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Modelling river-aquifer interactions

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Modelling river-aquifer interactions

Edited by P. L. Younger

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Modelling river-aquifer interactions

Editor's Preface:

RIVER-AQUIFER MODELLING IN PERSPECTIVE

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River-aquifer interactions are at the very core of the hydrological cycle. Nevertheless, process descriptions and engineering analyses of river-aquifer interfaces have traditionally faced two hindrances, one cultural, the other practical.

The cultural hindrance is the traditional sub-discipline boundary between hydrogeology and surface water hydrology, with researchers in either camp feeling slightly uneasy at approaching the "territory" of the other. (This tendency has been less marked in the USA, where the US Geological Survey has long held national responsibility for both sub-disciplines). Such territoriality is gradually eroding as hydrology continues to evolve into a science in its own right (a process fostered by the British Hydrological Society).

The practical hindrance to river-aquifer studies arises from the differing response times of surface waters and groundwaters: While a river might respond to an input of rainfall by fluctuations of stage over a period of minutes, significant changes in groundwater levels after recharge may take days or weeks to develop. It is for this reason, of course, that water levels are usually logged at 15-minute intervals at river gauging stations, while groundwater levels (other than during pumping tests) might be recorded at daily (or even monthly) intervals.

Until recently, this fundamental mismatch in response times has led to gross simplifications in the formulation of models for coupled river-aquifer systems. Hydraulic routing models for river systems have often reduced the groundwater components of total runoff to simple specified boundary fluxes. Similarly, regional groundwater models have often represented rivers as time-invariant specified head boundaries (thereby implicitly ignoring the complex flow patterns which develop where aquifers meet partially-penetrating, sediment-lined river channels). Models of this kind, in which either element of a river-aquifer system is reduced to a simple source or sink term (which will accept fluxes arising from detailed simulation of the other element without prior satisfaction of an internal boundary condition), are referred to as "externally coupled" river-aquifer models (Dillon, 1983). In externally-coupled models, the simulation of one element of a river-aquifer system is entirely dependent upon the prior simulation of the other element. In many hydrological circumstances, this approach can be readily justified. For instance, recharge of aquifers by infiltration from hydraulically disconnected ephemeral rivers (Alderwish and Dottridge, *this volume*) does not depend on a minute-by-minute comparison of heads in the river and the deep underlying aquifer, and therefore an externally coupled solution is perfectly reasonable.

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At the other end of the spectrum, it is possible to formulate "internally coupled" river-aquifer models, in which simultaneous solution of physically-based equations for groundwater flow and open channel flow are sought, with river-aquifer exchanges being obtained by iterative comparison and updating of the boundary fluxes until convergence is achieved. Such models are not new (the earliest published example is apparently that of Pinder and Sauer, 1971), yet because they require solution of distributed groundwater models at very short timesteps (to allow complete and stable analysis of river flows) they are expensive in terms of computer run-time and storage. Hence internally-coupled models have not been as widely used as might have been hoped.

In practice, there are a large number of mathematical models which fall somewhere between the two end-member conditions of external vs internal coupling. Many of the papers in this volume exemplify the application of such models, and illustrate the range of creative solutions which have been developed for complex practical applications.

In a review of river-aquifer modelling techniques completed eight years ago (Younger, 1987) it was concluded that water quality dynamics in river-aquifer systems were ripe for further exploration. Since then, river-aquifer models have been developed to include representations for solute transport (eg Younger *et al*, 1993; Clark *et al*, this volume) and sediment erosion and deposition (Mwaka *et al*, this volume). However geochemical processes characteristic of river-aquifer boundaries (eg abrupt changes in redox conditions) have yet to be integrated with flow and transport codes.

References

Alderwish, A., and Dottridge, J., *this volume*, Modelling Infiltration from Ephemeral Wadi Flows in the Sana'a Basin, Yemen. In Younger, P.L., (editor), **Modelling River-Aquifer Interactions**. British Hydrological Society, Occasional Paper No. 6. pp 4 - 16.

Clark, K.J., Oakes, D.B., and Wilkinson, D.E., *this volume*, Application of a Coupled River-Groundwater Model (ROSANN) to the Prediction of River Nitrate Concentrations. In Younger, P.L., (editor), **Modelling River-Aquifer Interactions**. British Hydrological Society, Occasional Paper No. 6. pp 55 - 71.

Dillon, P.J., 1983, Stream-Aquifer Interaction Models: A Review. **Institution of Engineers of Australia, Civil Engineering Transactions,** 25, 107 - 113.

Mwaka, B.M.L., O'Connell, P.E., Nalluri, C., and Younger, P.L., *this volume*, Modelling the Effects of Variable Sedimentation Rates on Stream-Aquifer Interactions. In Younger, P.L., (editor), **Modelling River-Aquifer Interactions**. British Hydrological Society, Occasional Paper No. 6. pp 72 - 85.

Pinder, G.F., and Sauer, S.P., 1971, Numerical Simulation of Flood Wave Modification due to bank storage effects. **Water Resources Research,** 7, pp 63 - 70.

Younger, P.L., 1987, Stream-aquifer interactions - a review. **NERC WRSRU Research Report** 5. Natural Environment Research Council, Water Resource Systems Research Unit, University of Newcastle Upon Tyne. 115pp.

Younger, P.L., Mackay, R., and Connorton, B.J., 1993, Streambed sediment as a barrier to groundwater pollution: Insights from fieldwork and modelling in the River Thames Basin. **Journal of the Institution of Water and Environmental Management**, 7, pp 577 - 585.

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MODELLING INFILTRATION FROM EPHEMERAL WADI FLOWS IN THE SANA'A BASIN, YEMEN

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Abstract

Recharge to groundwater from surface flows in ephemeral streams (wadis) in the Sana'a Basin, Yemen, was calculated using the streamflow routing package in MODFLOW. Models of the shallow alluvial aquifer in two wadis were constructed and calibrated in steady state, using the available data on aquifer geometry and hydraulic properties, observed piezometry, abstraction and direct recharge. The transient calibration used measured and estimated floods from the first rainy season of 1993, with adjustment of the quantities of groundwater recharge, specific yield of the alluvium and the hydraulic properties of the stream bed, until the hydrographs observed in wells during and after the rainy season were reproduced. The calculated values of infiltration were in good agreement with the results of a channel water balance model. Sensitivity analyses demonstrated that the recharge volumes are strongly influenced by flood duration, but that the flood discharge and distribution of flood intensity are of minor importance. The models will be used predictively to compute recharge over a 10 year period, from estimated wadi flows.

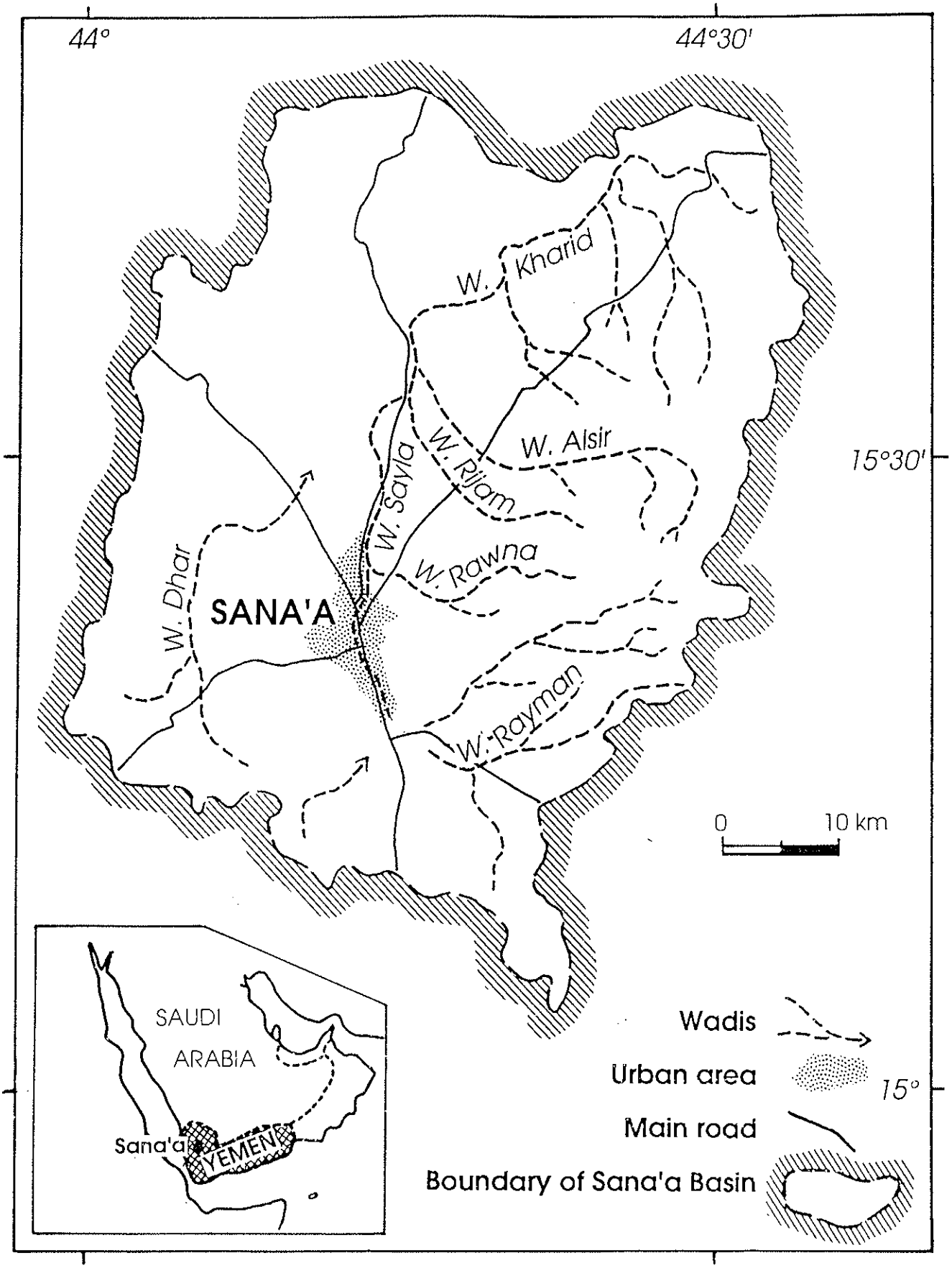
Introduction

Infiltration of surface flows through the beds of ephemeral wadis is the most important mechanism of groundwater recharge to unconfined aquifers in many arid areas. However, quantification is often problematic due to the lack of existing data and the impracticality of widespread, accurate flow gauging. It is therefore necessary to develop indirect methods for estimating recharge. An indirect approach for estimation of recharge by calibration of a groundwater model is presented here, with examples of its application to two wadis in the Sana'a Basin, Yemen (Figure 1).

Wadi recharge is controlled by flow and channel characteristics and by subsurface conditions (Rushton, 1988). The streams in the study area are normally dry, with infrequent floods of very short duration. When runoff occurs, a large proportion of the flow is absorbed into the bed and banks, causing the stream to disappear before reaching its mouth. Collection of complete and reliable data on flood hydrograph characteristics is a difficult and expensive task, but fortunately it is easier to measure two essential items:

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Figure 1. The Sana'a Basin and Wadis



(i) the maximum flow width (which controls the opportunity for infiltration in space). This is related to maximum stage by cross sectional area.

(ii) the duration of flow (representing the infiltration opportunity in time).

Where data are sufficient, all of the available information for the aquifer can be used in constructing and calibrating a groundwater model. Then modest information on the flow and channel characteristics are combined with the groundwater level measurements (over at least one distinct wet and dry season) to estimate recharge. This approach provides estimates of recharge without the need for difficult measurements of unsaturated zone characteristics. Two quantities can be calculated; the cumulative infiltration ("recharge volume", from the rise in groundwater levels), and the specific yield (using the falling limb of the groundwater hydrograph). However, the two parameters are not independent, and if the specific yield is underestimated, an apparently good calibration can be obtained by underestimating the recharge by a proportionate amount.

A similar approach has been used by Besbes et al (1987) to establish a deconvolution equation for estimating recharge in Tunisia, but without using flood measurements. The availability of the surface flow characteristics for the present study made it possible to assess the importance of flow parameters which are often uncertain. Although several previous studies (Reeder et al, 1980; Freyberg, 1983; Parissopoulos and Wheater, 1991) deal with infiltration and recharge through ephemeral wadi beds, most are either one dimensional or based on a hypothetical cross section.

The Study Area

The work presented in this paper forms part of a comprehensive investigation of recharge rates and processes in the Sana'a Basin, Yemen. The location of the area and its main features are shown in Figure 1. The Sana'a Basin is a high altitude, intermontane basin, surrounded by steep mountains, which are eroded by deeply incised wadis. The main aquifer, the Tawilah Sandstone outcrops in the central part of the basin, but elsewhere it is overlain by thick Tertiary and Quaternary Basalts, as illustrated in Figure 2. Unconsolidated alluvium covers the central plain and lower reaches of the wadis. The annual rainfall of 200 to 300 mm is concentrated in two rainy seasons, April-May and August-September, causing occasional floods in the main wadis. The water that recharges the alluvium is collected from a larger catchment than the recharge surface, thus runoff values are essential to define potential recharge. Recent measurements of soil moisture profiles (Alderwish, 1994) have shown that direct infiltration of rainfall is significant only in irrigated areas.

Two catchments with sufficient subsurface information for groundwater modelling were selected for detailed study (Figure 1):

- Wadi Dhar (Figure 3), with a catchment of 13.4 km^2, where flood flows were measured in 1993;

- Wadi Alsir (Figure 4), with a larger catchment area of 182.7 km^2, where a

rainfall-runoff model was used to estimate the volume of runoff. (The hydrographs for the first rainy season of 1993 were constructed using information about the duration of the flood from local residents and a relationship between the peak discharge and the shape of the flood hydrograph, developed by Mosgiprovodkhoz (1986) for the western wadis of Yemen).

In both valleys, the thickness of alluvial deposits varies from a few metres at the upstream end to several tens of metres at the mouth of the main wadi. Similarly, the depth to the water table also increases in the direction of flow.

Modelling of the Alluvial Aquifers

Digital models of the unconfined alluvial aquifers in Wadi Dhar and Wadi Alsir were constructed in order to quantify the groundwater recharge. The models used the USGS MODFLOW package (McDonald and Harbaugh, 1988), which is a transient, finite difference groundwater flow model, with a range of options capable of representing the essential features of the water balance and groundwater flow in the study area. The Streamflow Routing Package (Prudic, 1988) simulates the interaction between surface and groundwater and calculates the quantities of streamflow. However, it is not a true surface water flow model, but rather an accounting program that tracks the flow in one or more streams interacting with the aquifer, and limits the amount of infiltration to the water available in the stream. As required by the Streamflow Routing Package, only the sections of the alluvial aquifer that have no hydraulic connection with the underlying formation have been simulated, thus only single layered models were required.

The groundwater models were based on all the geological, hydrogeological and hydrochemical information available for the modelled area, comprising: Geophysical data from borehole logging and vertical electrical soundings; topographic, soil and geological maps; hydraulic properties of the channel, unsaturated zone and the shallow aquifer (Mosgiprovodkhoz, 1986; Selhozpromexport, 1987); a detailed well inventory of the present abstraction from the shallow aquifer (SAWAS, 1993, personal communication), and irrigation return flows estimated from simulation of the unsaturated zone (Alderwish, 1994).

#The steady state calibration focused on re-creation of the general patterns of groundwater flow as shown by the reference piezometric map of 1985 (Mosgiprovodkhoz, 1986). This is the most recent complete map, which seems to represent approximately average conditions, and indicates flow of groundwater from the sides of the wadis towards the main channel and in a downstream direction, as illustrated by Figures 3 and 4. As part of the present study, regular water level measurements started in March 1993. The depths to water from the March measurements were combined with the 1985 map to produce piezometric maps for March 1993, used for both the steady state calibration and as initial conditions for the transient modelling. The transmissivity, boundary conditions and average recharge values were adjusted until the steady state models reproduced the piezometry for March 1993.

Figure 2. Simplified Geological Map of the Sana'a Basin

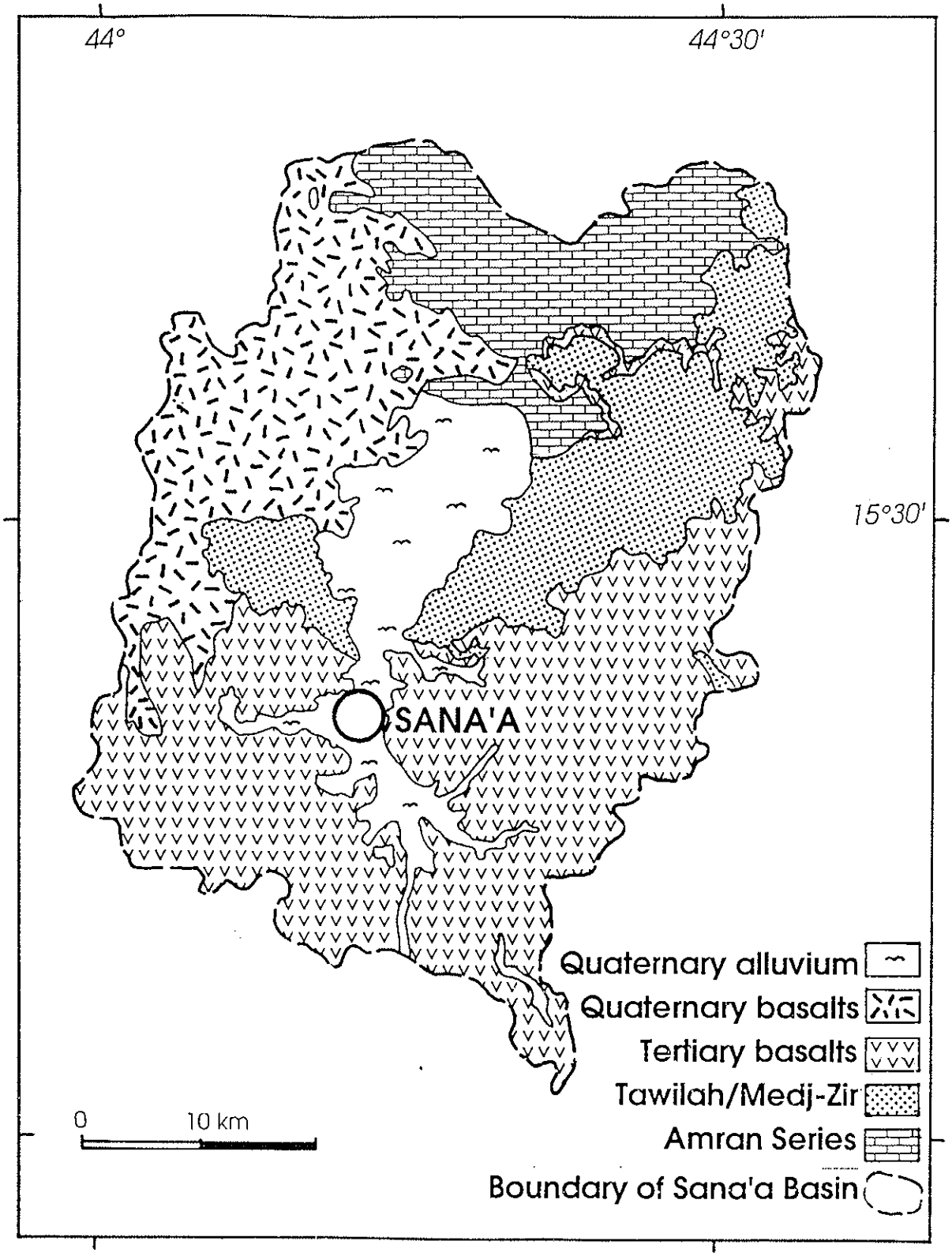


Figure 3. Detailed Model of Wadi Dhar

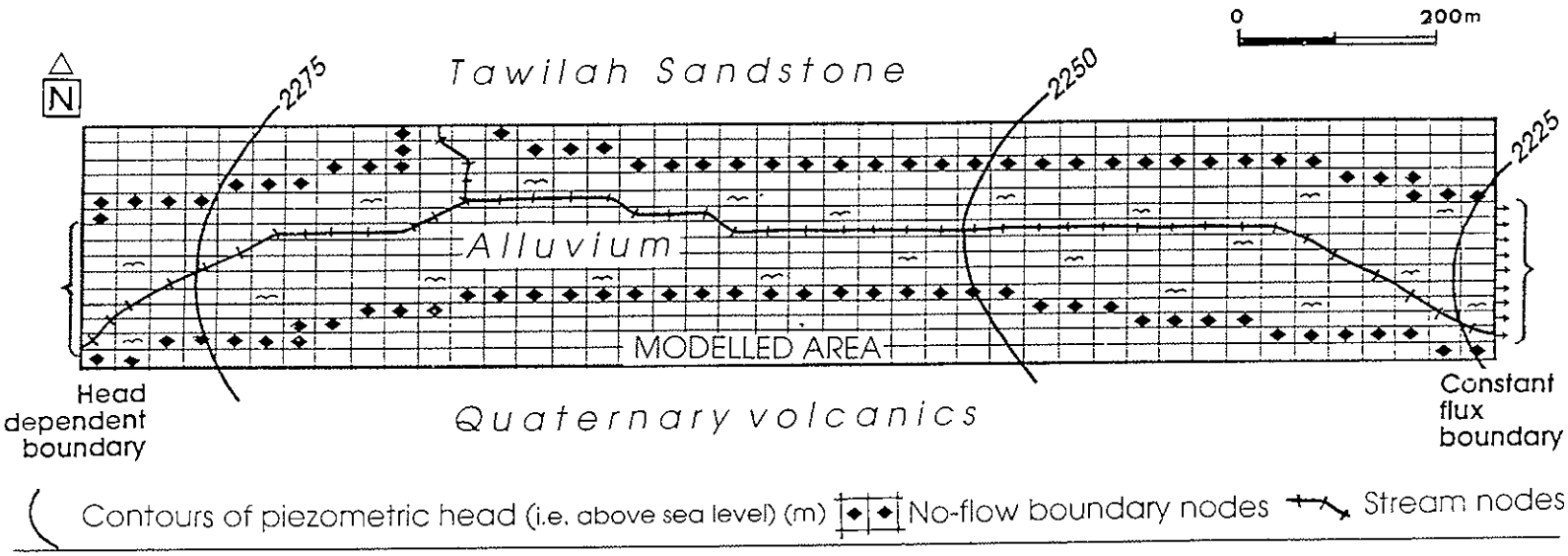
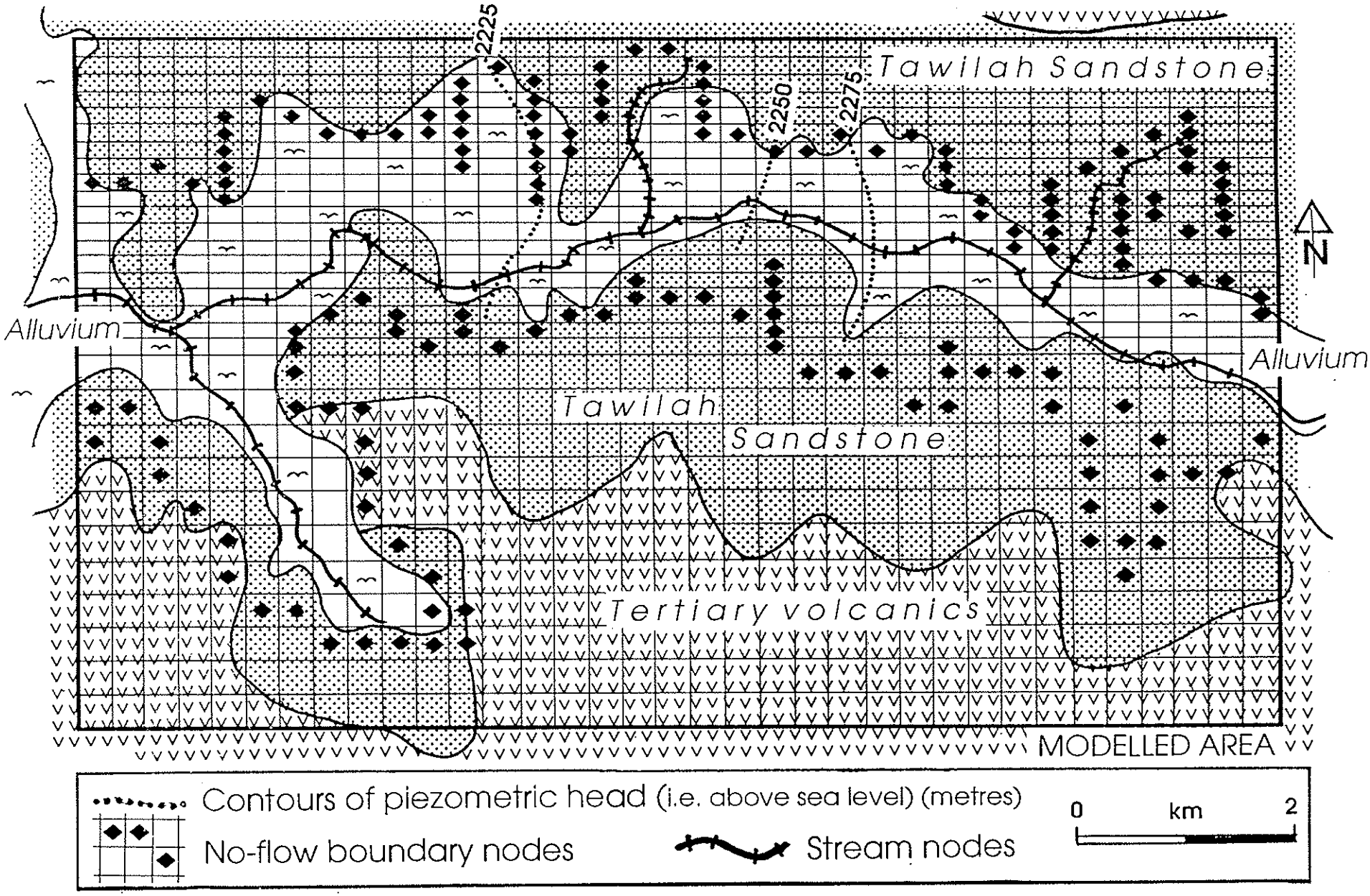


Figure 4. Detailed Model of Wadi Alsir



Estimation of recharge from transient calibration

Using the steady state calibration as the starting point, the transient calibration was performed over the period of the first rainy season (April-May), followed by a dry season with no storms until 8 August 1993, thus using both the recovery period during the rainy season and the recession during the dry season. The four floods which occurred during the first rainy season, on 30 March, 7 April, 15 April and 14 May 1993, were all simulated for both sites. The boundary conditions, aquifer thickness and hydraulic conductivity values from the steady state calibration were not changed. The values of the streambed hydraulic conductance (which represents the degree of hydraulic connectivity between stream and aquifer), and the specific yield of the alluvium, were adjusted simultaneously, by a trial and error process, until the observed water level fluctuations in the wells during and after the floods were reproduced. This approach could be very time consuming in the case of a very long series of floods.

The process of transient calibration included the following steps:

(1) estimation of the specific yield. The starting values were 8% and 6% for Wadi Dhar and Wadi Alsir respectively, based on analysis of past hydrographs (Mosgiprovodkhoz, 1986);
(2) assigning values for streambed hydraulic conductance, derived initially from calculation of the volume of water under the recharge mound;
(3) running the model in transient mode from 30 March to 8 August 1993, using measured or simulated flood data;
(4) comparison of the simulated and observed groundwater hydrographs, and repetition of steps (1) to (3) until a satisfactory simulation was achieved.

Figures 5 and 6 show the computed and observed heads at the end of the calibration. Although the agreement is good for both models, the results for Wadi Dhar are a better fit and were obtained more easily than those for Wadi Alsir. This is attributed to the fact that the MODFLOW package assumes that leakage from the stream to the aquifer is instantaneous, which is more realistic for Wadi Dhar (where the unsaturated zone is thinner) than for Wadi Alsir.

The calculated volumes of recharge for the modelled areas are shown in Table 1. The results compared well with independent estimates of recharge (also given in Table 1) from computation of the surface water balance for the modelled reaches over the same time period. This method used the same rainfall and runoff data as the groundwater models, with evapotranspiration rates derived from climatic data and measured soil moisture profiles (Alderwish, 1994).

Sensitivity Analysis

The aim of the sensitivity analysis was to assess the significance of some parameters, whose values are commonly uncertain. The analyses were based on the flood on 7 April 1993 in Wadi Dhar.

Figure 5. Measured and Simulated Water Levels, Wadi Dhar

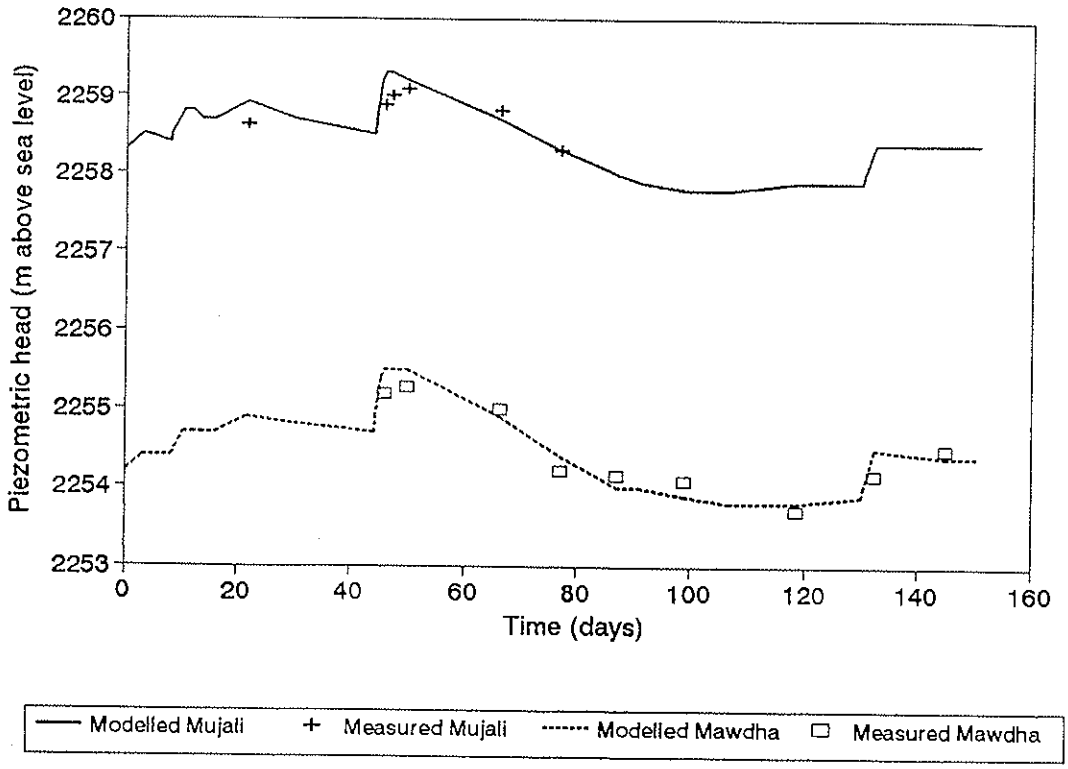


Figure 6. Measured and Simulated Water Levels, Wadi Alsir

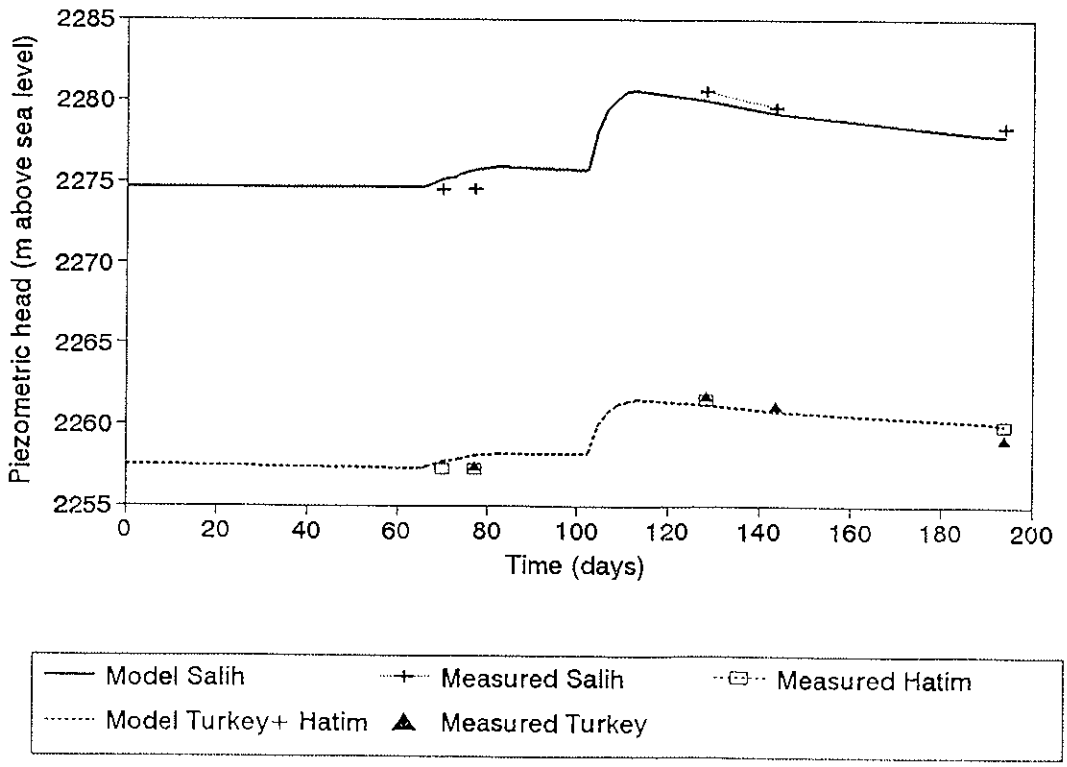
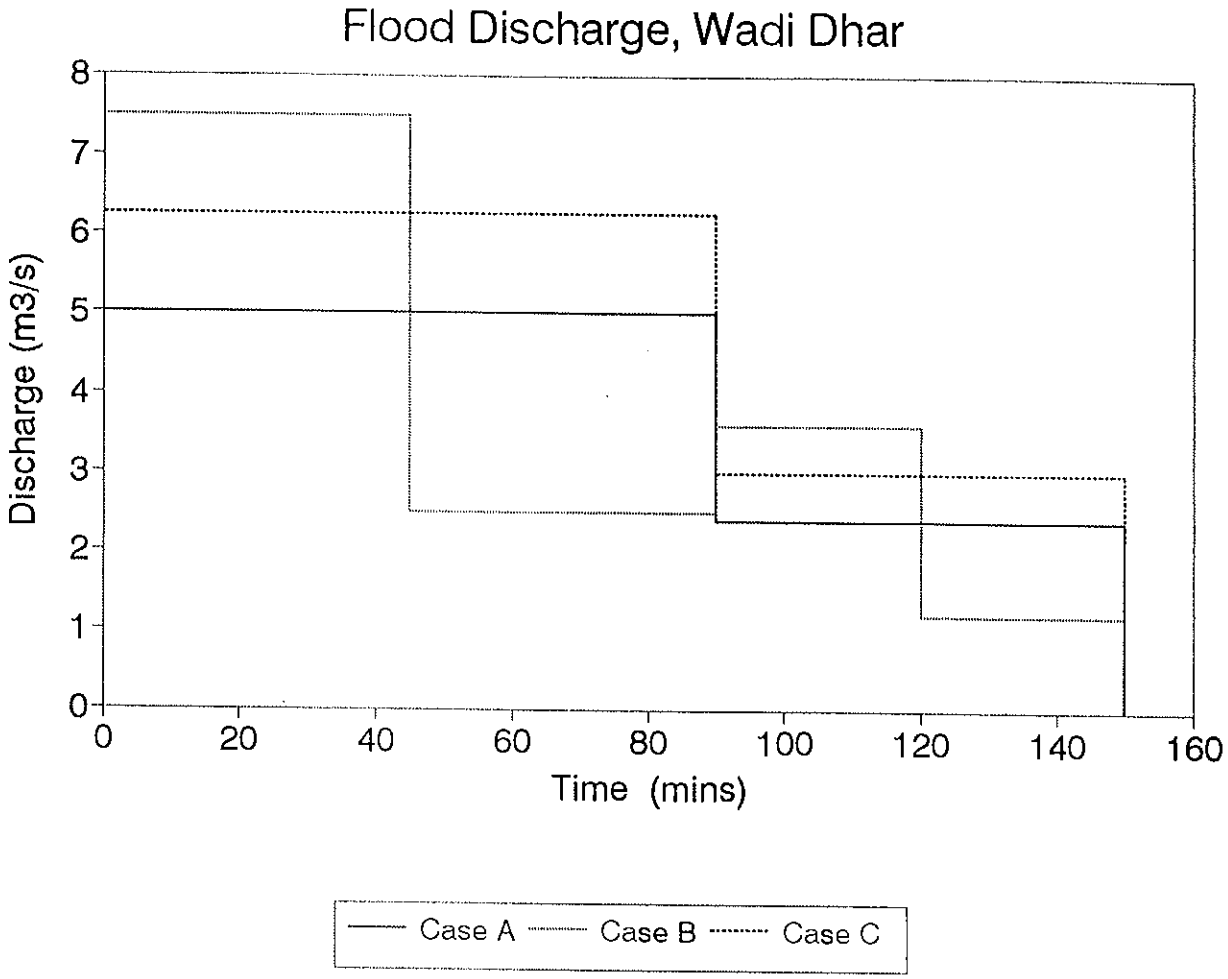


Figure 7. Variations of Flood Discharge in Wadi Dhar, used for Sensitivity Analyses



The ephemeral flows in minor streams in arid areas are characterised by a rapid rise in the stream stage and a very short duration of a few hours. Consequently, simulation of the flood requires the use of extremely short time steps in the stream-flow routing package. This is time consuming and tedious work because each stream reach has to be entered for every time period.

Therefore the first parameter tested was the temporal distribution of the flood intensity. This was done by holding the volume and the duration of the flood constant, but changing the distribution of the flood, as illustrated by Figure 7. In case A, two time intervals were used, of 90 mins with a flood discharge of 5 m^3/s, followed by 60 mins at 2.4 m^3/s. In case B, the time intervals were divided by two, so the flood discharge for each time interval became 7.5 m^3/s (for 45 mins), 2.5 m^3/s (for 45 mins), 3.6 m^3/s (for 30 mins) and 1.2 m^3/s (for 30 mins). The results, in Table 2, show a difference between the cumulative infiltration from the two cases of only 0.6%, which indicates that inaccuracies in the temporal distribution of the flood intensity do not affect the model's estimate of cumulative infiltration.

The second parameter tested was the flood discharge volume, which was estimated from observations of stream stage in surveyed cross sections using Manning's formula. One of the uncertainties of this method is the estimation of the roughness coefficient, Manning's n. For the modelling, the duration of the flood was held constant at 150 mins, and the discharges used for case A, 5 m^3/s for 90 mins and 2.4 m^3/s for 60 mins, were multiplied by a factor of 1.25 for case C, as shown in Figure 7. The difference of the recharge volume for the two cases is less than 2%, as shown in Table 2, far less than the 25% increase in the volume of flood discharge. Therefore, the cumulative infiltration can be considered to be relatively insensitive to the magnitude of flood event.

Table 1 Calculated Recharge Volumes (30.3.93 to 8.8.93)

Location	Recharge (m^3) from groundwater model	Recharge (m^3) from channel water balance
Wadi Dhar	35718	37635
Wadi Alsir	594720	548575

Table 2 Results of Sensitivity Analyses

Parameter	Multiplication Factor	Change in recharge volume	Change in heads (m)
Distribution of flood intensity	--	+ 0.6%	no change
Discharge	1.25	+ 1.7%	+ 0.1
Duration	1.25	+ 23.0%	+0.1 to + 0.2

The third test concerned the duration of the flood. An increase in the flood duration of 25% caused an increase in recharge volume of 23 %, as shown in Table 2. This agrees with the conclusions of Parissopoulos and Wheater (1991) for a hypothetical wadi, and clearly indicates the strong influence on cumulative infiltration of the infiltration opportunity in time.

Conclusions

Application of the method described above to two wadis has demonstrated that estimation of groundwater recharge from ephemeral wadis by calibration of a groundwater model can give useful results, even for areas such as the Sana'a Basin where there are few measurements of floods. A valid estimate of groundwater recharge can be obtained without a complete representation of the unsaturated zone. The accuracy of the method is confirmed by the good agreement between the values computed from the groundwater models and those from a channel water balance approach. The models will be used predictively to compute recharge over a 10 year period, using estimated wadi flows derived from the rainfall-runoff model which was applied to Wadi Alsir.

The present data set, together with transient calibration of the groundwater models, has allowed analysis of some general factors which affect recharge in ephemeral wadis in arid zones such as the Sana'a basin. The sensitivity analyses showed that the magnitude of the cumulative infiltration through the wadi beds is largely dependent on the duration of the flood, but that flood discharge and the distribution of flood intensity have only minor influence.

The method requires an extensive set of subsurface data to allow successful calibration of the initial steady state model, which forms the basis for the subsequent transient phase. The essential items for transient calibration are measured or calculated floods and groundwater hydrographs spanning both the recovery during the rainy season and the recession during the following dry season. A unique calibration would be impossible without use of the falling limb of the hydrographs to estimate the specific yield of the aquifer. The presence of a distinct rainy season, or a series of separate floods is therefore essential.

The time required for implementation of this approach depends on the number of floods, as each must be simulated individually. Due to the amount of work involved and the requirements for extensive subsurface data, this approach is not suitable in all cases. In addition, the limitations and assumptions of MODFLOW's streamflow-routing package may become significant, for example in areas with a very deep water table.

Acknowledgements

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References

Alderwish, A.M., 1994, **Quantitative aspects of groundwater recharge of Sana'a basin, Yemen.** Unpublished report for transfer to Ph.D degree, University of London.

Besbes, M., Delhomme, J.P, and De Marsily, G., 1987, Estimating recharge from Ephemeral streams in Arid regions; A case study at Kairouan, Tunisia. **Water Resources Research,** 14, 281-290.

Freyberg, D.L., 1983, Modelling the effects of a time-dependent wetted perimeter on infiltration from ephemeral channels. **Water Resources Research,** 19, 559-566.

Mosgiprovodkhoz, 1986, Sana'a Basin Water Resources Scheme Volume 1, Hydrology; Volume 2, Geology and Hydrogeology; Volume 3, Soil Zones; Volume 4, Agriculture.

McDonald, M.G., and Harbaugh, A.W., 1988, **A modular three-dimensional finite-difference ground-water flow model.** USGS open file report 83-875, Book 6.

Parissopoulos, G.A., and Wheater, H.S., 1991, Effects of wadi flood hydrograph characteristics on infiltration. **Journal of Hydrology,** 126, 247-263.

Prudic, D.E., 1989, **Documentation of a computer program to simulate stream-aquifer relations using a modular, finite-difference, ground-water flow model.** USGS Open file report 88-729.

Rushton, K.R., 1988, Numerical and conceptual model for recharge estimation in arid and semi-arid zones. In Simmers, I., (editor), **Estimation of natural groundwater recharge.** Proceedings of the NATO Advanced Research Workshop, Turkey, March 1987. Reidel Publishing Co, Part 2, 205-223.

Reeder, J.W., Freyberg, L.D., Franzini, J.B, and Remson, I., 1980, Infiltration under rapidly varying surface water depths. **Water Resources Research,** 16, 97-104.

Selhozpromexport, 1987a, Contract for design of a diversion weir and canal at Bani Hushish, Wadi Alsir. Vol 1. Tender documents. Ministry of Agriculture and Fisheries, Sana'a, Yemen.

Selhozpromexport, 1987b, Contract for design of a dam at Bayt al Jamal, Wadi Dhar. Vol 1. Tender documents. Ministry of Agriculture and Fisheries, Sana'a, Yemen.

OBSERVATIONS AND MODELLING OF INFLUENT SEEPAGE FROM THE RIVER WOLF, DEVON.

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Abstract

This paper describes observations of influent water flow through the bed and banks of the River Wolf during a controlled release of water from Roadford Reservoir in Devon. Changes in hydraulic potential were observed by monitoring five dipwells and two tensiometers located in a transect beside the river as discharge increased from 0.5 to 8.8 cumecs in three hours. Hydraulic head only increased within 15m of the river bank; however, the data illustrate the temporal sequence of influent river flow, and provide evidence of a bank storage effect. Variations in hydraulic gradient between the dipwells and tensiometers are described, and the hydraulic characteristics of two stratigraphic layers are estimated: a confining silt/clay unit and an underlying gravel aquifer of unknown thickness. These results are used to devise a two-dimensional groundwater model, using the MODFLOW package, and to demonstrate the bank storage effect during the progression of a flood hydrograph. The model is calibrated for an initial period preceding the flood release, and is tested by comparing model predictions of hydraulic head response to influent seepage with field data. The paper demonstrates the potential of field monitoring to derive an *in situ* estimate of the hydraulic conductivity of floodplain sediments.

Introduction

The exchange of water between rivers and aquifers has a significant effect upon certain hydrological processes, influencing, for example, the river regime, and temporal variations in solute concentrations. Quantification of the direction and magnitude of riverbed seepage is becoming increasingly necessary to improve the accuracy of numerical models. Temporary storage of river water within aquifers may attenuate flood hydrographs and must therefore be incorporated within hydraulic routing models (Bathurst, 1988). Furthermore, infiltration of polluted river-water may contaminate an alluvial aquifer (Heij, 1989). Nutrient flow associated with varying groundwater fluxes also has ecological implications. For example, Gibert *et al* (1990) have described how the dynamics of varying vertical fluxes of water, and their associated nutrients, provide the conditions under which distinctive aquatic fauna may develop.

Theoretically, seepage of water may be predicted easily, as the water flux is proportional to the gradient between the river level and the hydraulic head of the aquifer, mediated by the permeability and thickness of river bed and bank sediments. Water flows can therefore occur in either direction between the river and aquifer depending upon the relative difference in hydraulic head. It is also possible for the direction of seepage to vary over short time periods when either the local water table or river level increase

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independently and reverse the previous hydraulic gradient. However, although the principles of subsurface seepage can be demonstrated readily, there are considerable practical and mathematical difficulties associated with isolating the component of water-table change which arises from river seepage. This problem arises because water-table fluctuations within floodplains are determined by a combination of: the hydraulic properties of the floodplain sediments; the geometry of the valley; flow of tributary streams; river stage; and distance from the river. Furthermore, the position of the water-table reflects the interaction and balance of a combination of other hydrological processes, notably infiltration of rainfall, evapotranspiration, and irrigation.

Numerical modelling provides one means of identifying the importance of individual water fluxes. However, modelling the exchange of water between aquifers and rivers ideally requires the simultaneous solution of two separate flow equations (describing open channel flow and subsurface flow through alluvial sediments). In general, studies of bank storage have required simplifying assumptions to represent the hydrogeological situation, some of which may not be justified (Sharp, 1977). For example, many models assume a homogeneous aquifer, while natural variations in floodplains such as palaeochannels, may produce local variations in subsurface flow (Sophocleous *et al.*, 1988). It is also difficult to determine the precise nature of river-aquifer interaction and the extent of river penetration of an aquifer. Analytical solutions for bank storage were obtained by Cooper and Rorabaugh (1963). More flexible numerical solutions were pioneered by Pinder and Sauer (1971) who produced a code to simultaneously solve the groundwater and open channel flow equations. Hunt (1990) avoided the difficulties in simultaneously solving the two equations by first solving the equation for open channel flow, and then using the modelled values of river level to solve the equation for groundwater flow by a perturbation method.

The sedimentological and hydraulic characteristics of most river beds and banks differ significantly and consequently in some instances it may be reasonable to assume that seepage only occurs through the river bed, and not over the whole wetted perimeter. This assumption considerably simplifies the mathematical treatment of aquifer-river interaction, and is used in groundwater models such as MODFLOW (Prudic, 1989).

While field observations of an association of fluctuating water-tables with river levels can provide confirmation of influent and effluent river flow (eg Hurr, 1983), it is frequently difficult to attribute water-table fluctuations specifically to river seepage. This paper describes preliminary results from a study of influent river flow during a release of water from Roadford Reservoir in Devon. This event enabled the water-table fluctuation arising from river seepage to be identified, and a numerical model (calibrated using the data obtained) indicates the varying quantity of water stored within the river bank.

Field investigation

The flow of the River Wolf has been regulated since construction of Roadford Reservoir construction in 1989, and currently has a compensation flow of 9Ml/day. The reservoir has a maximum surface area of 3km^2, with a catchment area of 31km^2. Occasional

Figure 1. **Cross-section showing the location of the five dipwells (D1-D5) and two recording tensiometers (T1 and T2). Initial and maximum river levels are indicated (S_{min} and S_{max}).**

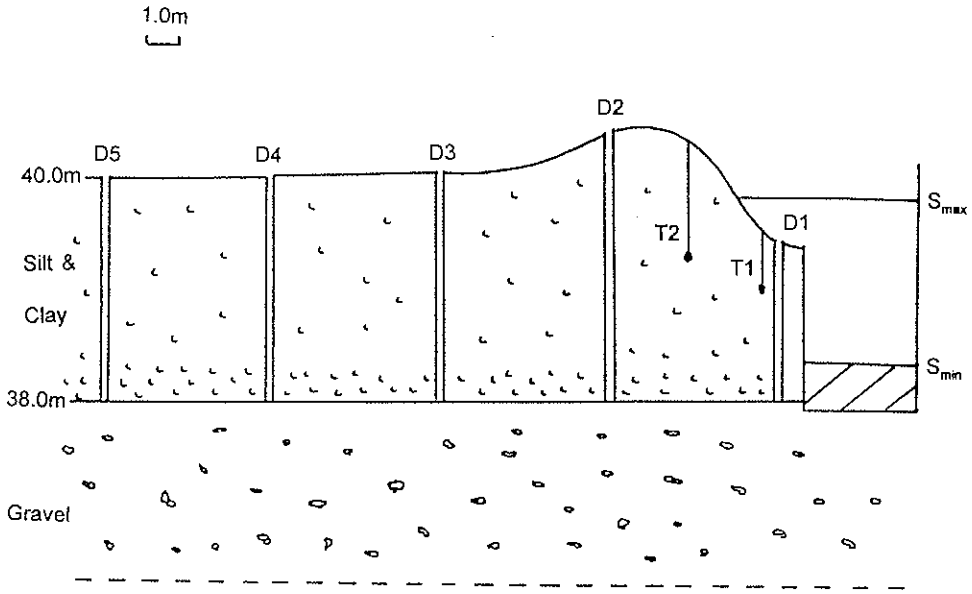


Figure 2. **Time-series of the variation of dipwells 1 and 2 in relation to river stage on 3 November 1994.**

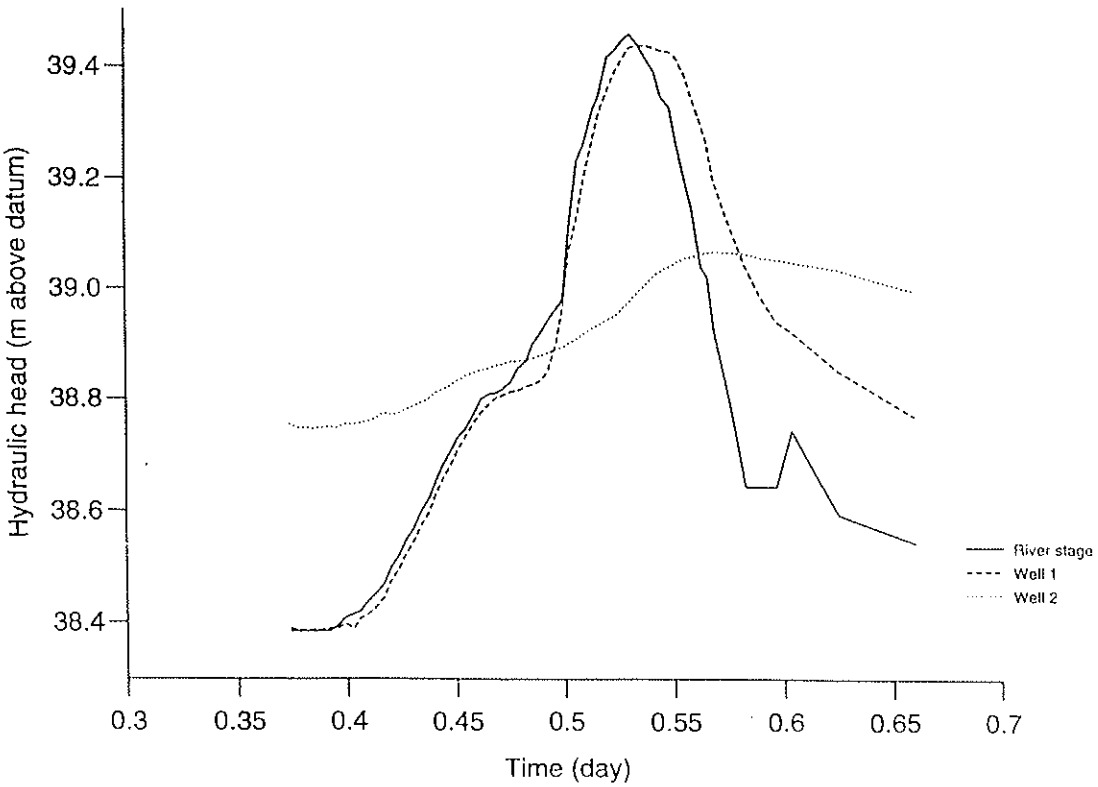
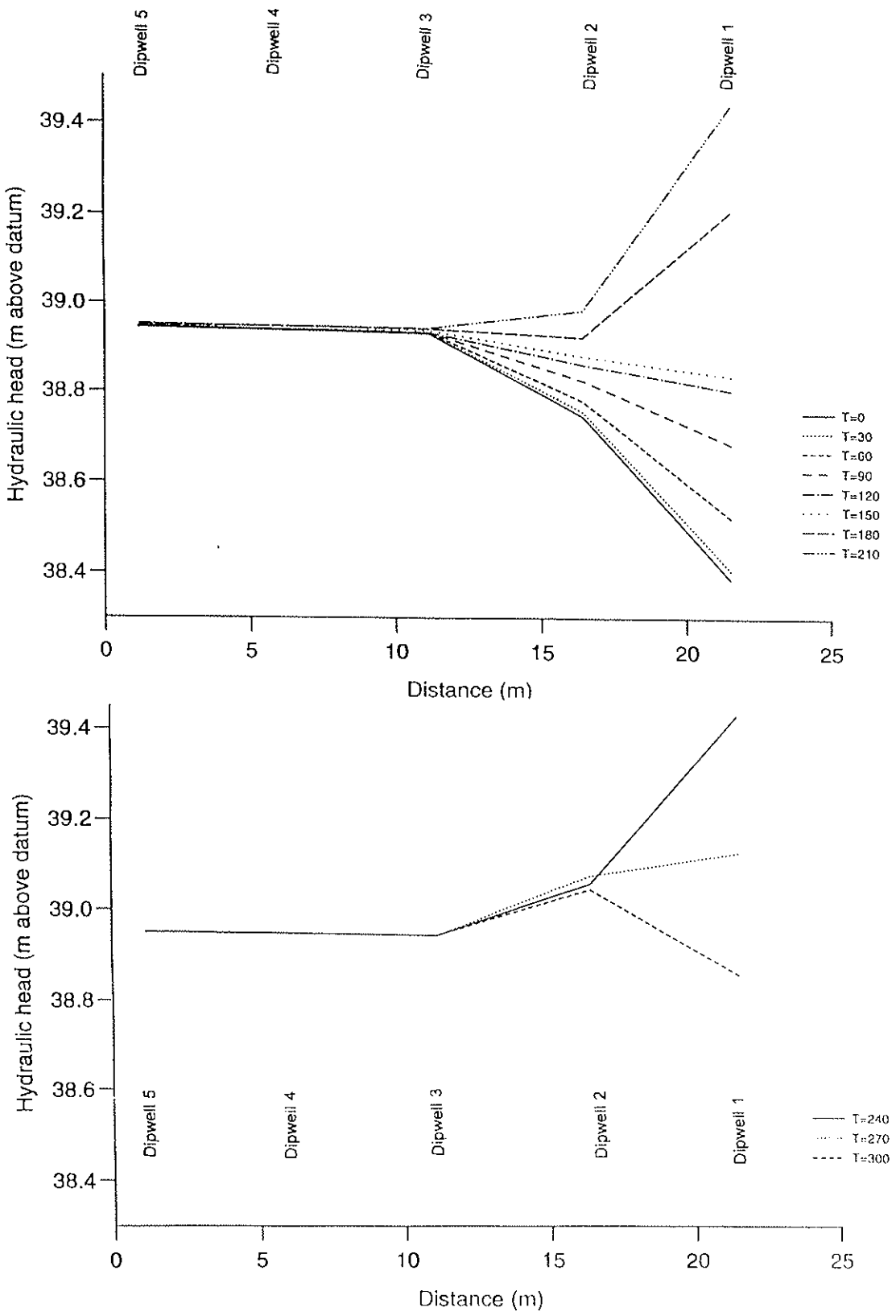


Figure 3. **Cross-sections at 30-minute intervals illustrating the development and initial decay of a water-table ridge through bank storage.**



releases of water from the dam are intended to mobilise bed sediments and prevent the sedimentation of fine deposits (which appears to have increased as a result of the modified hydrological regime after dam construction). The catchment is underlain by largely impermeable Palaeozoic mudstones, shales and sandstones. Within the valley localised gravel deposits of varying thickness are in close contact with the river.

