A study of soil moisture controls on streamflow behaviour: results for the OCK basin, United Kingdom

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ABSTRACT

In humid temperate areas ground wetness plays a key role in storm runoff generation, but until recently there have been no instruments capable of providing continuous, reliable, records of changing soil moisture conditions in the field. A new instrument, the IH capacitance probe, can provide continuous measurements of soil water contents. Together with rainfall records these data have been used to study the variations in river flow response of a medium sized (234 km2) rural catchment. Daily flows were simulated, firstly using a standard rainfall runoff model (IHACRES) with conventional hydrological and climate data and, secondly, by replacing the net rainfall calculation by a simple functional relationship to the measured soil moisture contents.

It was found that incorporating soil moisture measurements in the runoff model:

a)Reduced the length of record required for model calibration,
b) Improved the simulation of streamflow.

INTRODUCTION .

Hydrologists require information about rainfall and soil conditions for flood warnings and for design flood estimation. Whilst the measurement of rainfall has received much attention, especially since the implementation of weather radars, there is generally little information on the changing status of ground wetness. Yet this is a key factor affecting the response of humid temperate zone catchments to rainfall; very different flood responses are observed for similar rainfall inputs onto different initial ground wetness conditions. At present, measurements of soil water content using manually read neutron probes are normally available only on a weekly or monthly basis. For changes over shorter periods an index based on weather data (rainfall and potential evaporation) is commonly used, as for example in the Flood Studies Report (NERC, 1975) which is the standard method for engineering flood design in the UK.

Early research work on flood flows was based on observations of widespread overland flow in the semi arid south western USA (Horton, 1933). However, in well vegetated humid temperate areas, overland flow is rarely seen. Infiltration capacity is generally far greater than most rainfall intensities, so that except on disturbed ground infiltration excess runoff will not normally occur. Some studies show stormflow results mainly from direct channel precipitation and saturation excess overland flow generated by rainfall onto saturated areas close to streams (Cappus, 1960; Dunne and Black, 1970). Other research has indicated the importance of subsurface stormflow from close to stream channels (Hewlett and Hibbert, 1967). Evidence of the importance of subsurface flows also comes from observations of saturated soil layers above an impeding horizon (Weyman, 1970) and from macropores (Beven and Germann, 1982). It is now recognised that even within a single catchment a range of runoff generation processes will be operating. These studies have, however, emphasised the general importance of near surface soil water conditions on

stormflow, whether through its control on the extent of saturation excess overland flow, or through its control on subsurface flows. Isotopic evidence also indicates the importance of subsurface flows (Pearce et al, 1986).

Spatial pattern of soil moisture

Once it became recognised that only the saturated parts of a catchment could contribute to quick flow, much work has been conducted to predict the location and extent of these zones. Dunne et al (1975) describe field survey methods for the recognition and mapping of the saturated zones in small catchments. In addition to contiguous channel-side areas, Kirkby & Chorley (1967) identified areas of subsurface flow convergence likely to lead to soil saturation. These comprise: concavities in plan (contour curvature) and in slope profile, and also areas of thin soils. To these situations may also be added soils in which porosity and permeability decrease with depth (especially, but not exclusively in layered soils) resulting in the building up of saturated layers, above any regional groundwater table.

Beven and Kirkby (1979) describe a semi distributed hydrological model, TOPMODEL, which uses a topographic index (Kirkby, 1975) to describe the propensity of any point in a catchment to develop saturated conditions. Calculations of the index for a number of catchments show it to have a skewed frequency distribution; some parts of a basin are subject to saturation more frequently, and for longer periods, than other areas. Some zones may rarely - if ever - become saturated except in the most exceptional conditions.

Problems of predicting storm runoff volumes

Much progress has been made in understanding the complexity of stormflow processes, but practical methods for estimating storm losses and runoff have yet to be developed (Pilgrim and Cordery, 1993). Tests of the US Soil Conservation Service method found large differences between observed and predicted peaks, with poorer results for dense vegetation cover than for bare soil or sparse vegetation; it was concluded that the assumed antecedent moisture condition had a major effect (Pilgrim and Cordery, 1993). The weakness of the estimation of storm runoff coefficients in UK flood design is also recognised (IH, 1987).

