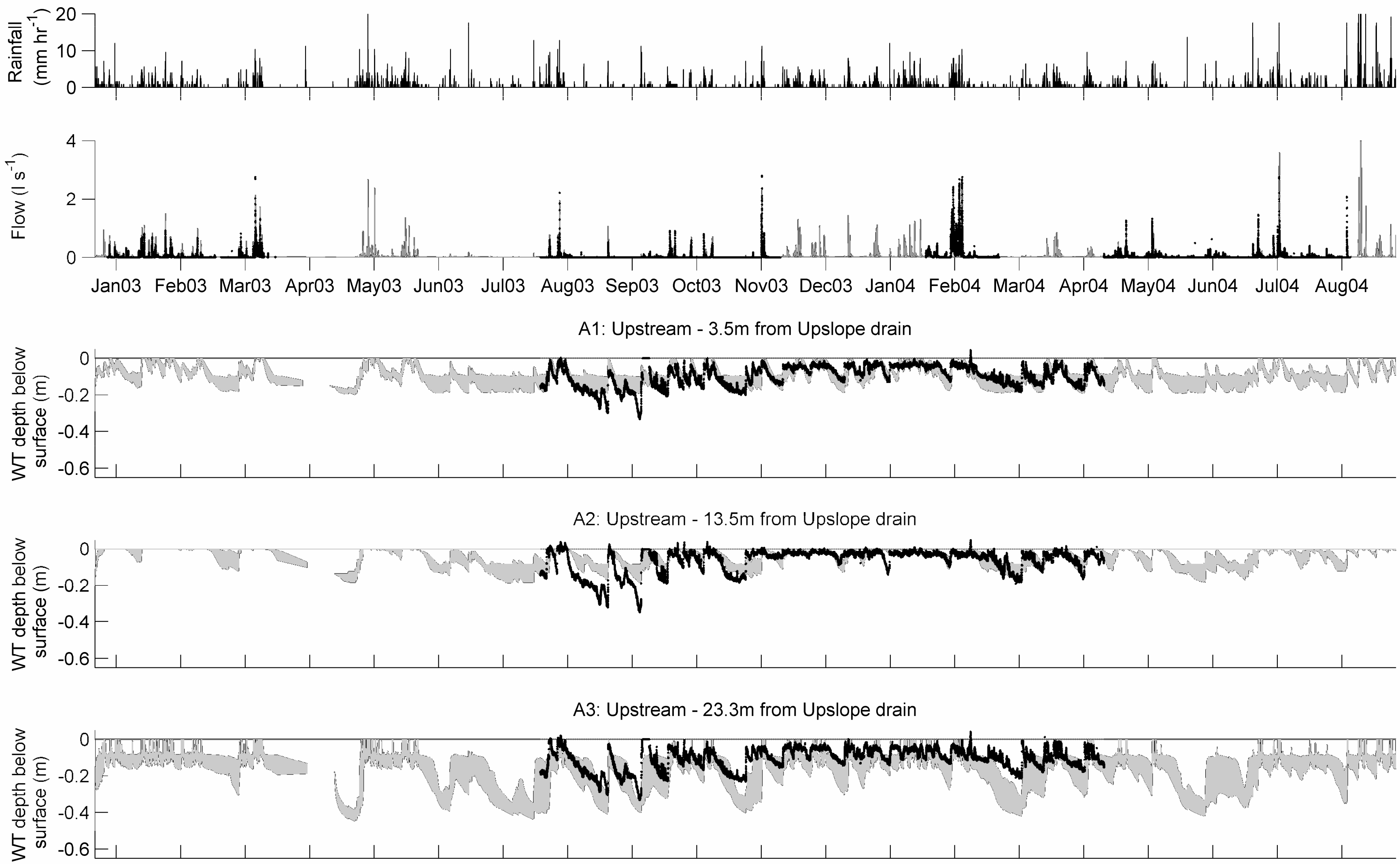


Figure 6: Rainfall, flow and upstream water table depth for the verification period. Grey area: 90% confidence interval of *behavioural* simulations; black line or black dots: observations.