Influent seepage from the River Wolf was monitored during a controlled release of water from Roadford Reservoir on 3 November 1994. Measurements were undertaken at a location approximately 3km downstream of the reservoir dam which had previously been the site of channel substrate surveys (Petts, unpublished). At this point the floodplain is approximately 50 m wide, while the channel itself is 3m wide. Five dipwells, numbered one to five, consisting of 2m lengths of plastic tubing, 4cm outside diameter, and slotted over the basal 25cm were installed by hand augering in a transect at 90° to the river channel at intervals of 5m (Figure 1). The dipwells all rested directly on gravel deposits (which was at least 70cm thick at this point). Overlying the gravel was a layer of heavy clay, with medium silt/clay above. Two tensiometers were positioned near the river, the first (T1) at a depth of 40cm on a small terrace in the river bank, and the second (T2) at a depth of 80cm, 2 metres from T1. The tensiometers were installed by hand auger, and were sealed from the top using pellets of bentonite clay. Variations in positive and negative pressures were measured using two pressure transducers with the output recorded by data loggers. Measurements of river stage was obtained by reading water levels adjacent to dipwell one. The relative heights of the tops of the dipwells and tensiometers were obtained by levelling relative to a temporary bench mark.

The reservoir release on 3 November produced a steady increase in discharge downstream, from an initial 0.5 cumecs at 08:15, to a peak flow of 8.8 cumecs at 12:15. The flow from the reservoir then decreased slowly until reaching the initial starting discharge at 01:30 the following day. Variations in the hydraulic head observed at individual dipwells were read manually using an electrical contact device at 5 minute intervals from 08:30 until 17:00 when measurements ceased. The results are shown in Figure 2 and 3 which illustrate in time-series form, and as cross-sections, the progressive change in hydraulic head. At the measurement point, river stage increased by just over 1 metre, beginning to rise at 09:30 (0.4 day) and increasing at a constant rate of 20 cm per hour until 11:15 (0.47 day), when the rate of increase slowed until 11:45 before increasing at a rate of 20 cm per 10 minutes to peak at 12:45 (0.53 day). Dipwell 1 initially varied in close association with the rising river level, however, on the falling limb the dipwell response was lagged due to the slower drainage of water from the dipwell. The response of dipwell 2 was strongly attenuated; here the hydraulic head increased by 30cm during the event producing a peak level at 13:30 (0.56 day) when the river level had begun to fall.

The cross-sections of water-tables shown in Figure 3 reveal how the water-table response to influent seepage varies with distance from the river. Only dipwells 1, 2 and to a lesser extent dipwell 3, responded to influent seepage thereby limiting the area of response to within 15m of the river bank. Hydraulic head near the river increased steadily until 210 minutes after measurements began, and as the river fell a water-table ridge developed centred on dipwell 2 which is most evident at 300 minutes. The ridge

then slowly decreased in height as water drained in two directions both towards and away from the river.

The tensiometer results, corrected for elevation head and atmospheric pressure, are given in Figure 4, together with river stage. There are clear differences in the timing and magnitude of the response of tensiometers T1 and T2. The former was evidently in close contact with the river, and although initially it recorded a higher hydraulic head illustrating effluent conditions, soil water pressures increased together with river stage, and by 12:00 (0.5 day) T1 was acting as a simple piezometer directly recording the change in river level. In contrast, the soil pressures recorded by T2, 2m from the river bank, display differences in the timing and magnitude of the response to infiltration. The initial readings indicate effluent seepage, however, pressures only increased 30 minutes after the river stage exceeded the head at T2, whereupon the soil pressure continued to increase steadily until after river stage had begun to fall.

These field data clearly illustrate the way in which near-bank hydraulic gradients change during the rise and fall of river level. The observations provide the basis for the modelling simulations described below.

Numerical Modelling

A transient two-dimensional profile model was devised to investigate variations in vertical water flow between the river and the aquifer, and also the relationship of hydraulic heads within the gravel aquifer and overlying layer of silt and clay. Given the limited amount of data available, the model simulation sought to replicate the observed variation in hydraulic heads with distance, and thus to consider what hydrological mechanisms might be responsible for the differences in water-table response. The approach assumes that groundwater flow parallel to the river axis is not significant, and consequently that all flow occurs parallel to the plane of the profile. The USGS groundwater model MODFLOW was used for the simulations; this is a quasi three-dimensional model which adopts a finite difference method to solve the equations governing head distribution (McDonald and Harbaugh, 1988).

The model area, which is shown schematically in Figure 1, was subdivided into 26 cells, each having a uniform length of 1.0m. The stratigraphy was represented in the model by simply distinguishing between an unconfined surface layer of silt and clay, and an underlying, confined, gravel aquifer, the top elevation of which could be obtained from the dipwell heights but the bottom elevation was unknown. Although MODFLOW includes packages which are specifically designed to simulate river and stream flow, these are not applicable to profile models, and therefore for this study the river was represented by introducing a General Head Boundary at cell no. 1, which acted as a constant head within the gravel of Layer 2. This constant head could either comprise a source of water when higher than the surrounding water-table; or an area of water-table drainage when below the local water-table. The cell in Layer 1 which was situated above the river cell was defined as an active cell, so that theoretically seepage could occur laterally in Layer 1, and not simply by transmission through the deposits of Layer 2. A second constant head was designated at cell no. 26 in Layer 2. This was necessary to maintain the lateral height of the water-table to conform with the field data.

Model simulations were undertaken using minutes as the unit of time, and data such as hydraulic conductivity were specified accordingly. A summary of the data required, and an indication of how the values were obtained, is given in Table 1. In Layer 1, yield was estimated as 0.1, while the hydraulic conductivity was obtained using the tensiometer data, and in particular the lagged response of T2 to the increase in river level. The Vertical Conductance parameter is a measure of the vertical hydraulic conductivity between Layer 1 and Layer 2. It is proportional to the distance between the two layers. For this simulation a low value of $5x10^{-4}$m was chosen to represent the effect of a layer of heavy clay directly above the gravel. The transmissivity of Layer 2 was obtained mainly by the calibration process described below, while the specific yield of Layer 2 was estimated from sedimentological considerations.

Two calibration periods were used in development of the final model; for the first period the model was run over a period of 1440 minutes to ensure that the initial distribution of hydraulic heads observed at the dipwells could be replicated. For the second period, the response of the model to a sequence of 22 stress periods (each of 15 minutes duration) was determined. During these stress periods the constant head of cell no. 1 was changed in proportion to the observed fluctuation of river level. At this time the transmissivity of the gravels in Layer 1 was varied so as to reproduce the timing of influent seepage indicated by the dipwell response. Between dipwells 4 and 5 a value of 0.9 m^2/min was used which decreased to $2.5x10^{-2}$ m^2/min near the river. The results for hydraulic heads in Layer 2 are shown in Figure 5 and demonstrate the degree to which the model is capable of recreating the bank storage effect.

Discussion

While the field data summarised above provide clear evidence of variations in the rate of water infiltration through the river bed and banks, the success of the numerical model in simulating the fluctuations in water-table is not surprising as this time period also represents the calibration period for the model. Consequently, the model is mainly useful in suggesting the importance of possible mechanisms of water flow, and in particular the relationship between Layers 1 and 2. The main assumption of the model is that the gravels of Layer 2 are effectively confined by overlying silt/clay deposits in Layer 1, and therefore the water-table distribution cannot be represented using the Dupuit-Forchheimer approximation. Therefore, in this case it was necessary to vary the transmissivity of the gravels to reproduce both the initial spatial distribution of hydraulic heads, and also differences in the timing of response to river-water infiltration. Lateral variations in transmissivity may represent either differences in the depth of the gravel deposits or variation in hydraulic conductivity. The decrease in transmissivity near the river may be explained if the river has incised into the gravel layer thereby locally reducing the thickness of gravels.

This interpretation of the hydraulic properties of Layer 2 provides one possible mechanism for the observations of a localised bank storage effect and the development of a groundwater ridge. However, it still remains to explain why dipwells 4 and 5 recorded no variations in hydraulic head if transmissivity was greatest between these two points. Although no significant differences in the silt/clay stratigraphy were recorded between these dipwells, slight variations in the surface topography adjacent to dipwell

Figure 4. **Hydraulic head recorded by Tensiometer 1 and 2 and river stage.**

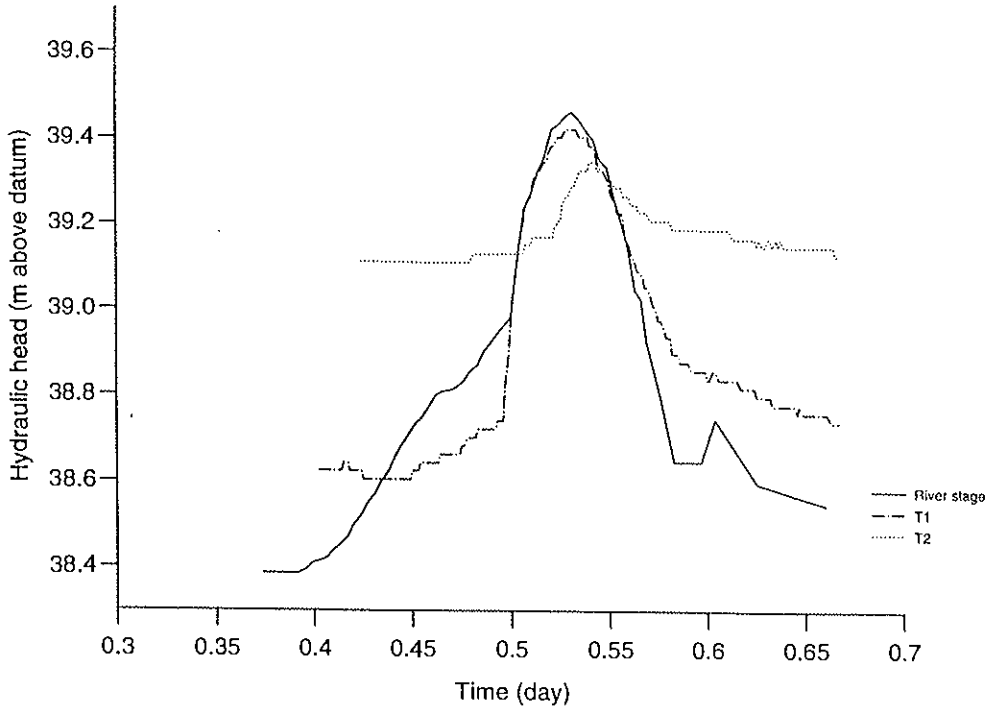
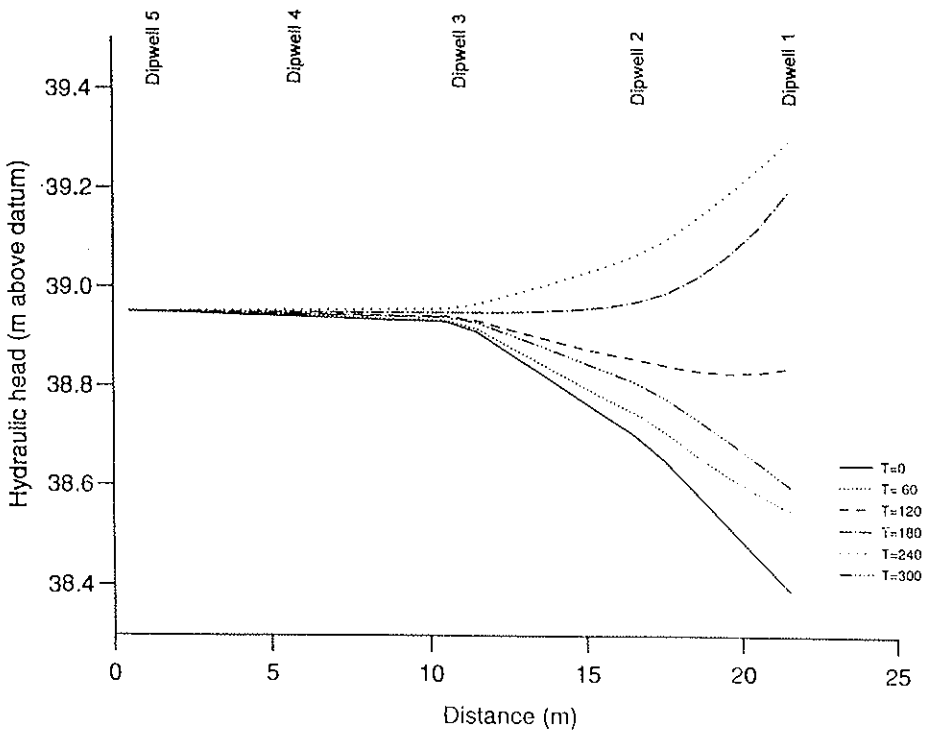


Figure 5. **The simulated response of water-tables to the fluctuation in river stage, shown at a 60 minute time interval.**



5 might represent the site of a flood channel or palaeochannel. The variation in permeability associated with this feature may enable the rapid transmission of increases in hydraulic head, such that the hydraulic head essentially remains constant.

The development of a numerical model simulating bank storage illustrates the possibility of estimating hydraulic properties by a combined monitoring and modelling approach. However, as the notes in Table 1 indicate, the model is still dependent upon estimated values for the yield of Layer 1, and the vertical conductance term, and therefore the modelling results are largely speculative. For example, it would be interesting to see if a second model could be developed by changing the vertical conductivity between layers 1 and 2, and considering the gravel layer to have constant transmissivity. In this case the configuration of the water-table would be envisaged as reflecting the Dupuit-Forchheimer theory for free-surface flow. Further development of the model is therefore necessary to investigate whether the simplifying assumptions are justified.

Conclusion

This paper demonstrates the possibilities of monitoring changes in subsurface seepage from rivers under controlled conditions, of which a reservoir release is an ideal example. Although the scale of the study is limited, the results indicate the potential of simple numerical models for investigating the mechanisms of water flow. This study also indicates the data requirements for developing a more complete model, which can then be used to investigate variations in influent and effluent flows under a variety of different conditions.

Table 1. Summary of the hydrologic parameters used to describe the model cross-section.

Parameter	Value (/min)	Value (/s)	Notes
Layer One - Unconfined aquifer of silt and clay.			
Hydraulic conductivity	0.067 m/min	1.12×10^{-3} m/s	Obtained from tensiometer response.
Storativity	0.1		Estimated.
Vertical conductance	5.0×10^{-4}	8.3×10^{-6}	Equivalent to K of a heavy clay.
Layer Two - Confined aquifer of gravel.			
Transmissivity	$2.5 \times 10^{-2} \rightarrow 4 \times 10^{-2} \rightarrow 0.9$ m^2/min	$4.2 \times 10^{-4} \rightarrow 6.7 \times 10^{-4} \rightarrow 0.012$ m^2/s	Obtained from dipwell response.
Storativity	0.02		Estimated.

References

Bathurst, J.C. 1988. Flow processes and data provision for channel flow models. In: **Modelling Geomorphological Systems**, M.G. Anderson (ed.), pp. 127-152. Wiley: Chichester.

Cooper, H.H., Jr., and Rorabaugh, M.I., 1963, Groundwater movements and bank storage due to flood stages in a surface streams. **USGS Water Supply Paper**, 1536J, pp. 343-366.

Gibert, J., Dole-Olivier, M.J., Marmonier, P., and Vervier, P., 1990, Surface water-groundwater ecotones. In **The ecology and management of aquatic-terrestrial ecotones**, R.J. Naiman and H. Décamps (eds), Ch. 10, pp. 199-225.

Heij, G.J., 1989, River-groundwater relationships in the lower parts of the Netherlands. **Journal of Hydrology**, Vol. 108, pp. 35-62.

Hunt, B., 1990, An approximation for the Bank Storage effect. **Water Resources Research**, Vol. 26, No. 11, pp. 2769-2775.

Hurr, R.T., 1983, Ground - Water hydrology of the Mormon Island Crane Meadows Wildlife area near Grand Island, Hall County, Nebraska. **USGS Professional Paper**, 1277H.

McDonald, M.G., and Harbaugh, A.W., 1988, **A modular three-dimensional finite-difference ground-water flow model.** USGS open file report 83-875, Book 6.

Petts, G.E., Unpublished, Roadford Environmental Investigations: sediment sampling at fish spawning grounds in the Upper Tamar catchment. Phase 7 (1993). University of Birmingham, Environmental Research and Management.

Pinder, G.F., and Sauer, S.P., 1971, Numerical simulation of flood wave modification due to bank storage effects. **Water Resources Research**, Vol. 7, No. 1, pp. 63-70.

Prudic, D.E., 1989, **Documentation of a computer program to simulate stream-aquifer relations using a modular, finite-difference, ground-water flow model.** USGS Open file report 88-729.

Sharp, J.M. Jr., 1977, Limitations of bank-storage model assumptions. **Journal of Hydrology**, Vol. 35, pp. 31-47.

Sophocleous, M., Townsend, M.A., Vogler, L.D., Mclain, T.J., Marks, E.T., and Coble, G.R., 1988. Experimental studies in stream-aquifer interaction along the Arkansas River in Central Kansas - field testing and analysis. **Journal of Hydrology**, Vol. 98, pp. 249-273.

RIVER-AQUIFER EXCHANGE IN THE SHE CATCHMENT MODELLING SYSTEM

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Abstract

Tracer studies show that subsurface flow from a hillslope can contribute significantly to fast response event flow in a channel, as well as to the longer term baseflow. Application of the Système Hydrologique Européen (SHE) catchment modelling system to a snowmelt event in a 40.4-ha subcatchment of the Reynolds Creek catchment in Idaho shows that it can model such a contribution on the basis of a rapid near-channel subsurface response. Application of SHE at the full 234-km^2 catchment scale demonstrates an ability to model contrasting partitions of surface and subsurface channel supply mechanisms on a spatially distributed basis within a catchment. However, the use of large model grid squares (to represent spatial variability) reduces the accuracy with which near-channel responses can be simulated.

Introduction

River-aquifer exchange provides the mechanism by which water, and accompanying solutes, are transferred between the subsurface hillslope and the channel components of a catchment. Its representation is therefore an important element in the simulation of catchment response by physically-based distributed models. This paper describes how the exchange is represented in the Système Hydrologique Européen (SHE) modelling system and illustrates its application at different spatial scales for an observed snowmelt event. The exchange is considered from the point of view of bulk water routing, without taking account of the needs of solute flow modelling.

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River-Aquifer Exchange in Catchment Response

Subsurface flow from an unconfined hillslope aquifer is recognized as having the potential to contribute significantly to fast response event flow, or stormflow, as well as to longer term baseflow. In particular, tracer studies have shown that a significant proportion of stormflow is composed of "old" water(resident in the soil matrix before a storm event), as opposed to "new" water (delivered to the soil during the event), in all but the biggest storms (eg Sklash and Farvolden, 1979; Sklash, 1990; Bonell, Pearce and Stewart, 1990; Ogunkoya and Jenkins, 1991). The exact mechanisms and pathways by which significant quantities of "old" water are forced into the channel during an event are still under investigation. However, current thinking suggests the following model.

At the hillslope base near the channel, soil moisture content is relatively high and a "tension saturated" capillary fringe is assumed to characterize the soil layers just above the phreatic surface. In this zone the moisture content is at or close to saturation but is still under a negative pressure, or tension. A small amount of percolating water can then convert this into a zone of positive pressure (pressure saturated), causing a rapid rise in phreatic surface and the development of a groundwater ridge along the bank (eg Ragan, 1968; Sklash and Farvolden, 1979; Blowes and Gillham, 1988; Sklash, 1990). The resulting increase in the hydraulic gradient to the channel causes displacement of the "old" water and its discharge into the stream.

The effect, and its disappearance, can be produced very rapidly when the capillary fringe already intersects with the ground surface (eg Gillham, 1984). In such cases complete saturation of the soil can also occur: groundwater in affected areas more distant from the stream then issues at the surface in a seepage face and reaches the channel as saturation overland flow (eg Blowes and Gillham, 1988). The area of saturation varies with the severity of the event.

Away from the channel, where the phreatic surface is likely to be deeper, the transmission of water to the saturated zone is much slower. This produces a longer term response, contributing to channel baseflow.

Système Hydrologique Européen (SHE)

SHE is a physically-based distributed hydrological modelling system, applicable at the catchment scale and providing a full integration of the surface and subsurface responses, on the assumption of an unconfined aquifer lying above an impermeable bed (Abbott, Bathurst, Cunge, O'Connell and Rasmussen, 1986a,b) (Figure 1). Spatial distribution of catchment properties, rainfall input and hydrological response is achieved in the horizontal through representation of the basin on an orthogonal grid network and in the vertical by a column of horizontal layers at each grid square. Each of the major hydrological processes of water movement is modelled either by finite difference representations of the partial differential equations of mass, momentum and energy conservation or by empirical equations derived from independent experimental research. Of relevance to this paper, overland flow is modelled in two dimensions by the Saint Venant equations, channel flow is modelled in one dimension also by the

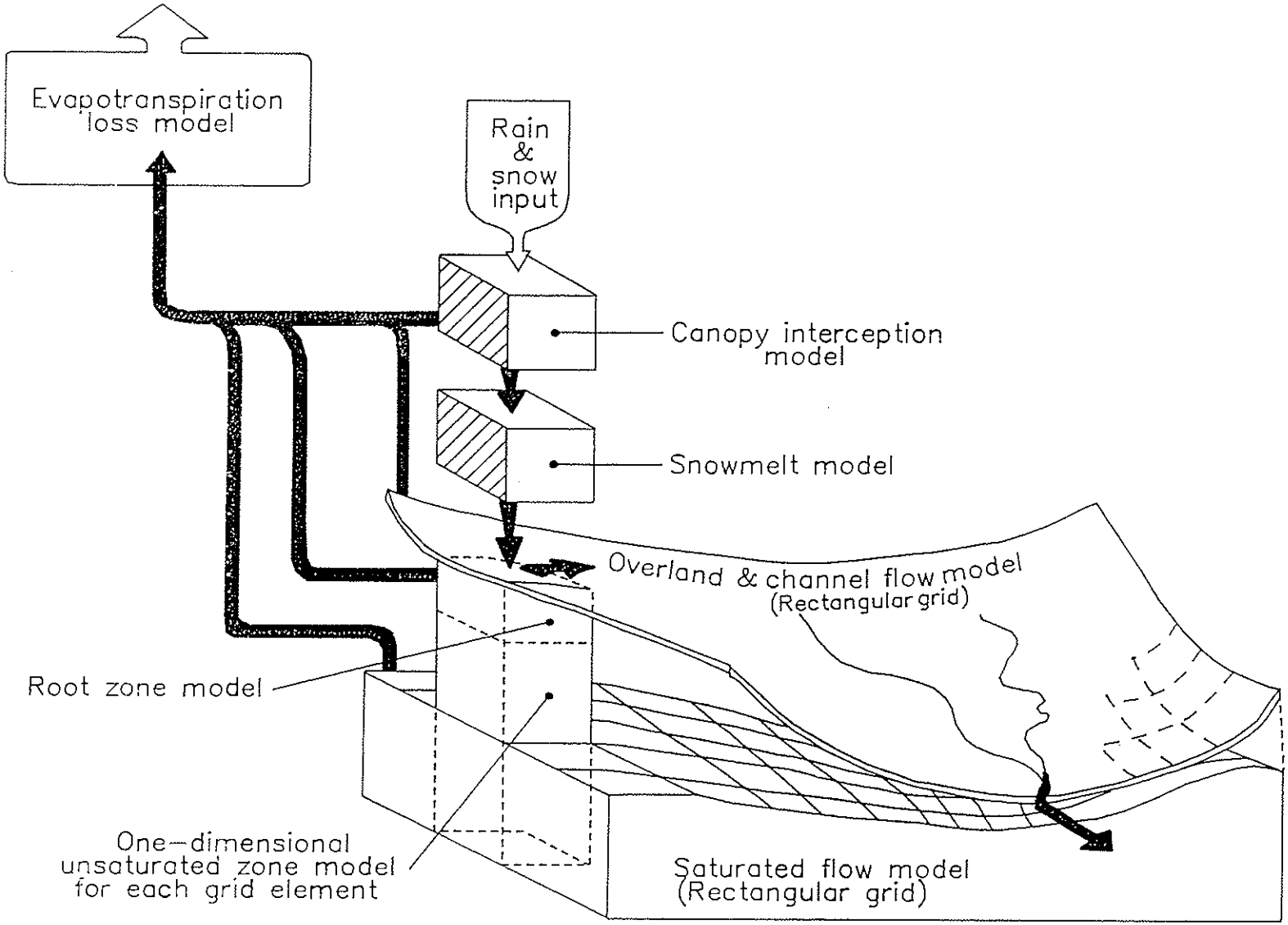


Fig. 1 Schematic diagram of the Système Hydrologique Européen (SHE) catchment modelling system

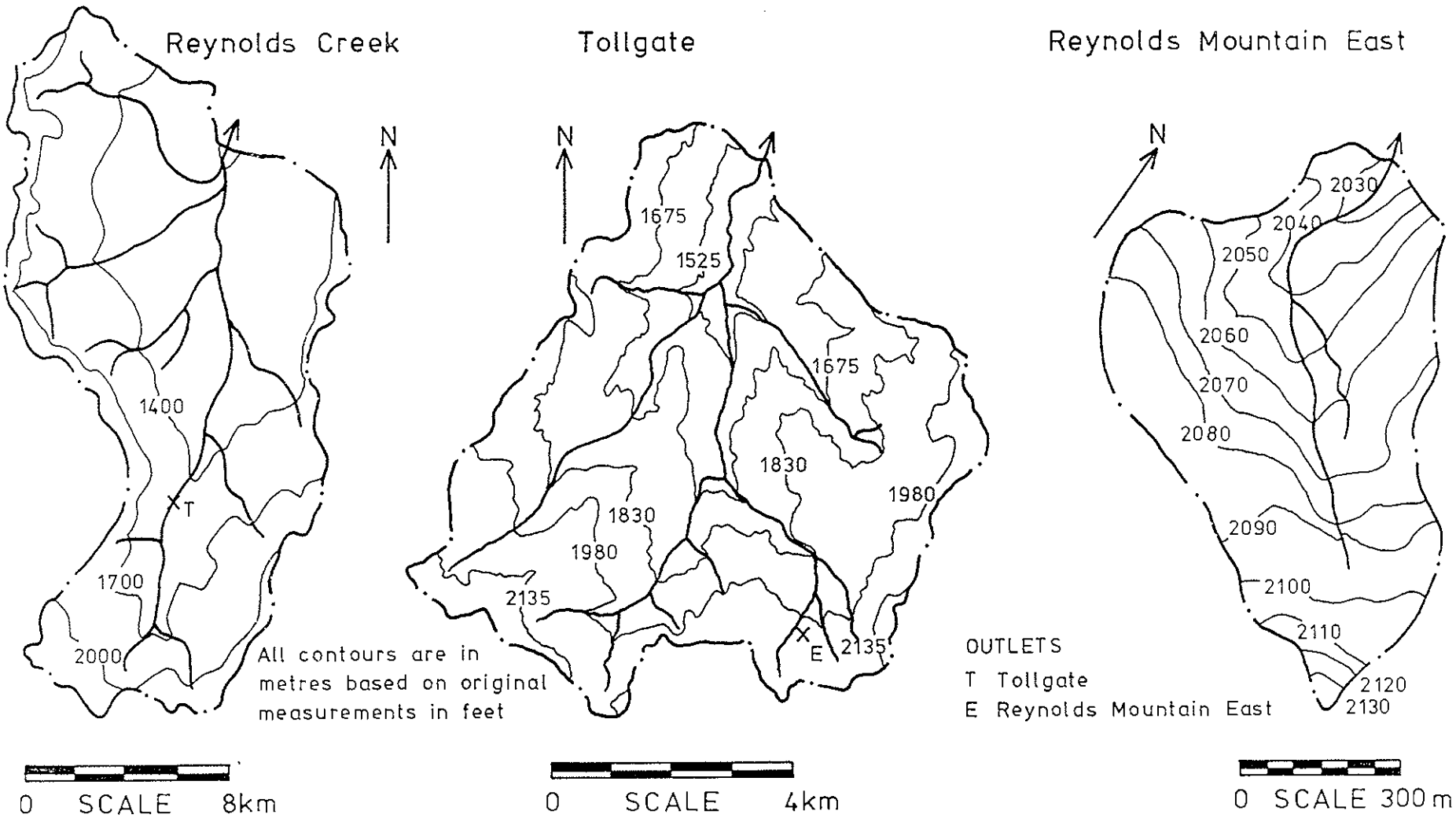


Fig. 2 Topographic maps of Reynolds Creek catchment and its Tollgate and Reynolds Mountain East subcatchments.

Saint Venant equations, unsaturated zone flow is modelled in one dimension (vertical) by the Richards equation and saturated zone flow is modelled in two dimensions (horizontal) by the Boussinesq equation. Water is therefore routed to the channel as either overland or saturated zone flow. Generation of overland flow is modelled by precipitation or snowmelt exceeding infiltration capacity and by upward saturation of the soil.

River-aquifer exchange is modelled as a function of the difference between the near channel phreatic surface elevation and the stream surface elevation, with a correction to allow for vertical saturated zone flow near the channel. The solution also allows for a channel which only partially penetrates the aquifer, for the phreatic surface to be disconnected from the channel and for a low permeability lining around the channel.

The SHE equations provide a description of small scale physics but are required to apply at the scale of a model grid square, which may vary typically from 50 m to 1 km. In order to account for the overall response at the grid scale, parameters are given "effective" values. It is recognized that this concept may not allow an accurate representation of the observed response in all circumstances (eg Binley, Beven and Elgy, 1989) but effective parameters may often represent the only practical approach given current computational and data provision constraints.

Reynolds Creek Application

SHE was applied to Reynolds Creek catchment in southwest Idaho, USA, in a joint research exercise between the Water Resource Systems Research Unit, University of Newcastle, and the Northwest Watershed Research Center of the United States Department of Agriculture, Agricultural Research Service. Reynolds Creek is a rangeland research catchment for which the scientific management is provided by the NWWRC. The overall aim of the simulation was to test the snowmelt component at the spatial scales of the 40.4-ha Reynolds Mountain East headwater catchment, the 54.5-km^2 Tollgate subcatchment and the full 234-km^2 Reynolds Creek catchment (Figures 2 and 3). Some of the results are presented in Bathurst and Cooley (in press): here only the river-aquifer exchange results are described.

Simulations were carried out for a 12½-day period in February 1982, when significant snowmelting was observed throughout the catchment. Figure 4 summarizes the meteorological conditions and shows the main outlet hydrograph. At the lower elevations the most marked, and a very flashy, response occurred during 14-16 February as a result of rain combined with snowmelt caused by raised temperatures (Fig. 4). At the higher elevations (eg Reynolds Mountain East) the most marked response occurred during 20-22 February, largely as a result of snowmelt caused by raised temperatures and turbulent transfer of sensible heat from the atmosphere (Figure 5).

Reynolds Mountain East

The subcatchment was represented by 161 squares of dimensions 50 m x 50 m (Fig. 3). Bathurst and Cooley (in press) examined four hypotheses of catchment response and concluded that, at least for the simulation period, the observed discharge record at the

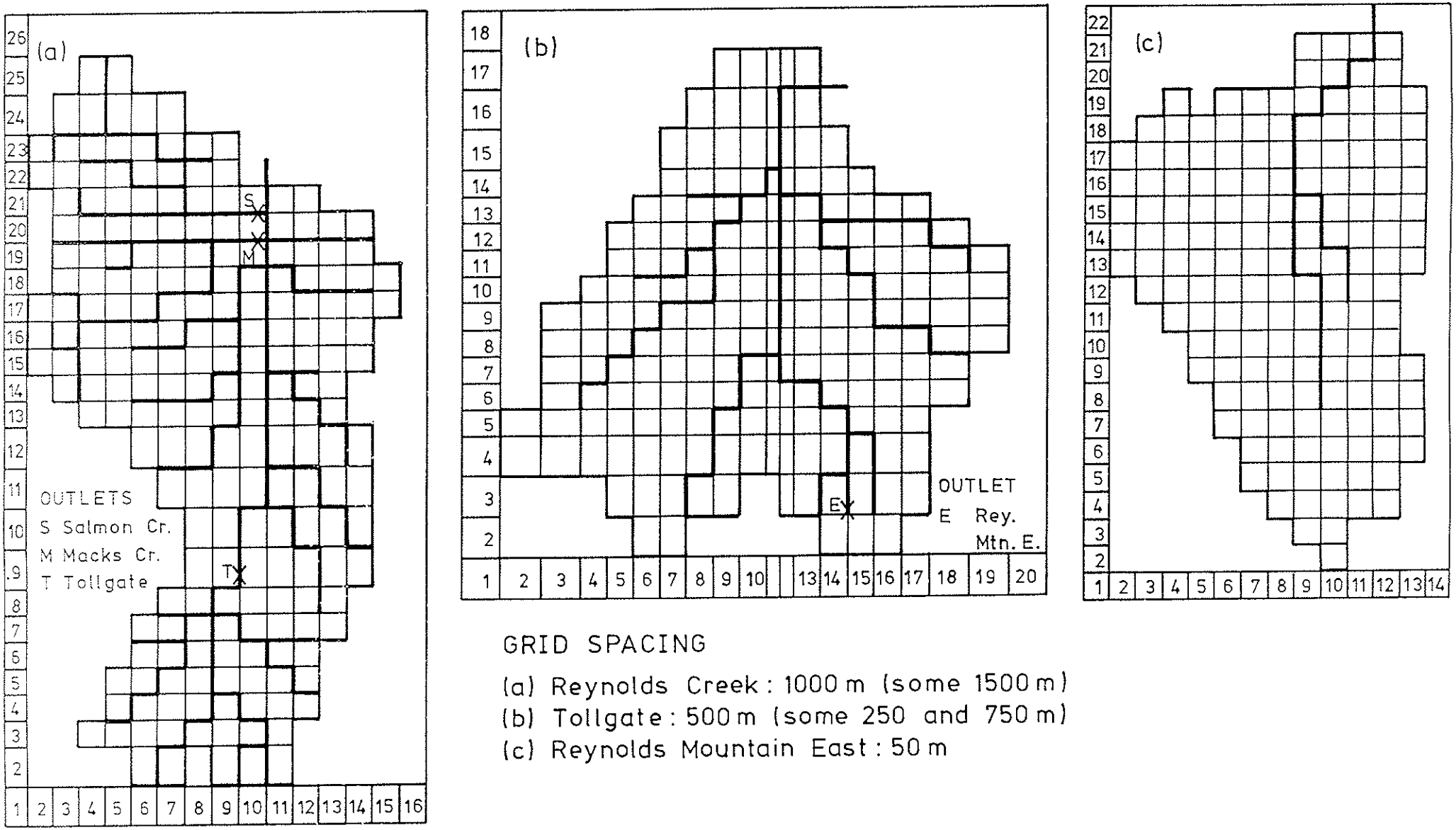


Fig. 3 Grid networks used in modelling Reynolds Creek catchment and its Tollgate and Reynolds Mountain East subcatchments.

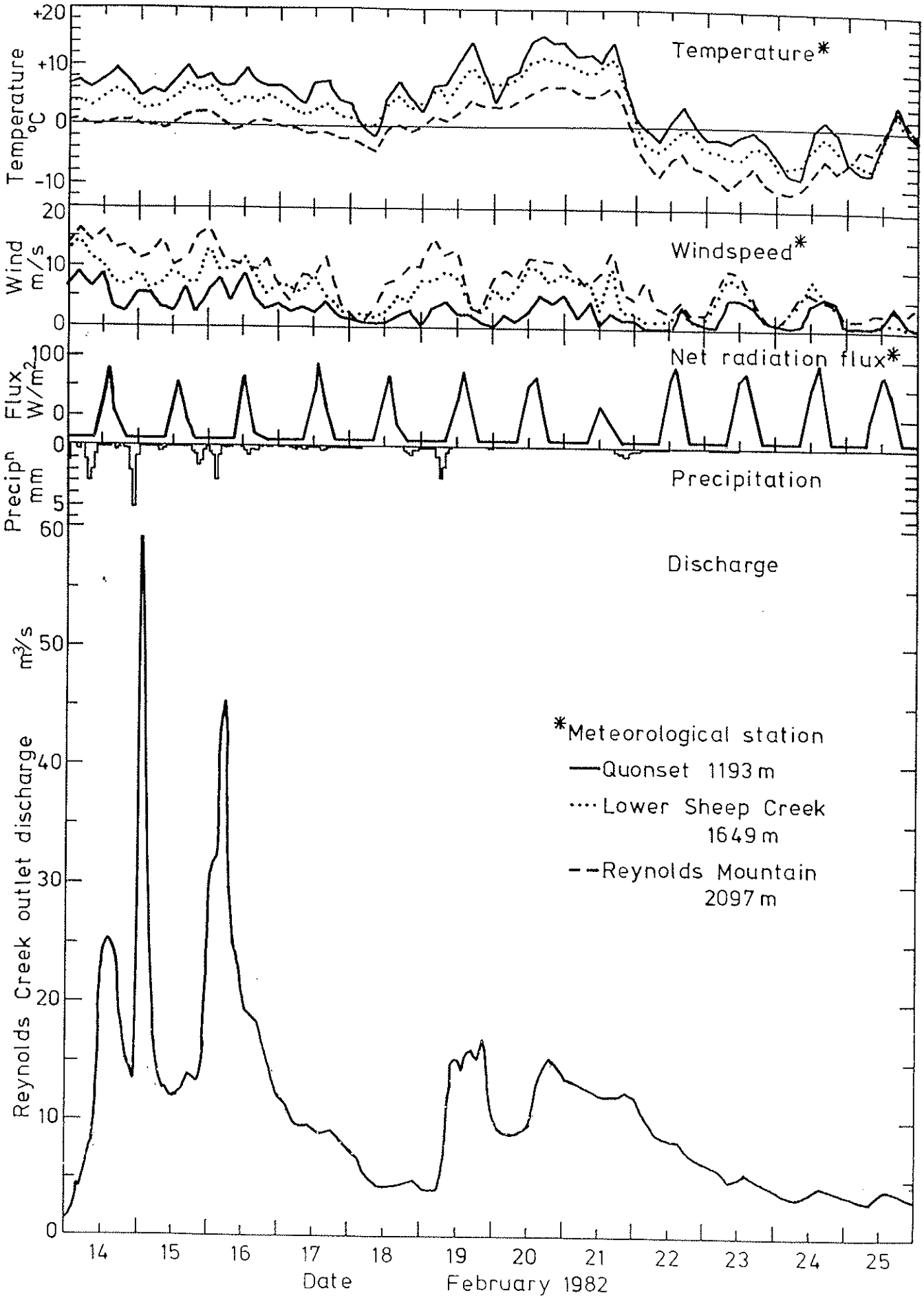


Fig. 4 Meteorological conditions at three stations in Reynolds Creek catchment and the measured outlet hydrograph for the simulation period.

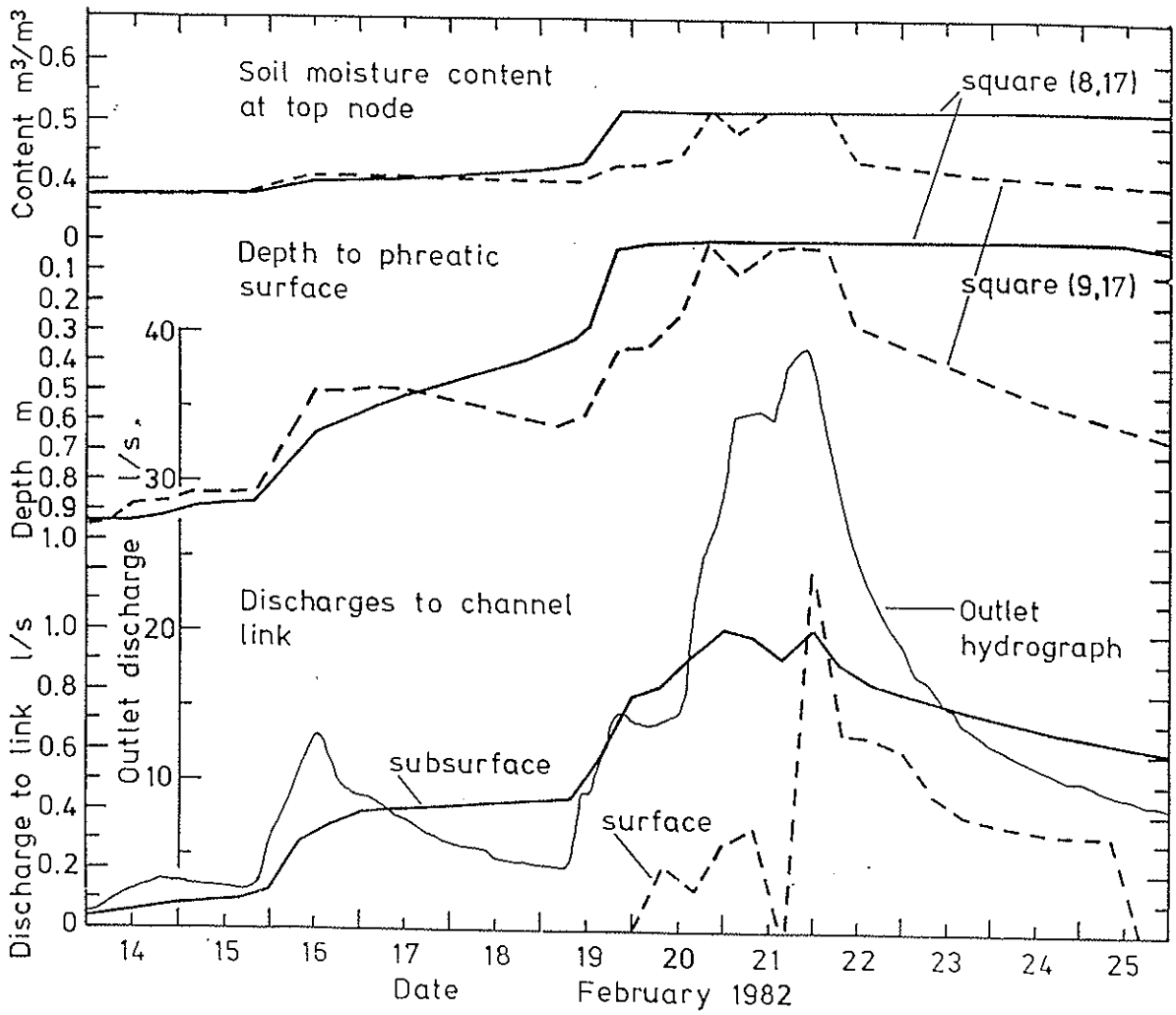


Fig. 5 Simulated variation in phreatic surface levels and surface soil moisture contents for squares (8, 17) and (9, 17), Reynolds Mountain East, along with the subsurface and surface inputs to the intervening channel link. The observed outlet hydrograph is shown for reference purposes. (Reproduced from Bathurst and Cooley (in press).)

outlet was best represented on the basis of a rapid, near-channel subsurface response. Away from the stream the meltwater was constrained to percolate to a deep phreatic surface in the fractured bedrock, on the assumption that it would appear over a longer period as a seasonal baseflow rather than in the observed hydrograph.

Figure 5 illustrates the simulated subsurface response in terms of the phreatic surface levels and surface soil moisture contents at squares (8,17) and (9,17), lying either side of a channel link (Fig. 3). Also shown are the subsurface and surface inputs to the channel. At square (8,17) there is a steady rise in phreatic surface level and a prolonged period of saturation with overland flow, probably produced by convergence of subsurface flow from neighbouring squares. The area is observed in practice to be marshy. At square (9,17) the response is more variable and exhibits rapid drainage, suggesting the moisture supply to be dependent on only the meltwater at that square. Comparison of the subsurface and surface discharges to the channel shows the channel response to be dominated by the groundwater discharge, itself varying according to the rapid change in phreatic surface levels. Only at the time of peak flow, when upward saturation results in complete saturation of the soil columns along the channel margins, does overland flow make a comparable contribution.

The simulated pattern mirrors almost exactly Ragan's (1968) field observations of a near channel subsurface response related to a groundwater "ridge". It demonstrates the ability of SHE to model a significant subsurface contribution to event flow according to the processes reviewed earlier (although without distinguishing between "old" and "new" water sources). However, given the limited nature of the application, it should not be assumed that the near-channel subsurface response provides a complete and adequate description of the behaviour of Reynolds Mountain East at all stages of the snowmelt season.

Tollgate and Reynolds Creek

Tollgate was modelled with 199 squares, mostly of dimensions 500 m x 500 m; Reynolds Creek was modelled with 213 squares, mostly of dimensions 1000 m x 1000 m (Fig. 3). The size of the squares prevented any special representation of the near-channel areas and it was not therefore possible to separate the responses from the near-channel and the more distant areas as in the Reynolds Mountain East simulation. The Reynolds Mountain East catchment is smaller than a single Reynolds Creek catchment grid square. Consequently the separate responses noted in Reynolds Mountain East occur at a subgrid level for Reynolds Creek and their integrated effect has to be represented through the use of "effective" parameter values.

Figures 6 and 7 illustrate the simulated channel supply responses for two contrasting areas in the Reynolds Creek catchment: a lowland area (squares (7,13) and (7,14)) and an upland area (squares (6,6) and (6,7)). The latter shows a good agreement with the Reynolds Mountain East simulation and emphasizes the importance of the subsurface contribution to event flow. Only at the peak runoff period is a surface contribution significant. The lowland results are consistent with the observed response. They show that prolonged saturation of the ground allows the early response to be dominated by overland flow, with consequent flashy hydrographs. A more subdued response occurs

around 19 February, reflecting the disappearance by then of the snowpack at lower elevations.

The simulation results demonstrate the ability of SHE to represent contrasting channel supply responses on a spatially distributed basis within a catchment. However, relative to the Reynolds Mountain East simulations, two problems are apparent.

(1) Partitioning of channel supply response into surface and subsurface components could be achieved but the simulations are a coarser reproduction of the observed response. In particular, it was not possible to reproduce accurately the very spiky shape of the stream hydrographs of the lowland areas at the start of the period. Generation of such a response requires a rapid runoff from a spatially limited, presumably near-channel, area. However, the use of large grid squares makes it difficult to simulate hydrological processes operating over small spatial scales. Likewise it is difficult to represent accurately the response of the smaller streams, such as Reynolds Mountain East, as this becomes a subgrid phenomenon.

(2) In order to reproduce the observed channel hydrographs it was necessary to increase the "effective" saturated zone conductivity with model grid scale. For Reynolds Mountain East the conductivity was 8.5 m/day, which is within the observed range. However, values of 17 m/day and 25.5 m/day were set for Tollgate and Reynolds Creek respectively, exceeding the range of measurements. It is likely that the increase compensates for the smoothing of the saturated zone hydraulic gradient which accompanies the use of large grid squares. In other words, the use of large grid squares makes it difficult to simulate the near-channel hydraulic gradient accurately, especially in catchments like Reynolds Creek where there is significant topographic variability.

Future Developments

The Reynolds Creek application demonstrates that SHE can represent the river-aquifer interaction with an accuracy sufficient for flow routing within a catchment, including a rapid near-channel subsurface contribution to event flow. However, there is increasing interest in modelling also the movement of contaminants within catchments. This requires a more detailed knowledge of the source of water flows and of the flow pathways than is required simply for flow routing: in particular, the contributions of "old" and "new" waters need to be correctly identified. Using SHE as a basis, staff at the WRSRU have developed SHETRAN, a water flow, sediment transport and contaminant migration modelling system. Several modifications to the original SHE subsurface model either have been made or are under way, which will allow the river-aquifer exchange to be represented with the detail needed to model contaminant migration. These include a river bank grid element (narrower than the standard grid square and designed to represent the near channel hydraulic gradient more accurately), an improved representation of soil layering (allowing for different hydraulic conductivities and flow rates) and a fully integrated (saturated and unsaturated zone) three-dimensional subsurface flow component (providing a more accurate depiction of flow pathways).