Whilst the critical importance of catchment wetness

for streamflow generation is now widely appreciated, soil moisture measurements have been limited by the instrumentation available. Until recently soil moisture measurements had to be made by manual methods - gravimetric sampling or by neutron probe - requiring a site visit for each set of readings.

SOIL MOISTURE MEASUREMENTS

Recent instrument developments based on the measurement of bulk soil dielectric constant by Time Domain Reflectometry (Topp et al, 1980) or capacitance probe (Dean et al, 1987) are capable of measuring changes in soil moisture continuously. This paper describes initial results of a project to test the feasibility of such an approach using the Institute of Hydrology (IH) designed Capacitance Probe. Manually operated variants have been used in the field over a number of years, to monitor changing ground wetness (eg Dean et al, 1987; Robinson and Dean, 1993). This is the first application of a continuous logging multichannel system, using capacitance probes buried at different depths to record soil profile water content changes for flood studies. Due to its simpler electronic circuitry the capacitance probe is considerably cheaper than the TDR for individual applications, whilst the ability to multiplex the TDR makes it more appropriate where measurements are required at a number of points. The two techniques have been reviewed (eg Gardner et al, 1991).

Capacitance probes, in contrast to other instruments such as pressure transducer tensiometers, record water content directly and function over the whole range of soil water content encountered in the field, and only require attention once per year for battery replacement. The application of the capacitance probe to soil moisture studies capitalises on the very large difference between the dielectric constant of water (approx 80) and that of air (unity) and soil (approx 2 to 4 depending on the material). This makes the dielectric constant of the bulk soil (ie soil, air and water) very sensitive to changes in water content. The capacitance probe measures the dielectric constant by inserting two stainless steel electrodes into the soil. The electrode rods and the soil between them form a capacitor, and together with the probe body which contains a battery powered oscillator they form part of the oscillator circuit. The oscillator operates at about 150 MHz in air, and the frequency of the whole circuit (electronic components, rods and bulk soil) varies from about 90 MHz in wet soil to 140 MHz in dry soil.

Field sites

The river Ock catchment (234 km2) is a rural catchment in S England, some 30 km south west of Oxford (Figure 1). It has natural flows which are measured by a weir at Abingdon. The catchment comprises permeable chalk uplands (which sustain summer baseflow) and a central valley covered with relatively impermeable clay soils, which are the source areas of storm runoff and result in the catchment having a long history of flooding. The long term average annual precipitation is approximately 650 mm and the streamflow averages 200 mm. Daily catchment average areal rainfall was calculated from a network of five long term gauges, using isohyets. The Penman potential short grass evaporation is about 500 mm with a pronounced summer peak in June to August and low values from October to March.

Soil water monitoring has been conducted at two sites; each has a raingauge, three capacitance probes buried in the soil at 5, 15 and 45 cm depths, and three purgeable pressure transducer tensiometers at 15, 25 and 50 cm depths. A dedicated interface controls and links the sensors to a solid state logger. Since electronic signals from the probe would affect its electrical field, and hence the dielectric constant, each probe is connected to the logger by fibre optic cables. The oscillation frequency of
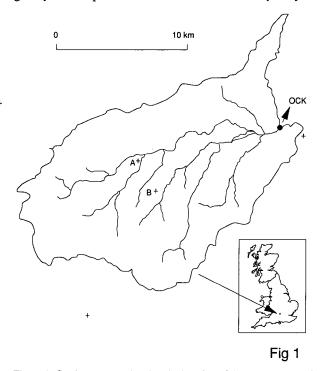
the probe is converted to an optical signal which is transmitted from the probes and reconverted to an electrical signal which is received by the logger.

The soil water stations were sited on surface water gley soils in the central clay soil part of the Ock catchment, which provides the bulk of the storm flow response. The instruments at both sites are on permanent pasture, away from localised effects of stream channels, ditches and field drains, and the ground is fairly flat. The Stanford Park Farm site, lies in the valley bottom and is on Kimmeridge Clay (Jurassic period clays and clay shales). It has permanent grass, cut once per year for hay in summer and then used for grazing. The second field site, about 2 km to the south, is situated on higher ground at Challow Hill Farm, and is on Gault Clay (Cretaceous period, mainly clayey).