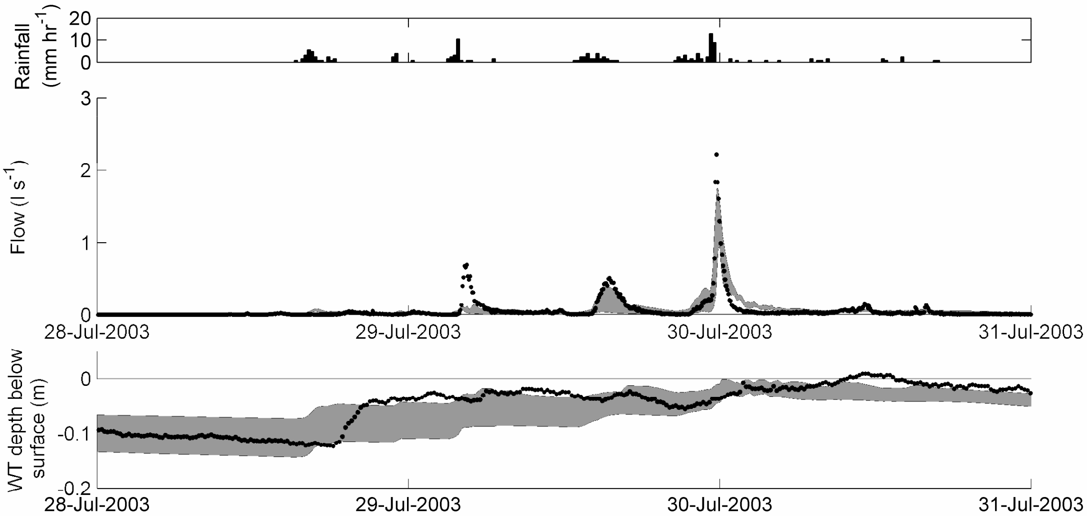


Figure 7: Flow hydrograph and water table (WT) depth for borehole A2 from verification period. Grey area: 90% confidence interval of *behavioural* simulations; black dots: observations.

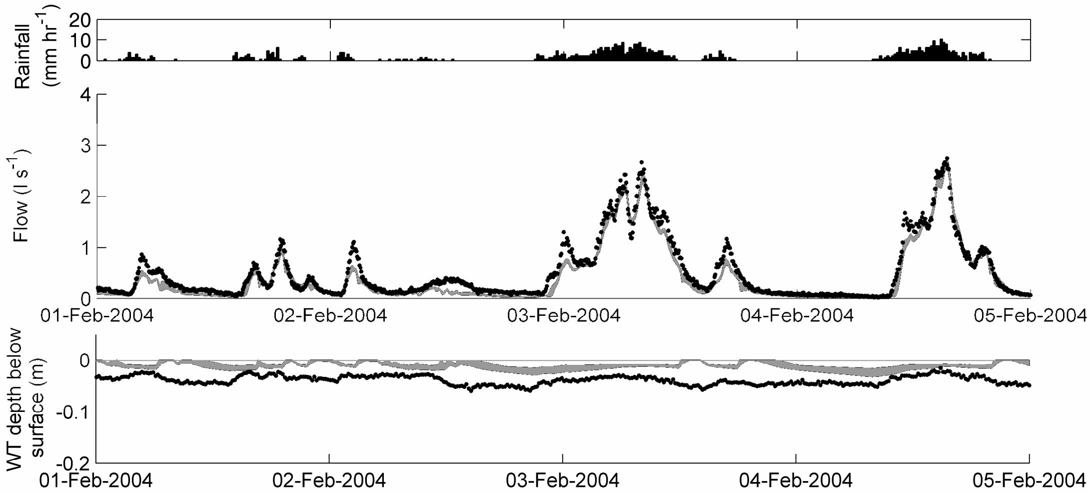


Figure 8: Flow hydrograph and water table (WT) depth for borehole A2 from verification period. Grey area: 90% confidence interval of *behavioural* simulations; black dots: observations