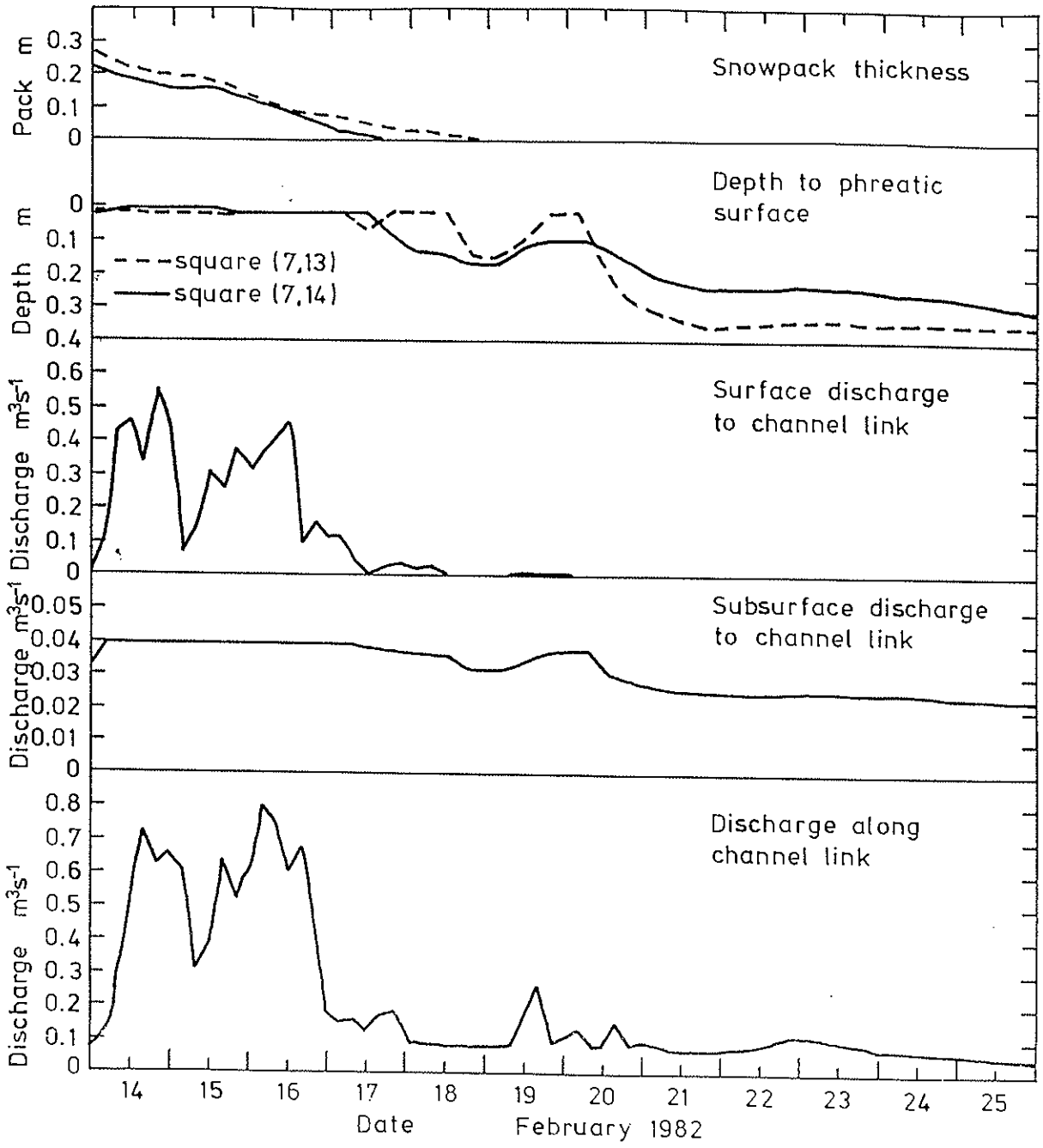


Fig. 6 Simulated variation in snowpack thickness and phreatic surface levels for lowland squares (7, 13) and (7, 14), Reynolds Creek, along with the subsurface and surface inputs to the intervening channel link and the discharge along the link.

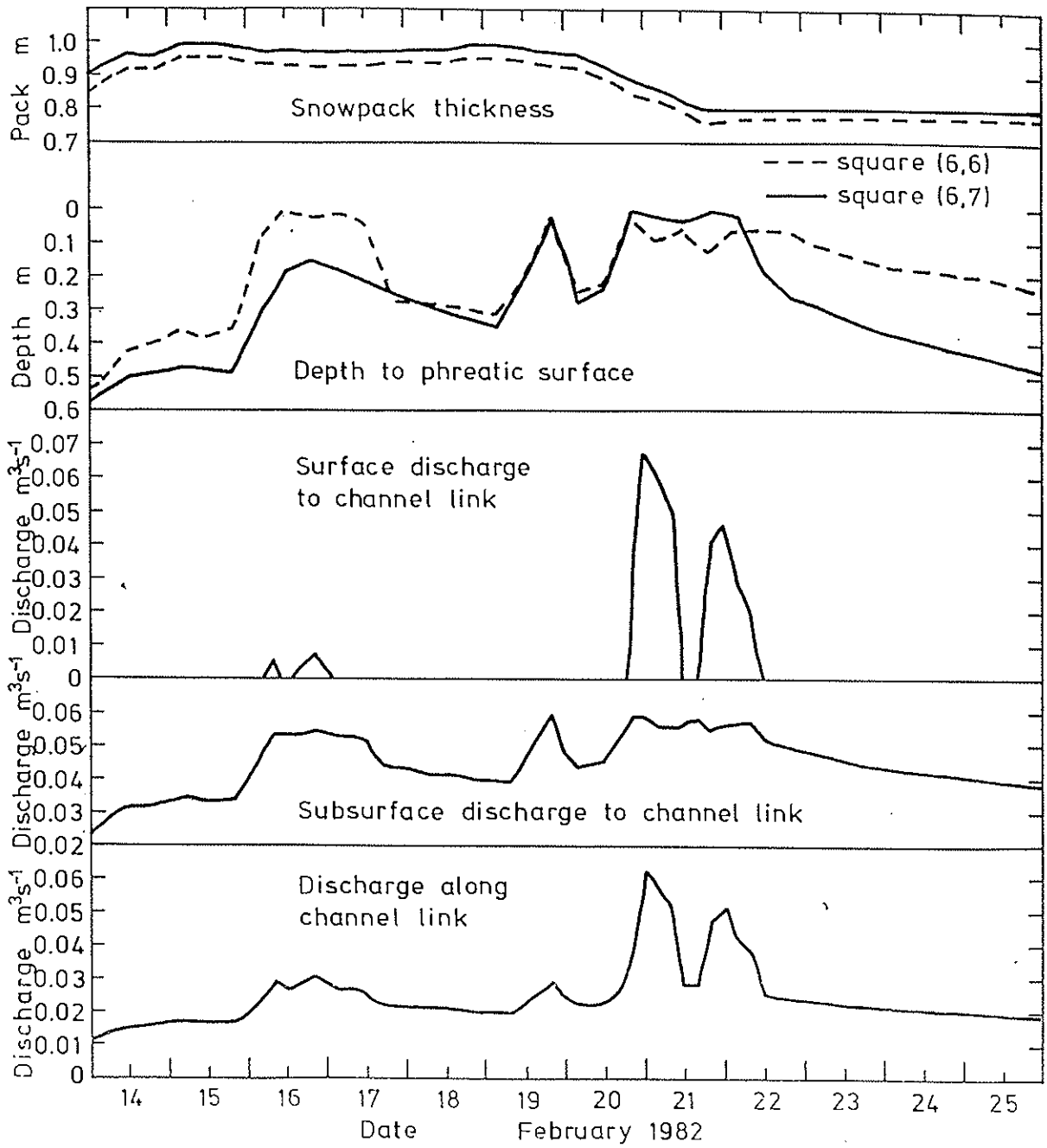


Fig. 7 Simulated variation in snowpack thickness and phreatic surface levels for upland squares (6, 6) and (6, 7), Reynolds Creek, along with the subsurface and surface inputs to the intervening channel link and the discharge along the link.

References

Abbott, M.B., Bathurst, J.C., Cunge, J.A., O'Connell, P.E. and Rasmussen, J., 1986a, An introduction to the European Hydrological System - Système Hydrologique Européen, "SHE", 1: History and philosophy of a physically-based, distributed modelling system. **Journal of Hydrology**, 87, 45-59.

Abbott, M.B., Bathurst, J.C., Cunge, J.A., O'Connell, P.E. and Rasmussen, J., 1986b, An introduction to the European Hydrological System - Système Hydrologique Européen,"SHE", 2: Structure of a physically-based, distributed modelling system. **Journal of Hydrology**, 87, 61-77.

Bathurst, J.C. and Cooley, K.R., In press. Use of the SHE hydrological modelling system to investigate basin response to snowmelt at Reynolds Creek, Idaho. **Journal of Hydrology.**

Binley, A., Beven, K. and Elgy, J., 1989, A physically based model of heterogeneous hillslopes 2. Effective hydraulic conductivities. **Water Resources Research**, 25(6), 1227-1233.

Blowes, D.W. and Gillham, R.W., 1988, The generation and quality of streamflow on inactive uranium tailings near Elliot Lake, Ontario. **Journal of Hydrology**, 97, 1-22.

Bonell, M., Pearce, A.J. and Stewart, M.K., 1990, The identification of runoff-production mechanisms using environmental isotopes in a tussock grassland catchment, Eastern Otago, New Zealand. **Hydrological Processes**, 4, 15-34.

Gillham, R.W., 1984, The capillary fringe and its effect on water-table response. **Journal of Hydrology**, 67, 307-324.

Ogunkoya, O.O., and Jenkins, A., 1991, Analysis of runoff pathways and flow contributions using deuterium and stream chemistry. **Hydrological Processes**, 5, 271- 282.

Ragan, R.M., 1968, An experimental investigation of partial area contributions. **Proceedings of the Berne Symposium of the International Association of Scientific Hydrology (IASH)**, IASH Publication No. 76, Institute of Hydrology, Wallingford, Oxon, UK, pp 241-249.

Sklash, M.G., 1990, Environment isotope studies of storm and snowmelt runoff generation. In: **Process Studies in Hillslope Hydrology**, M.G. Anderson and T.P. Burt (eds), John Wiley & Sons Ltd, Chichester, pp 400-435.

Sklash, M.G., and Farvolden, R.N., 1979, The role of groundwater in storm runoff. **Journal of Hydrology**, 43, 45-65.

AN INVESTIGATION OF THE MALMESBURY AVON CATCHMENT IN THE COTSWOLDS OF SOUTHERN ENGLAND

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Abstract

The Malmesbury Avon Catchment occupies an area of some 300km^2 on the southern margins of the Cotswolds and is underlain by an alternating sequence of dipping oolitic limestones and clays. The limestone aquifers have been highly developed for public supply, with most of these resources exported from the catchment. Over the last 5 to 10 years there has been increasing concern from the local community that the character of the river is being adversely changed, particularly by groundwater abstraction. WS Atkins were commissioned by the National Rivers Authority (NRA) to investigate the full impact of groundwater abstractions upon the river environment and to examine the relevance of other river management changes. Two of the fundamental assumptions underlying the most recent groundwater developments were that (i) the two main aquifers, the Great Oolite and Inferior Oolite, are not significantly hydraulically connected and (ii) the confined Great Oolite has virtually no connection with the river system in the lower part of the catchment. It was proved at an early stage in this study that these assumptions were not valid and the hydraulic interaction between the aquifers and the surface water regime is significantly more complex than was previously envisaged. The involvement of the general public has been an integral part of this project. An Open Day proved very valuable and identified, for example, certain river reaches where strong baseflow used to be present throughout the year, and the suggested influence of the change from movable mill sluices to fixed low level weirs in transforming river appearance, particularly algal bloom developments. River diarists were also recruited, and provided suprisingly consistent assessments of the acceptable (and unacceptable) flow regimes on various river reaches. The MIKE SHE model was able to simulate accurately the hydrological and hydraulic conditions within the catchment and was used to test a variety of resource management options. These included the reduction of abstraction from specific groundwater sources and the transfer of abstraction from supply to stream support.

Introduction

The Malmesbury Avon is the name generally given to the upper reaches of the Bristol Avon Catchment, which in this study was defined as the catchment upstream of Great Somerford gauging station, covering an area of approximately 300km^2. The catchment is located in the Cotswolds of Wiltshire in the south of England, about 40km to the east of Bristol. The catchment location is shown on Figure 1 with a study area plan on Figure 2.

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Figure 1 - Location of Study Area

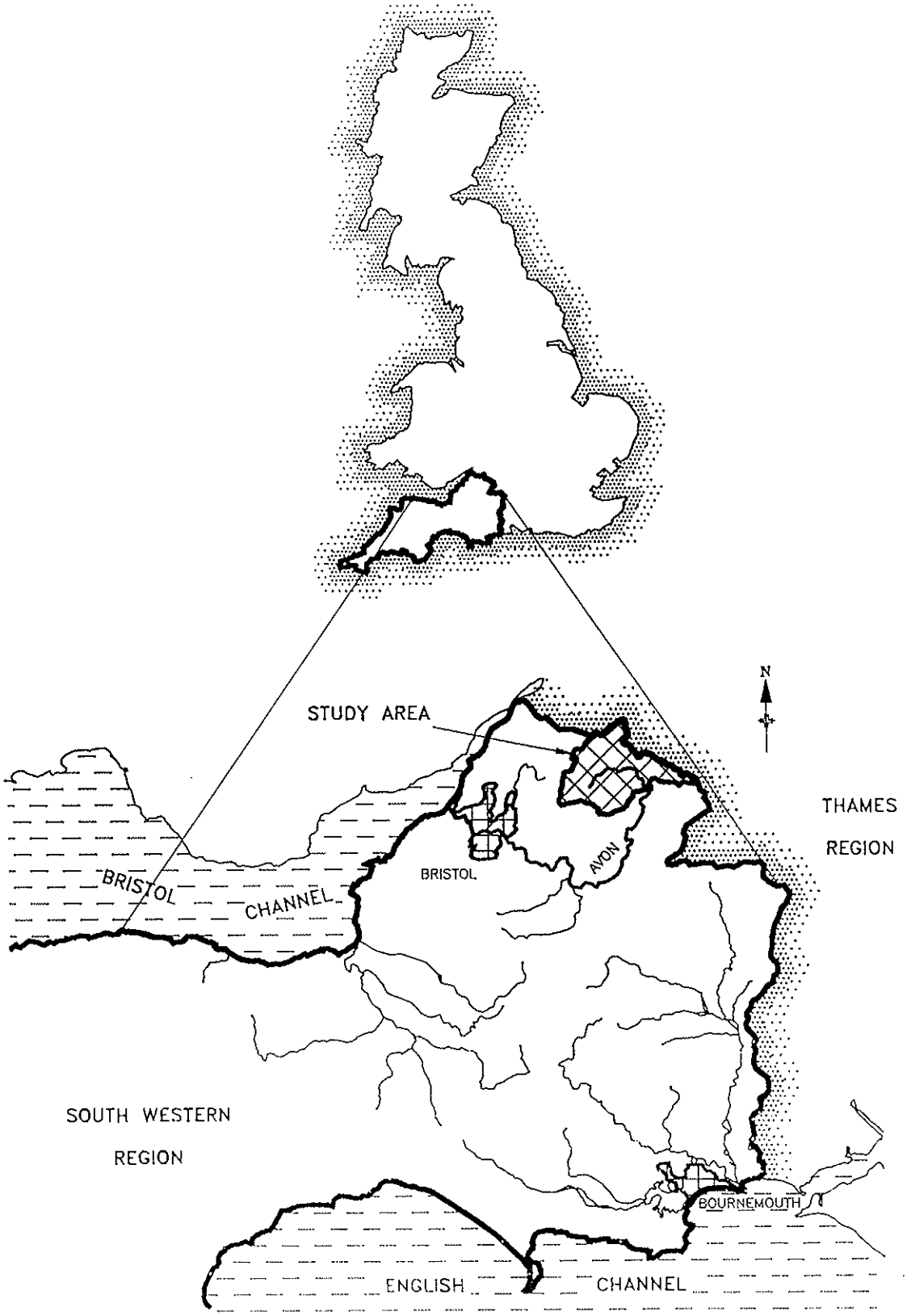
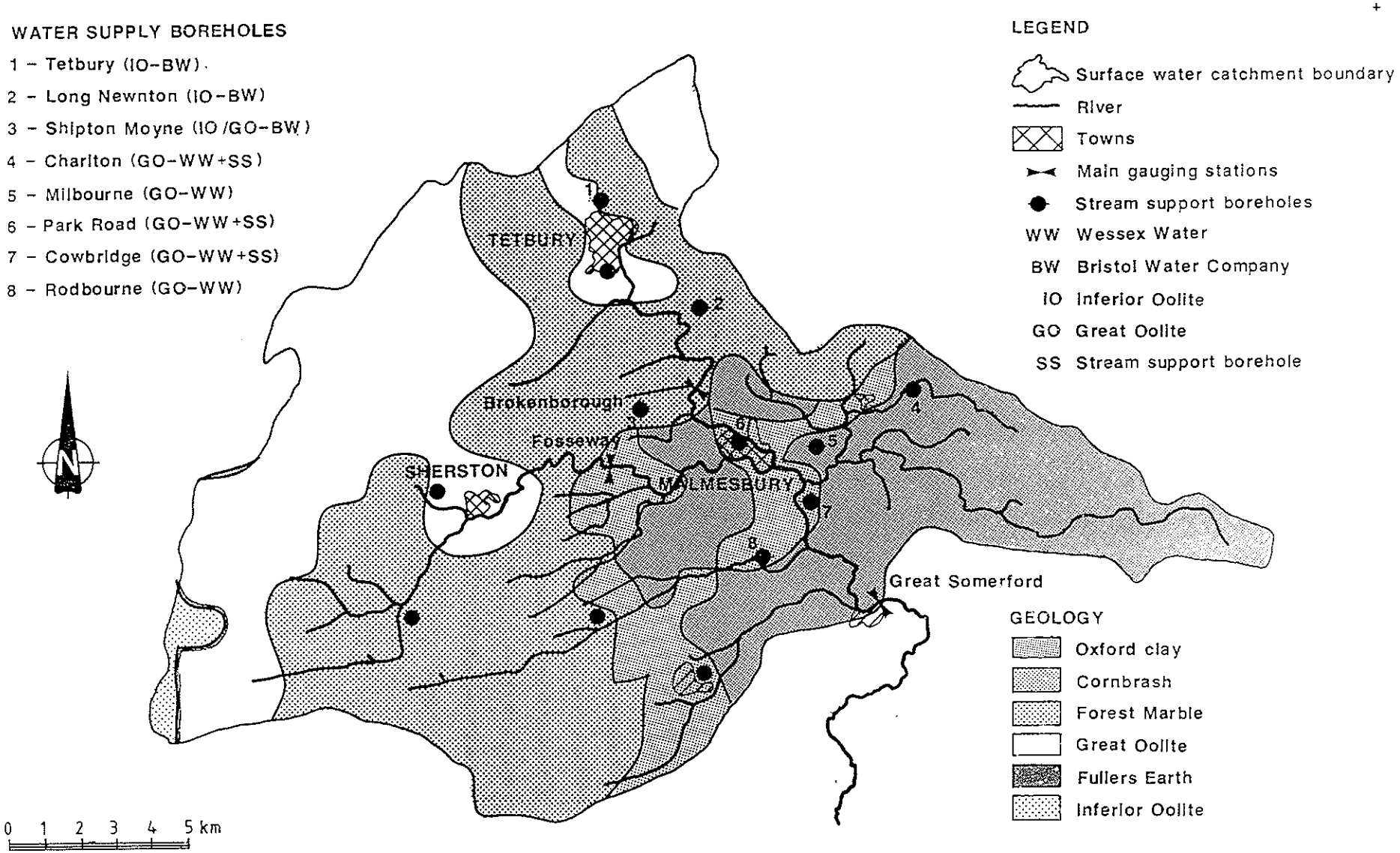


Figure 2 - Map of Study Area



The river consists of two main tributaries, known as the Sherston Avon and the Tetbury Avon after the towns through which they flow. These tributaries meet in Malmesbury from where the river flows south east and is joined by three smaller tributaries (Woodbridge Brook, Gauze Brook and Rodbourne Brook). The catchment topography slopes gently to the south east, reflecting the dip in the underlying geology. The river system flows within narrow valleys incised into the catchment and 20 to 30 metres below the surrounding land.

Catchment Geology

The catchment geology (as shown in Figure 2) consists of interbedded layers of fractured oolitic limestones, marls and clays, dipping to the south east at a slightly steeper angle than the topography. The geology can be simplified as follows:

Oxford Clay - thick low permeability clay sequence, only found in the south east of the catchment.

Cornbrash (3 to 6m) - alternating sequence of marl and limestone, not considered significant in catchment water supply terms.

Forest Marble (10 to 20m) - calcareous and occasionally sandy clays with lenses of limestone. Acts as a leaky aquitard and will transmit groundwater, particularly where thin or strongly faulted.

Great Oolite (20 to 40m) - flaggy and massive oolitic and reef limestones with some calcareous siltstone. Groundwater flow is through fractures and fissures. This sequence provides the main groundwater baseflow to the river system and outcrops across much of the central part of the catchment.

Fullers Earth (20 to 30m) - thick and extensive sequence of clays with some limestone. Previously considered as an effective barrier to flow between the Great and Inferior Oolite, it now appears that flows do occur across this layer.

Inferior Oolite (12 to 45m) - a relatively homogenous sequence of fine to medium grained shelly and oolitic limestones. This sequence outcrops as a thin band along the western scarp boundary and underlies the entire catchment.

Groundwater Resources Development

The Malmesbury Avon catchment has a long history of groundwater abstraction for public supply, and wells and boreholes have been installed in both the Great Oolite and Inferior Oolite aquifers by Wessex Water and Bristol Water Companies.

Conventional groundwater sources had been developed at regular intervals up until 1980 when the most recent scheme, known as the Malmesbury Groundwater Scheme, was licensed. This scheme combined increased public supply with 8 new or re allocated surface water augmentation boreholes, abstracting from the Inferior Oolite in the upper reaches of each tributary and operated at or below trigger flows on downstream gauging stations. The complete supply and augmentation system is shown on Figure 2 and the average daily

yield from each aquifer in 1992 was as follows:

	Supply (Mld^{-1})	Augmentation (Mld^{-1})
Great Oolite	40	0.2
Inferior Oolite	10	2.6

Public concerns as to the impact of groundwater abstraction on the river environment grew in the 1980's. The concerns were principally related to low summer flows and focussed on (i) the amenity and environmental value of the lower reaches of the Sherston and Tetbury Avon, particularly around Malmesbury and (ii) the fisheries value of the main Avon downstream of the confluence. In November 1991 NRA appointed WS Atkins to undertake a detailed investigation into the catchment, the results from which are presented in this paper.

Conceptual Flow model

A conceptual understanding of the dominant hydraulic processes in the catchment was developed at an early stage of the project, the main features of which are:

(i) The catchment responds rapidly to winter rainfall and the baseflow index for the catchment to Great Somerford is only 0.58. This is considered to be due to (a) rapid flow through the fissure system, and (b) low specific yield of the aquifers resulting in annual head variations in the order of 20 metres and a rapid baseflow response to recharge.

(ii) The groundwater table falls to low levels in the summer and there is little or no recharge in response to summer rainfall. Low flows in the Sherston and Tetbury Avon through Malmesbury are commonly in the order of $5Mld^{-1}$, compared with mean flows in the order of $100Mld^{-1}$.

(iii) The Great and Inferior Oolite aquifers had been thought to be essentially discrete. Over much of the catchment there is a significant head difference between the two aquifers with the head in the Inferior Oolite being up to 40 metres below that in the Great Oolite.

A dye tracer test carried out in the 1970's was intended principally to assess the flow directions in the Great Oolite. However, the Inferior Oolite directly below the Great Oolite release point was also monitored and recorded a dye trace around 60 days following release, indicating a distributed downwards seepage flow between the aquifers.

In addition, when the groundwater heads in the Great and Inferior Oolite were studied closely, mounds were identified in the Inferior Oolite where the groundwater level had risen, in some cases up to the Great Oolite level. These mounds have been interpreted as zones of discrete downward flow, related to fault and fissure development.

(iv) Monthly flow accretion data revealed zones of stream flow losses on both the Sherston and Tetbury Avon, just upstream of the confluence. This was surprising as the Great Oolite was considered to be confined along these reaches.

Boreholes installed by the NRA in 1991 along the lower reaches of the Sherston Avon revealed that the Great Oolite is within a few metres of the ground surface along a 250m reach in Malmesbury. There is therefore potential for interaction between the aquifer and the river in this area.

In summary, the basic conceptual model is of a catchment where groundwater flow is dominant but which, nevertheless responds rapidly to winter rainfall with very low summer flows. There is also evidence of connections both between the river and the confined Great Oolite and between the two aquifers. The section in Figure 3 illustrates this.

Public Open Day

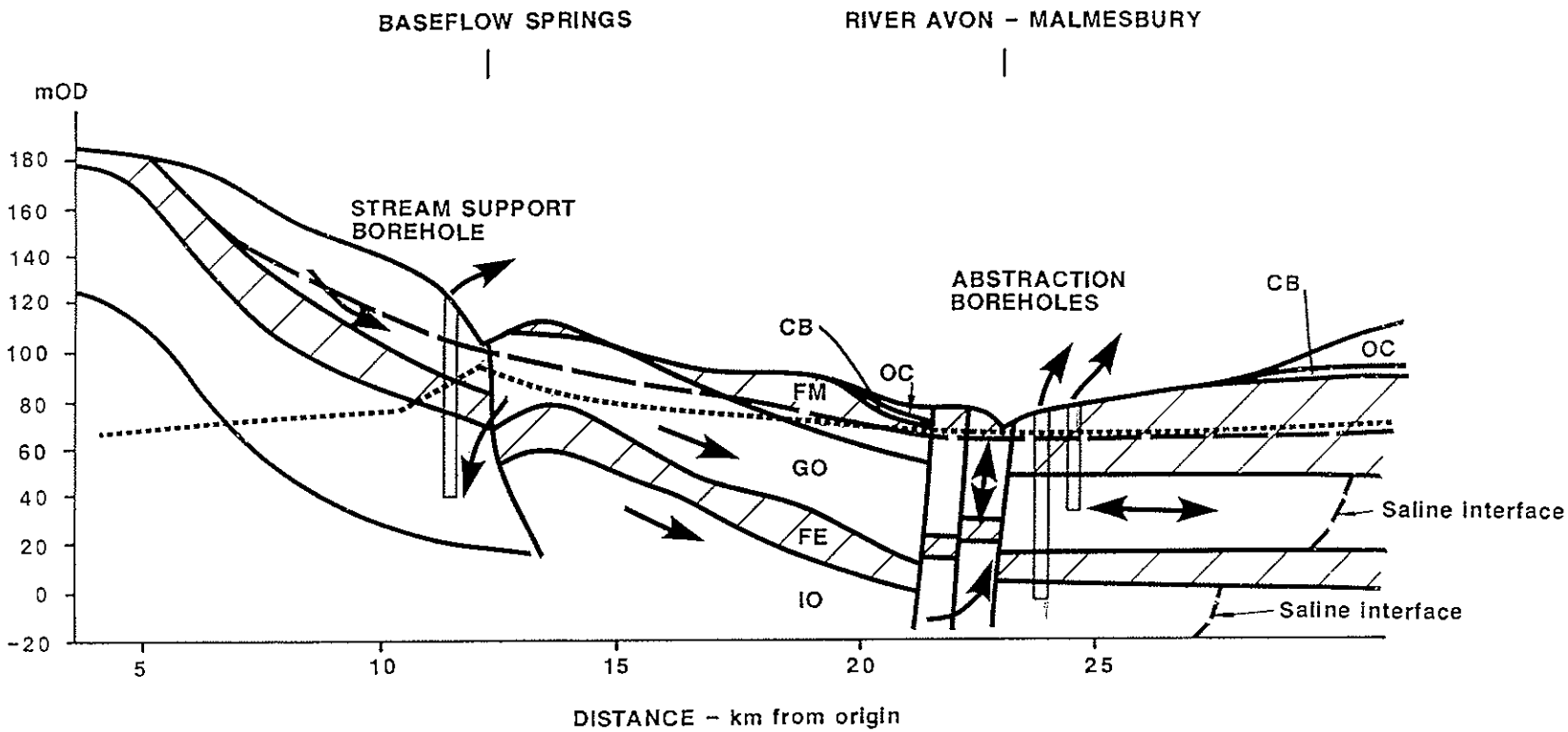
In June 1992 an Open Day for the project was held in Malmesbury. This had a dual purpose of (a) publicising the project to the public and (b) recording their recollections of the river. The main findings are presented below.

(i) In the years before the last war a swimming race was held in Malmesbury every August, racing along the lower reaches of the Sherston Avon. One competitor recalled a considerable variation in river temperature along the route, which was generally interpreted as due to the strong spring flows into the river. Around the same period a spring, known as Daniel's Well, was used by locals as a drinking water supply. These spring flows have not been present in recent summers.

(ii) The river tributaries around Malmesbury used to be controlled by moveable sluice structures allied to old mills. By the 1950's these were no longer operated properly and the area was regularly flooded. The structures were progressively replaced by low level static weirs, substantially reducing the flooding risk. One interviewee had moved away from the lower Tetbury Avon in the 1930's, returning in the 1980's. He was very surprised to see, each summer, bright green algal blooms develop over the river surface. Another interviewee had recorded the movement of a lemonade bottle in the river, travelling a total distance of around 200 metres in three weeks. The causes of eutrophication and algal blooms are generally considered to be increased nutrient loads. However, in most lowland rivers algal blooms are not nutrient limited and more important requirements are residence time (at least 5 days), low water velocity and high water temperature. The replacement of movable sluices, which provided a regular flushing flow to the river, with static weirs appears to have combined with reduced river flows to produce conditions favourable to algal bloom development.

(iii) The loss or reduction of submerged aquatic plants along the river, and replacement with emergent and marginal flora, was also recorded by a number of respondees. There were other general comments on the reduction in, for example, invertebrate fauna and wild trout spawning. These features are common to many of the lowland streams suffering from reduced river flows.

Figure 3 - Conceptual Hydrogeological Section Showing Hydraulic Interactions Between Aquifers and Surface Water Features.



River Diaries

A River Diary is a regular record of the condition of a particular river reach. In addition to general observations the diaries provide a subjective view of whether a river flow is acceptable as a summer minimum flow. Diarists recruited largely from the general public provided a weekly record of river conditions during the summers of 1992 and 1993.

A typical comparison of Diary records and river flow conditions is shown on Figure 4. The records proved very consistent, both in time and between diarists. However, in each case, the minimum flow requirements of the same diarist were significantly higher in 1993 than in 1992. Two main reasons for this can be envisaged as follows:

(i) Whereas the summer flow recession in 1992 was gradual, in 1993 the flows reduced rapidly accompanied by a rapid deterioration in river appearance. The low flows in 1993 were therefore contrasted unfavourably with conditions only a few weeks previously.

(ii) The summer of 1992 was reasonably dry with sunny weather and drought conditions throughout southern England were receiving nationwide publicity. The summer of 1993 was dull and damp and, nationally, the drought was considered to be coming to an end. As noted previously, summer rainfall has little actual effect on the low river flows but but may nevertheless influence the acceptability of river conditions.

Algal blooms were a feature of the river in 1992 and 1993. In late summer 1992 a brief spate resulted in the flushing of these blooms from the river system. Following this spate, river flows reduced to a lower rate than recorded previously but were considered more acceptable.

These results indicate that the public can provide a coherent and reasonable assessment of the acceptable flow requirements for a river. Perceptions can however be influenced by a number of related, and sometimes unrelated, factors and should not be the sole means of defining acceptability.

Fieldwork

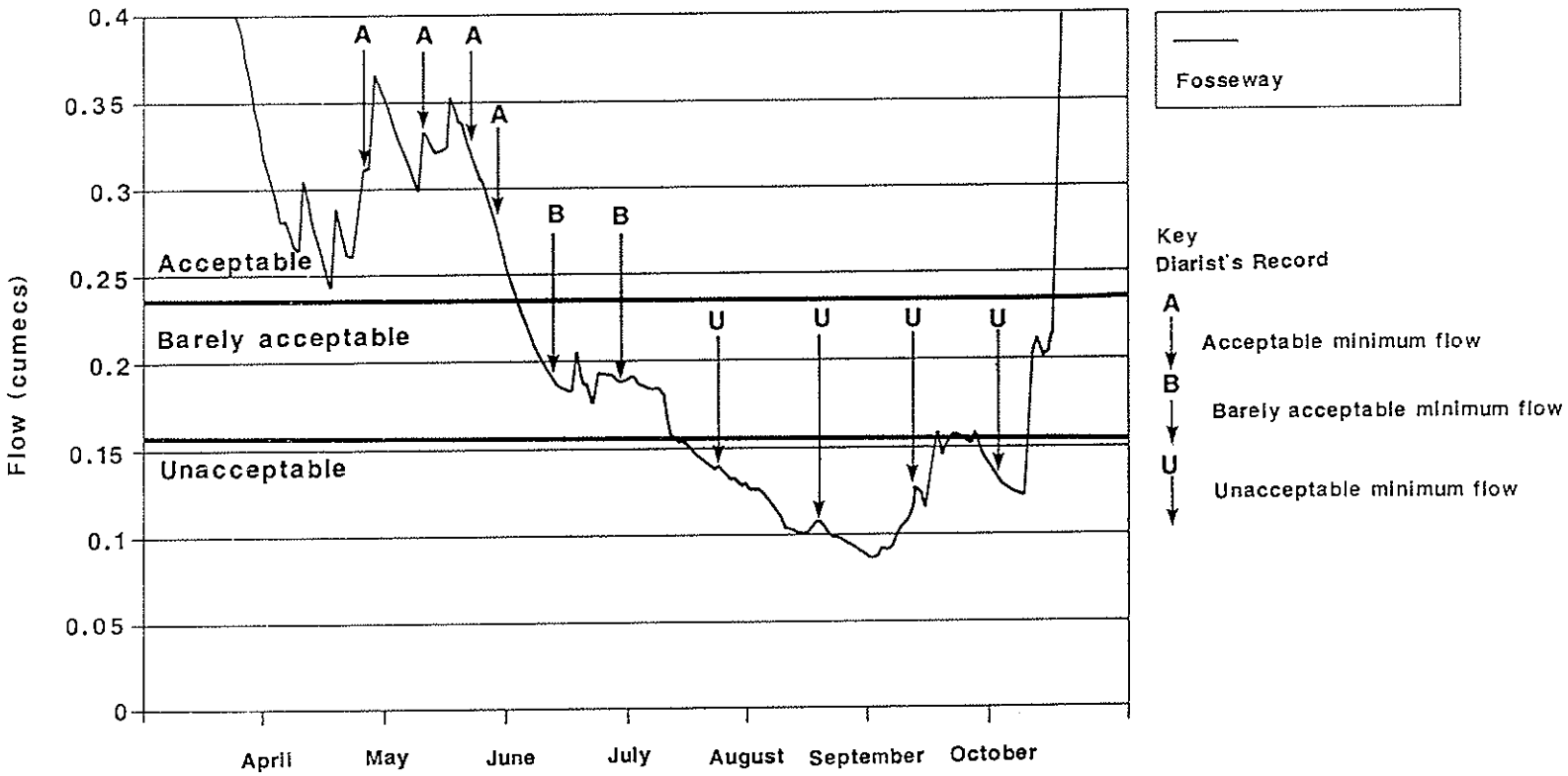
There are six public supplies from the confined Great Oolite and three from the Inferior Oolite. The fieldwork was designed to assess the impact of abstractions on groundwater levels by organising, with the co-operation of Bristol and Wessex Water Companies, to vary abstractions at each major supply source and monitor the impact on groundwater levels in the Great Oolite aquifer. Two investigations were carried out, in Summer 1992 and Spring 1993, and groundwater levels were monitored along the lower reaches of the Sherston and Tetbury Avon (the losing reaches) and upstream in both sub-catchments.

The main findings were as follows:

(i) The impact of Cowbridge and Milbourne sources is largely local to Malmesbury.

(ii) Robourne and Charlton sources yield from a wider area of the confined and

Figure 4 - A Typical Comparison of River Diary Records with Gauged River Flows



unconfined Great Oolite.

(iii) Shipton Moyne yields directly from the Great and Inferior Oolite and Long Newnton yields from the Inferior Oolite only. Both sources appear to induce downward flow from the Great Oolite however, which appears to be the ultimate resource for both sources. The sources yield largely from the unconfined Great Oolite with little impact on groundwater levels around Malmesbury.

The fieldwork results were subsequently compared with the model output as presented below.

Model Set-up and Calibration

The model was set up using the MIKE SHE finite difference modelling system. Separate modules were linked for:

(i) recharge through the soil zone and unsaturated zone flow
(ii) groundwater flow through a four layer system
(iii) 2 dimensional overland flow
(iv) 1 dimensional river flow

An initial model was set up at the start of the project and run in order to test the basic conceptual understanding of the flow system and to identify the fundamental areas where additional information was required. The model output was compared with actual records on the basis of:

(i) temporal and spatial variations in groundwater head in the Great and Inferior Oolite and

(ii) surface flow records at gauging stations, flow accretion curves and flow duration curves

The model was developed, as the conceptual understanding improved, and the following features were included in the final model:

(i) a four layer groundwater model. Layers 2, 3 and 4 represent the Great Oolite, Fullers Earth and Inferior Oolite respectively. The unconfined Great Oolite was divided into two layers with Layer 1 representing an upper high conductivity layer. This was necessary to reproduce the groundwater head variation in the aquifer and is consistent with other studies on the Great Oolite (Rushton et al, 1992). Elsewhere in the model, layer 1 represents the strata overlying the Great Oolite, viz the Forest Marble, Cornbrash and Oxford Clay.

(ii) A soil bypass mechanism, by which upto 75 per cent of the rainfall to the Great Oolite Catchment bypasses the soil zone through fissure systems. This mechanism is necessary to explain the rapid reponse to recharge in the winter period. The effectiveness of the bypass mechanism reduces to zero as the soil moisture deficit increases. This accounts for the reduced response to summer rainfall, with some

surface runoff but little or no groundwater recharge.

(iii) The specific yield in the Great Oolite was set at 0.5 to 0.7 per cent to account for the large head variations (20 to 30m) in response to winter recharge.

(iv) Zones of higher conductivity were set in the Fullers Earth (Layer 3), corresponding with the location of the mounds in the Inferior Oolite groundwater head.

(v) Heterogeneities were introduced into the Great Oolite to account for the spatial variation in groundwater heads and the flow accretion patterns in the river system.

Examples of the calibrated output from the model are shown on Figure 5.

Model Simulations

The calibrated model was used for both scoping simulations and to assess specific management options.

Scoping simulations were undertaken initially to assess, in broad terms, the impact of various abstraction sources on the surface flow record. The impact of, for example, naturalising the system (no abstraction), removing or doubling augmentation, and operating only specific groups of sources was tested over the 10 year run period, comparing the modelled actual output with the output from each simulation run. Figure 6, for example, shows the difference between historical and naturalised flow at the Great Somerford gauging station. The flow differences are expressed in absolute terms and as a percentage of the historical flow and show that

(i) Historical flows have been, on average, about $0.4m^3s^{-1}$ below the natural flow.

(ii) The flow reduction was at a minimum during summer low flow periods, due to the positive impact of augmentation. Nevertheless, actual flows still show a reduction of between 30 and 70 per cent over low flow periods.

Further model runs, to test specific management options, included transferring boreholes from supply to augmentation, increasing augmentation yields and lining specific river reaches.

The main findings from the modelling work are summarised below:

(i) under naturalised conditions there is relatively little lateral groundwater flow from the unconfined to the confined aquifers and downward flow from the Great Oolite to Inferior Oolite. Groundwater abstraction has induced (a) increased lateral flows into the confined aquifers (b) increased downwards flow to the Inferior Oolite and (c) reduced baseflow from both the unconfined aquifer and the partly confined aquifer around Malmesbury.

(ii) Abstraction is modelled as reducing surface flows to both Brokenborough and Fosseway gauging stations, at the downstream limit of the unconfined Great Oolite,

Figure 5 - Model Output after Calibration Compared with Observed Values

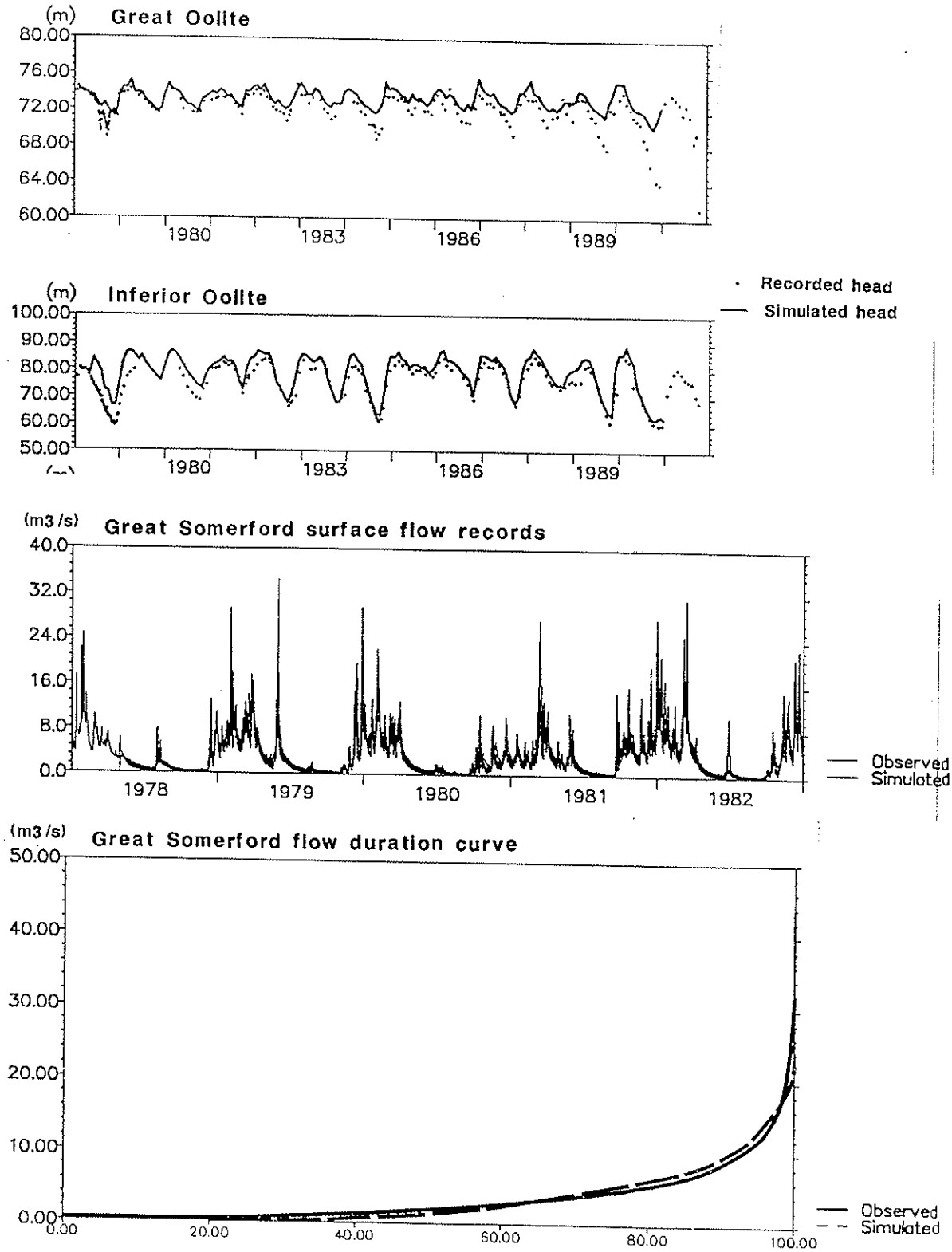
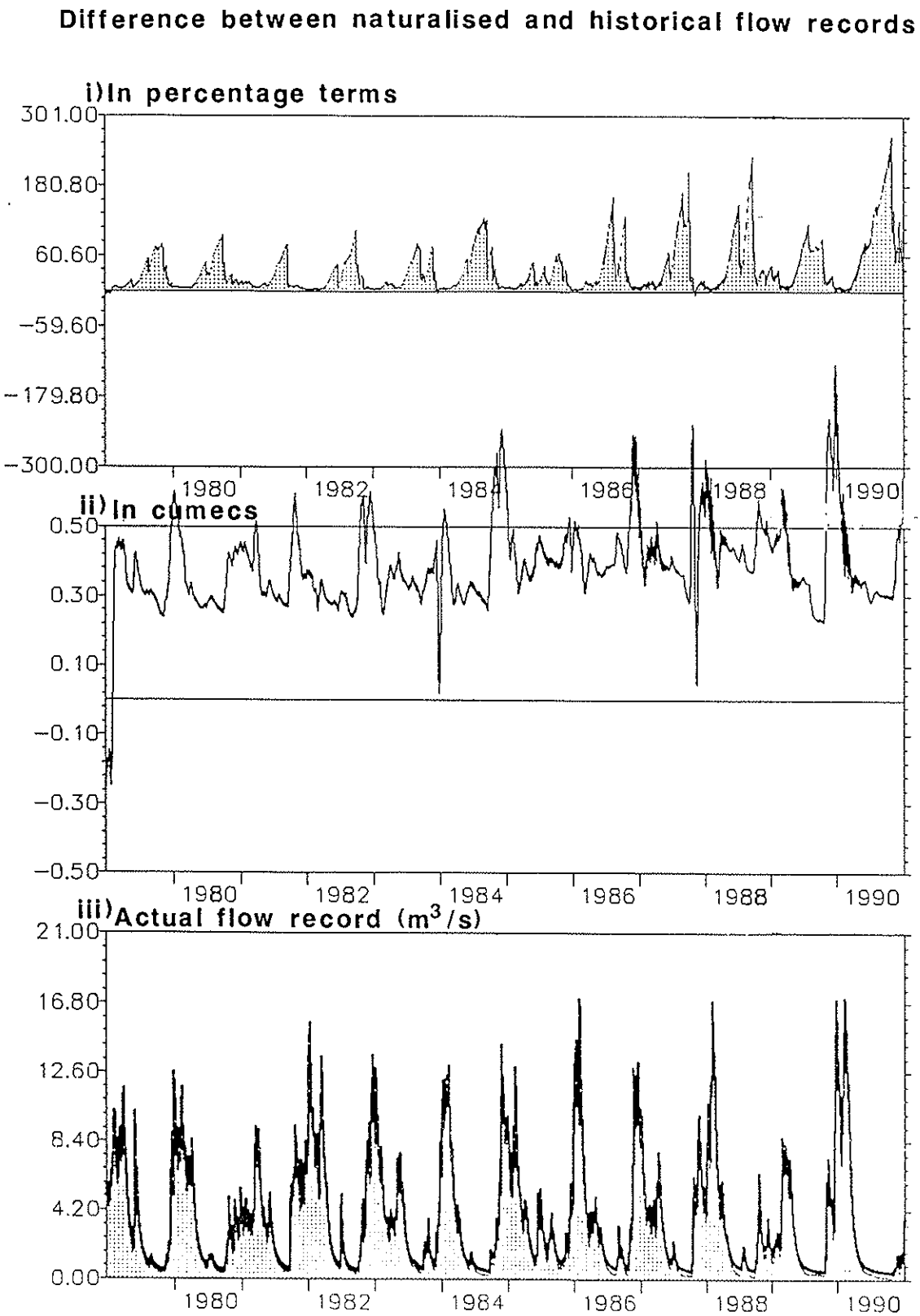


Figure 6 - Comparison of Historical and Naturalised Flows at Great Somerford



by around 5Mld^{-1}. During low flow periods, however, augmentation provides a small net benefit in comparison with natural conditions.

(iii) The average flow reduction to the confluence at Malmesbury is modelled as 14Mld^{-1} and this decreases to around 6Mld^{-1} under low flow conditions. Nevertheles actual low flows are only around 40 per cent of the natural flow. In addition, augmentation provides around two thirds of the actual flow and without augmentation these reaches would virtually dry up.

(iv) The impact of abstraction at the base of the catchment is not as severe, although minimum flows are still only around 60 per cent of the natural flow in drought years. The beneficial impact of augmentation is still evident, providing up to half the total flow in drought years.

(v) The augmentation scheme is very beneficial to the catchment over the summer low flow period. A nett benefit of 50 per cent of borehole yield is provided. Increasing augmentation from existing sources and/or transferring groundwater from supply to augmentation was shown to provide significant additional benefits, particularly in the lower Sherston and Tetbury Avon.

Summary

The Upper Avon is a very complex hydraulic system with an intimate interaction between the surface and groundwater flow regimes. Storage is low in the aquifers and groundwater levels respond rapidly to winter recharge. Conversely, summer recessions are steep and surface flows are particularly vulnerable to groundwater abstraction over the summer period. Groundwater abstraction, from both the Great and Inferior Oolite, has had a particularly adverse impact on the lower reaches of the Sherston and Tetbury Avon around Malmesbury where, from the model results, actual minimum flows are only around 40 per cent of the natural flows. Local inhabitants recalled that these reaches used to receive year round spring flows whereas flow losses can now occur in the summer. The appearance of these reaches has been compounded by the installation of fixed weirs which, whilst they retain a high water level, also provide ideal conditions for the development of algal blooms. The lower reaches of the Avon to Great Somerford are also adversely affected and lowest flows are reduced by around 40 per cent by abstraction.

The public have kept diaries over the last two years. The diary records revealed an impressively consistent assessment of acceptable flows for each year but also showed that other related and unrelated features can influence public opinion as to the acceptability of river flows. The augmentation scheme is providing a significant proportion of the river flow during the lowest flow periods. Various proposals are being considered by the NRA for developing the augmentation system further as part of the overall management plan for the catchment. These include (i) increasing the yield from the existing scheme and (ii) transferring a proportion of the public groundwater supply to augmentation. These and other management options are currently being developed, in consultation with all the interested parties. A definitive Action Plan for the catchment will be prepared at the end of this process.

Acknowledgement

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Reference

Rushton K.R, Owen, M., and Tomlinson, L.M., 1992, The Water Resources of the Great Oolite Aquifer in the Thames Basin, UK. **Journal of Hydrology, 132**, 225-248.

APPLICATION OF A COUPLED RIVER-GROUNDWATER MODEL (ROSANN) TO THE PREDICTION OF RIVER NITRATE CONCENTRATIONS

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Abstract

A model of the River Gipping catchment, NW of Ipswich, was required to predict the effect of possible land use changes on river nitrate concentrations. A PC based, coupled runoff, river and groundwater model (ROSANN) was developed and applied to the catchment to predict both river flows and water quality. Readily available hydrological and agricultural data for the Gipping catchment were obtained from a variety of sources. The catchment was represented on a regular 350 m grid. A GIS digital terrain model was used to derive catchment boundaries, surface runoff networks and river networks on this grid. The surface runoff module was based on the soil moisture deficit concept, and included denitrification and a delay mechanism for runoff. It provided weekly diffuse nitrate inputs to the river modules and recharge to the groundwater module. The groundwater module was based on a lumped parameter model relating baseflow to storage. It provided weekly baseflows and base nitrate loadings for the river module. The river flow module then calculated weekly mean flows and the river nitrate transport module solved the advection dispersion equation for the weekly mean nitrate concentration along its length. Point sources of nitrate such as sewage works, and abstractions from the groundwater and river were also included. Data on river flows and nitrate concentrations at various points within the catchment were obtained for the five year period 1988-1992 and used for calibration.

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The model replicated both the rapid response to winter rainfall events of the river flow with associated high nitrate concentration peaks and the pattern of declining summer baseflow over the five years. It is concluded that lumped parameter aquifer models are suitable for modelling river-aquifer interaction on the catchment scale used. It is also concluded that the use of both flow and water quality determinants enhanced the calibration of the whole model.

Introduction - the Gipping

This paper describes the ROSANN suite of models and their application to the River Gipping catchment. The Gipping catchment covers about 320 km^2 to the northwest of Ipswich upstream of the tidal limit at Constantine weir (Figure 1). The town of Stowmarket lies in the centre of the catchment. The catchment geology is Chalk overlain by boulder clay in higher relief areas, and by sands and gravels along valley bottoms. About 80% of the catchment is agricultural, with 53% used for the cultivation of winter and spring cereals.

Rising peak nitrate concentrations in reservoir intake water from the Gipping and the resulting possible impacts on drinking water quality prompted this project. It was felt that changing land use practises in this largely agricultural catchment could be a major cause of this trend. The requirements were therefore first to construct a general purpose mathematical model to predict monthly mean nitrate concentrations in a river, taking land and fertiliser usage and surface and groundwater transport routes into account, and then to apply this to the Gipping. The catchment-wide scale of this model, and the requirement that it be easily applicable to other catchments, required the use of readily available data. The groundwater-river interaction would be an important part of this model, since, as is typical for chalk rivers, aquifer-fed baseflow provides almost all the Gipping's summer flow.

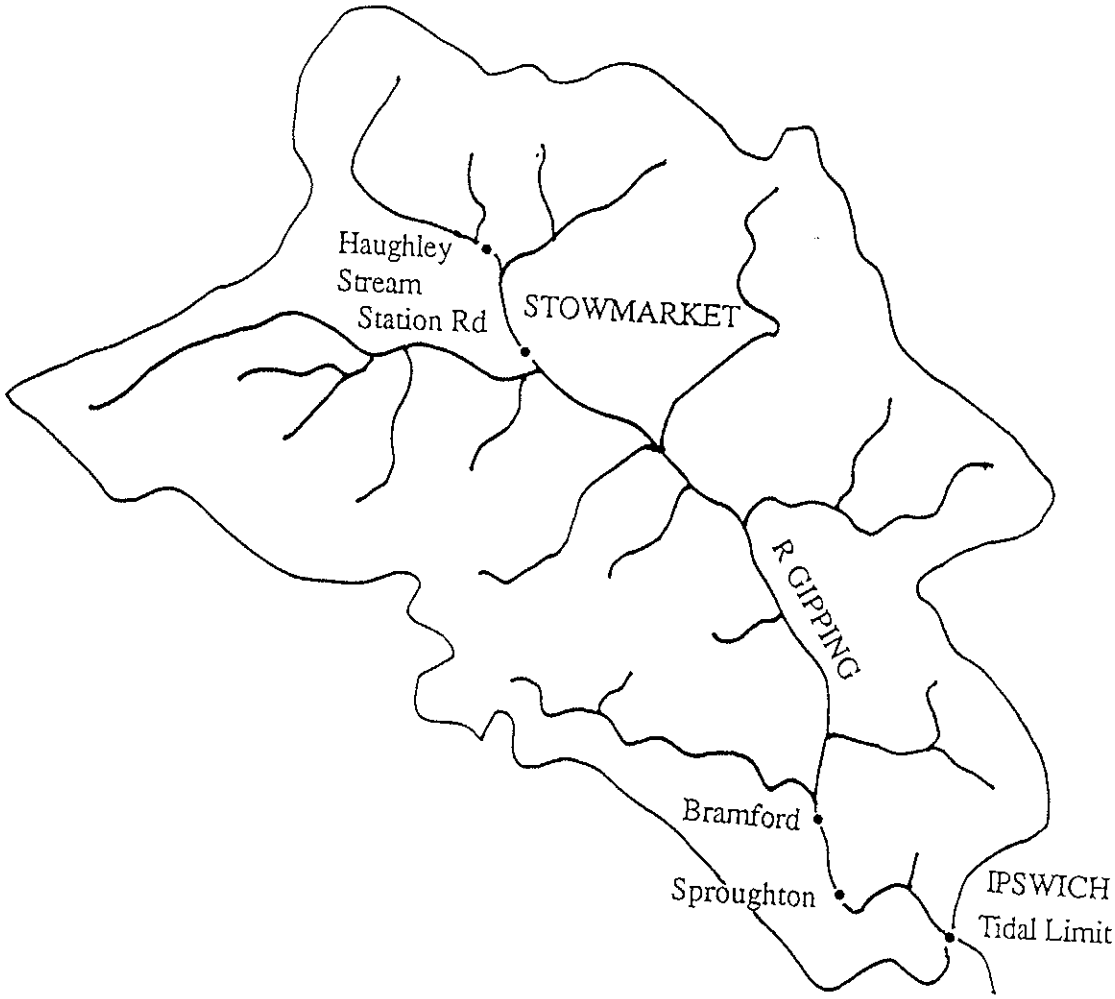
The Models

The leaching of nitrate from the soil and the consequent transport of the nitrate to and in the river were treated separately.