Due to the spatial variability of soil water, the global soil water storage of a catchment is unknown. However, manual readings at a network of sites in the catchment over a year (Hasnip, 1993), indicated that there is considerable similarity in the time varying seasonal patterns of near surface soil moisture across the catchment. Water content changes were well correlated and this finding indicates that each soil water measurement station can provide an index of the changing state of ground wetness over the catchment.

Calibration of the capacitance probes

Field calibration of all the capacitance probes (frequency to water content) was carried out. The sites are visited twice each month to download the loggers and check the instruments. The tensiometers are purged if required, and independent measurements made of soil water content (by neutron probe and gravimetric analysis). For the 5 and 15 cm probes, soil samples were taken to determine the volumetric water content by the thermogravimetric technique. For the deepest probes (45 cm) correlations have been made with water content measurements using a neutron probe. To avoid site destruction during the gravimetric sampling, the surface soil water contents at the location of the buried probes were measured using a hand held Surface Capacitance Insertion Probe (SCIP) (Robinson and Dean, 1993). Further SCIP readings were then taken at 5 to 10 metres distance until a similar value was obtained, indicating a location where the soil water content was the same. Here the soil sample was taken.

Fig 1

Figure 1. Catchment map showing the location of the measurement sites, including the streamflow gauge (•), raingauges (+) and soil water stations (A, Stanford Park; B, Challow Hill).
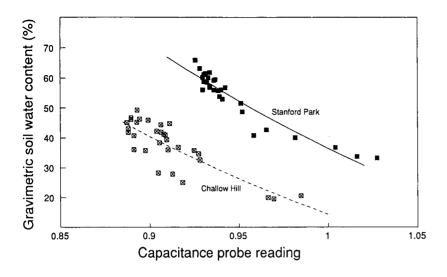
Figure 2. Field calibration curves for the 15 cm depth capacitance probes.
Upper curve - Stanford Park R2=90%
Lower curve - Challow Hill Farm R2=68%

Separate calibrations were obtained for each site and for each depth. The calibration curves show a high correlation despite the necessary transposition in space between the buried probes and the location of the gravimetric sample. For 5 cm depth and 15 cm depth probes 70 to 80% of the variance in water content was explained. The fit was poorer for 45 cm, where the very limited range in observed water contents (5%) was of similar magnitude to the errors of the neutron scattering method. Calibration curves (between the capacitance probe reading and the gravimetric soil water content) are given in Figure 2 for the two capacitance probes at 15 cm depth, since that depth proved to be the most useful in the streamflow modelling described later. The difference between the two sites is due to factors including soil density, texture, and organic matter content which influence the dielectric constant (eg Jacobsen and Schjønning, 1993). The soil at Stanford Park has a higher clay content than that at Challow Hill (50 vs 35%) and a lower bulk density (0.8 vs 1.2), factors which act to reduce the bulk soil dielectric constant and so increase the reading at a given water content.

APPLICATION OF SOIL WATER DATA TO STREAMFLOW PREDICTIONS

The benefit of using measured soil moisture data for streamflow prediction was investigated using a lumped black box time series rainfall runoff model, IHACRES (Jakeman et al, 1991; Littlewood and Jakeman, 1994). The model structure comprises a non-linear loss module

which generates rainfall excess from areal rainfall and air temperature, followed by a linear rainfall excess streamflow module (Figure 3). This separate treatment of net rainfall calculation and subsequent routing makes it relatively easy to introduce a new net rainfall module based on soil water content. In normal application of the model the loss module calculates catchment wetness s, as a function of rainfall r, and temperature t. To calibrate this component of the model the optimum values of two parameters were determined. These comprise w, a time constant describing an exponential decrease in s due to evapotranspiration in the absence of rainfall, and f, a temperature modulation factor which quantifies how w changes per degree Celsius change of temperature. Net rainfall u, at time k, is then calculated by:

$$u_k = r_k * 0.5(s_k + s_k - 1) \qquad (1)$$

The computed net rainfall is then input to a system of linear storages in any series and/or parallel configuration to estimate streamflow. The standard configuration of the rainfall excess-streamflow module was used, which comprises two parallel flow components (see Figure 3). The model calibration ensures that the computed rainfall excess equals the volume of observed flow.