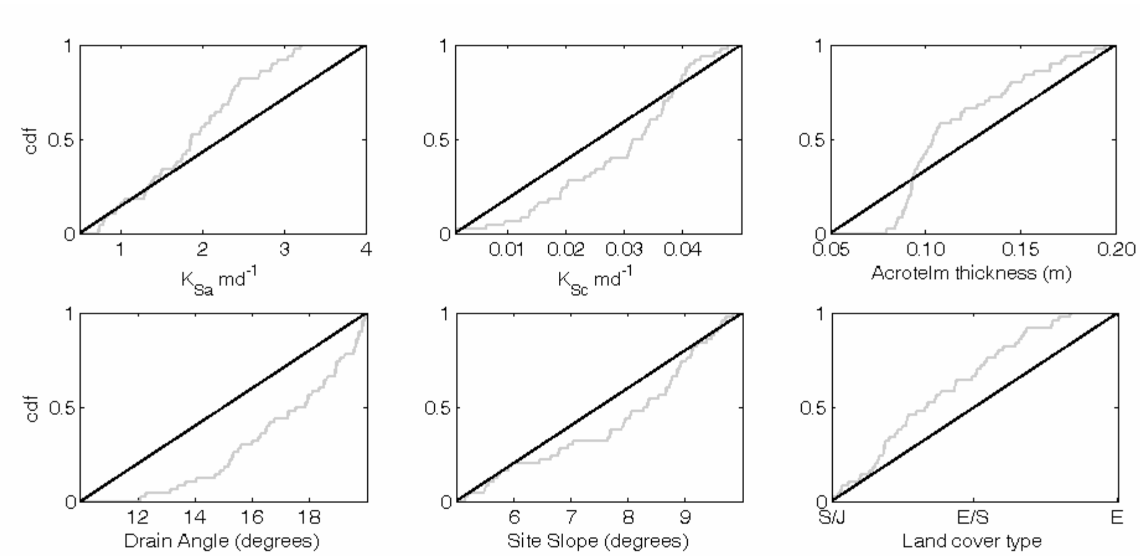


Figure 9: Cumulative density plots of the *a priori* and behavioural parameter distributions. Black line: *a priori* parameter distribution; grey line: behavioural parameter distribution

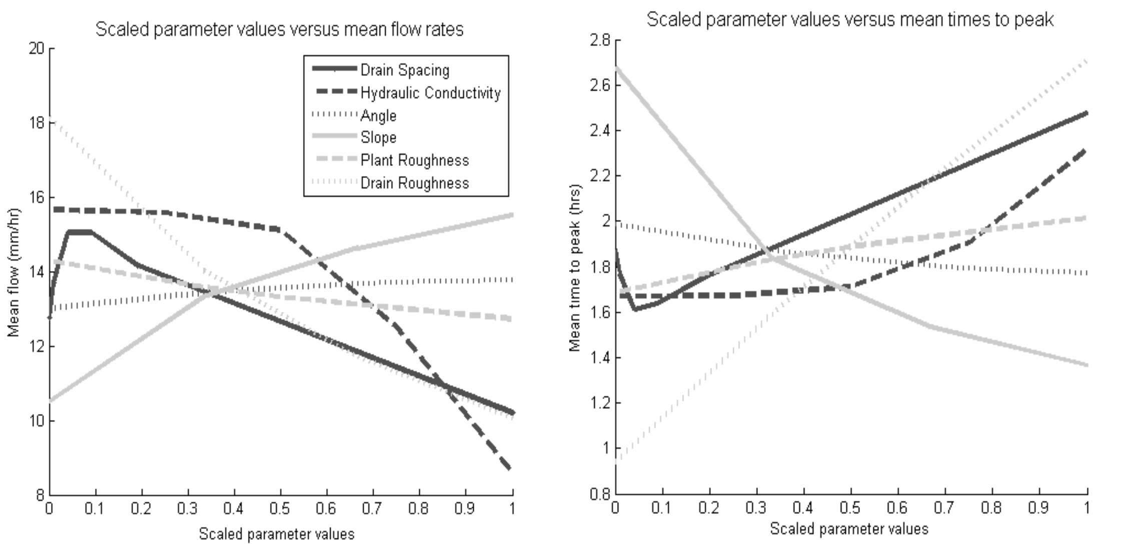


Figure 10: Mean flow rates and mean times to peak versus scaled parameter values