Leaching Model

A distributed leaching model DISLEACH was developed based upon a previous lumped leaching model MODLEACH (Oakes and Clark, 1993). A set of subcatchments is defined by the user along a hydrological basis. Separate annual time series of nitrate leaching are then calculated for each subcatchment, using the same rules as MODLEACH. These rules relate nitrate leached to crop type and fertiliser input, and to animal type, and have been derived from extensive drilling, sampling and field experimental programmes (Young and Gray, 1978; Lawrence Gould, 1985; Whitmore and Addiscott, 1986; Young, 1987; SSLRC, 1993). They represent a greatly simplified nitrogen cycle.

Figure 1 - Gipping catchment, Suffolk, UK.



For each subcatchment, the total nitrate leached, N kg ha^{-1} a^{-1} is given by

$$N = N_C + N_A \tag{1}$$

where the contribution from crops N_C is calculated by

$$N_C = \sum_{c=1}^{12} A_c f_c L_c \tag{2}$$

and the contribution from animals N_A is calculated by

$$N_C = \sum_{a=1}^{5} \frac{N_a}{R_a} P_a^{LU} \tag{3}$$

where
c = crop type (12 categories)
a = animal type (5 categories)
A_c = proportion of subcatchment occupied by land use *c*
f_c = fertiliser application rate (kg N ha^{-1} a^{-1}) for land use *c*
L_c = leaching rate or rule for land use *c*
N_a = number of animals of type *a* in subcatchment
R_a = number of animals of type *a* per Livestock Unit (LU)
$P^{LU}a$ = nitrogen production (kg N LU^{-1} a^{-1}) for animal type *a*

Flow and Transport Models

The flow and transport models for surface runoff transport, groundwater transport and river transport all operate on a weekly timestep, which was determined by the timescales of the physical processes, the availability of weekly meteorological data, and the output required from the models.

The meteorological and leaching time series are input for the surface runoff component, which provides recharge inputs to the groundwater module and runoff inputs to the river module. The groundwater module provides recharge inputs to the river module which combines these with the runoff and point sources and abstractions to predict river flow and quality time series along the river network. The nitrate calculations are carried out in parallel with the flow calculations.

The runoff and river models are based on a uniform square grid, and a GIS derived flow direction network, in which each node has a single downstream node to which runoff or river flow and hence nitrate can travel. Each node in the model can potentially have different hydrological and geological properties, derived from sources such as soil and geological maps. Thus these modules are spatially distributed in nature. The aquifer model takes the alternative lumped parameter approach for two reasons. The first was that issues of model run times and data requirements were important, and a spatially distributed groundwater module would affect both significantly. The second was that the aquifer generally provides a slowly varying component of the overall response, so the less detailed approach for this component was physically consistent.

The groundwater model used a relationship between baseflow discharge B and storage V developed for the Great Ouse resource model (Oakes and Keay, 1990; Oakes and Wilkinson, 1972):

$$B = cV \tag{4}$$

$$c = \frac{3.2T}{SL^2} \tag{5}$$

where

T = average aquifer transmissivity ($m^2\ d^{-1}$),
S = average aquifer storage coefficient,
L = mean catchment dimension from river to groundwater divide (m).

Solving mass balance for the aquifer gives the time variation in storage and hence baseflow as:

$$V_t = \left(V_0 - \frac{R}{c}\right) e^{-ct} + \frac{R}{c} \tag{6}$$

$$B = \frac{(V_0 - V_t)}{t} + R \tag{7}$$

where

t = time step (7 days),
R = net groundwater recharge rate ($m^3\ d^{-1}$),

V_t = aquifer storage at time t (m^3),
V_0 = aquifer storage at time 0 (m^3).

The equations are solved recursively as the model steps through time. The initial aquifer storage state is estimated as the long term average of R/c. The baseflow is uniformly distributed between each river node.

The groundwater nitrate model was based on conservation of mass within an active aquifer zone consisting of the storage volume V_t plus an additional active zone of depth *D*. Thus:

$$C_B = \frac{M_N}{V_T + AD} \tag{8}$$

$$M_{N_t} = M_{N_0} + \sum_{\text{catchment}} Q_R C - \sum_{\text{abstractions}} Q_A C_B - \sum_{\text{river}} BC_B \tag{9}$$

where:

M_N = mass of nitrate in aquifer (g),
C_B = nitrate concentration in aquifer (and baseflow) (mg l^{-1}),
A = catchment area (m^2),
D = active depth of aquifer beneath river level (m),
C = nitrate concentration of recharge (mg l^{-1}),
Q_R = recharge to groundwater (m^3 d^{-1}),
Q_A = abstraction rate from well (m^3 d^{-1}).

A delay mechanism on a proportion of the recharge was included to model twin speed routes for recharge: most percolating slowly through the unsaturated zone, and the remainder finding fast routes via fissures.

The surface runoff model is based upon a field capacity or soil water saturation capacity *SWS* and a minimum soil water holding capacity *SWH* for each model node. The runoff and recharge from a node decreases as the soil water holding capacity is approached. All water content surplus to the soil water saturation content is assumed to runoff immediately. (Alley, 1984; University of Birmingham, 1984; Mance, 1981; Oakes and Keay, 1990). A similar derivation to that for the groundwater model leads to the following equations for the runoff *SRO*, which is then split between surface runoff to the next node, and recharge to the aquifer in a ratio dependent on the geology beneath that node.

$$SWC_t = SWO_0 + \left(SWC_0 - SWH - \frac{R}{c}\right) e^{-ct} + \frac{R}{c} - X \qquad (10)$$

$$X = \max\left(0.0, \left(SWC_0 - SWH - \frac{R}{c}\right) e^{-ct} + \frac{R}{c}\right) \qquad (11)$$

$$SRO = \frac{SWC_t - SWC_0}{t} + R\ X \qquad (12)$$

where:

SWC = soil water content (at time t and time 0) (m^3),
R = HER (hydrologically effective rainfall) plus upslope surface runoff (m^3),
c = parameter reflecting rate at which runoff is generated (d^{-1}),
t = timestep (7 days),
X = surplus water content above saturation capacity (m^3).

The equations are solved recursively, with the initial soil water content set to the long term average value of R/c. In this way the surface runoff from each node is variously routed through the catchment until a river node is reached. Then the whole of the surface runoff so far accumulated enters that particular river node.

The nitrate in surface runoff and recharge are calculated in parallel to the flow calculations and are based upon conservation of mass, allowing for denitrification processes within the soil. The annual nitrate leaching is split in quarterly amounts in user defined ratios, then split evenly amongst the thirteen weeks of each quarter. However the actual mass of nitrate leached *WN* is dependent upon the hydrologically effective rainfall (HER) and a prescribed maximum nitrate concentration; if there is insufficient rainfall, then nitrate remains immobile and denitrifies within the soil. The nitrate concentration *C* (mg l^{-1}) and soil nitrate content (g) are given at the end of each timestep *t* (7 days) by

$$C = \frac{SN_0 + WN + RN}{SWC_0 + R} \qquad (13)$$

$$SN_t = SN_0 + WN + RN - C.SRO.rer - C.SRO.(1 - rer)e^{-at} \qquad (14)$$

where:

RN = nitrate from upslope runoff (g),
WN= nitrate from leaching for current week(g),
a = denitrification rate (wk^{-1}),
rer = the proportion of runoff which recharges groundwater.

Since mean weekly nitrate concentrations were the required output, and since the residence times and process timescales within the river were much shorter than this, steady state river flow and nitrate transport calculations were made for each weekly timestep. The flow calculations proceeded down the river network. The procedure took advantage of the fact that each node has only one downslope node, to define an ordering for the nodes such that flows or concentrations at every upstream node had been calculated before those for the current node. The surface runoff diffuse inputs, groundwater baseflow inputs and any point source inputs or discharges had also been calculated for the current timestep. Thus the flow at the current node is given by:

$$Q = \sum_{U} Q_U + \sum_{P} Q_P + Q_{SRO} + Q_B \qquad (15)$$

where:

Q_U = flow from an upstream node ($m^3\ wk^{-1}$),
Q_P = flow from a point source in current reach ($m^3\ wk^{-1}$),
Q_{SRO} = surface runoff ($m^3\ wk^{-1}$),
Q_B = baseflow ($m^3\ wk^{-1}$).

The steady state advection diffusion equation for the nitrate mass transport, including a first order decay model for denitrification processes and the appropriate source terms, can be written as

$$v\frac{dM}{dx} + M\frac{dv}{dx} = -KM + Q_B C_P + Q_{SRO} C_{SRO} + \sum_{P} C_P Q_P \qquad (16)$$

where:

M = mass of nitrate(g),
$K = K_0 10^{0.0293T}$, the in-river denitrification rate at temperature T (d^{-1})
C_P= nitrate concentration from point sources in current reach (mg l^{-1}),
C_{SRO}= nitrate concentration in surface runoff (mg l^{-1}),
C_B= baseflow concentration (mg l^{-1}),
v= river velocity at node (m s^{-1}).

which can then be discretised over the length dx associated with the current node to give an expression for M in terms of the known upstream flows and concentrations Q_U, C_U, and the nodal velocities v, v_U. A power law relationship $v = aQ^b$ is used to relate the velocity

to the flow rate (Nixon, 1959; Whitehead and William, 1984). This equation is solved in parallel with the flow calculations as the model proceeds down the river network from the catchment boundaries to the single outlet of the river network (at the tidal limit).

Implementation

The models described above were implemented as a PC based FORTRAN code, with graphical displays of flows and concentration as the run progresses. Text output files could then be further processed to give animated map displays of the catchment and river flows and nitrate concentration.

The Data

The data required by the models described above can be divided into **catchment data**, required to specify the extent, topography, geology, and soil characteristics of the catchment, and **scenario data**, required to describe the situation to be modelled. This latter category includes data such as historical and future land-use and fertiliser regimes, point sources and abstractions, meteorological time series data and also run details such as length and output requirements.

For the present application to the Gipping, a GIS digital terrain model was used to define the catchment extent and to derive a flow direction network on a 350 m grid. The GIS was also used to map parish boundaries onto this model grid. The 53 parishes within the catchment were grouped into nine subcatchments for the leaching model calculations. The GIS was finally used to allocate one of three geologies - boulder clay, chalk or sands/gravels - to each node, based on a simplified digitised map of the local drift geology.

Agricultural census data (parish statistics) for the groups of parishes comprising each subcatchment were obtained for 1988 and 1992 from ADAS. Typical fertiliser application rates for arable farms in Eastern England were derived from the 1974, 1979 and 1991 Surveys of Fertiliser Practise (ADAS/FMA). A five year time series (1988 - 1992) of HER for the MORECS grid square containing the Gipping catchment was obtained from the Meteorological Office.

Comparing the parish statistics for the catchment as a whole for 1992 and 1988 shows a 19% drop in livestock units, an apparent replacement of about 4000 ha of spring cereals by winter cereals and a 55% increase (720 ha) in the area growing oil seed rape.

Point source abstraction and discharge flow and quality data, along with flow gauge and river quality monitoring data were obtained from the NRA and the Anglian Water Services archives for the five years modelled. There were daily flow data from the Bramford gauging station, which is situated about 7 km upstream of the tidal limit. Nitrate or total oxidised nitrogen (TON) concentration time series were available for 21 sites spread along the length of the Gipping, from Haughley Stream near its head to Sproughton Weir

downstream of Bramford.

The Results

Table 1 gives the calibrated values for the various model parameters. Figure 2 shows the predicted and gauged discharge time series at Bramford. The root mean square error for the model fit to observed, which is dominated by the peak flow values, is 0.56 $m^3 s^{-1}$. The flashy winter peak flows are closely correlated to rainfall events shown in the HER time series. The slowly declining low summer flows are almost entirely baseflow, showing the strong influence of the groundwater on the river. The baseflow can be seen to decline from year to year during the model run. This reflects the very dry period experienced from 1989 - 1992. Figure 3 shows the HER and mean river temperature weekly time series from 1983 to 1992. The low and brief winter rains in the latter part of the graph are readily apparent.

The river nitrate predictions were calibrated against twelve monitoring sites to gain the best catchment wide fit. The mean groundwater nitrate concentration from the four abstraction sites was observed to vary little around a value of 8 mgN l^{-1}. The predicted concentration rose slowly from 8 mg l^{-1} to 8.3 mgN l^{-1} over the five years, reflecting the declining storage volumes but continued nitrate inputs over the prolonged low recharge period. The measured river nitrate concentrations are observed to peak during winter at about 25 mgN l^{-1} (compared to a drinking water limit of 11.3 mgN l^{-1}), to drop during summer months to about 5 mgN l^{-1} in the upper catchment and to 2 mg N l^{-1} in the lower reaches, and to be closely correlated to river flow and hence rainfall events. Figures 4 and 5 show examples of predicted and observed concentration time series at Station Rd Bridge, in the upper catchment at Stowmarket, and at Spoughton in the lower catchment. Note that the Station Rd Bridge observed data are monthly spot samples, rather than the weekly mean values modelled. The modelled fits for the twelve sites were generally good, capturing the main features of the observed time series noted above, and the timing of individual nitrate peaks. Poorer fits obtained immediately downstream of Stowmarket sewage works were attributed to insufficient knowledge of the nitrate inputs from this source. A consistent overprediction of 1988/89 winter nitrate concentration was eventually attributed to (anomalous and unexplained) low observed nitrate concentrations during that winter, compared to previous and following winters.

Discussion

There are several points of interest arising from this study, which are discussed below.

For this study the available water quality (nitrate) data covered far more of the catchment modelled than the single gauging station. This is likely to hold for other catchments. The close correlation of observed nitrate concentration variation at each site with the HER variation, together with the correlation of nitrate concentration and flow at the single gauging station, led to the use of the nitrate data to add confidence to the predictions of flow rates in the upper catchment. The correlation between peak HER, river flow and river nitrate concentration modelled (and measured) indicate that these river nitrate peaks are the result of surface runoff carrying leached nitrate from the catchment soils.

Table 1: Calibrated Values for Model Parameters

Parameter	Value
Initial soil water content SWC_0(-)	0.2
Runoff-soil water content parameter c (d^{-1})	0.5
Recharge:runoff ratio *rer*- Clay	0.10
Recharge:runoff ratio *rer*- Chalk	0.95
Recharge:runoff ratio *rer*- Sand/gravel	0.5
Soil denitrification rate a (d^{-1})	0.01
Quarterly leaching split Spr:Sum:Aut:Win	0.4:0.1:0.1:0.4
Maximum nitrate concentration in soil (mgN l^{-1})	250
Initial soil nitrate concentration (mgN l^{-1})	8
Initial aquifer volume V_0 (m^3)	5.6E6
Baseflow-Storage parameter c (d^{-1})	0.0015
Aquifer depth beneath river level D(m)	10
Slow recharge lag (y)	7
Proportion recharge 'quick'	0.1
Initial groundwater nitrate concentration (mgN l^{-1})	8
River velocity-flow parameter a (m^{-1})	0.61
River velocity-flow parameter b (-)	0.33
River denitrification rate K (d^{-1})	0.9
Initial river nitrate concentration (mgN l^{-1})	5

Thus the impact of varying land use and fertiliser practise could be usefully studied. A longer time series of groundwater concentrations, stretching back say 30 years would be required to establish the groundwater nitrate trend and its impact on future summer (minimum) river nitrate concentrations. It was found that the groundwater abstraction rates had a significant impact on the baseflow predictions and hence river flows under the very dry conditions being modelled. Groundwater abstractions resulted in negative net recharge rate during most summer months, but total predicted baseflow remained constant.

The summer river nitrate concentrations must be derived almost entirely from baseflow, yet drop from 60% to 25% of the groundwater concentration as one progresses down the system. The residence time within the Gipping, at low summer flow of 0.1 $m^3 s^{-1}$, is less than two days. The drop along the river system therefore suggests that rapid in-river denitrification or nitrate uptake processes are occurring. Similar uptake processes seem to occur in the river bed, to account for the 40% drop from groundwater nitrate concentration even in the upper reaches of the catchment.

Figure 2 - Predicted and observed weekly time series of mean discharge at Bramford gauging station, 1988-1992

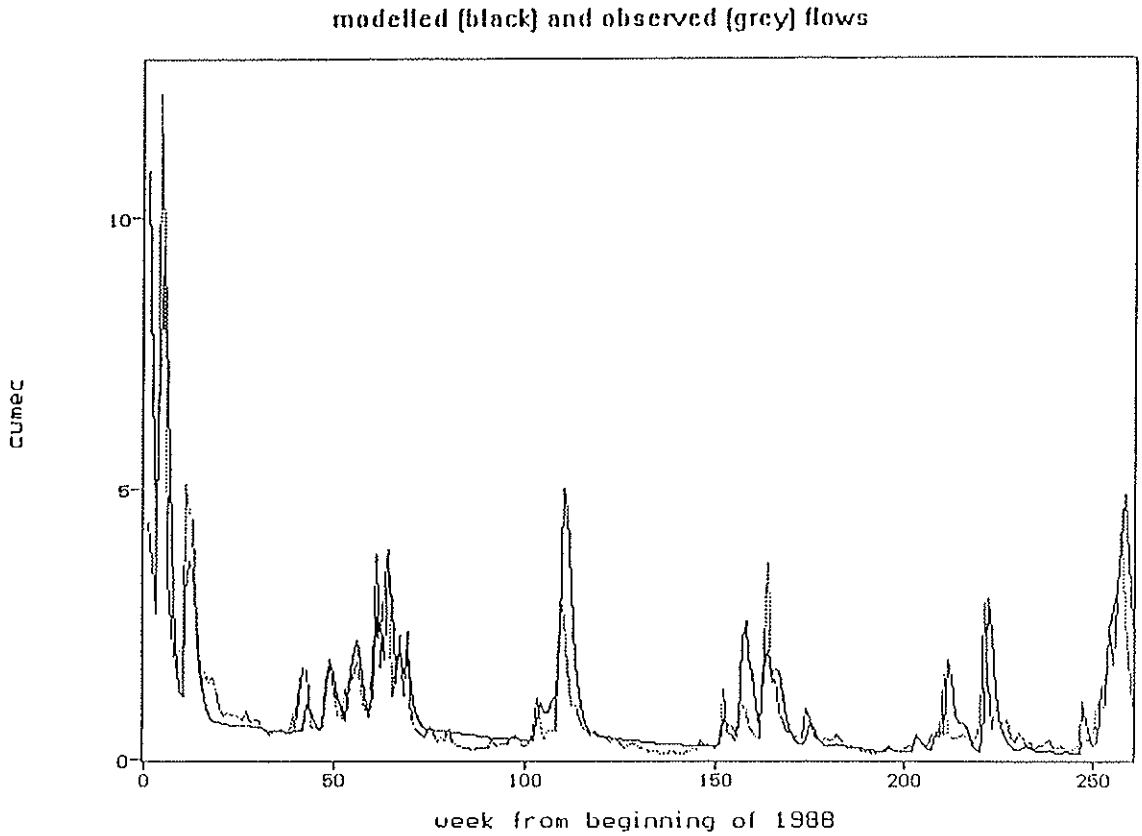


Figure 3 - Weekly HER (hydrologically effective rainfall) and temperature time series, 1983-1992

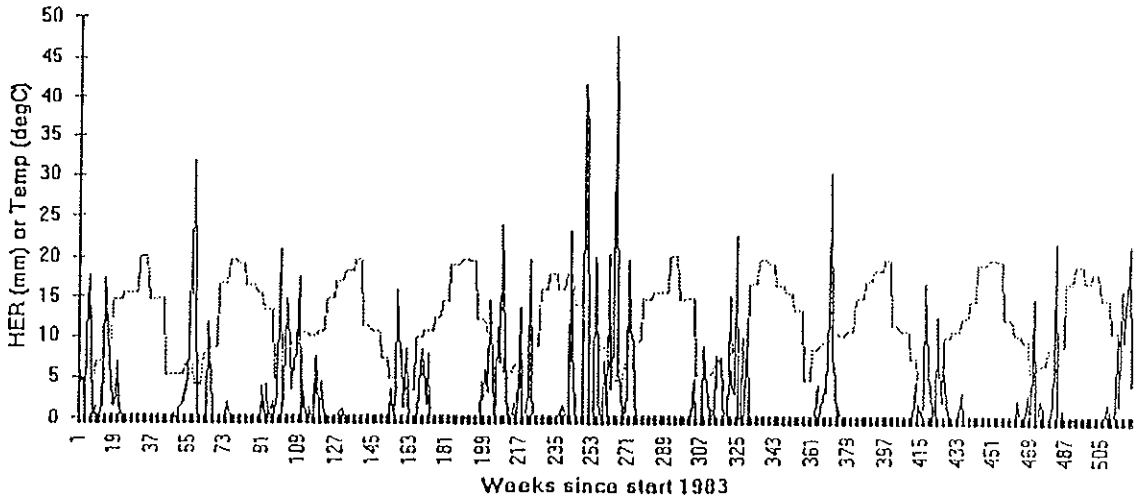


Figure 4 - Weekly mean predicted nitrate concentration (mgN l^{-1}) at Station Rd bridge, Stowmarket, with measured monthly spot values

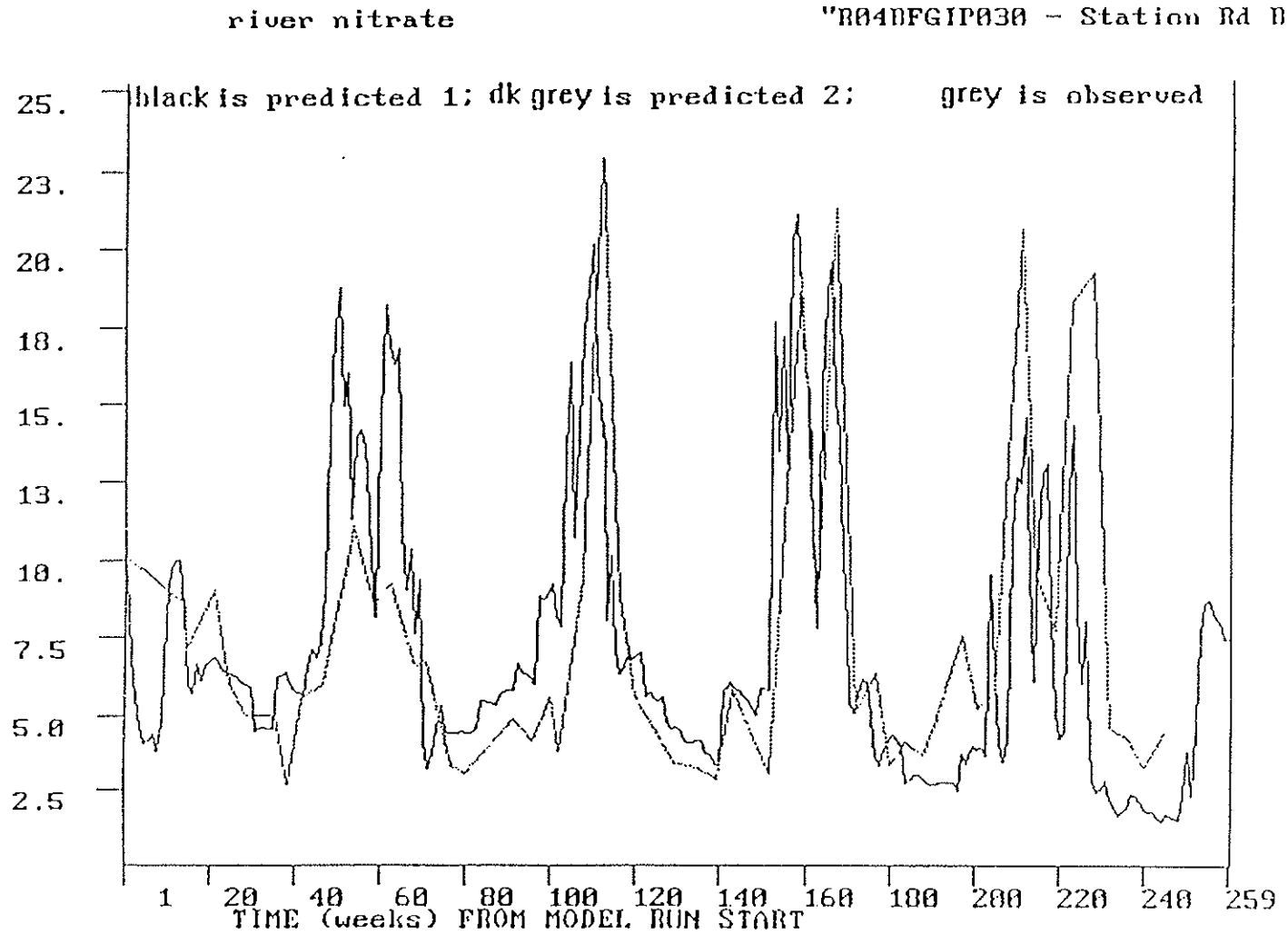
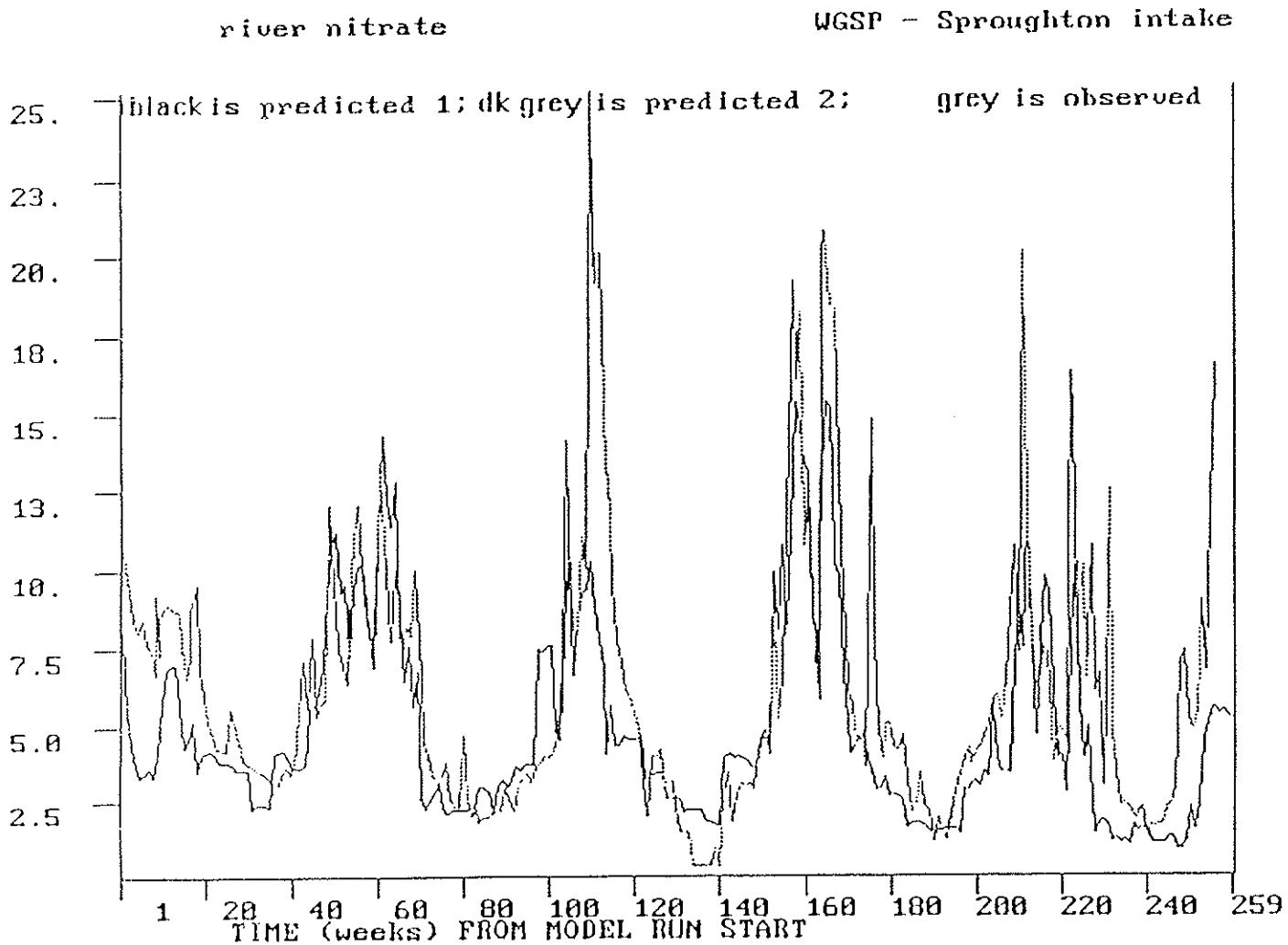


Figure 5 - Weekly mean predicted nitrate concentration (mgN l^{-1}) at Spoughton, with weekly mean measured values



Conclusions

1. For the Gipping catchment, the rapid reaction of river nitrate concentration and flow to rainfall events means river nitrate concentrations are very dependent on relative timing of nitrate application to land and rainfall events.

2. Low nitrate concentrations in river in summer compared to groundwater, despite groundwater providing all summer river flow, implies rapid denitrifying or nitrate uptake mechanisms are occurring in the Gipping and its bed.

3. The greater number and spatial spread of measured nitrate time series compared to flow gauging data, together with strong correlation of river flow and nitrate concentrations, meant that nitrate data played a valuable part in the calibration of and the increased confidence in the river flow model. Thus quality and flow data together can both be useful for model calibration.

4. The relatively simple lumped parameter groundwater model still was sufficient to model the interaction of aquifer, distributed catchment recharge and river processes so as to simulate the summer flow and quality processes in a small river catchment over a period of five years.

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References

ADAS/FMA (annual publication) **Survey of fertiliser practise** (fertiliser use in farm-type regions of England and Wales).

Alley, W.M., 1984, On the treatment of evapotranspiration, soil moisture accounting and aquifer recharge in monthly water balance models. **Water Resources Research. 20,** 1137-1149.

Clark, K.J., and Oakes, D.M., 1994, Final Report: ROSANN - River Nitrate Model, WRc Report UC 2344, June 1994.

Keating, 1992, A lumped parameter model of a chalk aquifer / stream system in Hampshire, United Kingdom. **Ground Water**, 20, 430-436.

Lawrence Gould Consultants Ltd, 1985, Draft working notes and analyses. Lincolnshire catchment area. Prepared for meeting with the DoE.

MAFF, 1989, (also 1973, 1979, 1983, 1985), Fertiliser recommendations. Reference Book 209, HMSO

Mance, G., 1981, A deterministic model of the quantity of nitrate released to a river directly from the catchment soil zone. Water Research Centre Report 190-M

Nixon, M., 1959, A study of bankfull discharge of rivers in England and Wales. **Proceedings of the Institution of Civil Engineers, 12,** 151-174.

Oakes, D.B., and Wilkinson, B., 1972, **Theoretical relationship between groundwater abstraction and baseflow.** Water Resources Board, Reading

Oakes, D.B., and Keay, D., 1990, A resource model of the Great Ouse river system, WRc Report CO2504-M

Oakes, D.B., and Clark, K.J., 1993, Prediction of groundwater nitrate concentrations, NRA R&D Final Report 110/11/ST

SSLRC, 1993, Prediction of groundwater nitrate: nitrate leaching studies, NRA Draft Final Report 110/8/ST

University of Birmingham, 1984, **Groundwater modelling investigation of River Gipping area of Suffolk.** Final Report, Departments of Civil Engineering and Geological Sciences, University of Birmingham.

Whitmore, A.P., and Addiscott, T.M., 1986, Computer simulation of winter leaching loss of nitrate from soils cropped with winter wheat. **Soil Use and Management, 2,** (1).

Whitehead, P.G., and William, R.J., 1984, Modelling nitrate and algal behaviour in the River Thames. **Water Science and Technology,** 16, 621-633.

Young, C.P., 1987, Nitrate in groundwater and the effects of ploughing on the release of nitrate. In **Effects of land use on freshwaters,** ed. J.F. del L.G. Solbe, Water Research Centre.

Young, C.P., and Gray, E.M., 1978, **Nitrate in groundwater.** Water Research Centre Technical Report TR69.

MODELLING THE EFFECTS OF VARIABLE SEDIMENTATION RATES ON STREAM-AQUIFER INTERACTIONS

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Abstract

When the dynamic equilibrium of alluvial rivers is disturbed, they tend to adjust to new pseudo-equilibrium conditions by depositing sediments or eroding the bed. Such adjustments may have significant effects on stream-aquifer interactions. A mathematical model (Model of Alluvial Stream Aquifer Interactions - MASAI) has been developed to enable *a priori* assessments of such interactions. Application of MASAI to a hypothetical stream-aquifer system reveals the complex relationships between individual elements of the system, and highlights the influence of rates of sediment erosion and deposition on river-aquifer exchange flows.

Introduction

Most alluvial rivers only partially penetrate their underlying aquifers, yet the transfer of water to or from the aquifers can have a significant effect on flow in the rivers. Furthermore, alluvial rivers normally transport sediments of various sizes, and if the dynamic equilibrium of such rivers is disturbed (for instance by dredging or sand and gravel mining from the bed), these rivers tend to adjust to new pseudo-equilibrium conditions by depositing the sediments or scouring the bed (Kornis and Laczay, 1988); either of these sedimentary processes can have additional effects on the interaction with a hydraulically connected aquifer. Many case histories of river adjustments to disruption, involving substantial changes in aggradation or degradation, can be found in the literature (eg Walling, 1983; Chang, 1988).

The water interchange between streamflow and groundwater is primarily a function of the difference in water levels between the aquifer and the river. The actual magnitude of the flow depends on the local geology, particularly the hydraulic conductivity of the interface boundary layer at the bottom of the river. Detailed information about the spatial variations of this boundary layer is difficult to obtain, and further difficulties arise for dynamic channels whose hydraulic and fluvial conditions are continuously changing.

Physically-based mathematical modelling affords the opportunity to explore the fundamental relationships between various elements of river-aquifer systems, and it also offers an opportunity to study short and long term hydrologic responses due to changes in the hydrologic systems (cf Abbott et al, 1986). To simulate the water interchange between alluvial rivers and floodplain aquifers, it is necessary to develop a mathematical model that estimates the variation of both groundwater and surface water

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levels, and the transient adjustments arising from sedimentation. The objective of this paper is to describe a model which incorporates these features, and to outline the insights into stream-aquifer interactions gained by application of the model.

Model Formulation

Prior to the inception of the work reported here, models of stream-aquifer interactions had previously been restricted to stable regime river problems, in which it was reasonable to assume fixed bed elevation profiles (eg Cunningham and Sinclair, 1979). Similarly, water and sediment routing simulations in erodible channels have previously been performed assuming groundwater contribution as a pre-defined lateral flow obtained as a calibrated parameter (eg US Army Corps of Engineers, 1977). When streamflow routing is performed for rocky basins, with river beds of very low permeability, or when baseflow occurs to rigid channel reaches (with very low sediment transport or storage changes), then errors introduced by adopting these standard assumptions will not be great. However, when a modelling exercise is directly concerned with conjunctive water use in alluvial stream-aquifer systems whose sediment transport/storage is not in equilibrium, these errors may sufficiently large as to invalidate the exercise. Considering that most hydrologic processes in stream-aquifer systems are highly sensitive to the streambed characteristics (Younger et al, 1993), then enhancement of existing modelling procedures to reflect the time-dependent streambed boundary condition is a worthwhile endeavour.

The mathematical model created during this study is known as MASAI (Model of Alluvial Stream Aquifer Interactions). MASAI is amenable to calibration and application for studies in planning, design and operation of riverine groundwater recharge, soil and water conservation, and other related transient alluvial stream-aquifer processes. The mathematical formulation of MASAI involves simultaneous numerical solution of the Saint Venant equations and the sediment continuity equation (describing stream flow and sediment transport processes), and the linearised Boussinesq equation (describing 2-D transient groundwater flow). The solutions for the surface water and groundwater systems are linked through a time-dependent interface boundary condition based on Darcy's Law (Mwaka, 1995). Further details are given below.

Streamflow Routing

Streamflow routing describes the temporal and spatial variation of the stage, discharge, energy gradient, and other hydraulic variables in a channel. The differential equations describing streamflow routing are based on the continuity and momentum equations of water (Equations 1 and 2 below). Their detailed derivation can be found in the literature (eg Strelkoff, 1969; French, 1986):

$$\frac{\partial Q}{\partial x} + \frac{\partial A}{\partial t} - q = 0 \qquad (1)$$

$$\frac{\partial Q}{\partial t} + \frac{\partial}{\partial x}\left(\frac{Q^2}{A}\right) + gA\frac{\partial H}{\partial x} - gAS_0 + gAS_f - qu_1 = 0 \qquad (2)$$

where Q is the discharge, A is the cross-sectional area of flow, t is the time, x is the distance along the discharge centreline, H is the depth of flow, q is the lateral discharge per unit length (which in this case is limited to groundwater seepage), u_1 is the velocity component of lateral inflow in the x-axis direction, g is the acceleration due to gravity, S_f is the energy gradient and S_o is the bed slope.

Solutions for these equations are usually obtained using numerical methods. The particular implicit finite difference formulation adopted in MASAI is described elsewhere in the literature (eg Amein and Fang, 1970; Chang, 1988). However, other simplified routing methods could feasibly be applied, so long as they described the variation of the river levels at the required spatial and temporal resolution (see Todini, 1991).

Sediment Routing

Sediment routing is applied to compute the changes in the sediment transport, bed elevation and bed-slope along the channel at each time step. The differential equation describing the sedimentation is based on the sediment continuity equation (Equation 3; derivation details are given by Chen, 1973).

$$\frac{\partial Q_s}{\partial x} + (1 - \lambda)\frac{\partial A_b}{\partial t} - q_s = 0 \qquad (3)$$

where Q_s is the sediment load, λ is the porosity of bed material, q_s is the lateral inflow rate of the sediments per unit length, and A_b is the cross-sectional area of the channel which results from accumulated erosion or deposition. The sediment load Q_s can be estimated by appropriate sediment functions (eg Ackers and White, 1973) and a numerical solution obtained (see Chang, 1988). However, noting the scarcity and uncertainty of data for calibration of models involving sedimentation, it may be appropriate to estimate the streambed conditions by stochastic methods, which can describe the variation of the streambed levels in both time and space (see Bathurst and Wicks, 1991).

Groundwater flow

The equation for mass continuity and Darcy's Law can be combined, and integrated over the saturated thickness of an unconfined aquifer (with an impermeable lower boundary), to give the Boussinesq equation for 2-D groundwater potential and flow in the floodplain aquifer:

$$T\left(\frac{\partial^2\psi}{\partial x^2} + \frac{\partial^2\psi}{\partial y^2}\right) = S_y \frac{\partial\psi}{\partial t} + I(x,y,t) \qquad (4)$$

and for the part of aquifer beneath the river, Equation 4 is modified to:

$$T\left(\frac{\partial^2\psi}{\partial x^2} + \frac{\partial^2\psi}{\partial y^2}\right) = S_c \frac{\partial\psi}{\partial t} + q\left(x,t,K_1,\Delta H,\Delta Z_1\right) \qquad (5)$$

where Ψ is the groundwater potential, T ($=K_2m$) is the aquifer transmissivity, K_2 and m being the hydraulic conductivity and the saturated thickness of the aquifer, respectively, S_c is the aquifer storativity, S_y is the specific yield of the aquifer, q is the seepage from the river, Z_1 is the thickness of the river-bottom layer, K_1 is the hydraulic conductivity of the river-bottom layer and I is the vertical infiltration into the unconfined aquifer. Detailed derivation of Equations 4 and 5 can be found, for example, in Bear (1972). The numerical solution applied to solve these equations is also widely known (eg Wang and Anderson, 1982).

Coupling the Streamflow and Groundwater Solutions.

MASAI is an internally coupled stream-aquifer model (in the terminology used by Younger, *this volume*). The formulation used builds logically upon earlier internal coupling techniques, first introduced by Pinder and Sauer (1971), by including an innovative representation for mobile river beds.

The equation for seepage through n layers of porous media can be obtained from Darcy's expression relating the total hydraulic head driving a flow (q) through the layers, as:

$$\sum_{i=1}^{n} (\Delta H)_i = q\sum_{i=1}^{n} \frac{Z_i}{K_i} \qquad (6)$$

where K_i = hydraulic conductivity of individual layer, Z_i = thickness of individual layer, $(\Delta H)_i$ = head difference across each layer and n = total number of layers.

Considering a losing stream, and assuming that seepage from/into the river is predominantly vertical through the channel bed (Figure 1), then Equation 6 can be rearranged as:

$$q = \frac{\sum_{i=1}^{n} (\Delta H)_i}{\sum_{i=1}^{n} \frac{Z_i}{K_i}} \qquad (7)$$

This assumption of predominant vertical flow is generally reasonable (Sharp, 1977), since:
(1) the river bed has a greater surface area than the banks,
(2) the bed is closer to the zone of higher hydraulic conductivity in the aquifer (since K increases with depth in most alluvial aquifers), and
(3) maximum pressure head acts on the bed.

In MASAI two porous layers are assumed: A bed sediment layer over an alluvial aquifer (Figure 1). Then,

$$q = \frac{k_1 k_2}{k_1 + k_2} \sum_{i=1}^{2} (\Delta H)_i \qquad (8)$$

where Σ (ΔH) is the total potential head across the two media, k_1 and k_2 are coefficients given as $k_1 = K_1/Z_1$, $k_2 = K_2/Z_2$, with the hydraulic conductivity as K_1 and K_2 and the thicknesses as Z_1 and Z_2 for the sediment layer and aquifer, respectively.

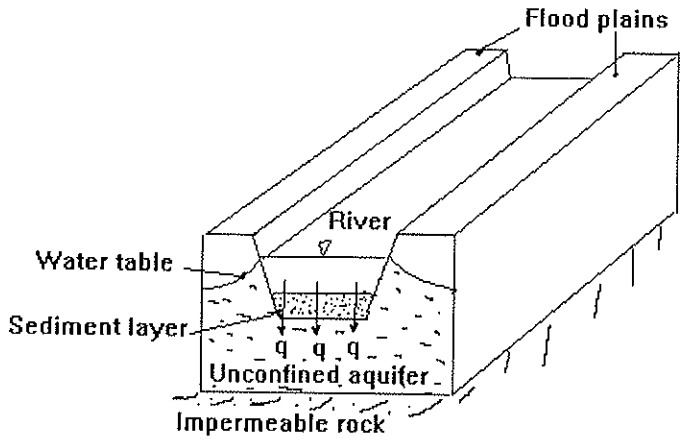
If the specific sediment seepage resistance is defined as $\zeta = 1/K_1$ so that $k_1 = 1/\zeta.Z_1$, and assuming that the pressure head in the aquifer at the reference point remains the same as ψ_a, so that $\Sigma (\Delta H)_i = (H + Z_1 + Z_2 - \psi_a)$, then:

$$q = \frac{k_2}{1 + \zeta Z_1 k_2} \left(H + Z_1 + Z_2 - \psi_a\right) \qquad (9)$$

Since the solution to this equation is dependent on contemporary values of the streamflow, streambed and groundwater conditions, it can be applied to account for the transient seepage flows and to link the open channel and groundwater hydrology. The equation summarises the modifications introduced in the theory of stream-aquifer interaction in order to relax the traditional assumption of boundary rigidity for alluvial channels that are hydraulically connected to their underlying aquifers. It is also worthwhile to notice that in this linkage mechanism, the new streambed thickness due to sedimentation (Z_1) is represented independently from the aquifer thickness (Z_2) so that in the event of aggradation, the different hydraulic properties due to the exotic sediment accumulations can still be accounted for exclusively, by assigning Z_1 and k_1 with the appropriate values. Even when the sediment deposit is washed away such that $Z_1 = 0$, the equation is still applicable, save that in this case the sedimentation variation will be affecting the aquifer thickness itself. The lateral groundwater flow into or out of the river can thus be evaluated according to the following procedure:

(a) The initial hydraulic variables H^o, ψ^o and Z_1^o at time t^{j+0} are used to compute a steady seepage flow q^o, which is applied in the routing solution to evaluate the new water conditions after a short time period Δt.

Figure 1 - A schematic sketch of a losing stream



(b) At this time, (t^{j+1}), the new computed values H^1, Z_1^1 and ψ^1 (provided by the water routing, sediment routing and groundwater flow modules respectively), are applied to compute the new groundwater transfer q^1, for routing in the next time step.

(c) The procedure is then repeated until the total time period under consideration is covered.

This modified solution procedure has more physical sense in erodible channels that carry sediment laden floods than the more usual internal coupling technique introduced by Pinder and Sauer (1971) for rigid-boundary channels.

Analysis of the Effects of Sedimentation on Baseflow.

Rationale for Analysis.

MASAI is suited to the evaluation of soil and water conservation practices on riverine plains, inasmuch as changes in land-use and/or river engineering can affect streamflow and sediment transport. In fact, streamflow and sediment loads (in their various aspects) are recognised diagnostic criteria for assessment of watershed conditions, and thus of the effectiveness of watershed and river management strategies. Total yearly amounts of both water and sediment loads, seasonal regularity, and frequencies and extremes of high and low discharges provide useful indications of the net result of conservation projects.

Analysis of baseflow and sediment discharges can provide a basis for predicting the

hydrologic and fluvial impacts of management practices on stream-aquifer systems. Variation of either water flow and/or sediment yield in turn affects sediment transport through a reach. Whenever the sediment load transported into a given reach amounts to less than its transport capacity, the channel scours, whereas when the incoming load is greater than its transport capacity, the channel bed fills. Simulations based on the MASAI concept are capable of tracking the scour/evolution of the channel bed, along with their influence on the baseflow contribution. In this study, however, only the transient evolution/regression of the river bed elevation and slope are tracked in the simulations. The other boundary parameters (eg hydraulic conductivity and Manning's roughness), which are also known to influence baseflow, have not been included explicitly in the simulations. It is nevertheless possible to correlate them as functions of the depth and nature of the alluvial accumulation (eg Sharp, 1977; Alam and Kennedy, 1969).

Simulation of the Effects of Sediment Yield on Baseflow Recession Curves

A baseflow recession curve is a plot showing the variation of baseflow contribution with time during periods of groundwater drainage into a river. In essence it is a measure of the rate of drainage of groundwater storage from the adjoining aquifers. Previous analyses of streamflow hydrographs have produced idealised recession curves according to the decay relationship of Equation 10 (Wilson, 1990):

$$Q_b = Q_0 \cdot e^{-\alpha t} \qquad (10)$$

where Q_b is the baseflow after a time period t, Q_o is the baseflow at the start of the time period, α is the recession constant representative of the system responses and e is the natural logarithm base. Any phenomenon that changes the pressure gradient between the aquifer and the river will cause the baseflow to vary. Passage of flood waves through rivers and diffuse recharge into aquifers are among the major causes of baseflow variability: the effects of both of these factors on baseflow recessions have been previously investigated by Todd (1955) and Singh (1968), respectively. Likewise, any sedimentary layer accumulating between the river and the aquifer media will increase the pressure gradient and thus affect the seepage resistance between the two domains. Clearly therefore, time-dependent channel sedimentation has the potential to cause not only evolution/regression of the riverbed boundary layer (cf Pearce, 1994) but also waxing and waning of river-aquifer exchange flows.

In this demonstration, baseflow recessions are generated in order to study the influence of sedimentation processes on the rate of groundwater depletion from the neighbouring aquifers. The experiments are based on the conceptual hydrologic system defined by Pinder and Sauer (1971), whose model was constructed to investigate the modification of a flood wave due to bank storage effects. The flood plain aquifer extended 42672 m along the length of the channel and 427 m across the valley. It was surrounded by impermeable materials on all sides. The initial saturated thickness ranged from 67 m at the upstream boundary to 27 m at the downstream boundary. The streamflow was along the axis of the valley through a straight channel with a constant cross-section and slope of 0.001. The hydraulic conductivity of the aquifer (assumed homogeneous and isotropic) was 0.003 m/s. The channel was 30m wide and the initial flow was 6m deep.

In this example, an inflow hydrograph was routed through the system by maintaining a flood flow of 500 m^3/s for three days in order to acquire steady conditions, and then suddenly reducing the flow to a baseflow of 200 m^3/s, where it remained throughout the experiment. The river was sub-divided into 10 smaller reach elements, and the interchange amounts were the net seepage flows into the river over each of the elements. By modelling the system responses under these conditions it was possible to isolate and study the behaviour of baseflow recession curves at any desired location. Results illustrated by the curves in Figures 2 and 3 are for the downstream boundary section, and are typical of those obtained using a variety of system parameters, save for the rates of the recessions which vary with every combination of the parameter in question.

It is observed that different sedimentation parameters of the pseudo-equilibrium channel produced different rates of baseflow recession (Figure 2). In addition, the experiments have demonstrated that the simulated curves, which in this case included sediment routing effects, did conform to the theoretical exponential decay relationship given by Equation 10. (Where a hydraulic conductivity of zero is assumed for the river bottom, baseflow is precluded and only channel flow and storage are considered).

The first set of results (Figure 2) was obtained by holding the upstream input sediment concentration at 5000 ppm, while varying the hydraulic conductivity of the river bed between zero and 1 x 10^{-4} m/s. In the second experiment (Figure 3), the hydraulic conductivity of the streambed sediment layer was held at 1 x 10^{-4} m/s, while the concentration of the upstream input sediments was varied between zero and 25000 ppm.

To interpret the nature of these relationships, it is necessary to recall Equation 9, which describes the coupling of the alluvial stream-aquifer system. For instance, with the hydraulic conductivity of 1 x 10^{-5} m/s (Figure 2), the baseflow recession is dissipated more slowly because of the higher seepage resistance across the interface sediment layer. In the presentations, the baseflow discharges are plotted against the time elapsed since the upstream flood flow was suddenly reduced from 500 to 200 m^3/s. The groundwater depletion is most clearly observed after the dissipation of channel storage effects, ie after the 10th hour. The slight rise in the recession curve at about the 15th hour indicates the time when the maximum baseflow contribution from bank storage arrived at the downstream station, coming after the channel storage had ceased in the 10th hour.