Model calibration without measured soil moisture data

Daily rainfall, streamflow and temperature data over the 10 year period (September 1982 to August 1992) were used for calibration. However, this proved to be
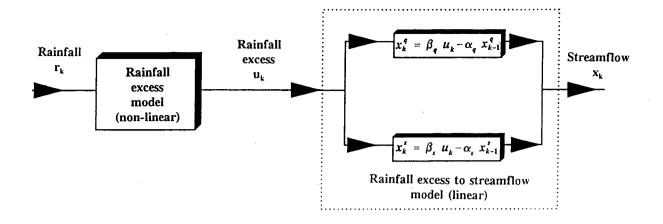
Figure 3. The IHACRES model structure using a parallel configuration of storages for quick and slow flow in the rainfall excess - streamflow module.
   xq is quick flow
   xs is slow flow
   x is total streamflow
   and ß are parameters for quick and slow flow which are determined by model calibration.
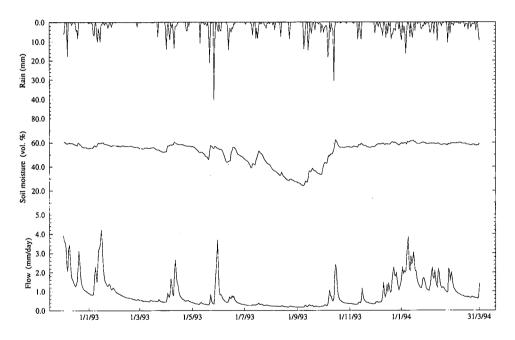

Figure 4. Daily rainfall, streamflow and measured soil water content (Stanford Park) 1992-1994.

too long a period for model calibration. Accordingly, the record was divided in four overlapping subperiods, each of three years length, which were independently calibrated. Experience with IHACRES has shown that three years is a suitable length for a satisfactory calibration of this model - it generally contains a sufficient variety in weather and flow conditions, but not too many data to prevent a model fit (Littlewood & Jakeman, 1994). Using subperiods enabled the stability of the model parameter values to be assessed. Due to the model 'volume forcing' the predicted flows to those observed

the calibration period should be chosen to begin and end with similar stream discharge (it being assumed this would ensure similar storage); and it is generally preferable to start the model during a time of low flow. The subperiods were chosen to start from 1st September and end on 31st August since the end of August was usually a time of low streamflow. Since some years did not have suitably dry summers to start or end the model calibration, there is an overlap between some of these subperiods.

Table 1. Calibration fits for each subperiod, showing the influence of flow conditions (mean and coefficient of variation) on net rainfall model parameters values, f and w .

Subperiod	f	τ_w hrs	R^2 (%)	ARPE (%)	Mean flow m^3s^{-1}	Coeff Var.
1) 1982-85	3.4	1.9	75.8	0.083	1.556	0.974
2) 1984-87	3.5	1.6	74.6	0.084	1.628	0.957
3) 1986-89	3.4	2.4	78.0	0.085	1.543	1.071
4) 1989-92	4.0	3.8	86.4	0.055	0.855	1.383

Table 2. Simulation fits for each subperiod and for the whole period September 1982 to August 1992, using the model parameter values derived for the four subperiods.