Because seepage resistance increases proportionally with the thickness of the sediment layer, then as the thickness is increased through sediment accumulations, the resistance to the stream-aquifer water exchange is also increased. The simulation results in Figure 3 provide more insights into the inter-relationship between baseflow recession curves and sediment yield, whose capacity is varied through the input sediment concentration. It is demonstrated that with higher bed load input, less groundwater is depleted because of faster builder-up of the seepage resistance by the evolving river bed. Increasing the sediment concentration correspondingly increases the rate of the bed layer evolution, which induces not only the seepage resistance, but also reduces the hydraulic gradient for baseflow towards the river. That is why the baseflow contribution decreases with

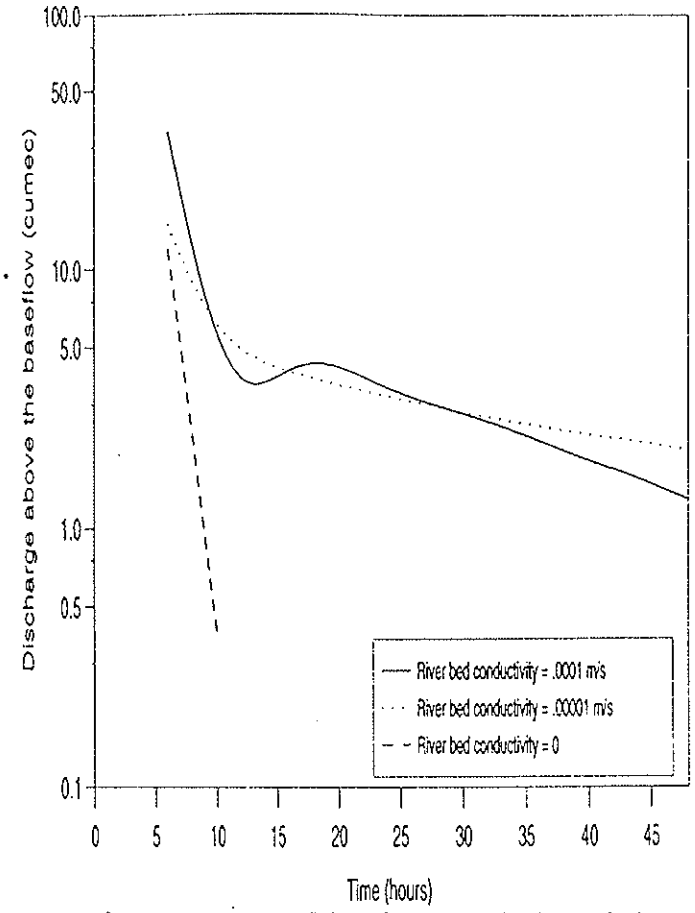


Fig. 2 Baseflow recession curves resulting from variation of river-bed conductivity

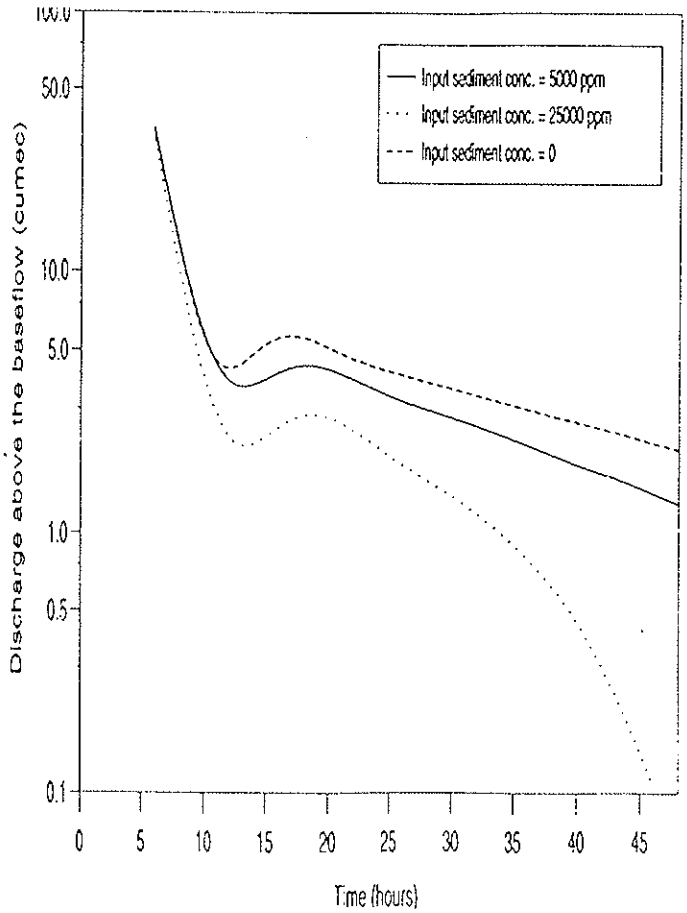


Fig. 3 Baseflow recession curves resulting from variation of sediment yield

increasing concentration of the sediment yield into the reach. It is also noteworthy that, at certain levels of sediment yield input, the baseflow recession curves no longer conform to the logarithmic decay curve relationship (Equation 10). Singh (1968) presented similar findings for a river that was continuously losing water into adjacent aquifers. He did analytical studies on the effect of aquifer leakage conditions (as a result of high groundwater withdrawal, evaporation or deep percolation leakage), whereby the traversing river became more and more influent, resulting in baseflow recession curves that were steeply deviating from the theoretical exponential relationship described by Equation 10. Here the deviation arises from the river bed evolution, which correspondingly increases the river stage and the hydraulic gradient away from the aquifer, making the river become more and more influent, and therefore presenting the same feature of steeply deviating recession curve from the theoretical exponential relationship. Problems associated with this type of change of streamflow hydrology include development of waterlogged riparian conditions, declining streamflow discharge in downstream reaches, etc.

With the assumed model conditions, the extent of bed scour was not very marked, and so the flattening deviation generated with streamflow of zero sediment concentration is not dramatic. Nevertheless, the deviation is still notable from the increasing gap between the zero and 5000 ppm curves. In the case of river bed degradation, it can be explained that the hydraulic gradient towards the lowered river stage is aggravated, and the baseflow recession curves deviate from the idealised exponential relationship, this time becoming flatter as the river becomes more and more effluent. Problems associated with this type of change of streamflow hydrology include excessive drainage of adjacent groundwater reserves, instability of river banks and neighbouring infrastructure, etc.

Based on these interrelationships between sediment yield and baseflow, it can be conjectured that rivers from highly erodible watersheds may have diminishing streamflow discharges due to in-stream aggradation, which consequently induces the rivers to become more and more influent. If the water leaving the influent streams is lost through evaporation, deep percolation etc, then the net amount of discharge arriving in downstream reaches will be reduced. Nonetheless, it seems this is the principle upon which groundwater recharge and flood control by check-dams is based. That is, the check-dams are made to decelerate the streamflow and cause the upstream river-stage to rise, and the corresponding bed elevation to aggrade; thereby promoting the river in the reach to become more influent (recharging) while downstream reaches may receive diminishing streamflow discharges.

Sensitivity Analysis

The purpose of conducting a sensitivity analysis on any hydrologic model is to establish relative influence of various parameters on model response. This information is useful particularly from the standpoint of determining the accuracy to which individual parameters must be known in order to ensure reliable model operation. For the purposes of this study, the term "influence" is formally defined as "a measure of how variation in a particular model input parameter is transmitted through and reflected in corresponding variation of the model output". It is convenient to quantify the influence

of a parameter by defining an influence index, X:

$$X = \left\{ \sum_{i=1}^{n} \frac{(Q_i - \hat{Q}_i)^2}{n} \right\}^{\frac{1}{2}} \qquad (11)$$

where:

Q_i = simulated output for the prevailing parameter value at time step i
$\hat{Q}_i$ = simulated output after some variation to the parameter at the time step i
n = number of time steps considered.

Thus, by systematically changing individual parameters while keeping all the others constant, the relative influence due to change in each of the system parameters can be examined by a comparison of the X values.

In carrying out this sensitivity analysis, an initial model solution was obtained for the parameter values defined by the Pinder and Sauer (1971) model. In order to analyse the baseflow contribution, it is necessary to separate it from the surface runoff component of the system hydrograph. This was achieved by routing the input discharge hydrograph through the channel reach; in one simulation the channel was assumed to be impervious, and in the other routing leakage through the wetted perimeter of the channel to the floodplain aquifer was permitted. The difference between the resulting discharge hydrographs for the two simulations at the same point in space and time indicates the effect of the interaction of the flood wave and the aquifer. The system was studied in terms of its response to the passage of a sediment laden flood wave with the following characteristics: peak flow was 792 m^3/s, taking 2 hours to reach the peak, and the hydrograph base time was 12 hours. The steady discharge flow prior to the introduction of the flood wave was 509 m^3/s, and the sediment concentration was 5000 ppm. Parameter sensitivities were then computed with respect to the model's baseflow output, by changing each parameter in question by a factor of 10 while keeping all other parameters at their original values. The results of this analysis are displayed in Table 1.

Table 1 - Results of sensitivity analysis of MASAI model baseflow output.

System parameter being changed	Influence index, X
Specific yield of aquifer S_y	13.2
Hydraulic conductivity of aquifer, K_2	14.3
Hydraulic conductivity of streambed sediments, K_1	2.3
Concentration of input sediments, C_s	12.5
Particle size of input sediments, D_{50}	1.6

The table provides some understanding of the relative influence on the model's baseflow output of changes in various system parameters. It is observed that, under these conditions, baseflow is most influenced by changes in S_y, K_2, and C_s. The influence due to change in the hydraulic conductivity of the streambed (K_1) is smaller, probably because of the thinness of the layer used, and because the contrast between K_1 and K_2 is not as marked as in some earlier river-aquifer modelling studies (eg Younger et al, 1993). However, its influence is certainly significant, noting that the initial thickness of the bed layer was only 0.1 m, compared to that of the aquifer (at least 27m).

By bringing to light the sensitivity of baseflow to sedimentation variability, and given the transient nature of sediment yield and storage in alluvial rivers (eg Walling, 1983), these results confirm the need for more efforts in the monitoring of sediment yields. The endeavour made in this study to include sedimentation effects in a river-aquifer model is also vindicated.

Conclusions

An enhanced concept of mathematical modelling for alluvial stream-aquifer interactions has been developed. The significant feature of the model is its extra capability to analyse transient stream-aquifer water interchange for eroding and/or evolving channels. This kind of analysis, relating the effects of sediment yield to baseflow hydrographs, could be applied to provide insights into possible impacts of proposed land-use and/or river engineering changes on the hydrological behaviour of alluvial river systems. The problem of how much a given land use will change sediment yield (or baseflow) is very complex and has not been addressed. However, a preliminary qualitative picture of the impact of such changes may be gleaned from the simulations presented here.

References

Abbott, M. B., Bathurst, J. C., Cunge, J. A., O'Connell, P. E., and Rasmussen, J., 1986, An introduction to the European Hydrological System - Système Hydrologique Européen, "SHE" 2: Structure of a physically-based, distributed modelling system, **Journal of Hydrology**, 87, pp 61 - 77.

Bathurst, J. C. and Wicks, J. M., 1991, Framework for Erosion and Sediment Yield Modelling. In Bowles, D.S., and O'Connell, P.E., (editors) **Recent Advances in the Modelling of Hydrologic Systems**. Kluwer Academic Publishers, Dordrecht, pp 1 - 27..

Ackers, P., and White, W.R., 1973, Sediment Transport: New Approach and analysis. **Journal of the Hydraulics Division, Am. Soc. Civil Engrs.**, 99, No. HY11, pp 2041 - 2060.

Alam, M.Z., and Kennedy, J.F., 1969, Friction factors for flow in sand-bed channels. **Journal of the Hydraulics Division, Am. Soc. Civil Engrs.**, 95, No. HY6, pp 1973 - 1993.

Amein, M., and Fang, S.C., 1970, Implicit flood routing in natural channels. **Journal of the Hydraulics Division, Am. Soc. Civil Engrs.**, 96, No. HY12, pp 2481 -2501.

Bear, J., 1972, Dynamics of Fluid in Porous Media. American Elsevier, New York.

Chang, H.H., 1982, Mathematical Model for Erodible Channels. **Journal of the Hydraulics Division, Am. Soc. Civil Engrs.**, 108, No. HY5, pp 678 - 689.

Chang, H.H., 1988, **Fluvial Processes in River Engineering.** John Wiley and Sons, New York.

Chen, Y.H., 1973, **Mathematical modelling of water and sediment routing in natural channels.** Ph.D. Thesis, Civil Engineering Dept, Colorado State University, USA.

Cunningham, A.B., and Sinclair, P.J., 1979, Application and Analysis of a Coupled Surface and Ground water Model. **Journal of Hydrology**, 43, pp 129-148.

French, R. H., 1986, **Open-Channel Hydraulics**. McGraw - Hill Book Company, New York.

US Army Corps of Engineers, 1977, HEC-6: Scour and Deposition in Rivers and Reservoirs - Users Manual. California.

Kornis, S., and Laczay, I.A., 1988, Effects of extensive dredging on the river regime. In Proceedings of an International Conference on River Regime. John Wiley and Sons, New York. pp 385-394.

Mwaka, B.M.L., 1995, **Analysis of the Hydrological Impacts due to Sedimentation Variability on Alluvial Stream-Aquifer Systems**. PhD Thesis, Department of Civil Engineering, University of Newcastle Upon Tyne, UK.

Pearce, F., 1994, Dam Truths on the Danube. New Scientist (17th September 1994), pp 27 - 31.

Pinder, G.F., and Sauer, S.P., 1971, Numerical Simulation of Flood Wave Modification due to bank storage effects. **Water Resources Research**, 7, pp 63 - 70.

Sharp, J.M., Jr., 1977, Limitations of Bank-Storage Model Assumptions. **Journal of Hydrology**, 35, pp 31 - 47.

Singh, K.P., 1968, Some factors affecting baseflow. **Water Resources Research** 4, pp 985 -999.

Strelkoff, T., 1969, The One-Dimensional Equation of Open Channel Flow. **Journal of the Hydraulics Division, Am. Soc. Civil Engrs.**, 95, No. HY3, pp 861 - 876.

Todd, D.K., 1955, Groundwater flow in relation to a flooding stream. Proceedings of the American Society of Civil Engineers, 81, pp 628-1 - 628-20.

Todini, E., 1991, Hydraulic and Hydrologic Flood Routing Schemes. In Bowles, D.S., and O'Connell, P.E., (editors), **Recent Advances in the Modelling of Hydrologic Systems**, Kluwer Academic Publishers, London. pp 389 - 406.

Walling, D.E., 1984, The sediment yield of African rivers. In **Challenges in African Hydrology and Water Resources**, Proceedings of the Harare Symposium, IAHS publ. no. 144. pp 265 - 283.

Walling, D.E., 1983, The sediment delivery problem. **Journal of Hydrology**, 65, pp 209 - 237.

Wang, H., and Anderson, M.P., 1982, **Introduction to Groundwater modelling: Finite difference and finite element methods**. W.H. Freeman, San Francisco.

Wilson, E.M., 1990, **Engineering Hydrology**. Macmillan Education Ltd.

Younger, P.L., *this volume*, River-aquifer modelling in perspective. In Younger, P.L., (editor), **Modelling River-Aquifer Interactions**. British Hydrological Society, Occasional Paper No. 6. pp 1 - 3.

Younger, P.L., Mackay, R., and Connorton, B.J., 1993, Streambed sediment as a Barrier to Groundwater Pollution; Insights from Fieldwork and Modelling the River Thames Basin. **Journal of the Institution of Water and Environment Management**, 7, pp 576 - 585.

THE IMPORTANCE OF MODELLING SURFACE WATER - GROUNDWATER INTERACTIONS IN CATCHMENT MANAGEMENT PLANNING

by

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Abstract

The recognition of the importance of surfacewater - groundwater interactions has recently attracted the renewed attention of catchment management planners as a result of the UK 1988 - 1992 drought. There is a need to fully consider these interactions when investigating optimum solutions to a particular set of water management problems in a catchment. A model has been developed, the Integrated Groundwater Surface Water Model (IGSM) which represents the hydrological cycle in a fully integrated manner. IGSM has a finite element structure and models surfacewater groundwater interactions where stream and groundwater nodes intersect. Flow simulations at these intersections are modelled in three ways; (1) flow out of the stream to subsurface under unsaturated conditions, (2) flow from stream to subsurface under saturated conditions, and (3) flow from the subsurface back into the stream. The significance of considering these interactions in the development of a catchment management plan using IGSM can be illustrated with reference to the Salinas Valley in California. Here, the yield of the groundwater basin was increased by inducing more stream recharge to the groundwater system using a streamside wellfield. Groundwater abstracted near the coast resulted in saline intrusion. This is replaced by water supplied from the new wellfield; producing a halt and reversal of the saline intrusion. Using IGSM, this was estimated to be on average, an extra 20 to 30 Mm^3/yr, depending on the alternatives under investigation.

Introduction

The recognition of the interactions between groundwater and surface water by hydrologists has been long standing. For instance, in traditional hydrograph analysis, attempts have been made to divide total runoff into quickflow and baseflow portions, with the baseflow component consisting mainly of groundwater contributions (Chow, 1964). However, the importance of understanding these interactions in broader environmental terms was highlighted in UK during the 1988-1992 drought (Marsh et al, 1994), where many predominantly groundwater fed streams and rivers experienced periods of record low flows and in some cases dried up completely along certain reaches eg the River Misbourne in Buckinghamshire (National Rivers Authority, 1993).

The example of the River Misbourne is also useful in that it illustrates the importance

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of understanding stream - aquifer interactions in developing the water resources of similar catchments, where groundwater is a major component of the available water resource. For example, about 65% of the water available to the river is pumped from groundwater for public water supply. Thus, the problems experienced on the River Misbourne were a combination of low rainfall and thus recharge of the aquifer systems with continued pumping of those aquifer systems to meet public water supply demands. The resulting fall in groundwater levels had a negative impact on surface water flows along certain reaches of the River Misbourne, especially in the reach between Amersham and Chalfont St Giles.

There are engineering as well as hydrological and environmental needs which have to be addressed in any catchment planning model. This means that full integration of the surface and groundwater portions of the hydrological cycle is beneficial. This allows the investigation of a wide range of solutions seeking to address the problems identified in the above example, where a fully integrated approach is essential to finding the optimum solution. This approach is essentially "holistic" in nature, and seeks to address all the conflicting demands imposed on a catchment to reach a solution that is both equitable and sustainable (Gardiner, 1994). Such a model has been developed and is called the Integrated Groundwater Surface Water Model (IGSM). This paper will therefore describe the salient features of IGSM, before going on to give a more detailed account of how IGSM models surface water - groundwater interactions. The importance of considering these interactions in developing a basin or catchment management plan will be further illustrated with reference to an example project from the USA, where a major component of groundwater recharge was from the main river flowing over the aquifer system of interest.

The Integrated Groundwater - Surface Water Model (IGSM)

IGSM is a comprehensive catchment planning model which includes groundwater, surface water, groundwater quality and reservoir operation simulation routines. The Model has a finite element quasi-three-dimensional structure, capable of simulating several aquifer layers. The groundwater portion operates on two basic scientific principles: (1) conservation of mass, and (2) Darcy's law of fluid flow through a porous media. The surface water portion includes hydrological basin analysis for rainfall, percolation, run-off and evapotranspiration. Stream flow simulation operates on a mass balance system. To integrate the surface and groundwater flow simulations, a soil moisture accounting and unsaturated flow system has been incorporated. This, coupled with runoff and percolation rates for given soils allow interaction between the groundwater and surface water flows. An additional reservoir operations module is included to derive reservoir releases for stream flow accounting. The basic time step used in the groundwater simulation is monthly. Surface water simulation can be performed on either a daily or monthly basis. Furthermore, water quality simulation is included which can track a contaminant through the processes of advection, dispersion, and dissolution.

The original version of the model was developed in the mid 1970's and since then it has undergone a number of revisions to include various hydrological and geological conditions in response to specific project requirements. The general philosophy behind

model development has been to use "tried and tested" techniques and methodologies, both in the methods of analysis and the numerical solutions adopted.

For modelling purposes, the hydrological system may be divided into three major components: (1) stream system, (2) groundwater system, and (3) soil and unsaturated zone. Figure 1 is a block diagram showing these components. The basic principle of the hydrological model is to track the movement of water from location to location in each of these components and to keep track of the interaction between the units using mass balance or water budget accounting procedures.

The groundwater and unsaturated zone components are simulated using a network of three-dimensional "volume elements" that have a triangular or quadrilateral surface shape, and depth represented by a series of layers to simulate the stratigraphy of the aquifer system. Model nodes are defined at each corner of these volume elements. Flux into the elements along the model boundaries can be defined as no flow, variable flow, or constant flow.

A surface water network is overlain on the groundwater network to simulate the surface water component and its interaction with the groundwater. Reservoir operations as well as lake-groundwater interactions can also be modelled within the surface water component. Land surface characteristics represented include land use types, crop mix patterns and soil types.

Elements are grouped together to define subregions. These subregions can be utilised for both input and output summaries representing individual water districts or other sub-areas. Input data requirements for simulation include not only the constant hydrogeological parameters, but also time-variant parameters including land use, water demand, rainfall, boundary inflows, and water supply. Many of these parameters can be assigned by subregion, while others must be assigned to individual elements.

Modelling of Groundwater - Surface Water Interactions within IGSM

Groundwater and surface water interactions within IGSM are represented through streamflow at the intersections of the streams with the model grid. Stream flow nodes are assigned to groundwater nodes at these intersection points. Three situations may occur at these points which are simulated differently. These include:

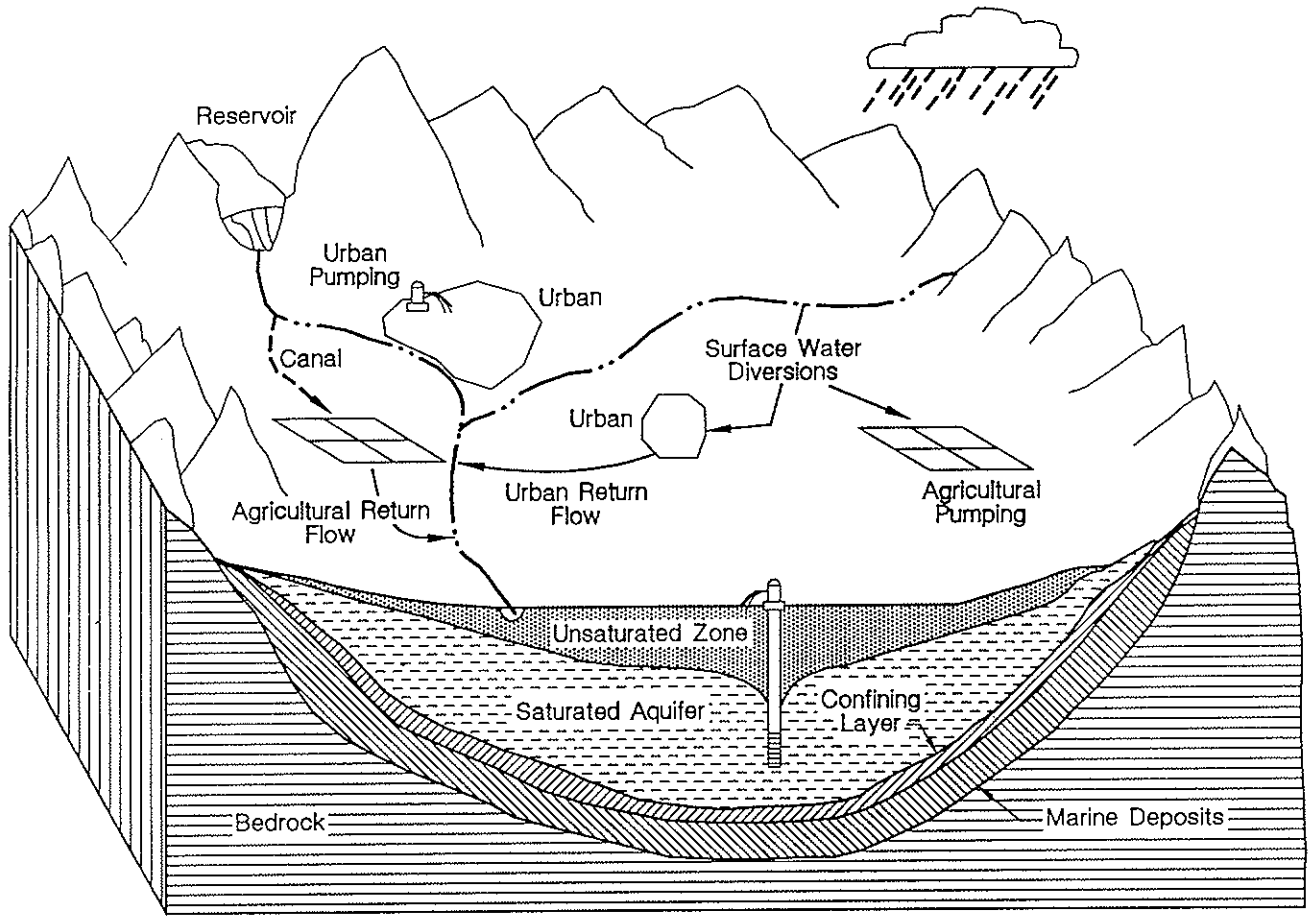
(a) flow out of the stream into the subsurface under *saturated* flow conditions;
(b) flow out of the stream into the subsurface under *unsaturated* flow conditions; or
(c) flow from the subsurface back into the stream.

Each of these situations is presented in Figure 2 and the formulas representing the relationships are given below.

Case A

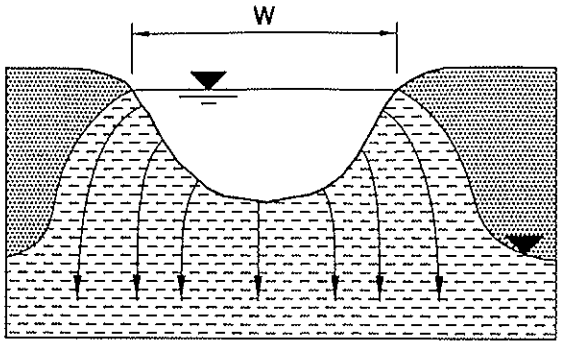
When the groundwater level is lower than the stream bottom, and the stream bed is comprised of similar materials to surrounding soils, the losing stream's seepage flow is

FIGURE 1 : HYDROLOGICAL COMPONENTS OF IGSM



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FIGURE 2 : GROUNDWATER - SURFACE WATER INTERACTION

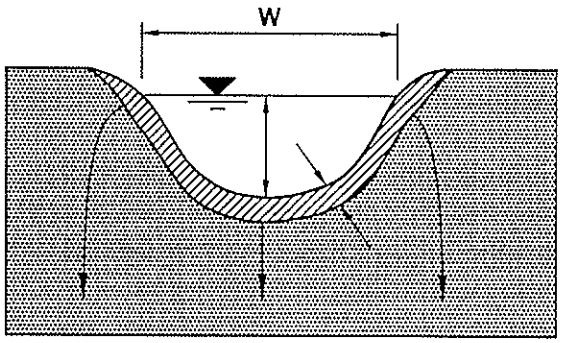


(a) LOSING STREAM,
SATURATED INFILTRATION

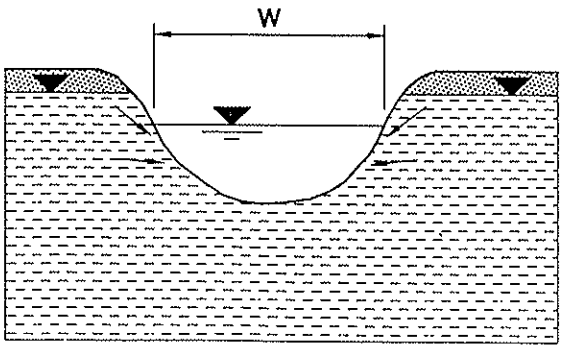
LEGEND

Unsaturated Subsurface

Saturated Subsurface



(b) LOSING STREAM,
UNSATURATED INFILTRATION



(c) GAINING STREAM,
SATURATED FLOW

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under saturated conditions as shown in Figure 2(a). Infiltration or seepage loss from the stream is:

$$Q_s = K (W + \propto D_s) \quad (1)$$

where:

Q_s = seepage from the stream per unit length,
K = estimated hydraulic conductivity of the stream bed material,
W = width of water in the stream,
D_s = depth of water in the stream,
$\propto$ = stream channel coefficient.

The above equation is described in detail in Polubarinova-Kochina (1962). The coefficient, $\propto$, generally varies from 0.5 to 3.0 depending on the shape of a channel cross-section. For a curvilinear shape, it becomes 2. For practical purposes, $(W + \propto D_s)$ may be replaced by the wetted perimeter at the node of interest.

Case B

In a stream bed which is comprised of finer materials than surrounding soils, as shown in Figure 2(b), the losing stream's seepage flow mechanism may become unsaturated. In this case, the seepage equation is:

$$Q_s = K W (D_s + b_s)/b_s \quad (2)$$

where:

Q_s = seepage from the stream per unit length,
K = estimated hydraulic conductivity of the stream bed material,
W = width of water in the stream,
D_s = depth of water in the stream,
b_s = thickness of the stream bed material.

As can be seen from equation (2), the properties of the stream bed materials are important in determining the seepage from the stream to the unsaturated zone beneath it.

Case C

When groundwater levels are higher than the stream bottom, the gaining stream is under saturated flow, see Figure 2(c), and the following equation is used:

$$Q_s = K \ P_w \ G \quad (3)$$

where:

Q_s = seepage loss from the stream per unit length. A negative value indicates groundwater seepage to the stream,
K = estimated hydraulic conductivity of the stream bed material,
P_w = wetted perimeter of the channel corresponding to the elevation of

the stream water surface or the groundwater table, whichever is greater,

G = gradient between stream stage and groundwater table calculated as $(h_g-h_s)/b_s$, but limited to a maximum value of 1,

h_s = elevation of the stream stage,

h_g = elevation of the groundwater table,

b_s = thickness of the stream bed material.

From these equations, the model data requirements include river node location, tributary and return flow details, stream bed elevations, wetted perimeter and discharge rating tables and stream bed thickness and hydraulic conductivity. This information can be collected from a variety of sources, including survey information from flood defence schemes, with the use of, for example, simple hydraulic models such as HEC-2 to work up rating tables for various nodes along the stream reach of interest. The values of stream bed hydraulic conductivity can be determined from available geological information or field tests. In contrast, it may not be possible to estimate stream bed thickness with any confidence and thus it becomes a parameter that is considered in model calibration.

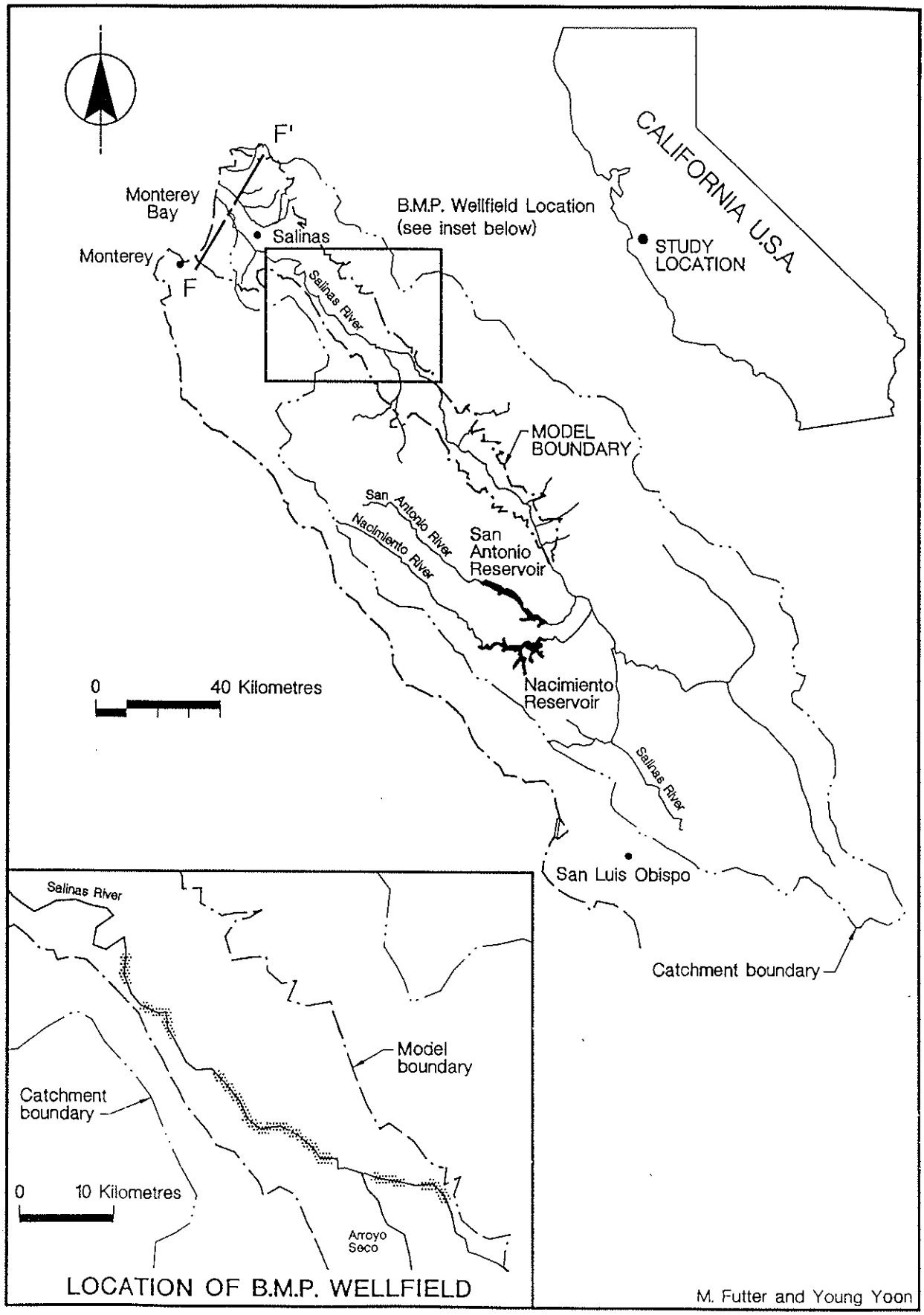
IGSM internally monitors the hydraulic situation at the stream nodes and adjusts the simulation of the stream aquifer interactions when needed in response to respective changes in groundwater levels. However, as the model simulates groundwater conditions on a monthly timestep and can simulate stream flows on a daily timestep, any changes in the nature of the stream-aquifer interactions occur at the end of each month, when the surface and groundwater timesteps are coincident. Within months, on a daily basis, seepage from or into the stream is then simulated, but with a fixed groundwater level at the stream groundwater node.

The importance of the fully integrated modelling of surface water - groundwater interactions and more specifically stream-aquifer interactions will be illustrated with reference to an example project from central California USA. In this case study, a major component of aquifer recharge is due to stream seepage, and this had important implications in the development of a basin management plan to meet future water resource needs.

A Case Study: The Salinas River Basin Management Plan

The Salinas Valley is located in central California, USA, approximately 160 km south of San Francisco Bay (see Figure 3). The catchment covers an area of 4600 km^2 and receives 95% of its water supply from groundwater. The main source of groundwater replenishment is by infiltrating surface water, predominantly along the Salinas River. This is aided by two reservoirs, Nacimiento and San Antonio, which were constructed in 1957 and 1965 respectively, to provide storage for releases to the Salinas River to increase groundwater replenishment. The Salinas Valley is the second most important agricultural area in California, after the Central Valley, and as agricultural production has increased, so has demand for irrigation waters, which make up 90% of total needs in the valley.

FIGURE 3 : LOCATION OF STUDY CATCHMENT AND B.M.P. WELLFIELD



With increasing industrial and domestic demands this has resulted in increased reliance on groundwater sources with the result that groundwater levels have fallen and saline intrusion has increased in the coastal portion of the basin, given that a large majority of the water resources demands are in this coastal area.

The water resources of the Salinas Valley are managed by the Monterey County Water Resources Agency (MCWRA), who recognised that the problems of groundwater overdraft and saline intrusion must be addressed. As part of this response, MCWRA set into motion the development of a Salinas River Basin Management Plan (SRBMP). The main aim of the SRBMP was to halt and reverse the saline intrusion while meeting future water demands. This was to be achieved by reducing groundwater pumping in the coastal part of the valley and replacing the pumped water from sources further upstream, predominantly from the aquifer in the main area of stream recharge by the Salinas River.

Thus, a Salinas Valley IGSM was developed for MCWRA to aid the development of the SRBMP. The application of a model that is capable of representing both the surface water (including reservoir operations) and groundwater portions of the catchment in a fully integrated manner is necessary to evaluate the various management scenarios developed as part of the SRBMP effort.

Study Area Characteristics

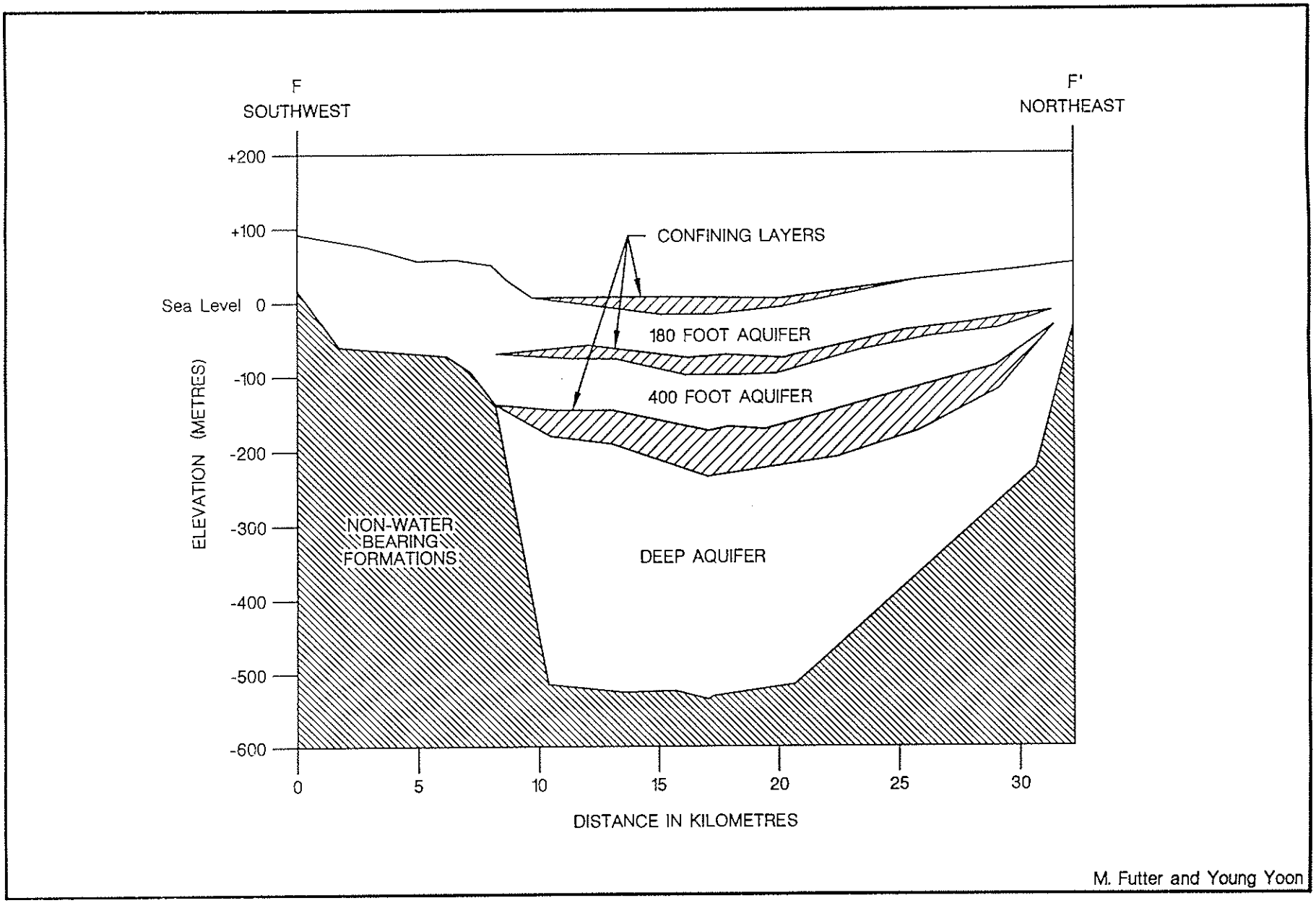
The Salinas Valley is an elongated catchment surrounded by the Diablo and Santa Lucia mountain ranges, and drains in a north westerly direction into Monterey Bay. Annual average precipitation ranges from 280 to 360 mm/year with 75% falling between November and March, while annual average flow in the River Salinas at the tidal limit is about 12 m^3/s. As mentioned above, the primary land use in the catchment is agriculture, most of which is irrigated. The main urban centre is the City of Salinas, with a population of about 110 000.

The stratigraphy of the Salinas Valley groundwater basin is characterised by the presence of three major water bearing strata termed the 180 foot aquifer, the 400 foot aquifer and the deep aquifer. The aquifer system is both confined and unconfined, as illustrated by the geological section presented in Figure 4. The 180 foot aquifer consists of Quaternary/Recent sand and gravel alluvium formations, and this is the main aquifer used for water supply. It is overlain by a blue clay layer aquitard of varying thickness. The 400 foot aquifer is predominantly sand and gravels, but also includes some conglomerates and is separated from the 180 foot aquifer by a leaky aquitard of marine blue clay. The deep aquifer consists of alternating layers of sand-gravel mixtures and clays, and is thus not such a distinct aquifer.

Development and Calibration of the Salinas Valley IGSM

The main features of the Salinas Valley IGSM are given in Table 1. The model area is concentrated on the Salinas groundwater basin, with contributions from outside the model area entering as peripheral watersheds. The model can include both rectangular and triangular elements, and thus has great flexibility in representing model area outline

FIGURE 4 : GEOLOGICAL CROSS SECTION F - F'



details. The basic analysis units are the model subregions, but model results from smaller areas of interest can also be output from IGSM.

Model calibration was carried out for the period October 1969 to September 1990 and involved a two stage process. The first is a water budget calibration to ensure that the model is accurately simulating the key components of the groundwater basin's hydrological water balance. Some of the key parameters adjusted during this stage included stream bed hydraulic conductivity and thickness. The second stage of calibration, involves the comparison of simulated and measured groundwater levels at specific wells. During this stage of calibration adjustments were made to model aquifer and confining layer parameters.

Calibration Period Groundwater Budget

The IGSM calibration period groundwater budget for the whole model area are summarised in Table 2, and indicates that the three main sources of aquifer recharge in the model area are deep percolation, stream/river seepage and subsurface boundary inflows. Of these, the stream/river seepage is the largest at 300.9 Mm^3/yr on average. The primary sources of this recharge are the Salinas River (including releases from the San Antonio and Nacimiento reservoirs and the Arroyo Seco).

However, smaller streams and creeks also contribute to the groundwater system. The results presented in Table 2 indicate that the amount of recharge from streams is not necessarily a function of the amount of stream flow occurring within a given year, although under normal circumstances that would be the case. In the Salinas Valley, water releases are made from both the Nacimiento and San Antonio reservoirs according to specific operating rules to augment streamflows in the Salinas River during dry years. As a result, even during the extreme drought years of 1976-77, the gain from streams is as high as 282.7 Mm^3, comparable to that during the wet years of 1982-83 (361.8 Mm^3). However, during the dry year of 1990, the gain from streams is only 49.7 Mm^3. In this particular year, the normal streamflow was low. Also, due to three previous consecutive dry years (1987-89), the available storage of both San Antonio and Nacimiento reservoirs had been depleted. Consequently, there was not enough water available in the reservoirs to augment low streamflows in the Salinas River, resulting in the significantly lower gains from streams.

Alternative Screening Analysis

Given the extra knowledge and understanding of groundwater and surface water interactions resulting from the calibration of the Salinas Valley IGSM, the model was used to investigate various possible SRBMP options. As stated previously, the main objective of the SRBMP is to halt saline intrusion at the coast by increasing the yield of the groundwater basin, while meeting 2030 water demands. All of the alternatives investigated were based on looking at various methods of enhancing the surface water recharge of the Salinas 180 foot aquifer. The options included combinations of various components. For example, percolation ponds, new reservoirs to store winter rainfall for release during the summer months and river well systems to induce extra recharge. The use of surface water diversions was also investigated as direct supply (via treatment

Table 1
Salinas Valley IGSM Features

Feature		Magnitude
Model Area		1680 km^2
Finite Elements	Number of Elements	1615
	Number of Nodes	1726
Number of Model Subregions		6
Number of Streams Modelled		5
Number of Peripheral Watersheds Modelled		109
Number of Rainfall Stations Used		9
Number of Reservoirs Modelled		2
Number of Aquifers Modelled		3

works) and to the proposed new reservoirs.

The methodology adopted in the alternative analysis was to develop baseline conditions (i.e. no action) under 2030 demands, with initial conditions based on those resulting from the final calibration run to 1992, and imposing the hydrology for the period 1949 to 1992 (44 years). The alternative analysis and comparison was then carried out by incorporating each of their components into the Salinas Valley IGSM and repeating the simulation run. As well as meeting the water resources/water quality objectives, the alternatives were also assessed in terms of cost and environmental impacts.

Based on the model analysis, the preferred option recommended to the MCWRA was the use of a well system along a reach of the River Salinas just above the tidal limit of the river (see Figure 3). The well system would induce extra stream recharge of the River Salinas flows, and in doing so, increase the yield of the groundwater basin. From the model simulation results, it was estimated that the operation of the well field would induce, on average, an extra 31 Mm^3/year of recharge. This option met all the objectives in terms of water resources, water quality and environmental impact, and had the lowest estimated capital cost.

Detailed Analysis of Preferred Options

Further detailed analysis of the preferred option was carried out in response to concerns expressed by the client about the impact of the well system on local groundwater levels and thus on local farmers groundwater pumping for irrigation.

This analysis attempted to modify the preferred option but reduce the impact on the

Table 2: Salinas Valley IGSM Calibration Groundwater Budget - (Units = Mm^3)

Year	Deep Percolation (+)	Gain From Stream (+)	Other Inflow (+)	Boundary Inflow (+)	Pumping (-)	Change in Storage (=)	End Storage 1 000 Mm^3
1970	240.3	359.9	3.7	73.9	701.3	-23.4	24.06
1971	270.3	299.1	3.5	73.9	697.5	-50.8	24.01
1972	213.4	314.4	3.9	85.5	754.9	-137.7	23.87
1973	416.8	353.9	3.2	71.8	681.6	164.1	24.04
1974	292.3	353.8	3.1	67.1	635.1	81.2	24.12
1975	266.6	328.3	3.2	68.7	642.8	24.0	24.14
1976	179.9	282.7	3.6	75.0	634.2	-93.0	24.05
1977	173.3	281.6	3.6	77.4	672.9	-137.0	23.91
1978	365.8	386.8	3.1	69.9	640.7	184.9	24.10
1979	282.6	333.8	3.5	73.1	705.0	-12.1	24.08
1980	288.6	343.9	3.4	72.5	634.6	73.7	24.16
1981	209.7	314.2	3.7	78.0	684,4	-79.0	24.08
1982	275.5	361.8	3.1	67.3	565.4	142.0	24.22
1983	422.5	327.0	2.3	55.4	577.5	229.7	24.45
1984	238.6	252.1	3.0	63.6	697.1	-139.7	24.31
1985	227.8	271.7	3.3	69.2	668.6	-96.6	24.21
1986	278.6	298.4	3.4	72.1	626.5	25.9	24.24
1987	10.3	266.5	3.6	76.1	692.2	-135.7	24.10
1988	207.5	280.2	3.8	80.0	642.0	-70.5	24.03
1989	199.0	299.3	3.9	82.5	654.0	-69.4	23.96
1990	163.8	49.7	3.9	84.4	618.1	-316.1	23.65
1991	259.8	241.0	4.2	91.7	690.0	-93.3	23.56
1992	266.1	320.0	4.2	94.2	667.9	16.6	23.57
Average Annual	258.6	300.9	3.5	74.9	660.2	-22.3	

Note: Deep percolation - the amount of water entering the aquifer system from the soil and unsaturated zones.
Gain from Stream - gains (+) or losses (-) to/from the aquifer due to stream aquifer water interactions.
Boundary Inflow - amount of groundwater entering the model area from surrounding areas.
Change in Storage - amount of change in groundwater storage.
End Storage - total groundwater storage within all aquifers in the study area.

local groundwater levels under operation of the scheme. During this analysis, a more detailed investigation of the additional stream recharge produced by operation of the scheme and its variants was undertaken. This involved quantifying the recharge induced by the scheme within the project well field area and downstream from it compared with the baseline situation. Results for two of the options investigated using IGSM are presented in Table 3.

Table 3
Annual Average Additional Stream Recharge Induced by Project (Mm^3/yr)

	Baseline	Alternative A	Alternative B
Wellfield Reach	0	41.9	34.5
Downstream of Wellfield	0	-11.1	-13.6
Net Effect	0	30.8	20.9

As can be seen from Table 3, a significant amount of additional stream recharge is taking place under both Alternatives A and B. However, this is concentrated in the wellfield area, where pumping by the wells increases the gradient from the river to the groundwater and thus induces additional seepage from the river. However, downstream of the wellfield area, less recharge occurs compared with the baseline situation because less groundwater pumping is taking place in this area under the BMP, and thus, the gradient between the Salinas River and the groundwater levels is reduced.

Differences between Alternatives A and B relate to the use of a surface water abstraction under Alternative B to meet some of the coastal irrigation demands, and thus a reduction in the pumping requirement from the wellfield area. This lower pumping requirement reduced the impact on local groundwater levels by on average 0.8 m and thus the amount of stream recharge taking place along the wellfield reach of the River Salinas.

As with the screening analysis, Alternatives A and B met all the criteria set out as part of the BMP and were recommended to the client as the apparent preferred options for their final decision,

Conclusions

The consideration of surface water-groundwater interactions and more specifically stream-aquifer interactions should not just be an area of interest for researchers in hydrology and hydrogeology. As illustrated by the example presented in this paper, it can also have a direct bearing on the need to find an optimum engineering and planning solution to pressing water resources and water quality problems. The development of the Salinas Valley Basin Management Plan also shows the importance of considering these interactions in a fully integrated manner, where integration of the surface and groundwater portions of the hydrological cycle recognises in a much more realistic way the actual processes taking place within a catchment of interest. The use of a model such as IGSM increases the level of understanding about these interactions in the catchment of interest. However, IGSM is not purely a research tool, but seeks to provide engineers and planners with the comprehensive information they need to develop optimum solutions as part of the catchment management planning process.

References

Chow, V T, 1964, Section 14 - Runoff. In **Handbook of Applied Hydrology: A Compendium of Water Resources Technology** edited by V T Chow. McGraw-Hill New York, USA.

Gardiner, J.L., 1994, Sustainable Development for Catchments. **Journal of the Institution of Water and Environmental Management**, 8, pp 308 - 319

Marsh, T.J., Monkhouse, R.A., Arnell, N.W., Lees, M.L., and Reynard, N.S., 1994, **The 1988-92 Drought**. Hydrological Data UK Series, Institute of Hydrology/British Geological Society, 79 pp

National Rivers Authority, 1993, **Low Flows and Water Resources: Facts on the Top 40 Low Flow Rivers in England and Wales**, 63 pp

Polubarinova-Kochina, P.Y., 1962, **Theory of Groundwater Movement**. Treatise by R.J.M. De Wiest, Princeton University Press, Princeton, USA.

INTERACTION BETWEEN RIVERS AND THE NOTTINGHAMSHIRE SHERWOOD SANDSTONE AQUIFER

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Abstract

Understanding the interaction between the Nottinghamshire Sherwood Sandstone aquifer and the rivers crossing the aquifer is of vital importance in catchment resource planning even though the magnitudes of the interactive flows are small compared to the total recharge. Flow gauging has been used to estimate the magnitude of the interactive flows and to aid the accurate incorporation of river-aquifer interactions in a mathematical model. Two rivers are studied in detail, and the influence of pumping on the river flows is examined. Areas of the aquifer contributing to the rivers are identified from the mathematical model results.

Introduction

In certain catchments, river-aquifer interaction is the dominant feature; typical examples include the Chalk of the Berkshire Downs (Rushton et al 1989) and the Great Oolite Cotswold aquifer (Rushton et al 1992). In the case of the Great Oolite Limestone, the rivers (which are mainly in deeply incised valleys) are very powerful in attracting water. Their ability to attract groundwater is enhanced by the increased transmissivities in times of high recharge.

For aquifers with more subdued relief and with a slower lateral movement of water, the interaction with the rivers may not be so dominant. When extensive abstraction occurs from the aquifer, the effect of the rivers may appear to be less significant. Nevertheless the rivers do have an important effect since they are a source of outflow from or inflow to the aquifer system. The Nottinghamshire Sherwood Sandstone aquifer is a typical example.

The interaction between rivers and aquifers can best be examined in detail using mathematical models. Methods of representing the interaction are well established (Rushton & Tomlinson 1979, Rushton et al 1989); however there is still only limited experience in obtaining the field information required for a successful simulation. When the flows involved in the interaction with the aquifer are only a small proportion of the total river flow, the collection of field data is especially difficult.

A further issue considered in this paper is the effect of river-aquifer interaction on

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groundwater protection issues. From a groundwater model it is possible to identify the flows within the aquifer system; consequently the area from which a gaining river collects water can be identified.

This paper concentrates on two rivers in the central part of the Nottinghamshire Sherwood Sandstone aquifer. The study of the aquifer was carried out on behalf of the National Rivers Authority, Severn Trent Region. Certain aspects of the study of the aquifer system will not be reported here; this paper concentrates on the river-aquifer interaction.

Brief introduction to Nottinghamshire Sherwood Sandstone Aquifer

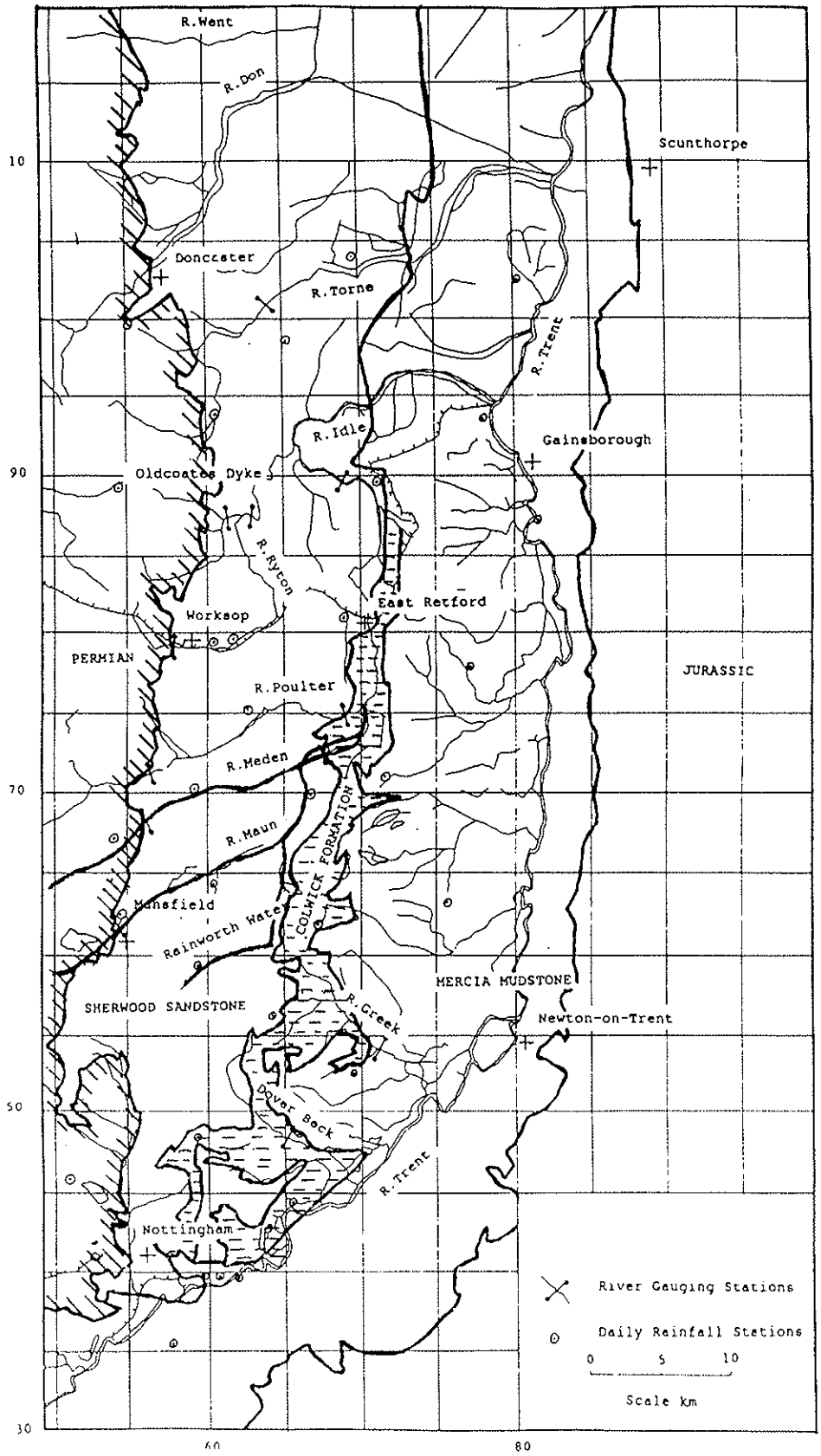
The Permo-Triassic Sherwood Sandstone aquifer in Nottinghamshire is currently used as a major source of groundwater. The outcrop of the aquifer is typically 10 km wide (in the west-east direction) and extends northwards from Nottingham to beyond Doncaster (Figure 1). The aquifer is underlain by Permian strata; to the east the Sherwood Sandstone is overlain by the Colwick Formation (which is not present in the north of the study area) and the Mercia Mudstones. The Sherwood Sandstone dips gently to the east and varies in thickness from zero at the boundary with the Permian strata to more than one hundred metres where the aquifer becomes confined. To the south of the study area the aquifer is almost 100 m thick but to the north under the confining strata, thicknesses of more than 300 m have been identified.

The outcrop area is largely drained by the tributaries of the River Idle which rise on the Permian strata to the west, cross the aquifer and move in a northward direction close to the eastern edge of the outcrop. The major tributaries of the River Idle are the Rivers Ryton, Poulter, Meden and Maun. In addition the southern portions of the Sherwood Sandstone aquifer are drained by the Rivers Greet and Leen and the Dover Beck which flow directly to the River Trent.

Pumping stations are distributed throughout the Sherwood Sandstone aquifer; some are located in the outcrop area, others are in the confined area. Abstractions from pumping stations can be as high as 20 Ml/d but this will be from several boreholes. A number of the pumping stations are located within 1 km of the rivers.

Prior to the development of abstraction from the Nottinghamshire Sherwood Sandstone aquifer, the main outlet from the aquifer was to the rivers. Following the onset of significant exploitation, the groundwater contribution to the rivers has decreased with a noticeable effect on the rivers in the south. Since the tributaries of the River Idle originate on the Permian strata there are significant flows as the rivers cross onto the Sherwood Sandstone. For the Rivers Meden and Maun the combined flow from the Permian strata is usually in excess of 60 Ml/d and averages 90 Ml/d Consequently any groundwater interaction with the rivers of less than 10 Ml/d is difficult to quantify from field measurements.

Figure 1 - General details of Nottinghamshire Sherwood Sandstone Aquifer.



Despite the difficulty in quantifying the river-aquifer interaction, the rivers still have a very important effect on the groundwater conditions in the aquifer. Either the rivers can feed the aquifer especially in locations where there is an influence from pumped boreholes or the rivers may drain away excess water.

Formulation and Field Information for River-Aquifer Interaction

Relationships for river-aquifer interaction

The basic approach to the representation of river-aquifer interaction is now well established (Rushton and Tomlinson, 1979); Figure 2 summarises the assumed mechanisms. These mechanisms can be described by the relationships:

$$Q_r = RC(h - z_r) \quad \text{for } h > z_b \qquad (1a)$$

$$Q_r = -RC \quad \text{for } h \leq z_b \qquad (1b)$$

where

Q_r = flow from aquifer to river ($m^3/d/km$)
RC = river coefficient (m2/d/km)
h = groundwater head (m)
z_r = elevation of river surface (m)
z_b = river bed elevation (m) taken as 1.0 m below river surface

Four possible conditions are illustrated in Figure 2:

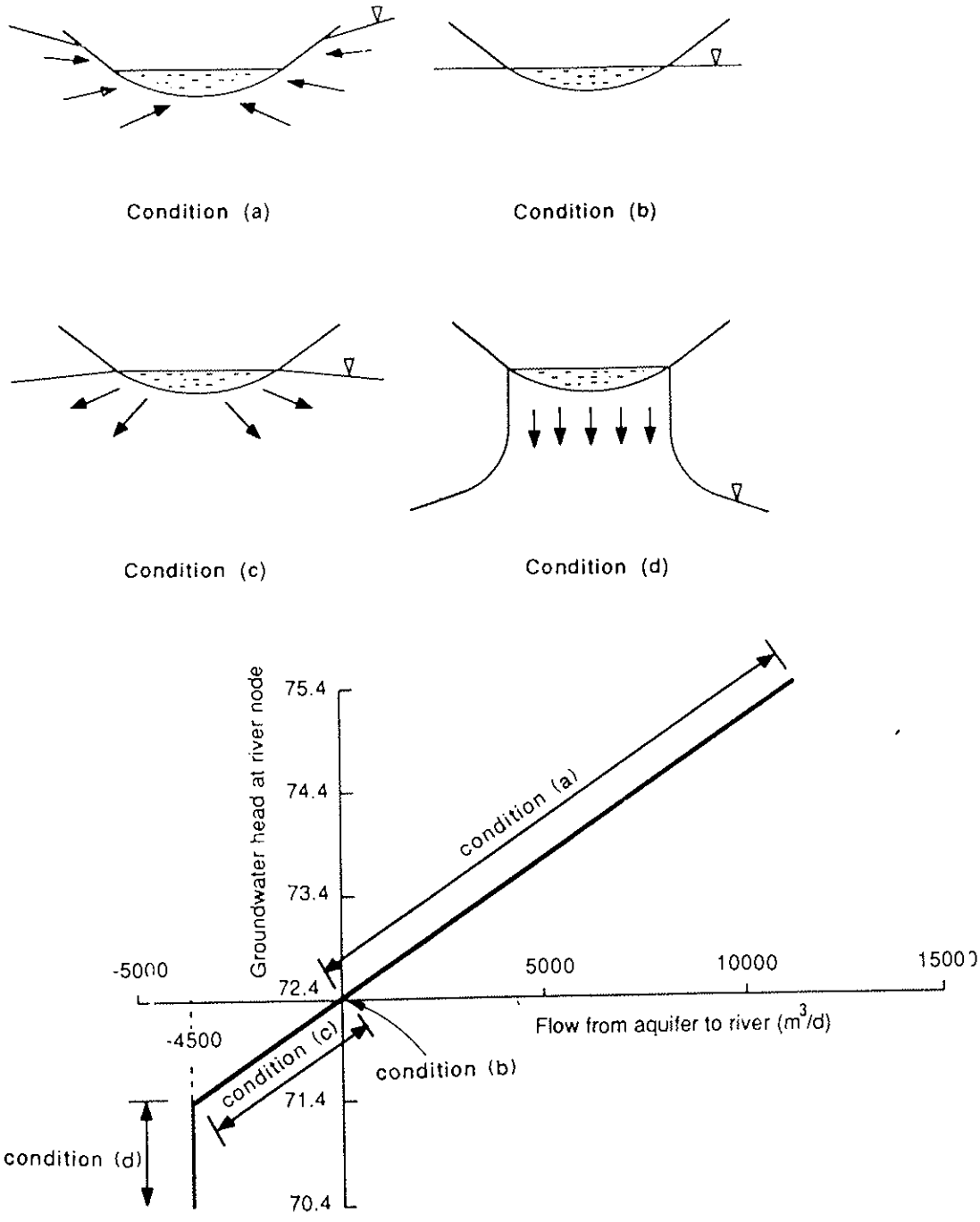
Condition (a), the groundwater head is above the elevation of the river surface hence the aquifer feeds the river and equation (1a) applies.

Condition (b), the groundwater head and river level are at the same elevation and there is no interaction; equation (1a) still holds.

Condition (c), the groundwater head is slightly below river level therefore there is a small flow from the river to the aquifer; equation (1a) holds although it probably over-estimates the river loss.

Condition (d), the groundwater head is significantly below river level, from equation (1b) the loss from the river to the aquifer equals the river coefficient, RC.

Figure 2 - Diagram defining river-aquifer interaction.



Information from permanent gauges

Detailed studies will be made of the Rivers Meden and Maun which originate on the Permian Strata and cross the outcrop of the Sherwood Sandstone aquifer from west to east before joining (Figure 3). There are gauging stations as the rivers cross onto the Sherwood Sandstone at Church Warsop and Mansfield; where the rivers join there are two channels which are gauged at Haughton and Bothamsall. To identity the net interaction between these rivers and the aquifer, differences are taken between the upstream pair and downstream pair of gauges. A further inflow occurs from the Rainworth Water and allowance must also be made for abstractions from rivers and mine, industrial or sewage discharges. Due to problems with the reliability of the gauges which have been affected by mining subsidence, reliable results are available for only a limited number of years.

An approximate water balance for the rivers based on available evidence for the Rivers Meden and Maun are given below:

Average inflow at Church Warsop	+ 48 Ml/d
Average inflow at Mansfield	+ 41 Ml/d
Average outflow at Bothamsall	- 53 Ml/d
Average outflow at Haughton	- 55 Ml/d
Other inflows	+ 6 Ml/d
Net gain from aquifer	13 Ml/d

Recognizing that there are tolerances in the gauging and uncertainties in the additional inflows and outflows, the conclusion of the above water balance is that there is a small gain to the rivers due to the river-aquifer interaction of the Rivers Meden and Maun.

Interpretation of spot gaugings

Since overall changes in flows do not provide sufficient information to develop an understanding of the river-aquifer interaction, spot gauging along the rivers was initiated. There are difficulties in obtaining reliable spot gauging readings using current meters:

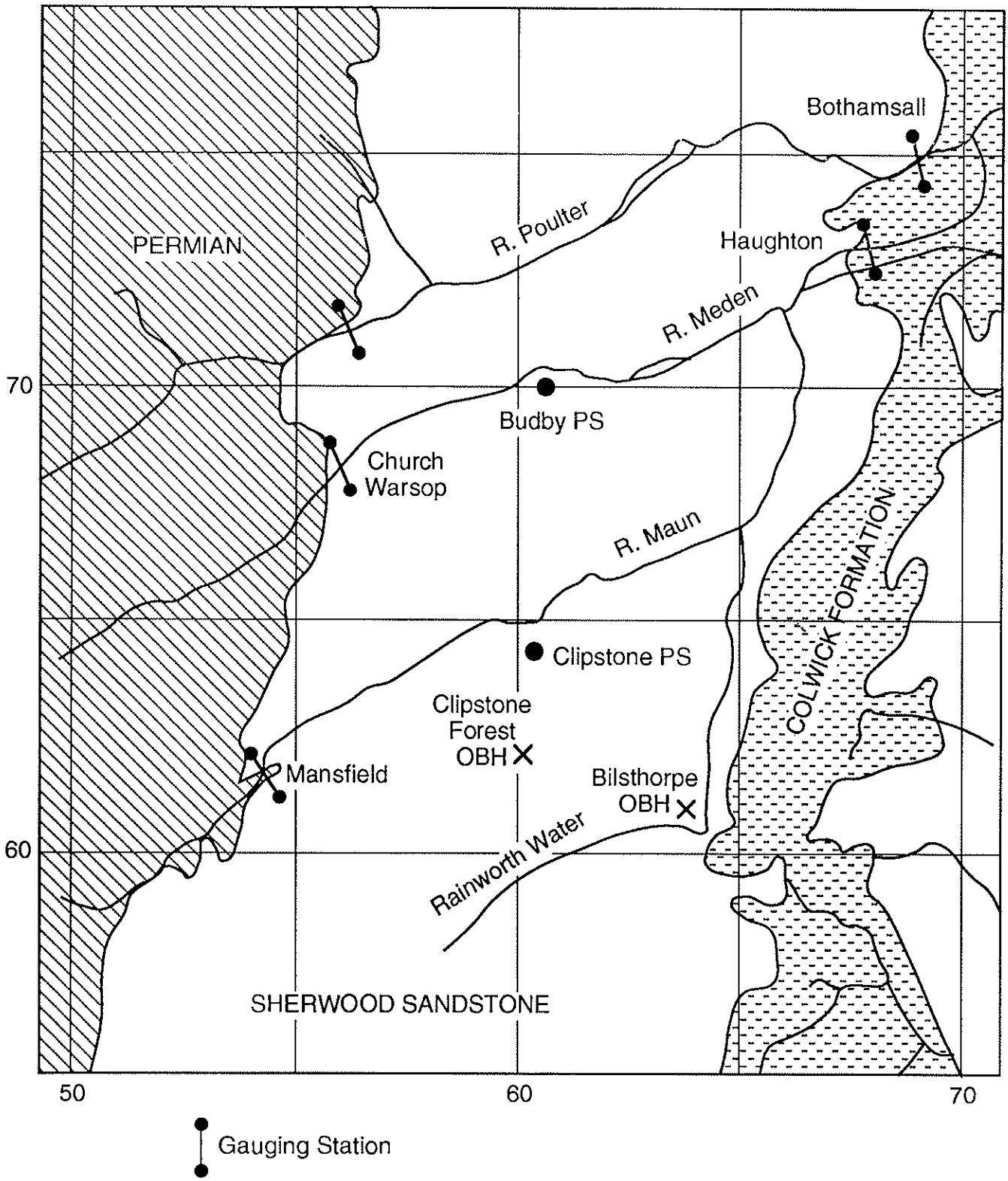
- gauging should preferably be carried out under stable low flow conditions since the object is to identify changes in flow as the river crosses the aquifer,

- gauging should only be carried out where there is access to the river and there is reasonably regular stable cross-section,

- even though multiple readings are taken on each cross section, there will be tolerances of up to 5% in the readings.

- to identify the changes in river flows caused by river-aquifer interaction, the spacing of the gauging sites along the river should be no more than 3 km with more

Figure 3 - Rivers Meden and Maun and associated features



detailed readings in the vicinity of pumping stations.

Results of the spot gaugings of the Rivers Meden and Maun are presented in Figure 4; for each stream a survey was made on two different occasions with eight gauging locations on the River Meden and seven on the River Maun; measurements were not made at every site for each survey. The field readings are not joined by curves since there are tolerances in the readings and information about abstractions from or inflows to the rivers may not be available. For example there is a discontinuity where Rainworth Water joins the River Maun. Nevertheless these readings do indicate the general trends of gains and losses.

The trends are clarified when note is taken of pumping stations which are in the vicinity of the rivers. For the River Meden, Figure 4(a), there is a distinctive fall in river flow upstream of Budby Pumping Station (which is 0.25 km from the river with an average abstraction of 14 Ml/d) and an increase in the river flows beyond Budby. Clipstone Pumping Station (which is 0.8 km from the River Maun with an abstraction rate of 7 Ml/d) may have a direct effect on the river, Figure 4(a), but the field measurements are insufficient to identify the effects.

Identification of the River Coefficients

Despite the limited information from the spot gaugings it is possible to obtain estimates of the river coefficients. The field results of Figure 4 are used to estimate the losses from the rivers to the aquifers. A first estimate of the magnitude of the river coefficient can be obtained from sections where the river is losing water to the aquifer; according to Figure 2 the loss from a river when the water table is several metres below the river level equals the river coefficient, RC.

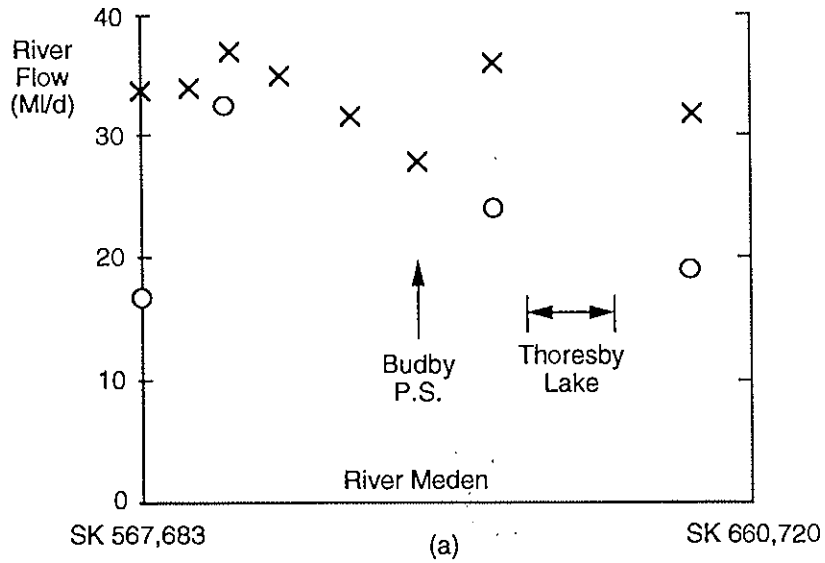
- for the River Meden, Figure 4(a), upstream of Budby Pumping Station there are four crosses which indicate a loss of 8 Ml/d in 4 km (the horizontal length of the river on this diagram is 11km) which is equivalent to 2 Ml/d/km,

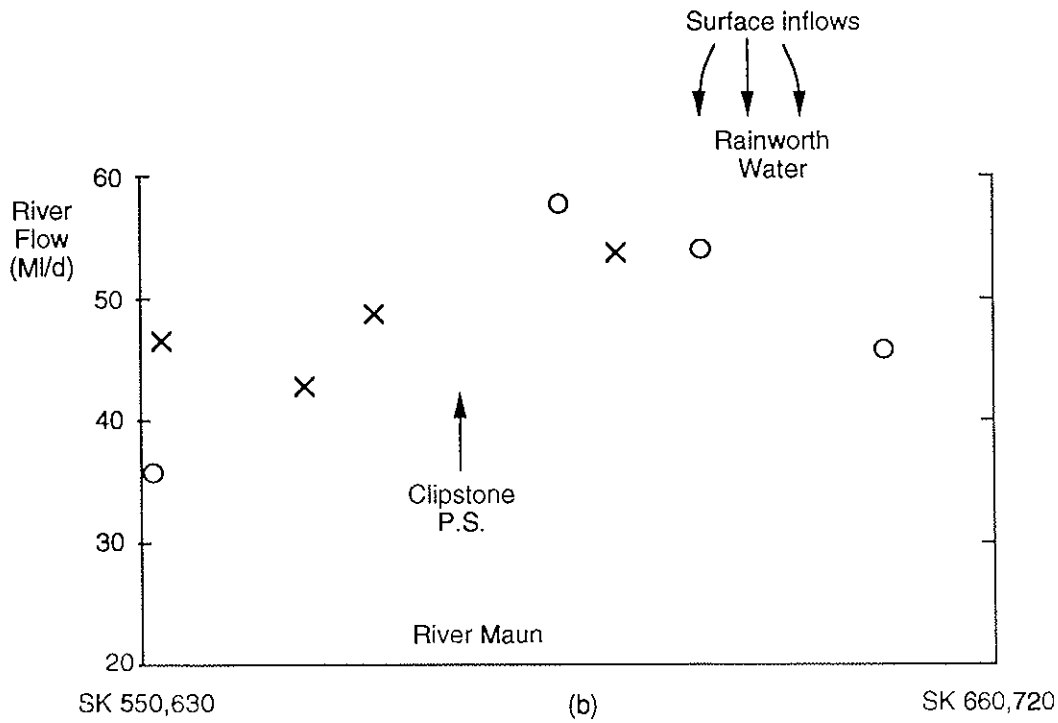
- the loss in the lower parts of the River Maun can be estimated from the last three circles of Figure 4(b), the loss is about 10 Ml/d in 7 km or 1.4 Ml/d/km.

Recognizing the tolerances in measurement, these values were used as a guide for the stream coefficients in the initial stages of the development of the model.

There is evidence from the quality of water in abstraction boreholes that some of the water drawn into the pumped boreholes enters the aquifer system from rivers. However, it is not possible to obtain reliable estimates of the quantity of water lost from the river using borehole chemistry due to the long flow times within the aquifer system. Water moving from a river to a borehole may take tens of years to reach the borehole; another important issue is that the three dimensional nature of the aquifer means that water entering the borehole at various depths will be of different ages.

Figure 4 - Results of spot gauging on the Rivers Meden and Maun: the symbol o refers to gauging in August 1992, the symbol x to gauging in September 1992.





Inclusion of river-aquifer interactions in a model

A groundwater model has been developed for the whole of the Nottinghamshire Sherwood Sandstone aquifer. Features represented in the model include:

- the whole of the Sherwood Sandstone Aquifer from Nottingham to Doncaster; the important parameters are the transmissivities and storage coefficients,

- recharge arising from precipitation and from water lost from water mains and sewers in urban areas,

- precipitation recharge and leakage through the Colwick formation,

- leakage through the Mercia Mudstones,

- abstraction from the aquifer; all abstractions greater than 0.1 Ml/d are included,

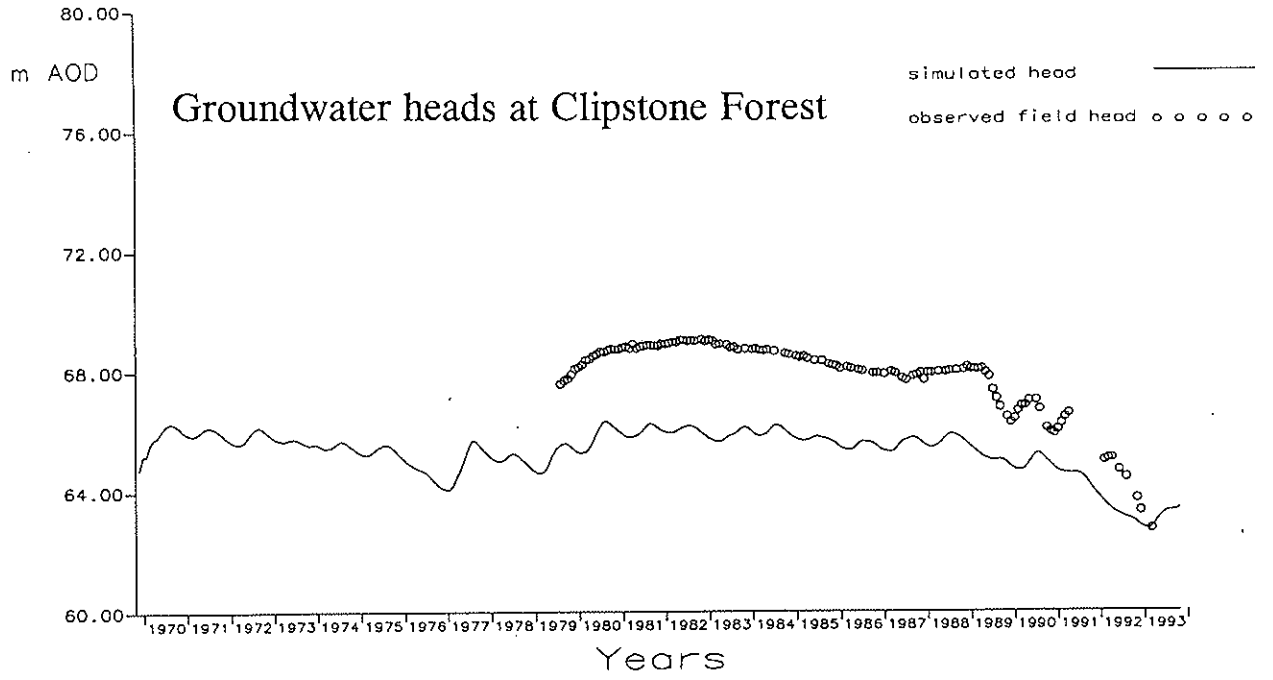
- river-aquifer interaction.

The model uses a finite difference backward difference formulation (Rushton and Redshaw 1979). River-aquifer interaction is included in the groundwater model according to equations (1). Initial values of river coefficients are quoted above but adjustments are made to the river coefficients and other parameter values to provide a best overall fit. The validity of the simulation is confirmed by comparisons with sixty observation boreholes and comparisons with stream flow records.

Two groundwater hydrographs which are in the catchments of the Rivers Meden and Maun are included in Figure 5. The Clipstone Forest observation borehole (SK 604 623) is 2 km south of the River Maun; there is a good comparison over a period of fifteen years between the field values shown by the circles and the continuous line representing the model. The decline in groundwater heads during the drought of 1989-93 is represented well. The second observation borehole at Bilsthorpe (SK 642 614) is on the eastern edge of the Sandstone outcrop; the long term small but steady decline in this area is reproduced by the model. When comparing field and modelled groundwater heads, the ability to reproduce changes in groundwater head may be more important than the absolute values. In this study the absolute values of groundwater heads depend critically on the river elevations which were deduced from contours on maps and are thereby liable to inaccuracies of a few metres.

Table 1 shows the river-aquifer interaction for the River Meden for August 1992, a similar table could be prepared for the River Maun. Information is included about the river and bed level, river coefficient, groundwater head as calculated from the groundwater model and the nodal river-aquifer interaction. River coefficients are lower than the values estimated from the gaugings of 2 Ml/d/km for the River Meden and 1.4 Ml/d/km for the

Figure 5 - Observation borehole hydrographs for Clipstone Forest and Bilsthorpe; circles represent field readings, model results are the continuous line.



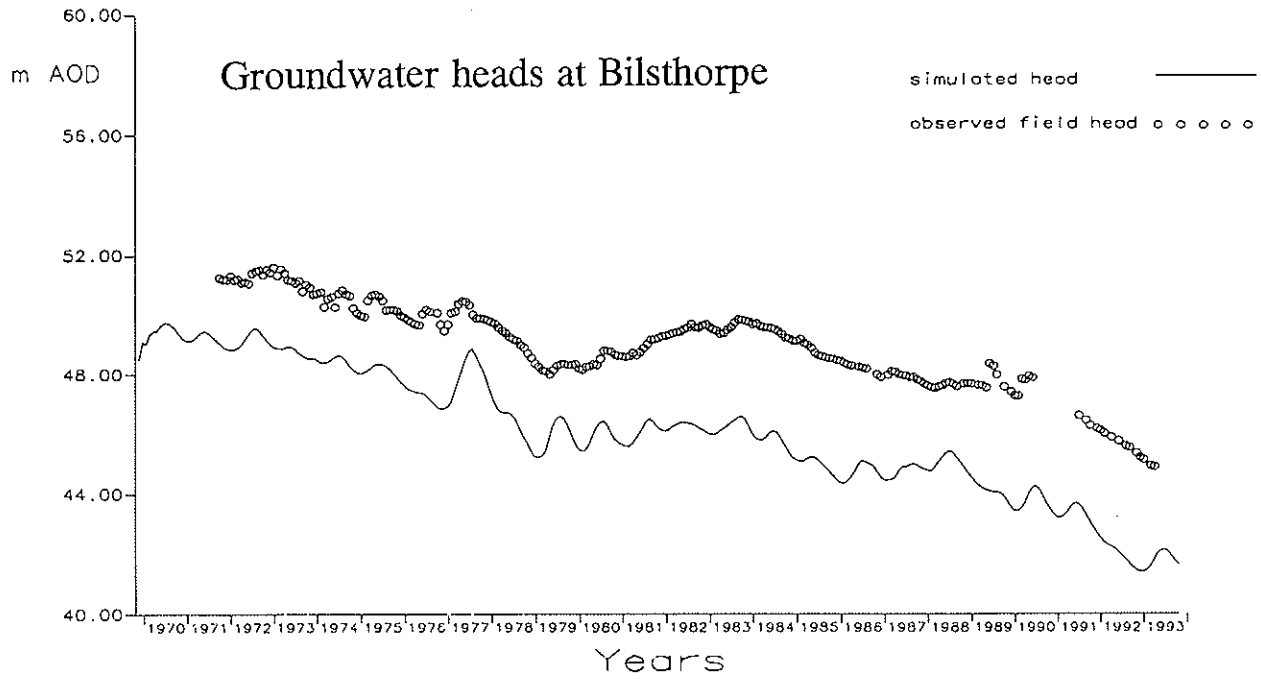


Table 1 - Model representation of stream-aquifer interaction for River Meden for August 1992

River Node Grid Ref	River Level (m)	Bed Level (m)	River Coeff (Ml/d/km)	Aquifer Head (m)	Aquifer - river flow (Ml/d/km)
55,66	53.0	52.0	0.925	54.28	1.19
56,68	51.0	50.0	0.925	52.01	0.93
57,69	48.5	47.5	0.925	49.63	1.04
58,69	46.0	45.0	0.925	47.29	1.20
59.70	43.5	42.5	0.925	41.35	-0.93
60,70	41.0	40.0	0.925	27.37	-0.93
61,70	39.0	38.0	0.925	38.33	-0.62
62,70	37.0	36.0	0.925	38.32	1.23
63,70	35.0	34.0	0.75	37.06	1.54
64,71	33.5	32.5	0.75	34.80	0.97
65,72	32.0	31.0	0.75	32.29	0.22

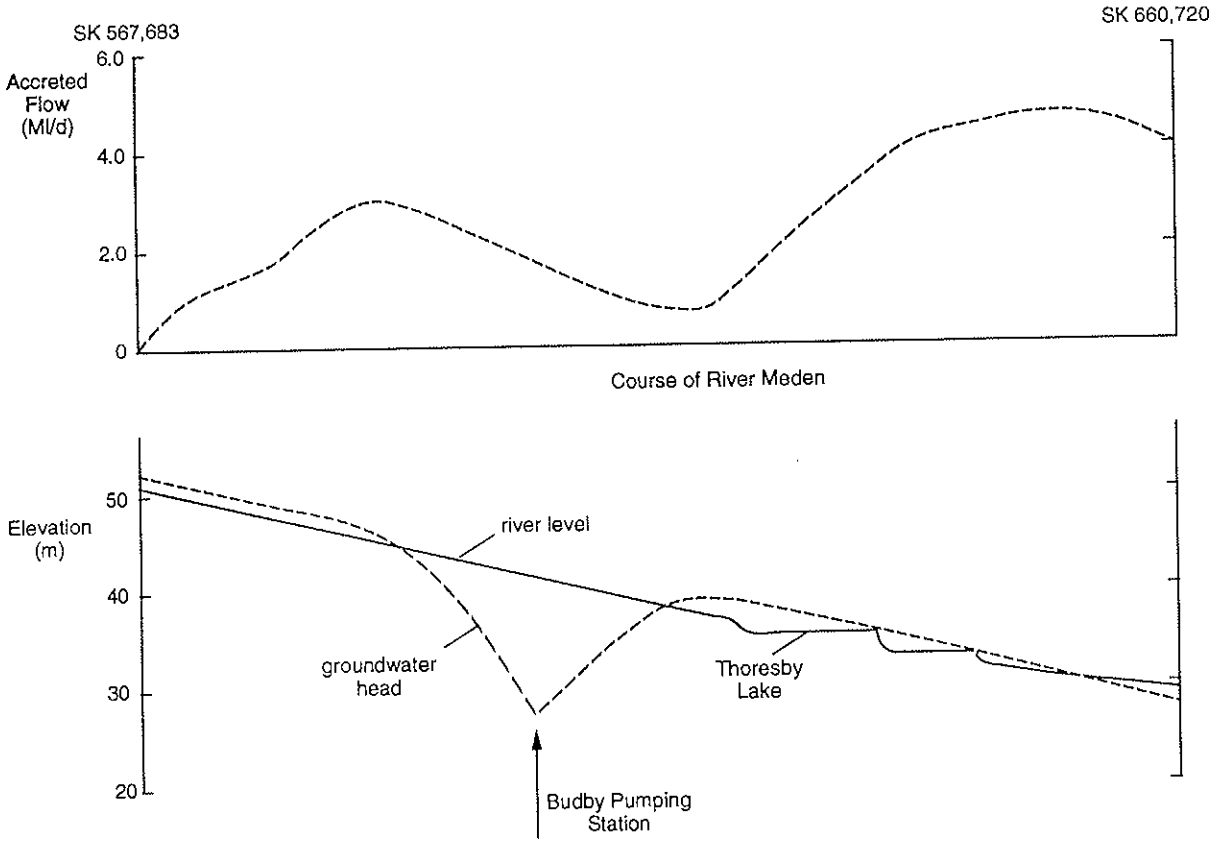
River Maun; values in the range 0.5 to 0.95 Ml/d/km were necessary to ensure that the model reproduced all the different aspects of the field observations of groundwater heads and flows.

Figure 6 contains two diagrams constructed from the model results. The lower diagram shows the relative elevations of the groundwater heads and the river or lake levels. The upper diagram shows the accreted flows. When the groundwater heads are above the river level, there is an increase in the accreted flows. However when the groundwater heads are below the river elevation, due to the effect of Budby Pumping Station, there is a loss from the river to the aquifer and a decrease in the accreted flows.

A comparison of the upper diagram of Figure 6 with Figure 4(a) shows that the model reproduces similar trends to the spot gaugings in the field. The field readings show greater variations than the model; this is partly due to the difficulties in gauging but a further effect is that the model gives averaged values over a month whereas the field readings reflect short term variations.

Significant changes do occur in the river-aquifer interaction, both seasonal and long term changes can be identified. Results derived from the model for August for five different

Figure 6 - Diagrams of calculated river-aquifer interaction for Rivers Meden and Maun, August 1992.



years are as follows:

August	Accretion (Ml/d)
1970	12.1
1978	8.6
1983	12.0
1988	9.4
1992	4.0

These variations occur because the rivers take away excess recharge.

Influence of rivers on flows in the aquifer

As indicated above, rivers take away the excess recharge from the aquifer. Consequently it is possible to identify areas of the aquifer from which water is collected by rivers. This approach is similar to the identification of groundwater protection zones associated with boreholes.

The areas which supply rivers can be identified by examining the inflows to the aquifer, the flows within the aquifer and the outflows into rivers. For the Rivers Meden and Maun there are three areas where the aquifer feeds the river, Figure 7; these results refer to February 1993.

Two of the areas, indicated by A and C on Figure 7, are on the upper reaches of the Rivers Meden and Maun. Recharge to the aquifer within these areas flows directly to the rivers. In each case the length of river gaining water from the aquifer is curtailed by the presence of the pumping stations at Budby and Clipstone.

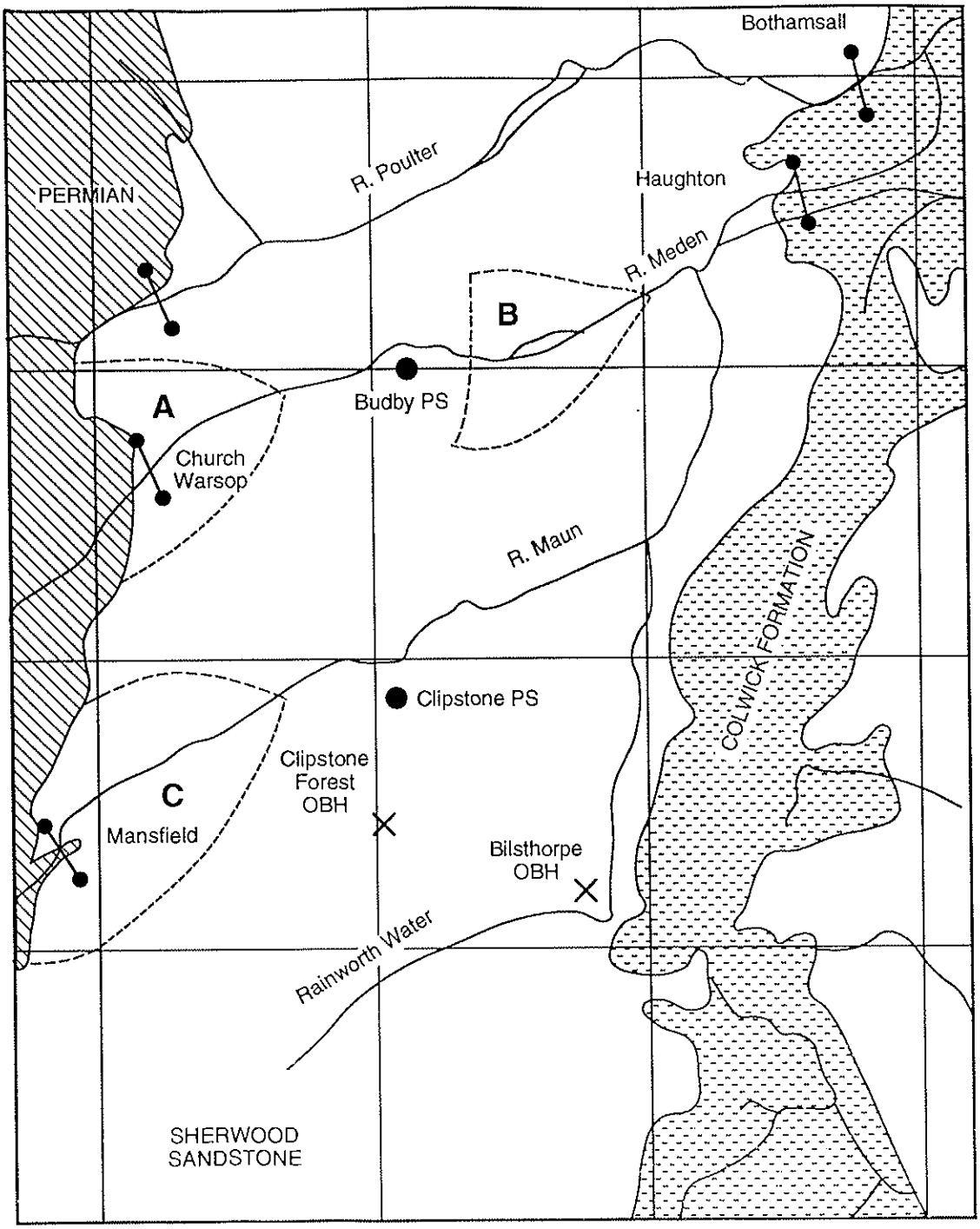
There is a second section of the River Meden below Budby Pumping Station which collects water from the aquifer; this is shown on Figure 7 as area B. This area coincides with Thoresby Lake; the unusual shape of the catchment is determined primarily by the influence of surrounding pumping stations.

Conclusions

This paper has examined the interaction of two of the rivers crossing the central portion of the Sherwood Sandstone aquifer of Nottinghamshire. There are no dramatic results with rapid changes in flow such as occur in many Chalk and Limestone aquifers. Nevertheless the rivers do have a crucial effect on the groundwater flow conditions. A significant proportion of the recharge to the aquifer system is pumped out from boreholes; this inevitably influences the river flows. In addition the rivers collect excess water. These two situations can apply to different reaches of a single river.

The difficulties encountered in obtaining field information about river-aquifer interaction

Figure 7 - Catchments for the sections of the Rivers Meden and Maun which gain water from the aquifer, catchments are denoted by A, B and C.



from continuous and spot gauging is highlighted. Uncertainties arise because the quantities involved in the river-aquifer interaction are small compared to the total flows in a river. Even detailed spot gauging results provide only an approximate indication of the interaction. Nevertheless the spot gauging results were used to estimate the river-aquifer interaction coefficients which could be incorporated in the numerical model of the aquifer system.

The brief review of the mathematical model indicates that the river-aquifer interaction is an important mechanism in the understanding of the aquifer responses. Whilst developing the model it was necessary to use values of the river coefficients which were smaller than those deduced from the field results. This was necessary to obtain a good match between hydrographs for groundwater head and changes in river flows.

Much information can be gained from the model; a typical example is the identification of the areas of the aquifer contributing to the river flows. The contributing area is influenced by many factors including the elevation and river-aquifer coefficients, the influence of other rivers, the magnitude of the recharge and the quantity withdrawn from pumping stations. It is the neighbouring pumping stations which have the strongest effect on the contributing area to the rivers.

Acknowledgements

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References

Rushton, K.R., Connorton, B.J., and Tomlinson, L. M., 1989, Estimation of the groundwater resources of the Berkshire Downs supported by mathematical modelling. **Quarterly Journal of Engineering Geology**, 22, 329-341

Rushton K.R., Owen, M. and Tomlinson, L. M., 1992, The water resources of the Great Oolite aquifer in the Thames Basin, UK. **Journal of Hydrology**, 132, 225-248.

Rushton, K.R., and Redshaw, S.C., 1979, **Seepage and Groundwater Flow**. Wiley, Chichester.

Rushton, K.R., and Tomlinson, L.M., 1979, Possible mechanisms for leakage between aquifers and rivers. **Journal of Hydrology**, 40, 49 - 65.

THE VALIDITY OF METHODS USED FOR MODELLING OF RIVER-AQUIFER INTERACTION

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Abstract

Modelling of river aquifer interaction is generally accomplished either by inclusion of the river within grid cells or by aligning the river as internal boundary elements along the sides of grid cells. The latter methodology has been successfully used by Mott MacDonald during a large number of catchment studies both in the UK and abroad. The effects of grid cell size on the accuracy of simulation of river/aquifer interaction was assessed during a modelling study in Bangladesh. Comparing the results of two strip models (one with coarse grid cell and one with a very refined grid) allowed for adjustment of parameter settings in the coarse grid model so that results from both models were comparable. The use of a constant value of river bed resistance, which results in a linear relationship between river-aquifer exchange and the hydraulic head difference between river and aquifer, was found to be invalid for flood flow conditions during a study of the Thames Gravel aquifer near Chertsey. Reduced river bed resistance values had to be adopted during flood flows in the river Thames. Further complexities arise where there is uptake of groundwater by vegetation in the riparian zone (resulting in a reduction in river baseflow) during extended dry summer periods.

Introduction

The interaction between rivers and aquifer systems is conceptually illustrated in Figure 1. Two commonly used methodologies for simulation of this interaction are:

- Inclusion of the river within grid cells with the exchange between the river and the aquifer (or aquitard) computed as vertical flow from:

$$q = A_r * (h_a - h_r) / C_r$$

where

A_r	-	river area within the grid cell [m^2]
h_a	-	piezometric level in the aquifer (aquitard) underlying the river (or the level allocated to the nodal point of the grid cell) [m]
h_r	-	river level [m]
C_r	-	river bed resistance [d]

In **Modelling River-Aquifer Interactions. (Edited by P L Younger). British Hydrological Society Occasional Paper 6. (1995). ISBN 0 948540 70 2**

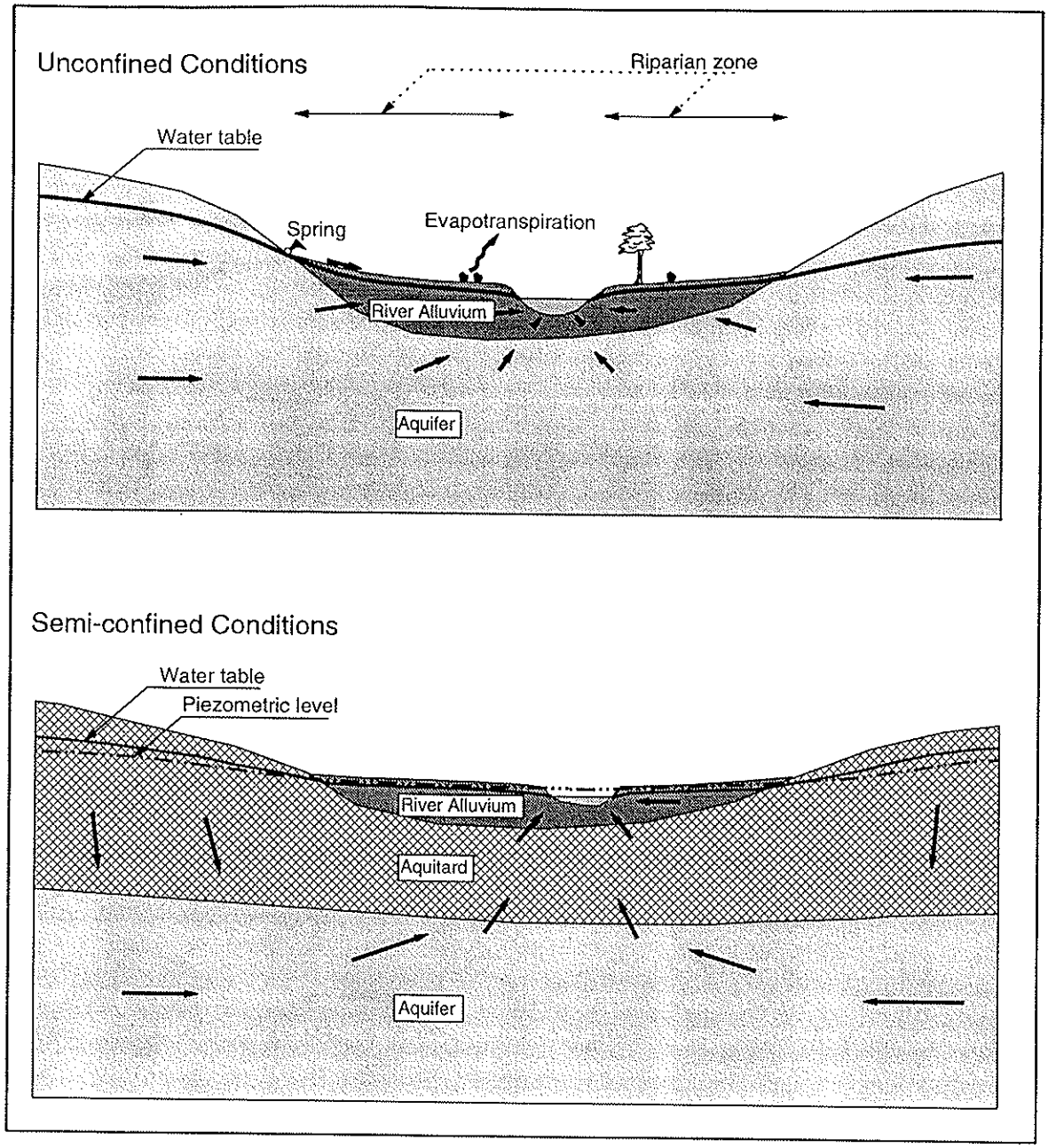


Fig. 1. Conceptual River-Aquifer Systems

If the piezometric level drops below the base of the river bed deposits, the downward flow becomes a constant flux, or

$$q = A_r * (B_r - h_r) / C_r$$

where B_r - elevation of the base of river bed deposits [L]

The above methodology is used in finite difference models such as MODFLOW.

For rivers which partially penetrate the upper layer of a multi-layer aquifer system, the influence of the river on the deeper aquifer is simulated as leakage flow over the full grid cell area (Figure 2). This may have a bearing on the selection of the size of grid cells that contain rivers.

- Inclusion of the river as an internal boundary element along the sides of grid cells.

  This methodology is used in Mott MacDonald's Integrated Catchment Model (ICMM). The internal boundary elements (IBE) are of small width in comparison to normal grid cells and are fully integrated within the catchment model. The transfer between IBE and the aquifer system fully conforms to the basic theories of the integrated finite difference method: continuity of mass and Darcy's flow equation. Different types of IBE can be used as illustrated in Figure 3.

The validity of the approximation of real river/aquifer interaction with model approximations is critically evaluated in the following sections. Use is made of findings from modelling studies carried out by Mott MacDonald in Bangladesh and the UK.

The Problem of Scale

The question of whether the use of coarse grid cells results in accurate simulation of river-aquifer interactions was addressed during the development of a regional groundwater model for the Dhaka Region of Bangladesh (Figure 4).

Large scale groundwater abstraction in Dhaka city has caused a decline in piezometric levels to the extent that part of this abstraction is balanced by inflow from the Buriganga river. The aquifer system in Dhaka city comprises two aquifers separated by a continuous clay layer. The upper aquifer is overlain by low permeability clays. The Buriganga river partially penetrates the upper clays and is locally in direct contact with the upper aquifer. Most of the groundwater abstractions are from the lower main aquifer.

The 200 m wide Buriganga river was represented in the model by internal boundary elements along the sides of 1.3 x 1.3 km grid cells. During the model calibration stage, two key questions had to be answered:

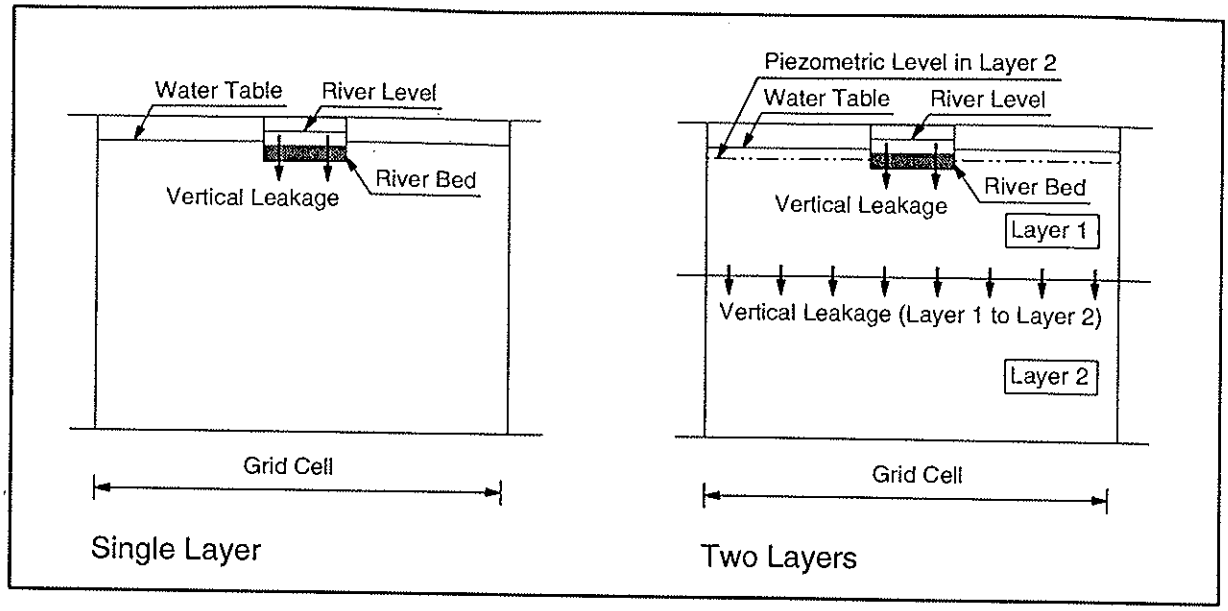
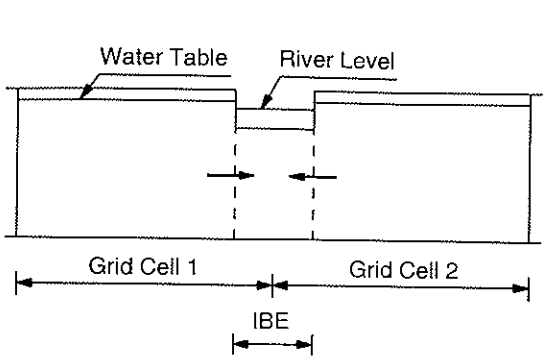
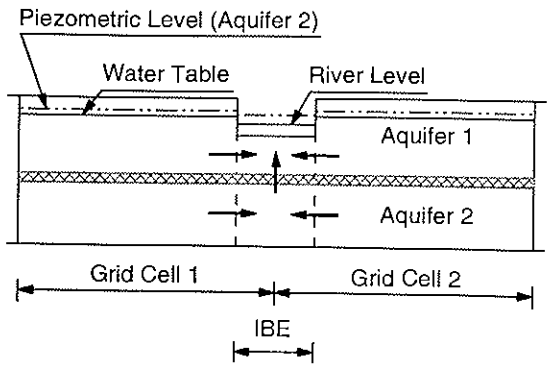


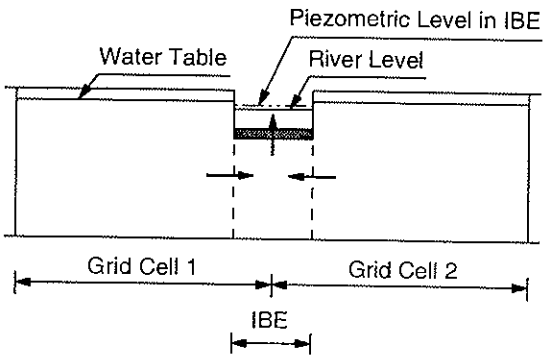
Fig. 2. Rivers Included Within Grid Cells



Type 1 - Fully Penetrating River
Influence of IBE relates to one aquifer
Increased resistance on IBE sides possible

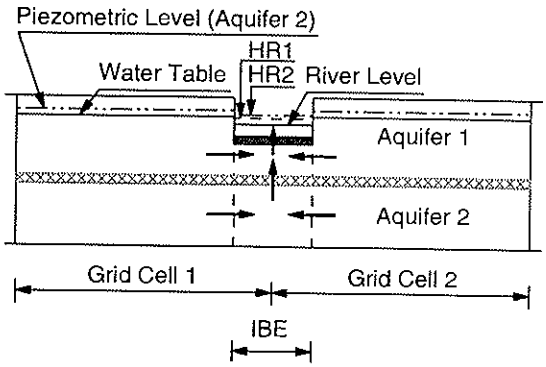


Type 3 - Fully Penetrating River in Upper Layer
Influence of IBE extends into 2 aquifer layers
Increased resistance on IBE sides possible



Type 2 - Partially Penetrating River
Influence of IBE relates to one aquifer

IBE - Internal Boundary Element



Type 4 - Partially Penetrating River in Upper Layer
Influence of IBE extends into 2 aquifer layers
Increased resistance on IBE sides possible
River influence extends into lower aquifer
HR1 - Piezometric level in upper aquifer within IBE
HR2 - Piezometric level in lower aquifer within IBE

Figure 3 Rivers Included on Grid Cells Sides

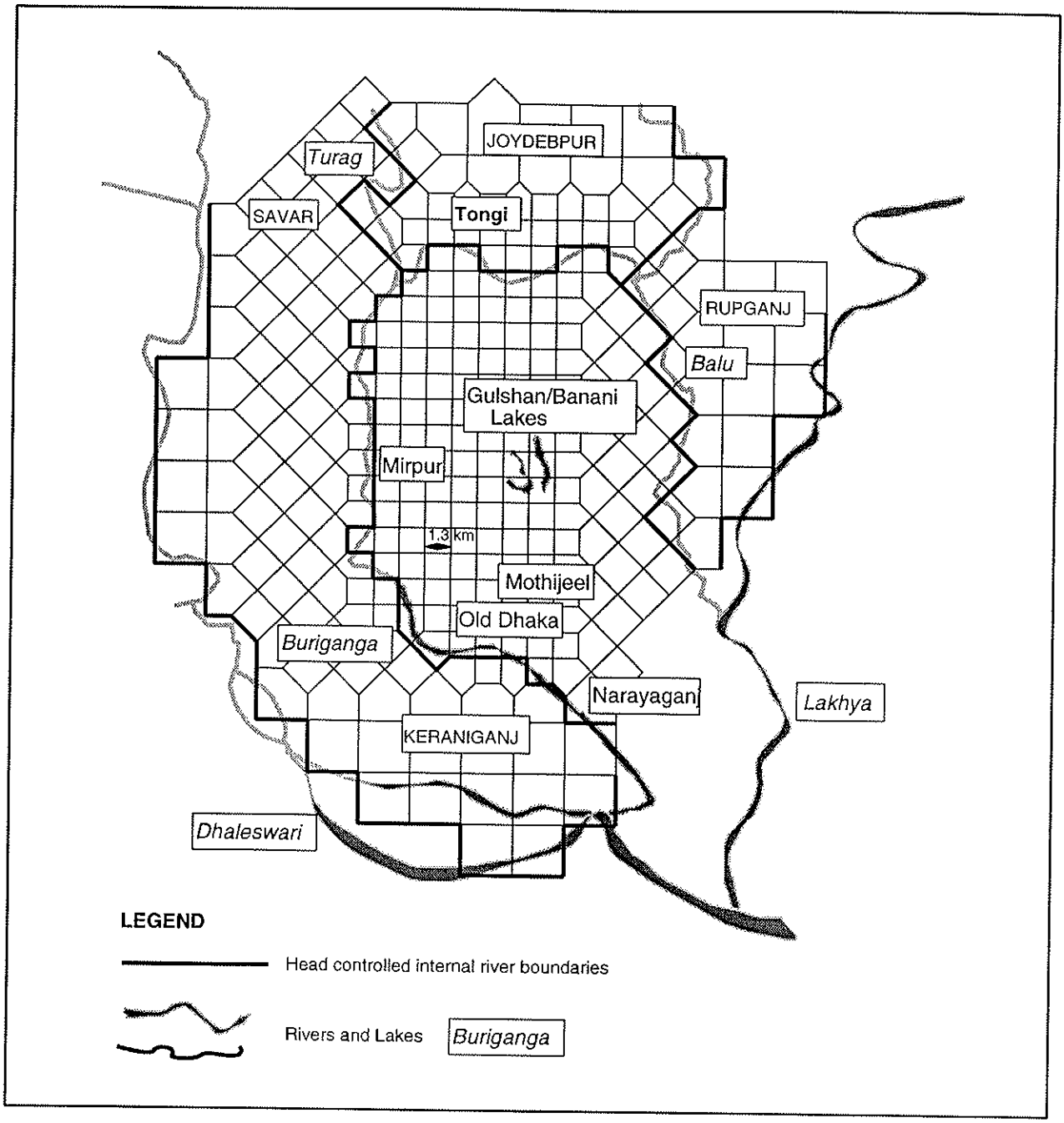


Fig. 4. Part of Regional Model Grid

- does the use of the relatively coarse model grid allow for accurate simulation of river-aquifer interaction, given that no river flow records were available for model calibration?