Subperiod	1982-1985		1984-1987		1986-1989		1989-1992		1982-1992	
	R^2	Bias	R^2	Bias	R^2	Bias	R^2	Bias	R^2	Bias
1	xxxx	xxxx	73.1	.00	74.1	.21	49.6	-.35	72.1	-.04
2	76.8	.00	xxxx	xxxx	74.2	.22	44.0	-.35	71.8	-.03
3	67.0	-.26	64.9	-.25	xxxx	xxxx	77.7	.00	63.3-	.25
4	57.6	.32	53.1	.37	54.7	.51	xxxx	xxxx	63.0	.29

Table 3. Goodness of fits for flows simulated March 1993- March 1994, with net rainfall model parameter values optimised using: a) Rainfall and runoff data in the 4 subperiods (Eq.1), b) Measured soil water data 1993-94 (Eq.4)

Period:	1982-85	1984-87	1986-89	1989-92	Soil water data
R^2(%)	80.3	81.6	81.5	61.9	87.1
Bias (m^3s^{-1})	0.23	0.23	-0.03	0.65	0.01

Table 1 shows the optimum combination of net rainfall parameters, for each subperiod and the goodness of flow fit determined by two measures, the coefficient of determination,R2 and the ARPE (Average Relative Parameter Error).

The optimum parameter values for subperiods 1 and 2 are very similar, while those for subperiod 4 are quite different, and those for subperiod 3 differ to a lesser extent. As in most models, the optimised parameter values are affected by the weather conditions during the calibration period (the number and magnitude of storms influencing the efficiency of fit of high flows, for example). The most recent subperiod (1989 to 1992) was particularly dry, which may account for the difference in the parameter values obtained compared with those for the first two subperiods.

Subsequently, these four sets of calibrated parameter values were used in simulation mode for the other three subperiods and for the whole period 1982-1992 (mean flow 1.353 m3s-1, coefficient of variation 1.203). This was to establish how well the values represent the long term catchment behaviour (Table 2). The fit is measured in terms of R2(%) and the volume bias (the difference between the observed and modelled mean flow in m3s-1).

Not surprisingly, the very similar parameter values from subperiods 1 and 2 gave similar fits when interchanged, and also when applied to the whole period, 1982-92.

The fit was poorer when the parameter values from the later subperiods were used, particularly subperiod 4.

These parameter values were then applied to the period for which soil water measurements were available (December 1992-March 1994). The initial period December 1992-February 1993 had to be omitted, however, since the model had problems with the `wet start' in December 1992. The period was reduced to 1st March 1993-31st March 1994, providing a period starting and ending with similar streamflow and, it would be assumed, catchment storage. This was a period of high flows (mean 2.13 m3s-1, and CV of 0.864), and it was fitted best using the parameter values derived from subperiod 3 (Table 3). Unfortunately it was not possible to obtain a calibration of the model on this year alone, since it was too short a period of time.

Applying measured water contents

The structure of the IHACRES model meant that it was relatively easy to replace the standard rainfall loss module based on rainfall and temperature to estimate catchment wetness with a net rainfall filter based on the measured water contents. Several functional relationships of varying degrees of complexity were investigated. In the first instance, the net rainfall at time k was taken to be the product of gross rainfall and the volumetric moisture content, w, in an analogous form to equation (1):

$$u_k = r_k * w_{k-1}$$

Since there was so little variation in moisture content for the 45 cm deep capacitance probes, this model was only applied using the 5 and 15 cm depth probes on each site. Previously, it had not been possible to calibrate IHACRES in its standard form on only one year of data, but by using this simple modification to the net rainfall calculation it was possible to obtain calibrations in three of the four cases (the exception being the shallowest probe at Hill Farm). This ability to achieve a model calibration over a much shorter length of data was the first practical benefit of soil moisture data for streamflow prediction to arise from the project.

Not surprisingly, given the simple nature of the filter the calibration fits were modest (with R2, values of 0.68 and 0.62 for Stanford Park 15 cm and 5 cm deep probes respectively, and 0.56 using Hill Farm 15 cm; the ARPE varied between 0.37 to 0.64). To improve the calibration fit further, a more realistic function is required: it may

be hypothesised that the relationship between storm runoff volume and soil wetness will not be linear, but rather of a sigmoid (s-curve) form. This may be justified as follows. However dry the catchment soil becomes, there will always be a small runoff response due to direct channel precipitation. As the catchment becomes wetter streamflow will slowly increase from riparian areas. Progressively more areas will contribute, and streamflow will increase at an increasing rate. Eventually the rate of increase will lessen and ultimately there will be an upper limit as some parts of the catchment (hill tops, very permeable soil, etc) will never contribute to stormflow. This picture is quite compatible with the catchment studies of stormflow generation and contributing areas, described earlier, and with the modelling work based on topographic form such as the approaches of Kirkby (1975) and O'Loughlin (1986).