- do the internal boundary elements have to penetrate into the main aquifer?

To answer both questions, two strip (ie vertical profile) models were established. A detailed strip model was constructed consisting of 52 100 x 100 m cells, with the river located in the central two cells. A second strip model consisted of 4 1.3 x 1.3 km grid cells with the river positioned in between the two central grid cells. The second strip was representative of the conditions used in the regional model. Both strip models included the four layers included in the regional model, with the river partially penetrating the upper clays.

Both strip models were run in steady state mode with two rates of abstraction (Figure 5) representative of typical rates of abstraction from wells located near the river.

The detailed strip model indicated that a large proportion of the leakage from the river migrates to the lower main aquifer over a limited width (Figure 5). This highlighted the need to extend the internal boundary element into the lower aquifer, with most of the leakage from the river moving into the lower aquifer within the IBE.

To obtain comparable flows from the river to the aquifer, only one modification had to be introduced in the four-cell strip model. This related to the hydraulic resistance of the aquitard which separates the upper and lower aquifers within the IBE. The value had to be reduced to take account of the effectively greater width over which most of the leakage between the upper and lower aquifers occurs (Figure 5).

Validity of the Use of a Constant River Bed Resistance

The use of a constant river bed resistance results in a linear relationship between the river/aquifer exchange and the hydraulic head difference between river and aquifer.

The use of such constant resistance is, however, not always valid, as was demonstrated during a modelling study of the Thames Gravels in the Chertsey area in west London.

Groundwater abstraction from the Thames Gravels at Abbeymead induces inflow of river water from the Thames. The general piezometric pattern is as illustrated in Figure 6. The river levels in the Thames are largely controlled and only show limited variations for most of the time. However, during extremely wet conditions in the Thames catchment, lock control causes the river level to rise to the extent that flooding of small strips of land along the east bank of the river occurs. Although the river level only rises by about 0.4 m, the groundwater level in the Thames Gravels shows a rapid rise of over 2.0 m in the area around the Abbeymead source. This large rise cannot be attributed to rainfall recharge to the gravels; a rainfall recharge of about 40 mm over the 10 day period of groundwater level rise would only cause a rise of some 0.3 to 0.6 m (assuming

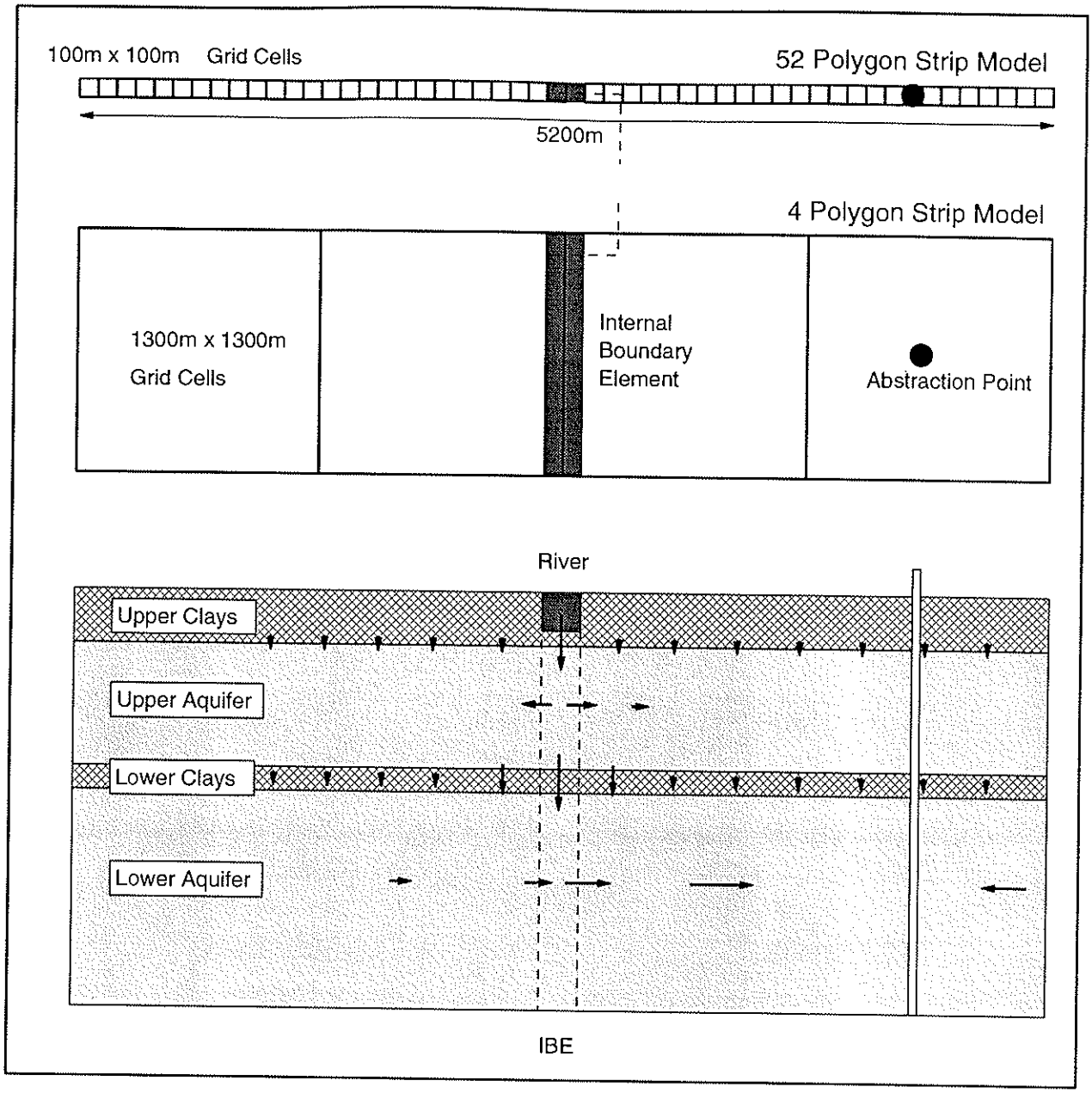
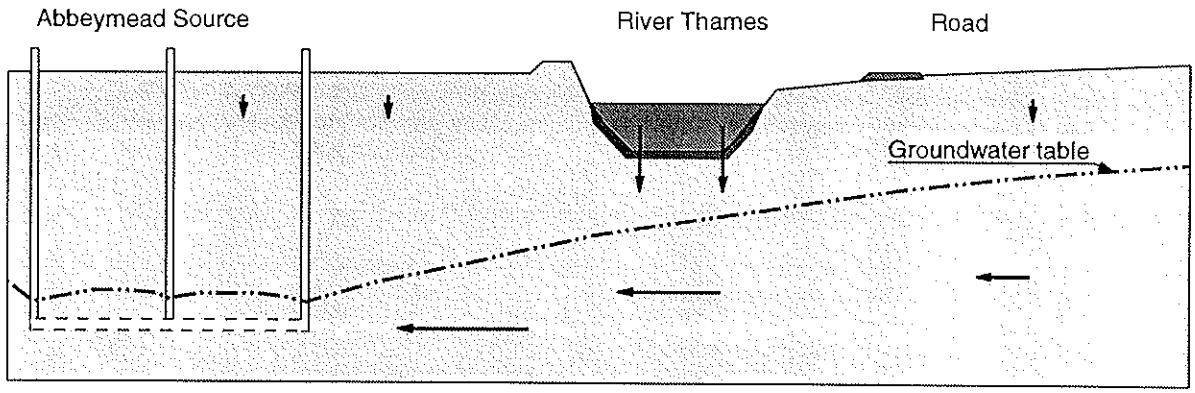
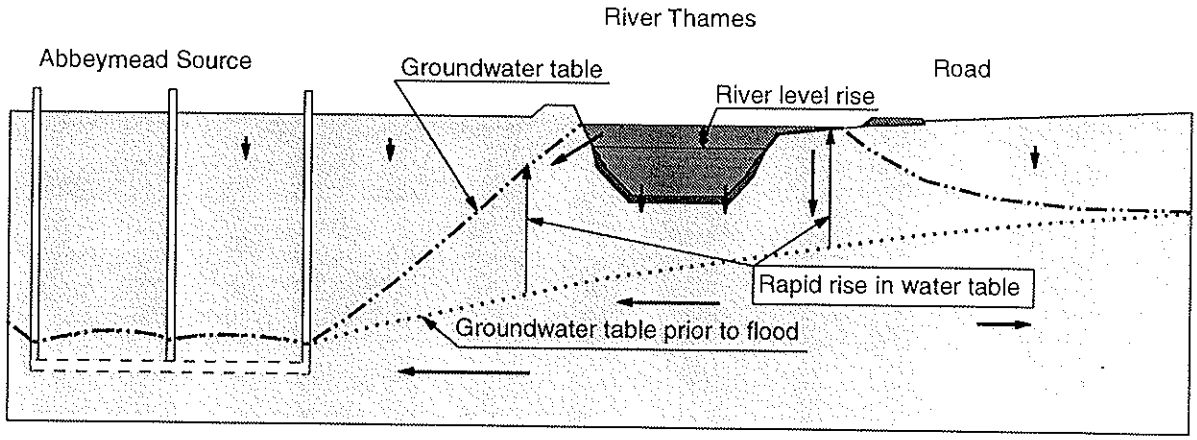


Fig. 5. Schematic of River Strip Models



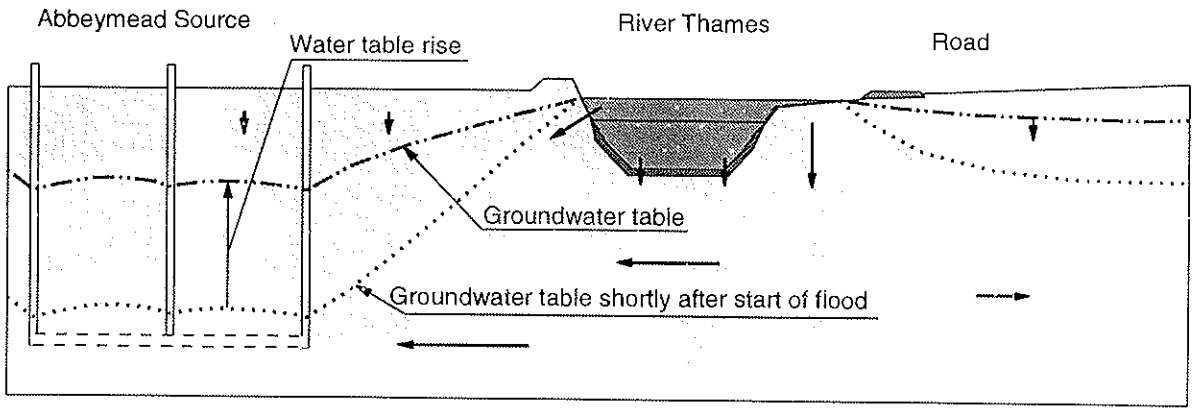
Note: Not to scale

Fig. 6. Conditions Prior to the Flood



Note: Not to scale

Fig. 7. Conditions Shortly After the Start of the Flood



Note: Not to scale

Fig. 8. Conditions at the End of the Flood Period

a specific yield value of the gravels in the range of 0.075 to 0.15).

The large rise in groundwater level can therefore only be attributed to the flood conditions of the river Thames. The effects of the flood on the aquifer was conceptualised as follows:

- For most of the time the river is contained within its banks and river levels are controlled and relatively constant. A continuous downward movement of river water to the underlying aquifer occurs through a moderately resistant river bed (Figure 6). Underflow from east to west occurs under these circumstances as was evident from water level monitoring.

- During a flood, river water covers a strip of land along the east bank of the river. The high permeability of the sub-soil in this flooded strip causes a high rate of infiltration and a groundwater level rise to approximate flood level (Figure 7).

  Although no flooding occurs along the west bank, the rise in river level will have caused a lateral inflow through silt free gravels of high permeability. Observed groundwater level rise in piezometers near the river bank supports this hypothesis (Figure 7).

- The sudden rise in groundwater levels along the river banks causes a sudden increase in hydraulic gradient and thus an increase in groundwater flow towards the abstraction source. With no change in output from the source, the aquifer responds with a rapid rise in groundwater level at the source in excess of the rise observed near the river bank. This is explained by the need for a shallower hydraulic gradient than prior to the flood period for the same amount of groundwater flow towards the source. This is due to the increase in saturated aquifer thickness and hence the aquifer transmissivity (Figure 7).

At the end of the flooding period there is a much reduced hydraulic gradient between the river and the source. After river levels have returned to their normal levels, a rapid decline in groundwater levels at the source occurs. Abstraction relies partly on groundwater storage release until pre-flooding gradients and water tables have re-established.

The observed piezometric response during the flood event could not be simulated by using a constant value of river bed resistance. The use of a reduced resistance during the flood resulted in a good agreement between modelled and observed aquifer piezometry.

The Influence of the Riparian Zone

Figure 8 illustrates the influence of the riparian zone on groundwater baseflow to the river. The occurrence of shallow water table conditions in the riparian zone causes

actual evapotranspiration to be at potential rate, even during dry summer periods. During wetter parts of the year this evapotranspiration is mainly supported directly from rainfall recharge.

During extended dry summer periods, uptake by vegetation is mainly by capillary upward movement from the shallow water table to the rootzone, while the position of the water table is maintained by groundwater flow. Part of this groundwater flow would otherwise have been directed towards the river as baseflow.

To put this interception of groundwater flow in the riparian zone into perspective, one may assume a 50 m wide zone and a potential evapotranspiration rate of 3 mm/d. If this is all supported from groundwater flow, it implies a reduction in baseflow contribution of 150 m^3/d/km (0.0017 m^3/s/km). Over a river section of 10 km this thus amounts to 0.017 m^3/s. Comparing this with low summer flows in some of the UK rivers places the significance of this potential baseflow interception into context. For example:

Location	95% Low Flow	Potential evapotranspiration from a 50 m wide riparian zone and actual river length upstream of the gauging station
River Darent at Hawley	0.08 m^3/s	0.04 m^3/s
River Little Ouse at County Bridge	0.13 m^3/s	0.03 m^3/s

Modelling of the effect of the riparian zone on river baseflow reduction firstly requires a correct understanding of the relationship between groundwater table, aquifer piezometry and the capacity for uptake of water from the groundwater table by vegetation. Water uptake from the riparian zone could be based on a moisture balance analysis, with groundwater uptake determined from the balance of potential evapotranspiration and effective rainfall (ie rainfall minus interception and surface runoff). The baseflow to the river calculated using the methodologies described above would then have to be reduced by the derived baseflow interception. The approach could be verified by the use of detailed strip models.

The alternative would involve the development of a very fine model grid in the riverine areas, so that the riparian zones can be accommodated into separate grid cells. However, this may make the computational effort of the model excessive.

Conclusions

- The lesson learned during a regional modelling study in Bangladesh is that one should firstly develop a correct conceptual understanding of the river/aquifer interaction and, secondly, convert this understanding into correct modelling concepts. Both modeller and model code require the ability to accommodate unusual situations.

Parameters which control the exchange between rivers and aquifers may be dependent on the size of grid cells associated with these rivers. When gauging data are not available for comparison with simulated river flows, verification of model simulation with detailed strip models is recommended.

- The use of a linear relationship between river-aquifer exchange and river bed resistance for a given hydraulic head difference is common in public domain software. Recent studies of the exchange between the River Thames and the Thames Gravels have shown that a linear relationship does not allow for correct model simulation. Again the importance of a correct conceptual understanding of the river-aquifer exchange mechanism, and the ability of the model to accommodate non-linear river-aquifer exchange mechanisms is important.

- The potential reduction in river baseflow by interception of groundwater flow in the riparian zone is of particular significance during extended dry summer periods. To accommodate the riparian effects in the model, two options are available. The first option is to refine the model grid to accommodate the riparian zones within grid cells. This would have repercussions on the computational requirements of the model. The second option would include the moisture balance of the riparian zone within the water balance of river elements, and would not require further refinement of the model grid.

- Experience has shown that successful modelling of unusual situations relies on:

  - a correct conceptual understanding of the river-aquifer interaction;
  - the correct translation of these concepts into modelling procedures;
  - the ability to adapt model code.

SIMULATION OF STREAMFLOW BEHAVIOUR IN CHALK CATCHMENTS

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Abstract

A groundwater model (SLAY) has been applied successfully to the simulation of river-aquifer interactions in two Chalk catchments in Southern England. Certain of the routines used in the two models (including the leakage flow simulation routine) have been widely used, but were adapted to meet specific requirements of streamflow prediction in the Chalk catchments investigated in these studies. A streamflow accretion routine developed in these studies uses runoff data calculated using the Penman-Grindley method, river-aquifer leakage flow, upstream flow and other flows to determine the total streamflow at a given model river node. The streamflow accretion routine also provides a reasonable simulation of non-perennial flow behaviour typical of Chalk aquifers. The use of longitudinal flow profiles along watercourses proved to be extremely valuable in determining the appropriate distribution of leakage coefficients to apply at model river nodes. Experience gained during model development, demonstrates the importance of incorporating the simulation of variable transmissivity with saturated depth, to the prediction of streamflow in Chalk aquifers. The completed groundwater models represent important tools for augmenting the full scope of activities involved in catchment management.

Introduction

Chalk aquifer sources account for over 70% of the groundwater supplies of England and Wales. These aquifers also sustain the baseflows of many major rivers and streams in southern and eastern England. The increasing potential for conflict between the need to meet water supply requirements whilst ensuring high environmental standards has, in recent years, placed a rising demand on the capabilities of methods used in the study of the interaction of Chalk aquifers with contiguous surface watercourses.

In the context of regional aquifer flow models, where the principal aim is to provide a means of predicting groundwater movement over a wide area, there can be difficulties of both space and time resolution in the simulation of streamflow. The simulation of river-aquifer interaction requires the representation of a phenomenon that may occur over a short period and within a small area, in a model which may be discretised into time steps of days and nodal elements of hundreds or thousands of square metres. An

In Modelling River-Aquifer Interactions. (Edited by P L Younger).
British Hydrological Society Occasional Paper 6. (1995). ISBN 0 948540 70 2

average head which applies to an entire element for the duration of each time step, becomes the only transient variable upon which influent and effluent leakage flows can be based.

In recent years, the environmental considerations related to Chalk streams have been the subject of a range of studies directed at the alleviation of low flows, with such problems identified in more than 40 rivers in England and Wales. Chalk streams have a unique ecology and are particularly important for the fisheries that they sustain. The imperative to find solutions is heightened by present public sensitivity, not only to general environmental concerns, but also to the specific issues that surround groundwater abstraction in Chalk aquifers.

The method of simulating Chalk river-aquifer behaviour described in this paper, is based primarily on a regional groundwater model, i.e. a model defined here to mean one that simulates the saturated zone of the aquifer only. Such a model is less demanding on data requirements than a full catchment model which may include sophisticated representations of the unsaturated zone and hydrodynamic simulation of surface flow.

Baseflow Analysis Methods

The study of the baseflow or groundwater component of a stream hydrograph is relevant to a wide range of surface water and groundwater problems. As a result, research into this area has been carried out from several different disciplines. Hall (1968) provides a valuable review of the historical development of the mathematical, statistical and graphical techniques that have been developed for use in analyzing baseflow.

Few generalisations can be made regarding the complex relationships that exist between surface and groundwater flow. This presents difficulties for deriving numerical schemes that would use field measurements of aquifer characteristics to predict baseflow. The problem is demonstrated by a study carried out to establish the variables that influence natural outflow from an aquifer. Parizek *et al* (1969) attempted to relate 18 parameters to baseflow using regression techniques. Few of these variables, which included land use categories, geological parameters, physical and geomorphological features, showed any reasonable correlation with leakage flow.

Nutbrown and Downing (1972) recognise the effect of the relationship between local storage coefficients (S) and transmissivities (T) on natural outflows. They define the aquifer response time, the ratio T/S, which governs the range of flows encountered in a river fed by a groundwater source. The flow range of a river is greater for higher ratios of T/S and affects the character of outflows in aquifers where field evidence suggests a sharp variation in storage coefficients and transmissivities with depth. These variations with depth are particularly important in Chalk aquifers which often include a layer of higher permeability towards the ground surface. Many of the considerations

faced in the study of problems involving parameter variation in an unconfined Chalk aquifer are discussed by Headworth *et al* (1982).

Computer simulations of aquifer-river leakage flow are generally based on a numerical scheme which includes the aquifer river head difference, a factor to correct for the difference between the nodal area and the wetted area of the river bed, and a coefficient to account for the impedance to leakage flow resulting from river bed sediment. Schemes based on this approach have been widely investigated by authors such as Herbert (1970), Prickett and Lonnquist (1971) and Miles (1987). In an important paper investigating possible river-aquifer exchange mechanisms, Rushton and Tomlinson (1979) show that baseflow recession is independent of the magnitude of the leakage coefficients.

A different approach is used to calculate the baseflow from the aquifer in a model described by Oakes and Pontin (1976). Outflows which occur when groundwater heads exceed river levels, are determined as a direct loss of storage equivalent to the difference between the groundwater head and a specified topographical height for selected nodes. The river is assumed to be fully penetrating and constant transmissivity conditions prevail throughout simulation. The approach also includes a routine for calculating the total baseflow accumulated to each river node.

Modelling Approach

Groundwater Simulation

The modelling approach adopted in the case studies described in this paper draws on various elements of the general schemes described in the previous section. The catchment groundwater model used is based on SLAY, a model developed by Dr Andrew Spink at the University of Birmingham. The SLAY model is two-dimensional and is based on the finite difference approximation of the time-variant governing differential equation for groundwater flow given in Equation 1.

$$\frac{\partial}{\partial x}\left(T_x\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(T_y\frac{\partial h}{\partial y}\right) = S\frac{\partial h}{\partial t} - q \qquad (1)$$

where:

h is the groundwater potential or head

T_x and T_y are the transmissivities in the x and y directions, such that

$T_x = f(h,x)$, $T_y = f(h,y)$

S is the storage coefficient

q is the recharge or abstraction per unit area, q = q(x,y,t). ie the sink or source term.

Groundwater flow behaviour is modelled in the X and Y directions (i.e in the horizontal plane). The third dimension, in the Z direction, is represented by the groundwater potential or head at each grid node, which is calculated at each time step.

Recharge

Recharge to the aquifer is entered at the start of a model run. It is calculated using the method based on the approach proposed by Penman (1949) for determining potential evaporation. This approach has been augmented in further research, by observation and in specific applications of the basic theory to many different field problems. The method was modified by Grindley (1967) who related vegetation and seasonal variations to evapotranspiration. This was an important development which allowed the temporal and spatial variation of catchment land use to be taken into account in recharge estimation. This method was revised further for the treatment of conditions relevant to Chalk catchments, in an approach proposed by Rushton and Ward (1979). This refinement allows a fraction of the effective precipitation to enter the aquifer directly, addressing a shortcoming of the former method which tended to predict lower than observed summer recharge in Chalk aquifers.

The model used to calculate recharge also determines the direct surface runoff. The gross recharge and runoff rates entered into the model for each time step are distributed across the modelled aquifer area by nodal factors.

River-Aquifer Exchange

The river-aquifer leakage is included as a part of q, the source or sink term using the numerical scheme described previously that is in wide use:

$$q_{riv} = l_{riv} * (h_{riv} - h) \qquad (2)$$

where:

h_{riv} is the river level
h is the groundwater head at the end of the time interval
l_{riv} is the river leakage coefficient (which includes the vertical permeability divided by

the vertical distance over which the head change occurs and an areal reduction factor to account for the difference between nodal area and river bed area).

It is clear from inspection of Equation 2 that the higher the leakage coefficient, the greater the potential for leakage flow (either in or out of the aquifer) to occur. The direction of this leakage flow is governed by the relative difference between groundwater and river levels.

In the scheme adopted here, the operation of the river leakage factor is constrained by groundwater head relative to the river bed level. It also depends on whether or not there is flow from upstream. Outflow from the aquifer will only occur if groundwater levels are above the river bed level. Inflow to the aquifer will only occur if there is streamflow from upstream entering a node, or if the groundwater level falls below the river bed during the time step.

Under ephemeral conditions, river-aquifer leakage flow is possible at a node for a given time step, if the groundwater level at the start of the time step is above the river bed or if there is flow from upstream entering the river node. If neither apply, then no river-aquifer flow is allowed at that node for the duration of that time step.

Streamflow Accretion

The various flow components assumed to apply at model river nodes are shown in Figure 1. The total streamflow at node n for a given time step is calculated from the following:

$$(Q_r)_n = (Q_{ra})_n + (Q_{ro})_n + (Q_{ae})_n + (Q_r)_{n-1} \qquad (3)$$

where:

$(Q_r)_n$ is the total streamflow at node n
$(Q_r)_{n-1}$ is the total streamflow at the adjacent upstream node to node n
$(Q_{ro})_n$ is the direct surface runoff to node n
$(Q_{ra})_n$ is the river-aquifer leakage flow at node n
$(Q_{ae})_n$ includes any augmentation or effluent return flows to the river at node n

Direct runoff to the catchment outlet is distributed regionally by a factor which is determined as the ratio of the catchment area associated with a river node to the total drainage area:

$$(Q_{ro})_n = rdf_n \times Q_T$$

where:

Q_T is the total direct runoff at basin outlet
rdf_n is the runoff distribution factor at node n

The runoff distribution factors are determined as:

$$rdf_n = \frac{A_n}{A_T} \tag{4}$$

where

A_n is the surface drainage area between node n-1 and node n
A_T is the total surface catchment area such that :

$$A_T = \sum_{n}^{nriv} A_n \tag{5}$$

where ***nriv*** is the total number of river nodes used to represent the watercourse within the catchment.

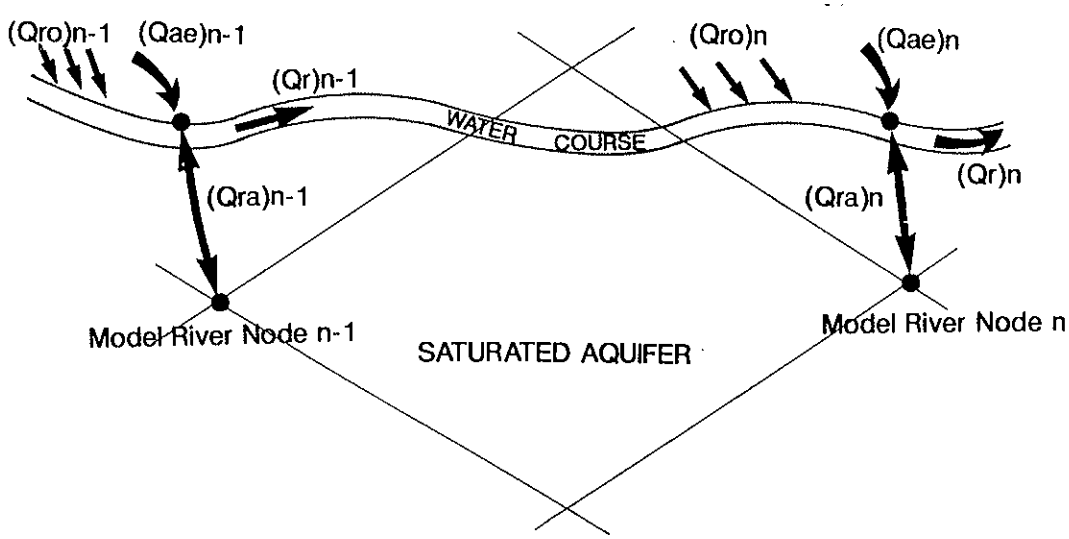
The streamflow accretion routine is carried out at the end of each time step, following the calculation of the groundwater head distribution. A further routine is also included to accumulate streamflows in catchments where there is branching of the watercourses.

Variable Transmissivity

The importance of transmissivity and its variation with saturated depth in Chalk aquifers has been discussed. This feature is represented in a variable transmissivity routine which has been incorporated into the SLAY model. In this routine, transmissivities are determined from generalised functions defined for each node used in representing the unconfined areas of the aquifer. These functions are used to evaluate the transmissivity at any given time step from the saturated depth resulting from conditions at the end of the previous time step. The form of the relationship used to represent the local hydrogeology can be varied to include step, linear and non-linear functions.

Application to Chalk Catchments

It is not the intention of this paper to provide a rigorous account of the case studies cited in this section, i.e. for the River Piddle and River Wylye. These studies are far



KEY

(Qr)n is the total streamflow at node n

(Qr)n-1 is the total streamflow at the adjacent upstream node to node n

$(Qro)_n$ is the direct surface runoff to node n

(Qra)n is the river-aquifer leakage flow at node n

(Qae)n includes any augmentation or effluent return flows to the river at node n

Figure 1 Schematic Diagram showing components of Streamflow Accretion Routine

wider in scope than the aspects considered here and are only referred to in order to illustrate major features of the groundwater modelling techniques described.

The streamflow and variable transmissivity routines were applied in two similar resource studies for the River Piddle and the River Wylye, each involving river-aquifer interaction along watercourses that display characteristics typical of Chalk catchments. The studies were carried out on behalf of the National Rivers Authority (South Western Region), by Sir William Halcrow and Partners. A groundwater model was developed in each study to provide a means of decision support for long term water resources management.

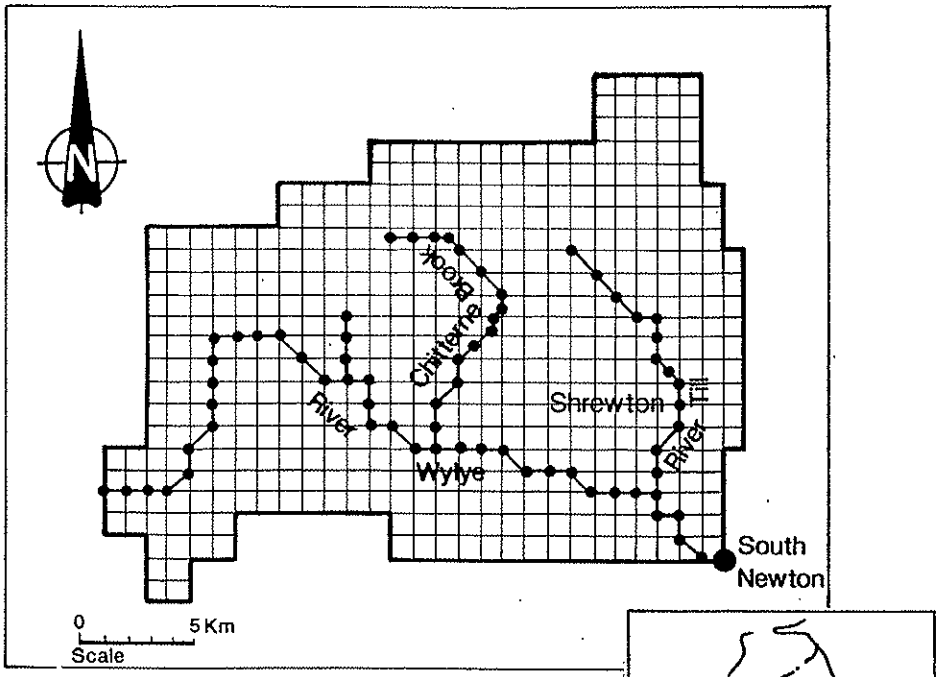
For the River Piddle groundwater model, a uniform mesh interval of 1 km is applied in a 24x19 node grid, with 65 active river nodes to represent regional flow over an area of 208 km^2. The groundwater model to simulate the River Wylye catchment of 472 km^2, uses a regular mesh interval of 500m in a 60x50 mesh interval, with 127 active river nodes. The location maps of the study areas and the finite difference grid boundaries for each of these models, including the overlaid river networks, are shown in Figures 2 and 3 respectively.

Model Development

Each of the models were developed using similar strategies. Following compilation of the data sets and construction of the finite difference grids, a series of sensitivity trials were carried out to establish the relationship between adjustments to input parameters and the response of the model in selected areas of interest. These relationships were used in conjunction with hydrogeological interpretation and field information to refine the model and simulate the historical behaviour of the aquifers for comparison with recorded data. The records used in these comparisons included time series data for streamflows and groundwater levels, and also spot gauging data monitored along watercourses.

The aspects of this exercise that were found to be critically important with respect to predicting river-aquifer leakage in Chalk catchments were recharge estimation, inclusion of variation of transmissivity with saturated depth and the choice of l_{riv}, the river leakage coefficients, along the watercourses.

Accurate runoff and recharge estimation is vital to the development of a river-aquifer simulation model. The model geometry and the nodal aquifer parameters simply distribute the recharge and runoff spatially. Difficulties in matching temporal trends are generally a consequence of inaccuracies in the recharge and runoff estimation and cannot be corrected simply by altering aquifer parameter distributions. Considerable effort had to be devoted to determinations of regional land use and to sensitivity trials to determine appropriate root constants as required by the Penman-Grindley method,



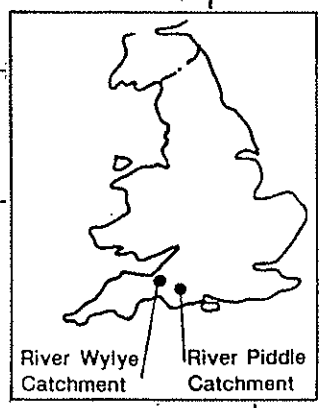


Figure 2 Finite Difference Grid for River Wylye Groundwater Model

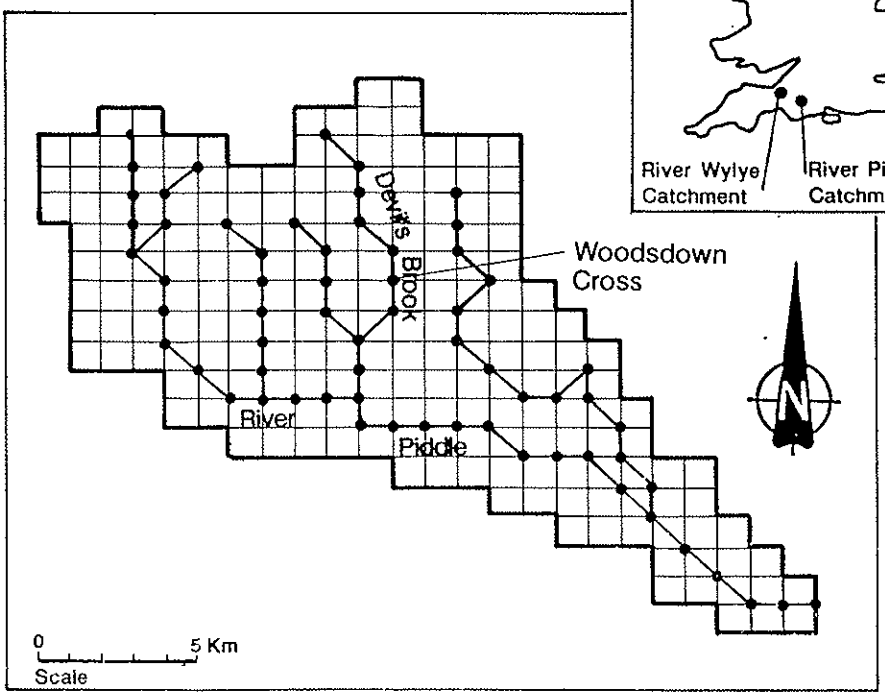


Figure 3 Finite Difference Grid for River Piddle Groundwater Model

for the different zones across the aquifer.

The sensitivity analyses to determine suitable relationships to describe the variation of transmissivity with saturated depth were supplemented by the results of packer tests carried out at selected sites across the aquifer. It was found that inclusion of variable transmissivity is particularly important to the simulation of aquifer baseflows. Figure 4 demonstrates the streamflow responses produced by different transmissivity relationships.

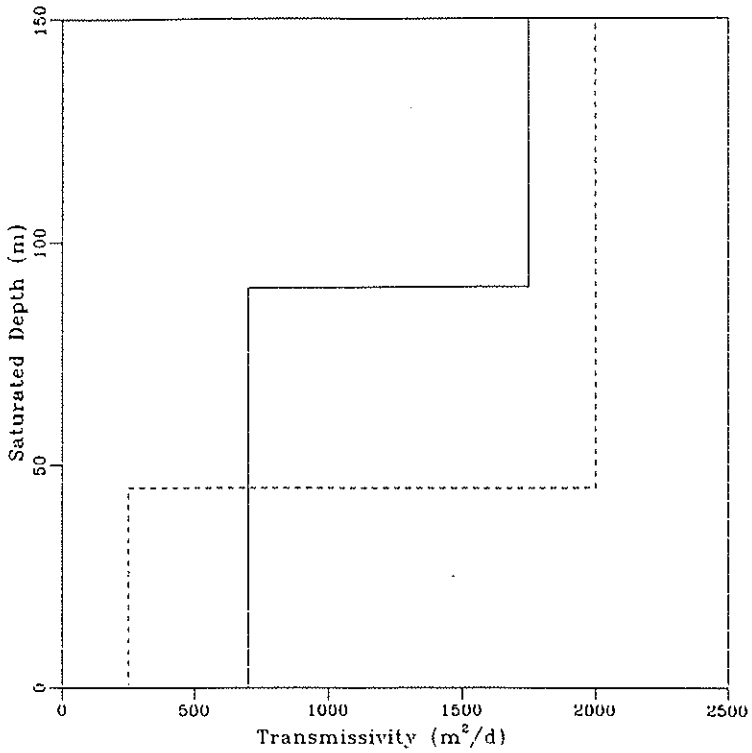
The river leakage coefficients, *lriv*, applied to nodes representing the watercourses have a major effect not only on the streamflow conditions, but also on the overall behaviour of local groundwater heads. Since these parameters cannot be measured directly through any field testing, they have to be determined in sensitivity trial exercises. A significant aspect of the model refinement carried out in establishing the river leakage coefficient distributions, was the use of longitudinal flow gauging surveys in comparisons with simulated profiles. These surveys were each carried out over a short time and provided a snapshot of the flow regimes along catchment watercourses.

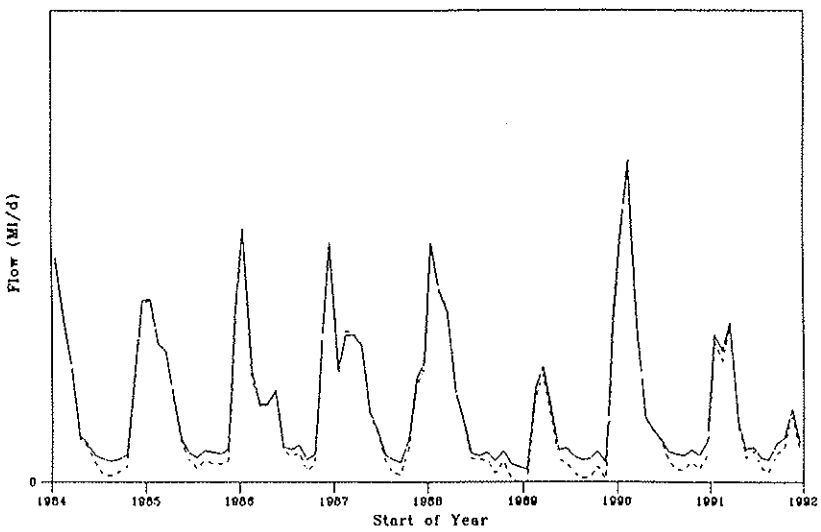
Model Results

These models were developed for use in aquifer management. For their verification, it was necessary to demonstrate that they are capable of reproducing historical responses over a wide range of conditions. The particular consideration of the fitness of these models to predict streamflow responses, is demonstrated in comparative plots of time series flow data and accretion profiles, where the simulated historical behaviour of the aquifer is compared to recorded data.

Long-term streamflow time series plots are predicted for comparison at sites where reasonable records are available. Output for time series observations monitored at Woodsdown Cross on the River Piddle and at South Newton on the River Wylye are shown in Figures 5 and 6 respectively. It was especially important to ensure that reasonable predictions could be made of low flows since these models were, amongst other objectives, to be made available to environmental studies investigating habitats and the in-river needs of aquatic life. The fact that these streamflow comparisons indicate good agreement, indicate that the overall flow balance of each catchment is represented adequately in each model.

The comparative results shown in Figures 7 and 8 for longitudinal flow profiles demonstrate that the models also provide reasonable accuracy in the distribution of leakage exchange flows between the river and aquifer. This demonstrates the value of using accretion profiles in model development. In fact, it was found that by seeking to achieve a suitable correlation between historical and simulated profiles, significant benefits followed in the time series predictions at sites along these watercourses.





KEY

Calibrated Transmissivity Profile
(Primary Transmissivity - 700m²/d
Secondary Transmissivity - 1750m²/d

Transmissivity Profile
(Primary Transmissivity - 250m²/d
Secondary Transmissivity - 2000m²/d

Figure 4 Effect of variable transmissivity on Streamflow response

The ability of these models to predict the ephemeral behaviour typical of Chalk streams was also tested in the development stages. Valuable qualitative data emerged during consultations conducted by the NRA with the local community on the nature of streamflow at Shrewton on the River Till in Wiltshire. Records were maintained over the period 1982 to 1990 for the number of days annually for which the River Till was flowing. Even though no flows were actually measured, these data were extremely valuable in verifying the ability of the model to provide a reasonable simulation of conditions over the non-perennial reaches of the River Till. A comparative plot of the simulated and observed flow periods between 1982 and 1990 is shown in Figure 9.

Conclusions

1 A groundwater model, SLAY, has been applied successfully to the simulation of river-aquifer interaction in modelling studies of two Chalk catchments in Southern England.

2 Certain of the routines used in the two models - eg the leakage flow simulation routine - have been in wide use, but were adapted to meet specific requirements of streamflow prediction in Chalk catchments investigated in these studies.

3 A streamflow accretion routine developed in these studies uses runoff data calculated using the Penman-Grindley method, river-aquifer leakage flow, upstream flow and other flows such as effluent return flows to determine the total streamflow at a given model river node.

4 The streamflow accretion routine also provides a reasonable simulation of non-perennial flow behaviour typical of Chalk aquifers.

5 The use of longitudinal flow profiles along watercourses proved to be extremely valuable in determining the appropriate distribution of leakage coefficients to apply at model river nodes.

6 Experience gained during model development, demonstrates the importance of incorporating the simulation of variable transmissivity with saturated depth, to the prediction of streamflow in Chalk aquifers.

7 The completed groundwater models represent important tools for augmenting the full scope of activities involved in catchment management.

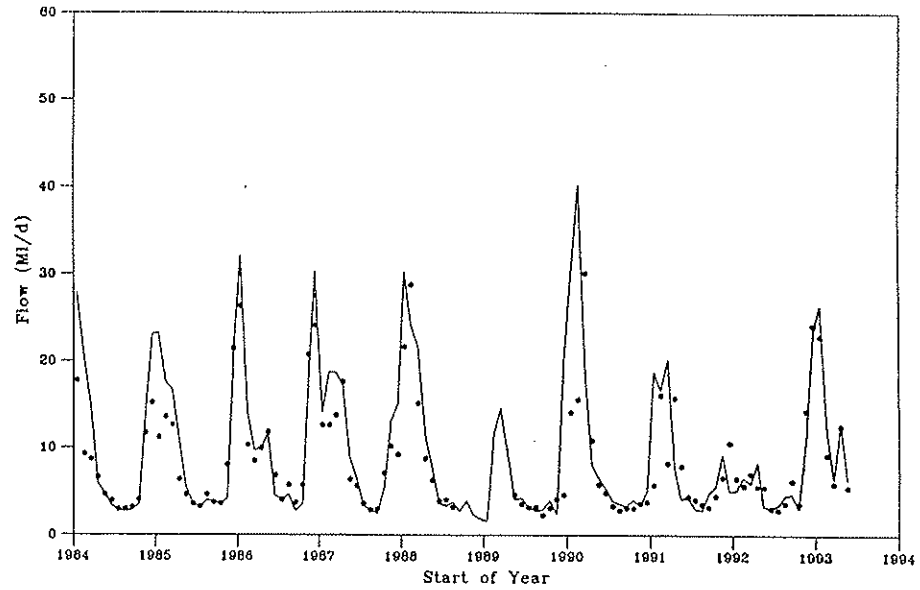


Figure 5 Woodsdown Cross - NGR ST 775002

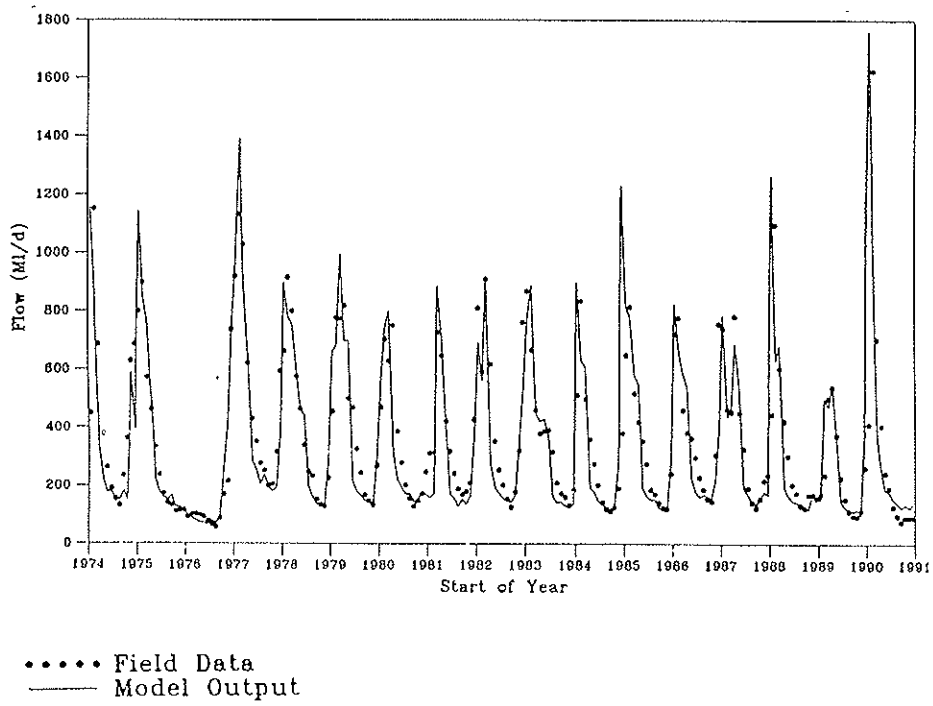
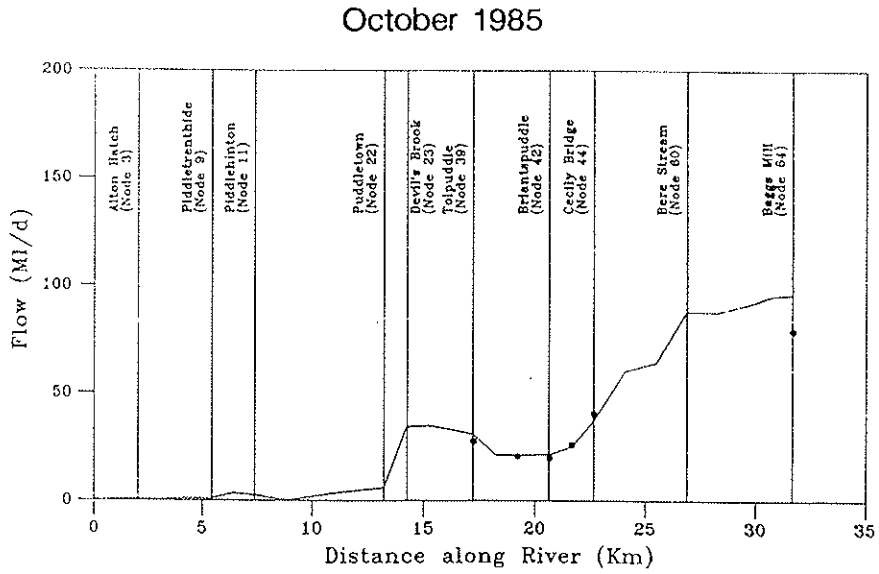
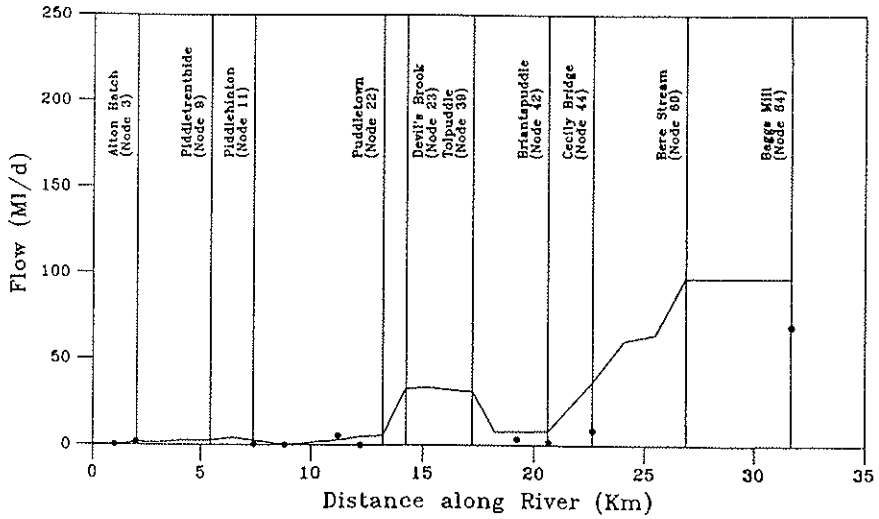


Figure 6 South Newton - NGR SU 086343



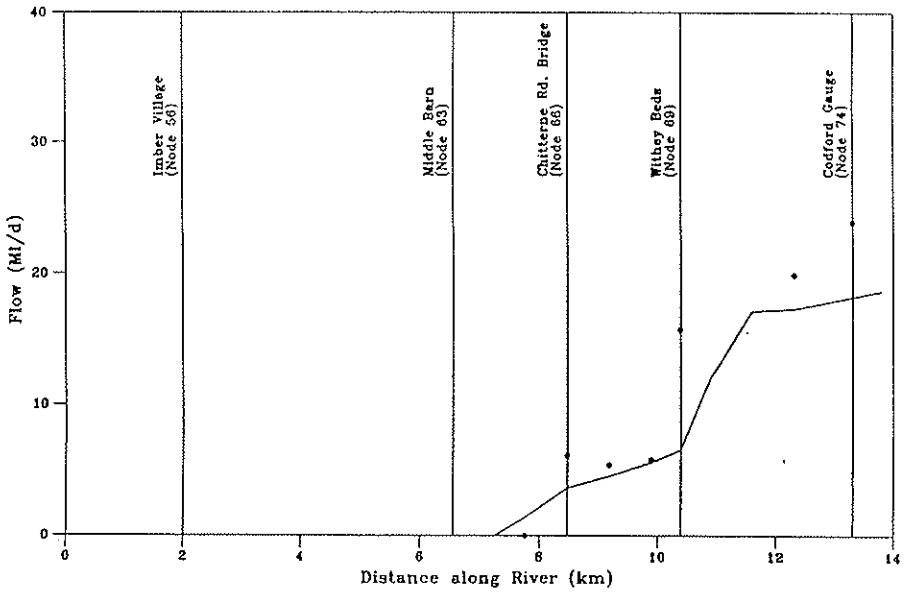
River Piddle
Streamflow Accretion



••••• Field Data
——— Model Output

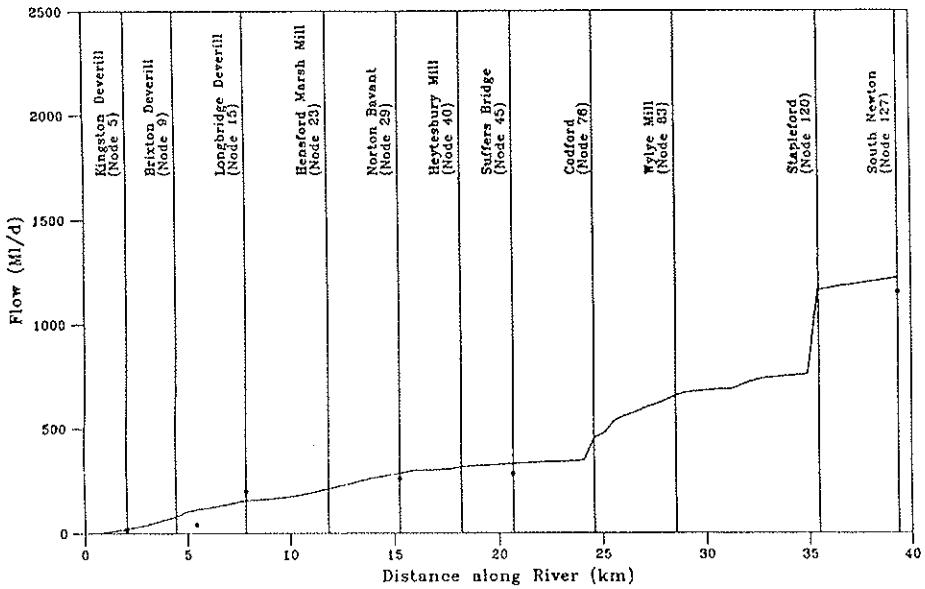
Figure 7 River Piddle
Streamflow Accretion

May 1972



Chitterne Brook
Streamflow Accretion

February 1974



••••• Field Data
——— Model Output

Figure 8 River Wylye
Streamflow Accretion

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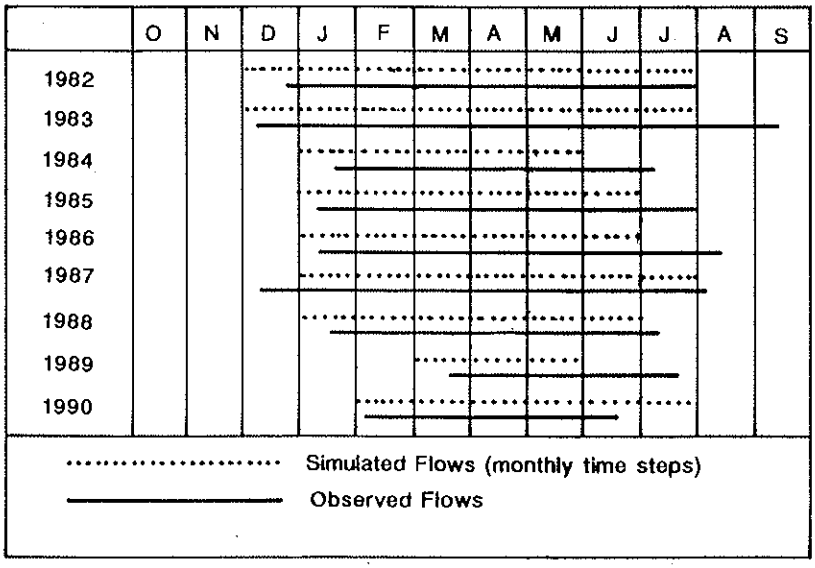


Figure 9 Comparison between observed and predicted ephemeral conditions on the River Till at Shrewton

References

Grindley, J., 1967, The estimate of soil moisture deficits. **Meteorological Magazine**, 96, no. 1137, pp. 97 - 108.