There are a number of exponential and logarithmic functions which incorporate a sigmoid shape for part of the range. Several forms were examined, and the one selected here is:

$$y = b \exp(-(ax)^4)$$

For a, b >0 this function monotonically increases from x=-∞ to x=0. The non-linear filter for the proportion of gross rainfall becoming excess rainfall, sk, has been thus been established in the form:

$$s_k = b \exp(-(a [w_k - 0.7])^4) + c$$

where a,b,c>0, and b+c=<1 ensure the excess rainfall never exceeds the gross rainfall. This filter has been applied to the 15 cm depth water contents at Stanford Park, since this probe had the strongest correlation to the streamflow. The maximum recorded moisture content was 0.7, so this value is subtracted from x (ie the measured moisture content) to ensure that the net rainfall proportion rises to a maximum value at this water content. There are three parameters to be optimised, of which two of them are semi-physical and their approximate values can be determined. Parameter b, is the maximum proportion of net rainfall when the water content is at its maximum, and parameter c represents the underlying baseflow component, and is necessary to make allowance for the groundwater contribution to total streamflow from the chalk areas of the basin.

The best calibration fit using soil water data for the period March 1993 - March 1994 had an R2 of 87.1% and an ARPE of 0.117. This is much better than using the
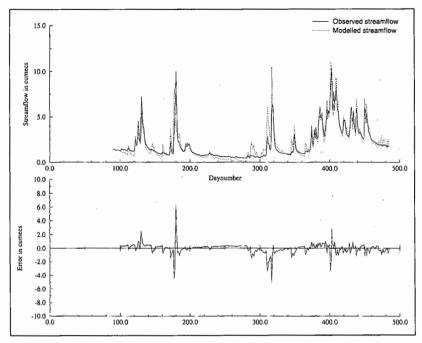
Figure 5a. Simulation for March 1993-March 1994, using IHACRES with the optimum parameters of subperiod 3.
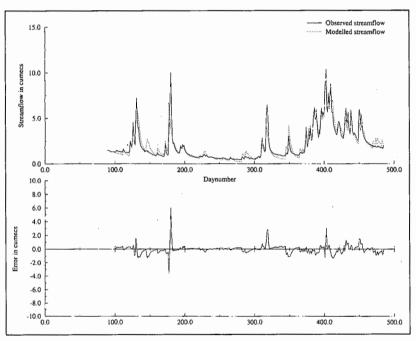

Figure 5b. Calibration for March 1993-March 1994, using the net rainfall filter based on measured soil water contents.

standard model with the best set of parameter values derived from the three year subperiods (Table 3). Figure 5 shows the fit between observed and modelled streamflow for a) standard model with the best parameter values obtained from subperiod 3, and b) calibration using measured soil moisture. It is obviously desirable to have a longer period of soil water data in order to be able to compare the predicted and observed flows for an independent period, and this remains the long term goal of this project. Additionally, it is hoped that advances with remote sensing of soil moisture (eg Cognard et al, 1995) may one day enable extrapolation of the point measurements of ground wetness across a whole catchment.

CONCLUSIONS

This study has demonstrated that the recent advances in instrumentation make it possible to collect continuous soil moisture data, and they may be used in rainfall runoff models to improve rainfall-runoff model performance in several ways:

Firstly, because of the extra constraints put onto the model behaviour it is possible to calibrate a streamflow model on a shorter period of observation.

Secondly, the extra information can improve the model calibration fit.

The time series of monitored soil water data currently available are too short to be used in an independent test of the model in simulation mode, but it is intended that the data collection will be continued long enough for this goal to be achieved.

The results presented here use a very simple non linear s-curve relation between the net rainfall proportion and the water content at a single depth in the soil profile. More sophisticated analyses using the depth varying water content profile may further improve streamflow predictions.

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