Hall, F.R., 1968, Base-flow Recessions - A Review. **Water Resources Research**, 4, pp 973 - 983.

Headworth, H.G., Keating, T., and Packman, M.J., 1982, Evidence for a Highly-Permeable Zone in the Chalk of Hampshire, UK. **Journal of Hydrology**, 55, pp 93-112.

Herbert, R., 1970, Modelling partially penetrating rivers on aquifer models. **Ground Water**, 8, pp. 29-36.

Miles, J.C., 1987, The representation of flows to partially penetrating rivers from layered and anisotropic aquifers. **Journal of Hydrology**, 95, pp 113-130.

Nutbrown, D.A., and Downing, R.A., 1976, Normal mode analysis of the structure of baseflow-recession curves. **Journal of Hydrology**, 30, pp 327 - 340.

Oakes, D.B., and Pontin, J.M.A., 1976, **Mathematical modelling of a Chalk Aquifer.** Water Research Centre Technical Report TR 24.

Parizek, R.R., Drew, L.J., and Bauer, J.W., 1969, Factors influencing streamflow variability of rivers draining carbonate terranes. **Proceedings of the Geological Society of America** (Meeting of the Hydrogeology Division), Atlantic City, NJ.

Penman, H.L., 1949, The dependence of transpiration on weather and soil conditions. **Journal of Soil Science**, 1, pp 74 - 89.

Prickett, T.A., and Lonnquist, C.G., 1971, **Selected Digital Computer Techniques for Groundwater Resources Evaluation.** Illinois State Water Survey, Urbana, Bulletin 55.

Rushton, K.R., and Tomlinson, L.M., 1979, Possible mechanisms for leakage between aquifers and rivers. **Journal of Hydrology**, 40, pp 49 - 65.

Rushton, K.R., and Ward, C., 1979, The estimation of groundwater recharge. **Journal of Hydrology**, 41, pp 345 - 361.

CONTRIBUTION OF RIVER FLOW SEEPAGE TO BOREHOLE YIELDS, AND ITS IMPLICATION FOR SOURCE PROTECTION ZONE DEFINITION

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Abstract

A modelling study was carried out, using MODFLOW/MODPATH, on the Gatehampton Chalk groundwater source situated adjacent to the River Thames northwest of Reading, Berkshire. The aquifer system was modelled as two layers, an upper one representing the Quaternary floodplain gravels, and a lower one for the Chalk. Interaction between the River Thames and the aquifer was via a low permeability streambed layer. Calibration was undertaken in both the steady and transient states. The paper describes the results of the model with emphasis on the effect of input parameters (particularly Chalk interfluve and streambed hydraulic conductivities) on river flow contribution, and their significance in determining source protection zones. A comparison is made between this and other approaches employing the FLOWPATH single layer model.

Introduction

The Gatehampton source is the largest groundwater source in the UK and is located adjacent to the northeast bank of the River Thames within the Goring Gap, in Berkshire. It comprises seven production boreholes into river gravels and chalk. The source is licensed to abstract a maximum yearly quantity of 25619.35 Ml corresponding to an average rate of 70.2 Ml/d. Near Gatehampton there are three other additional licensed abstractions: Cleeve 16.2 Ml/d; Woods Farm 3.73 Ml/d; and Pangbourne 3.87 Ml/d. This paper is based on a study, commissioned by Thames Water Utilities Ltd, which involved the hydraulic modelling of the Gatehampton source using MODFLOW, and the subsequent definition of protection zones using the particle tracking program, MODPATH.

Previous Studies

Robinson and Banks (1987), in their evaluation of a long duration (122 days) Group Pumping Test of the source concluded that abstraction was largely supported by river

In **Modelling River-Aquifer Interactions. (Edited by P L Younger).**
British Hydrological Society Occasional Paper 6. (1995). ISBN 0 948540 70 2

water. However, comparison of chemical analyses of the pumped and river waters showed no evidence of ingress of river water into the aquifer. Modelling studies by Younger (1990) and Younger et al (1993) indicated that the contribution to the aquifer from the River Thames was only 8% of abstraction and unlikely to be greater than 10%. Aspinwall and Co. (1992), reported a river contribution of no more than 50% of abstraction whilst Hydrogeological Services International (1993), in a similar exercise, indicated a range of between 55% and 75%, depending on the value of the river/aquifer leakance adopted.

Geology

Figure 1 shows the geology of the Gatehampton area. Chalk is exposed in the interfluves on either side of the River Thames floodplain. Within the valley, overlying the eroded chalk surface, are the alluvial deposits and gravels of the River Thames. In the Goring Gap, the river alluvium and gravels reach a thickness of 13m, thinning to 6-7m away from Gatehampton.

Little is known of the nature, distribution and thickness of the streambed sediment of the Thames. Limited sampling by Younger (1990; see also Younger et al, 1993) suggested a lithology of dark grey, organic silt containing aragonitic shell fragments and angular clasts of flint, and a thickness of 0.5m with perhaps some thinning (0.40-0.45m) at sharp river bends.

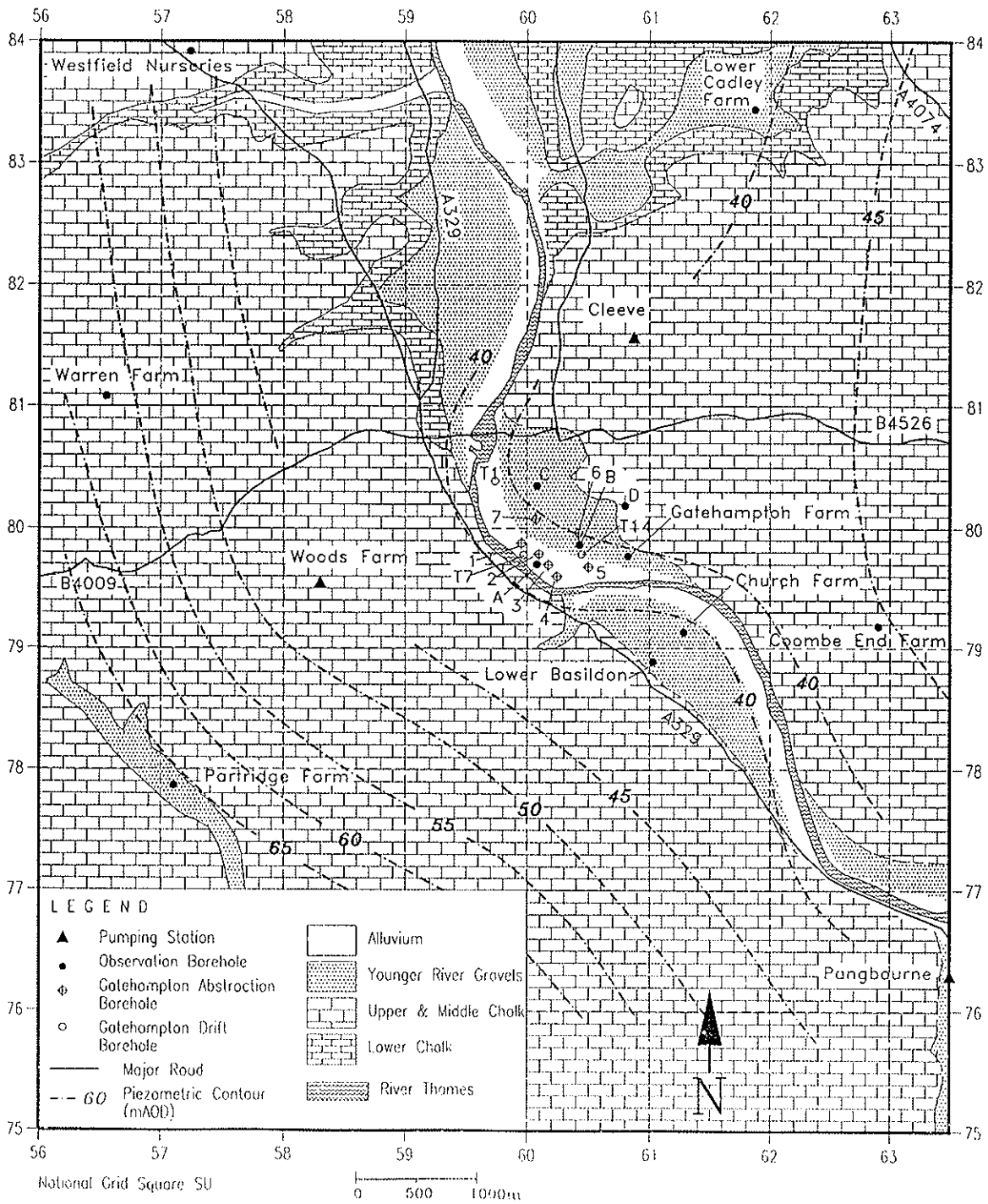
Hydrogeology

The aquifer system comprises river gravels in hydraulic continuity with the underlying Chalk. The gravels are fully saturated at and close to the river and become unsaturated near the edge of the floodplain. Fissuring dominates the hydraulic character and productivity of the chalk. Geophysical and flow logging (Robinson and Banks, 1987) showed that the main flows occur in the Middle Chalk and in the upper 15-20m of the Lower Chalk. Below a depth of 45-50m bgl, the Lower Chalk is stiff and very argillaceous and (for practical purposes), forms a basal aquiclude. Thus, the effective thickness of the chalk aquifer is probably 40-50m and of the combined gravel-chalk aquifer system 45-55m. The piezometric map in Figure 1, shows that, in general, the direction of groundwater flow is from the elevated interfluves into the low lying River Thames valley. Whereas average hydraulic gradients within the interfluve areas are on the order of 10^{-3}, in the floodplain at Gatehampton they are more gentle, approximately 10^{-4}.

In the interfluves, water levels fluctuate in response to seasonal variations of rainfall and exhibit recession characteristics. In the valley, close to the River Thames, well hydrographs are characterised by abrupt rises and falls of water levels in response to river stage.

All reported values of hydraulic parameters have been based on experience from other similar areas (Robinson and Banks, 1987) or from modelling calibrations (Younger, 1990). Transmissivities for the chalk in the interfluves have been generally taken as $50m^2/d$ and for the gravel-chalk system in the valley, more than $3000m^2/d$. Similar values were obtained by Younger (1990) during steady state calibrations. Analysis of time-

Figure 1 - Gatehampton Area: Geology and Piezometry (19-9-1986). (After Robinson and Banks, 1987).



drawdown data from the Group Test indicated a valley chalk transmissivity of 8580m^2/d, corresponding to a hydraulic conductivity of 172m/d. The specific yield of the gravels has been normally taken as 20-25% whereas that of the chalk as 1-5%. Storage coefficients in the confined chalk are generally in the order of 10^{-3} to 10^{-4}. Very little information is available on streambed hydraulic conductivity. Laboratory determinations on a few samples obtained by Younger (1990) indicated values at Gatehampton of 0.002-0.0024m/d. However, Younger (1990) noted that these values should be considered an absolute lower limit, since macropores can be expected to result in a higher hydrauiic conductivity under field conditions. From model calibration, Younger (1990) finally accepted a value of 0.2m/d and a thickness of 0.5m, corresponding to a hydraulic conductivity/thickness ratio of 0.4d^{-1}.

Behaviour of the Aquifer System at Gatehampton during the 1986-1987 Group Test

A detailed description and interpretation of the Group Test has been given by Robinson and Banks (1987). In summary, the test comprised the pumping of the seven boreholes at Gatehampton in four phases, at increasing rates of discharge, as follows:

Phase No	No of Days	Combined Discharge (m^3/d)
1	7	50,500
2	21	70,300
3	80	86,300
4	14	104,900

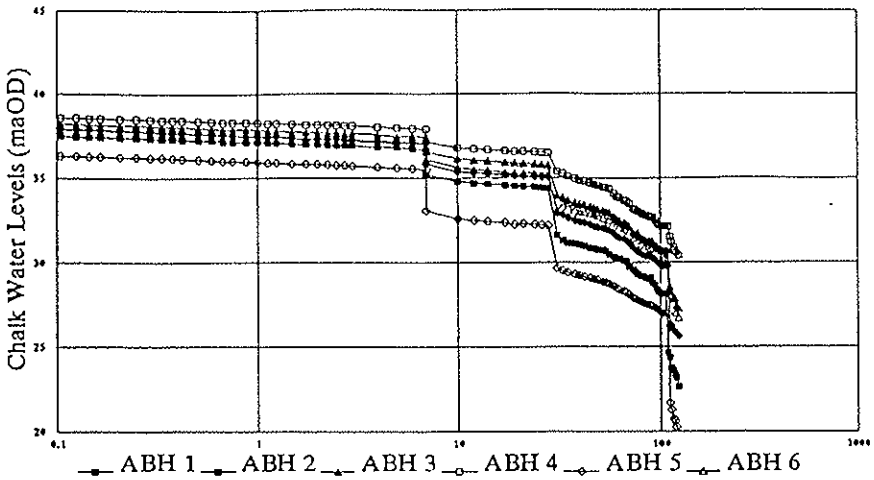
Each of the seven boreholes was pumped at approximately the same rate.

The time-drawdown curves of the abstraction boreholes and a selected number of observation boreholes (in the chalk and gravels) are shown in Figure 2. For the first two phases of pumping, the rate of change of drawdowns is small and a near equilibrium state appears to have been reached. In the third phase of pumping, the persistent fall of water levels indicates that abstraction exceeds inflows whilst the sudden steepening of the time-drawdown curves (the gradient more than doubles on day 58) suggests barrier boundary effects related to low chalk permeabilities in the interfluves, and dewatering of the gravels within the pumped area. In the final phase, a rapid acceleration in the rate of change of drawdown is observed indicating that these pumped rates could only be sustained for limited time periods.

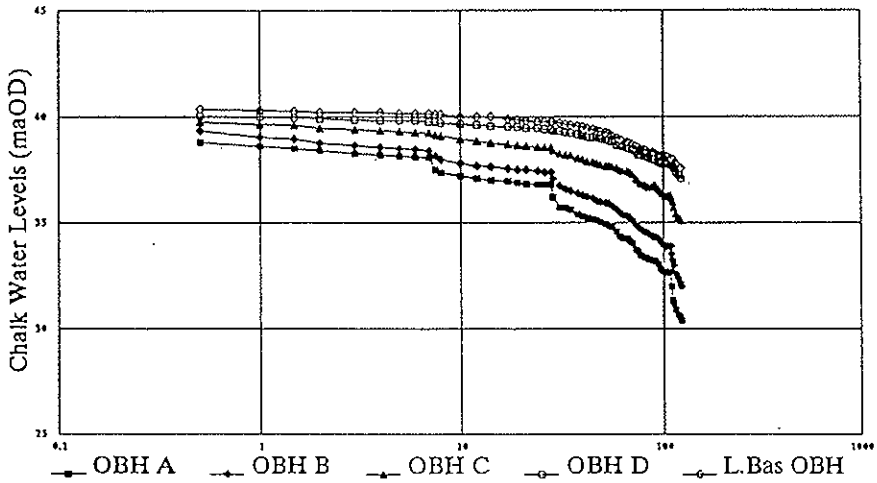
Hydrochemistry of Abstracted Groundwater and River Water

Plots of major ions, ionic ratios and nitrate and fluoride with time of pumping during the Group Test and comparison of these with the River Thames water chemistry did not indicate a significant contribution of river water to abstraction (Robinson and Banks, 1987). The pumped groundwater generally had a higher nitrate concentration, a lower chloride, a lower potassium/sodium ratio, a higher calcium/magnesium ratio and higher

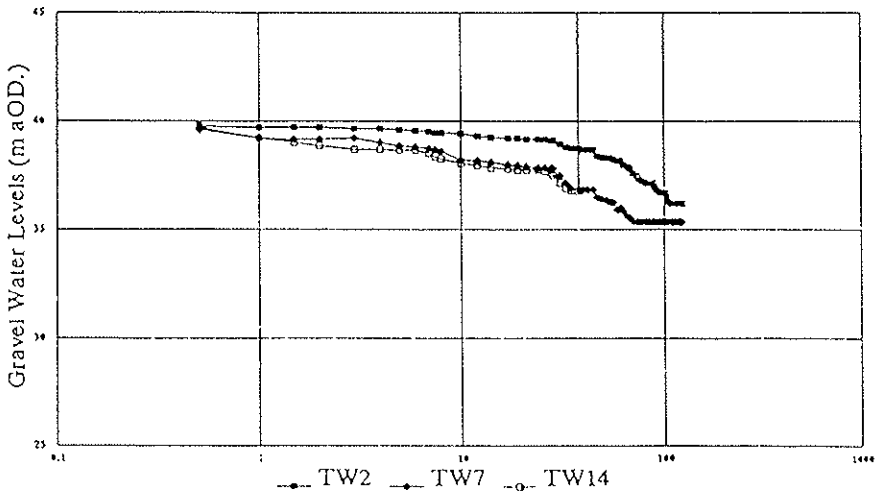
Figure 2 - Borehole Water Levels During 1986-87 Group Test



Abstraction Boreholes



Chalk Observation Boreholes



.Gravel Observation Boreholes

Time (Days) →

fluoride than the river.

Conceptual Model

The aquifer system was represented by two layers, the gravels (Layer 1) which occur within the valley (floodplain) and the chalk (Layer 2) which occurs in both the valley and the interfluve areas. Flow between the river and the aquifer is in the vertical direction through a thin (approximately 0.4m) low permeability layer representing the streambed sediment. The hydraulic connection between the gravels and the chalk is probably imperfect due to the likely existence of stratification or low permeability beds at the junction of the two aquifer units. Resistance to vertical flow between the gravels and chalk was represented by a retarding layer.

Model Description and Code

The numerical model used was MODFLOW (McDonald and Harbaugh, 1988). This is a block-centred variable grid finite difference code that can simulate layered aquifers (quasi-three dimensional). River-aquifer interaction is handled by the RIV Package. This is a specific module within MODFLOW designed to simulate river leakage into and out of an aquifer. Flow out of the river is independent of stream discharge but dependent on stage. The rate of leakage depends on the relative heads within the aquifer and the river, such that, for a given cell, as the head in the aquifer declines, the rate of leakage increases while above the bottom of the streambed layer. Further declines in aquifer head below this level produce no increase in seepage from the stream to the aquifer.

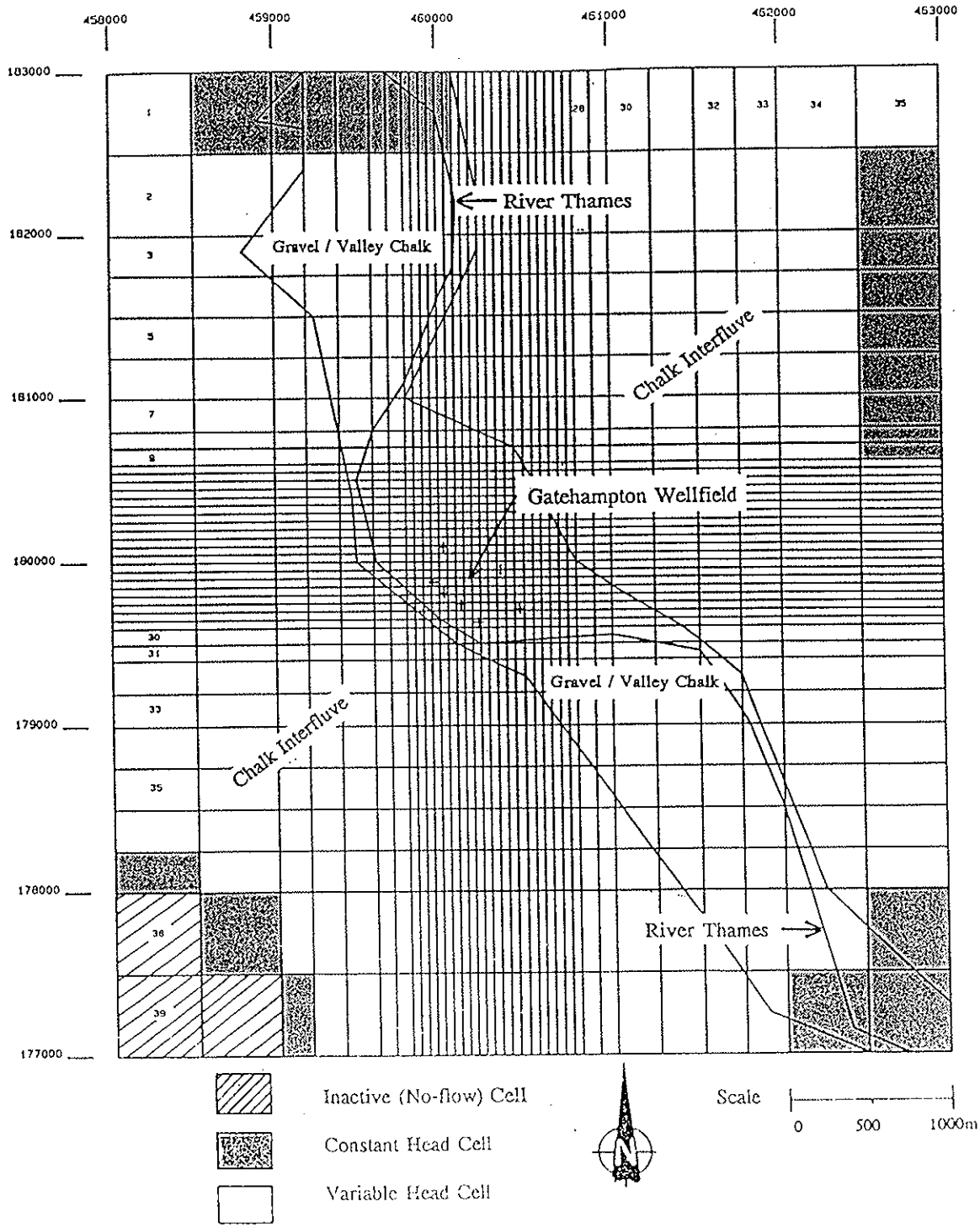
Model Geometry and Input Parameters

Figure 3 shows the model grid, which has a width of 5km (east-west) and a length of 6km (north-south), and comprises 35 columns and 39 rows. Grid spacing is finer in the Gatehampton wellfield area (50m x 50m) to allow for a more precise representation of heads and input parameters but coarser elsewhere (500m x 500m at maximum). Interaction with the River Thames is achieved by 68 river nodes located within the river channel. Constant head nodes were assigned to the model area, as shown, to simulate the observed steady-state piezometry. Elsewhere within the model itself, heads have been allowed to vary. The gravel nodes outside the valley are rendered inactive (no-flow). Some simulations were carried out using constant flux in place of constant head boundaries. However, as the cone of depression was found not to encroach significantly the model boundaries, a constant head approach was adopted. The river-gravel aquifer interaction was modelled as a head dependent boundary. The river stage at each node was fixed.

Model Calibration

Calibration of the model was carried out in two stages. A steady-state calibration aimed at providing a piezometric surface and hydraulic parameters matching observed average conditions. Using the data sets from this calibration as the initial condition, transient calibrations were carried out aimed at simulating the observed hydrographs in selected boreholes during the 1986-1987 Group Test.

Figure 3 - Model Grid



Steady-State Calibration

Calibration of the model in the steady-state was achieved by setting all storativities equal to zero. A number of simulations were run using different hydraulic parameters:

K_g	=	permeability of gravels (Layer 1), m/d
$K_{chk(v)}$	=	permeability of valley chalk (Layer 2), m/d
$T_{chk(i)}$	=	transmissivity of interfluve chalk, m^2/d
k	=	vertical permeability of streambed sediment layer (m/d)
m	=	thickness of streambed sediment layer retained constant in all simulations at 0.40m
$k_{g/chk}$	=	vertical leakage factor between gravels and chalk

In the range of interfluve transmissivities of 50-250m^2/d, the shape of the piezometric surface showed little change. However, at values of 1000m^2/d, groundwater contours showed little agreement with observed piezometry.

Changes in streambed permeability did not significantly modify the shape of the piezometric surface, nor did changes in gravel permeability or changes in the hydraulic conductance between the gravel and chalk. Increases in the chalk interfluve transmissivity resulted in an increase in total inflow, primarily from the interfluves. Changes to the streambed permeability affected both the inflow into the aquifer and outflows into the river; both becoming larger with increasing streambed permeability.

The shape of the piezometric surface was found to be sensitive to river stage. Reasonably acceptable simulations were obtained for a river stage at the top of the model of 43 to 42.5mAOD and at the bottom of 39 to 38.5mAOD (Figure 4). The hydraulic parameters used were K_g = 200m/d; $K_{chk(v)}$ = 100m/d; $T_{chk(i)}$ = 100 m^2/d; k/m = 0.2 d^{-1}; $k_{g/chk}$ = $5x10^{-5}d^{-1}$. The global flow balance for these parameters showed a flow of approximately 17000 m^3/d into the model, mainly from the interfluves. The outflow was mostly into the river as baseflow.

Transient Calibrations

During transient calibrations, storativity values were assigned to each cell of the model. Abstraction was simulated by utilising seven nodes in the Gatehampton area representing the production boreholes. Table 1 summarises the input parameters used.

It was found that the shape of the predicted drawdowns matched those observed, only within a narrow range of combinations of gravel and valley chalk permeabilities. Once acceptable values of permeability were determined, a more exact match of the simulated hydrographs was obtained by changing specific yield values of the gravels and, of both the interfluve and valley chalk (Figure 5, Run
TC16). The resulting piezometric pattern, shown in Figure 6, assumes an elliptical shape with the long axis trending northwest-southeast parallel to the high permeability valley gravels and chalk. The simulated piezometry is similar in general shape but, as expected, not in detail, with that observed.

Table 2 shows the contributions to abstraction derived by the model for selected simulations. At the lowest pumping rate, abstraction is mainly derived from aquifer storage with some contribution from lateral flow. Leakage from the river is negative, meaning that on balance more groundwater discharges into the river from the aquifer rather than from the river into the aquifer. The contribution of river leakage to the aquifer increases with increasing pumping rate and time of pumping, but it reaches a stable condition when the water-table in the gravels is below the streambed. For the accepted calibration (TC16) the river leakage into the aquifer is on average 41%, ranging from initially zero to a maximum of 48%.

Table 1 - Input Parameters During Transient Calibrations

Run No	Storativity				Permeability (K m/d) Transmissivity (T m^2/d)			Leakance (d^{-1})	
	Sy(g)	Sy(v)chk	Sy(i)chk	Schk	Kg	K(v)chk	T(i)chk	(k/m)riv	kg/chk
TC1	0.20	0.05	0.02	0.001	200	100	100	0.2	$5x10^{-4}$
TC2	0.25	0.05	0.02	0.001	200	100	100	0.2	$5x10^{-4}$
TC3	0.20	0.03	0.01	0.001	200	100	100	0.2	$5x10^{-4}$
TC4	0.20	0.05	0.02	0.001	200	100	100	0.2	1.0
TC5	0.20	0.03	0.01	0.001	200	100	100	0.2	$1x10^{-3}$
TC6	0.20	0.03	0.01	0.001	200	100	100	0.2	0.01
TC7	0.20	0.03	0.01	0.001	500	100	100	0.2	0.01
TC8	0.20	0.03	0.01	0.001	500	200	100	0.2	0.01
TC9	0.20	0.03	0.01	0.001	500	150	100	0.2	0.01
TC10	0.20	0.03	0.01	0.001	500	175	100	0.2	0.01
TC11	0.20	0.03	0.01	0.001	500	160	100	0.2	0.01
TC12	0.20	0.03	0.01	0.001	1000	150	100	0.2	0.01
TC13	0.20	0.03	0.01	0.001	750	150	100	0.2	0.01
TC14	0.20	0.03	0.01	0.001	625	150	100	0.2	0.01
TC15	0.20	0.03	0.01	0.001	700	150	100	0.2	0.01
TC16	0.25	0.05	0.02	0.001	700	150	100	0.2	0.01
TC17	0.25	0.05	0.02	0.001	700	150	100	0.2	1.0
TC18	0.25	0.05	0.02	0.001	700	170	100	0.2	1.0
TC19	0.25	0.05	0.02	0.001	800	170	100	0.2	1.0

Note: S refers to storage coefficient; Sy to specific yield

Figure 4 - Calibrated Piezometry (Steady-State Model)

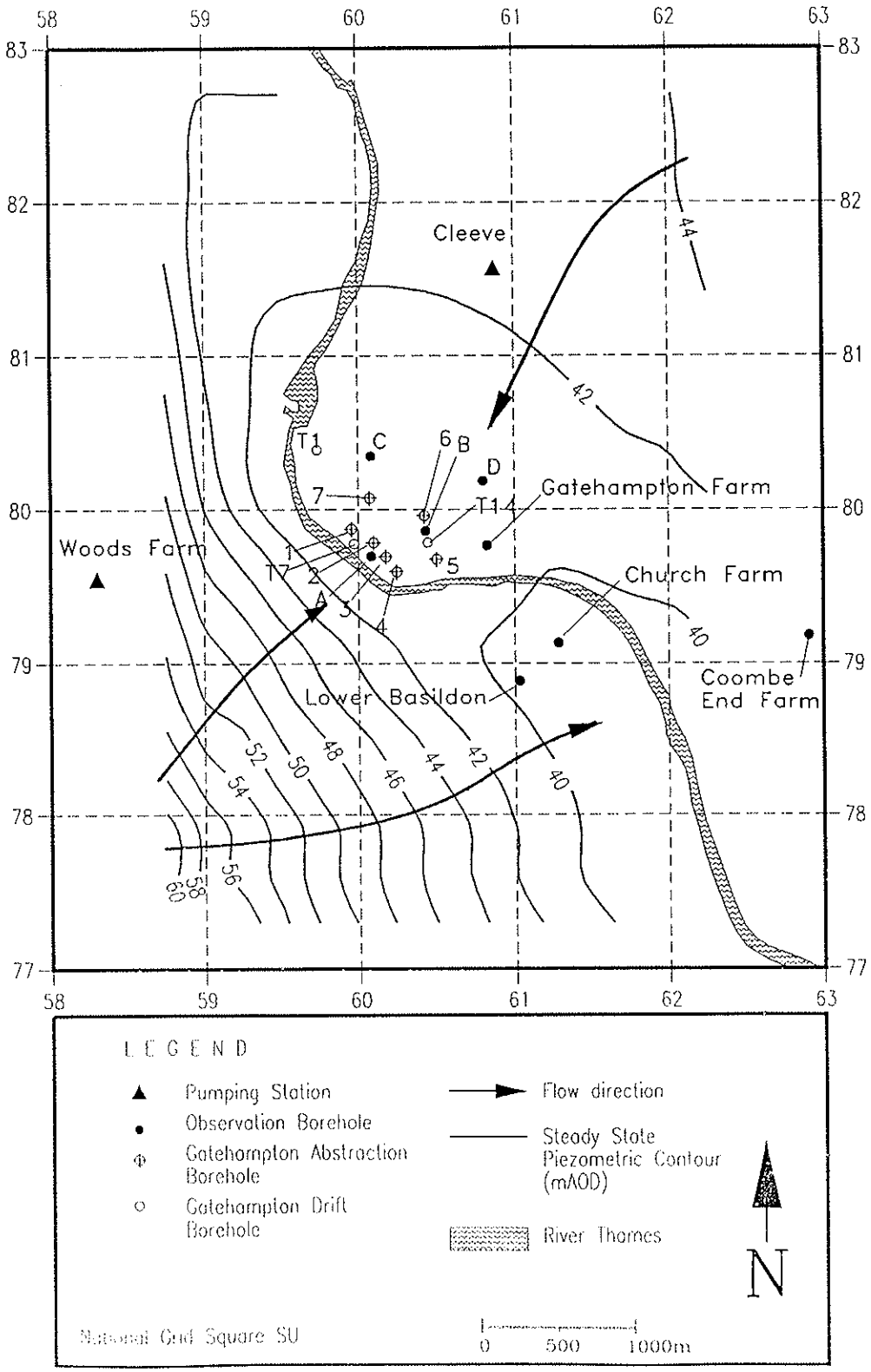


Figure 5 - Observed and Simulated Water Levels

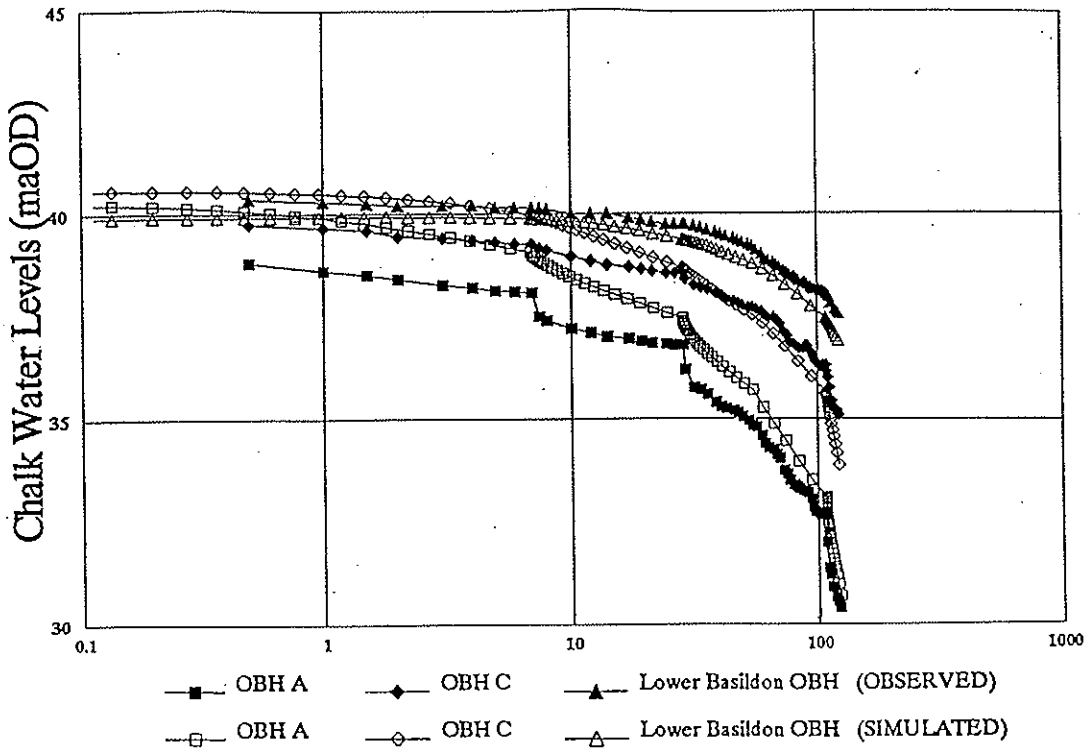


Figure 6 - Calibrated Piezometry, Transient Model

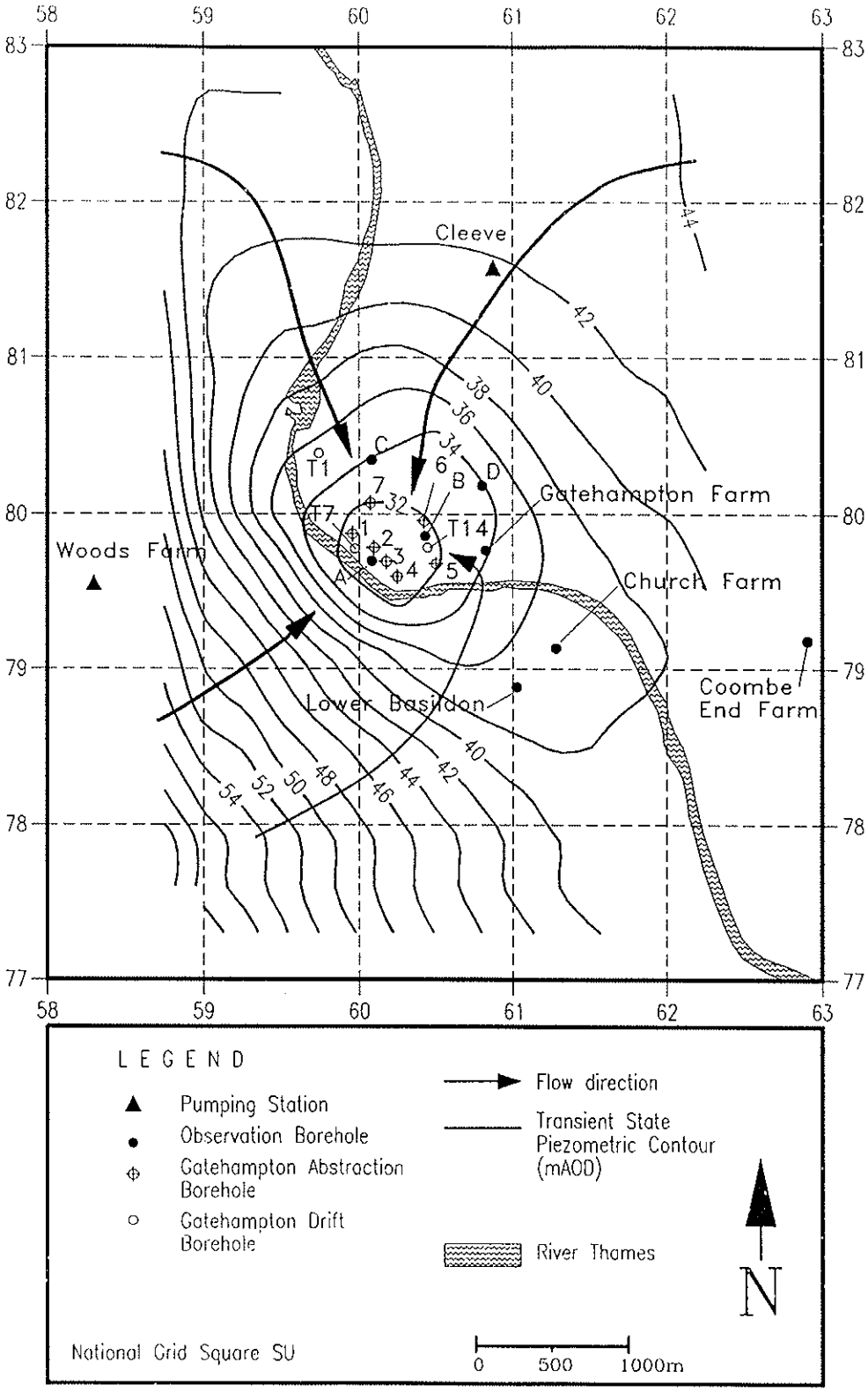


Table 2 - Proportion of Groundwater Contribution (per cent) from Different Sources during Transient Calibrations

Run No	Lateral Flow (%)	River (%)	Storage (%)	Abstraction m^3/d	Period of Abstraction (Days)
TC5	24.2	-9.6	85.4	50510	7
	18.4	21.6	59.8	70280	21
	20.4	55.0	24.6	86280	80
	19.8	57.1	23.1	104800	14
Average	20.1	48.3	31.6	83599	
TC16 (Accepted)	34.0	-17.4	83.4	50510	7
	25.4	22.5	32.0	70280	21
	26.2	47.6	26.2	86280	80
	25.0	44.9	30.1	104800	14
Average	26.20	41.30	32.50	83599	
TC19	36.6	-20.3	83.2	50510	7
	27.6	21.5	50.9	70280	21
	29.8	39.9	30.3	86280	80
	29.4	34.9	35.7	104800	14
Average	29.7	34.4	35.9	83599	

Predictive Pumping Simulations

Predictive pumping simulations were carried out using the hydraulic parameters obtained from the transient calibration TC16; each aimed at defining aquifer behaviour for different abstraction scenarios. At an abstraction rate of 70 Ml/d, contribution from the river was 32% of abstraction after 50 days of pumping; 61% after 400 days; and 65% after 1000 days. At the lower rate of 45 Ml/d, river contribution after 1000 days was 52%.

Source Protection Zones

In 1992, the National Rivers Authority (NRA) published a document on "Policy and Practice for the Protection of Groundwater" which sets out a policy statement on a number of aspects of groundwater protection. In support of this, the NRA has been engaged on a programme for the definition of protection zones using modelling and other techniques around selected groundwater sources in England and Wales.

MODPATH, which is directly compatible with the output of MODFLOW, was used to define source protection zones on the basis of the calibrated model described. The capture zones delineated at the abstraction rate of 70 Ml/d after 1000 days of pumping are shown in Figure 7a and 7b. The 50-day (Zone I) and 400-day (Zone II) zones are both elliptical in shape and aligned along the valley axis where the chalk permeability is highest.

Capture zones were also determined by means of the two-dimensional steady-state FLOWPATH model, using similar hydraulic parameters. For obvious reasons, direct comparisons between the two model outputs are not strictly appropriate.

Fig 7a

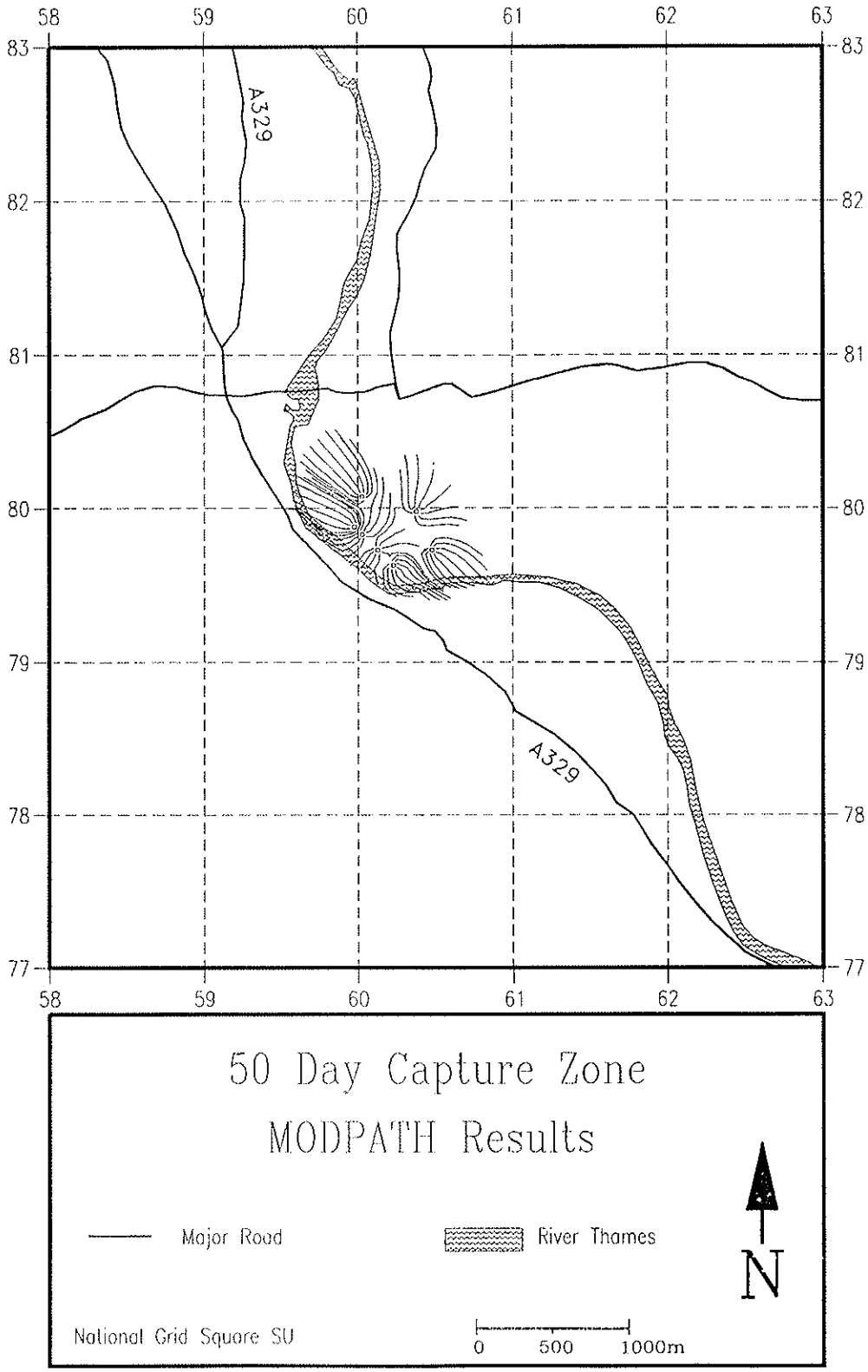

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fig 7b

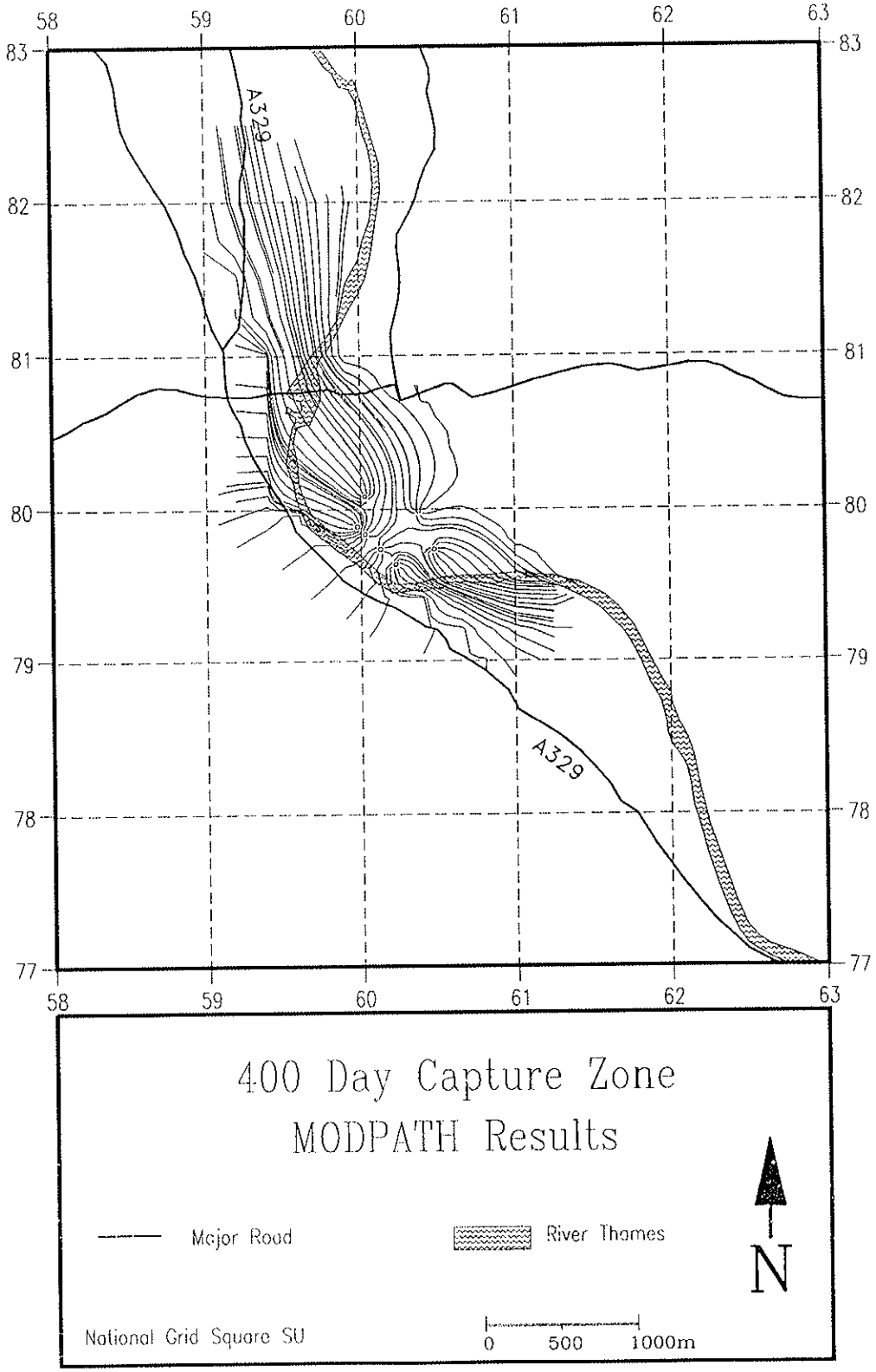

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Fig 8a

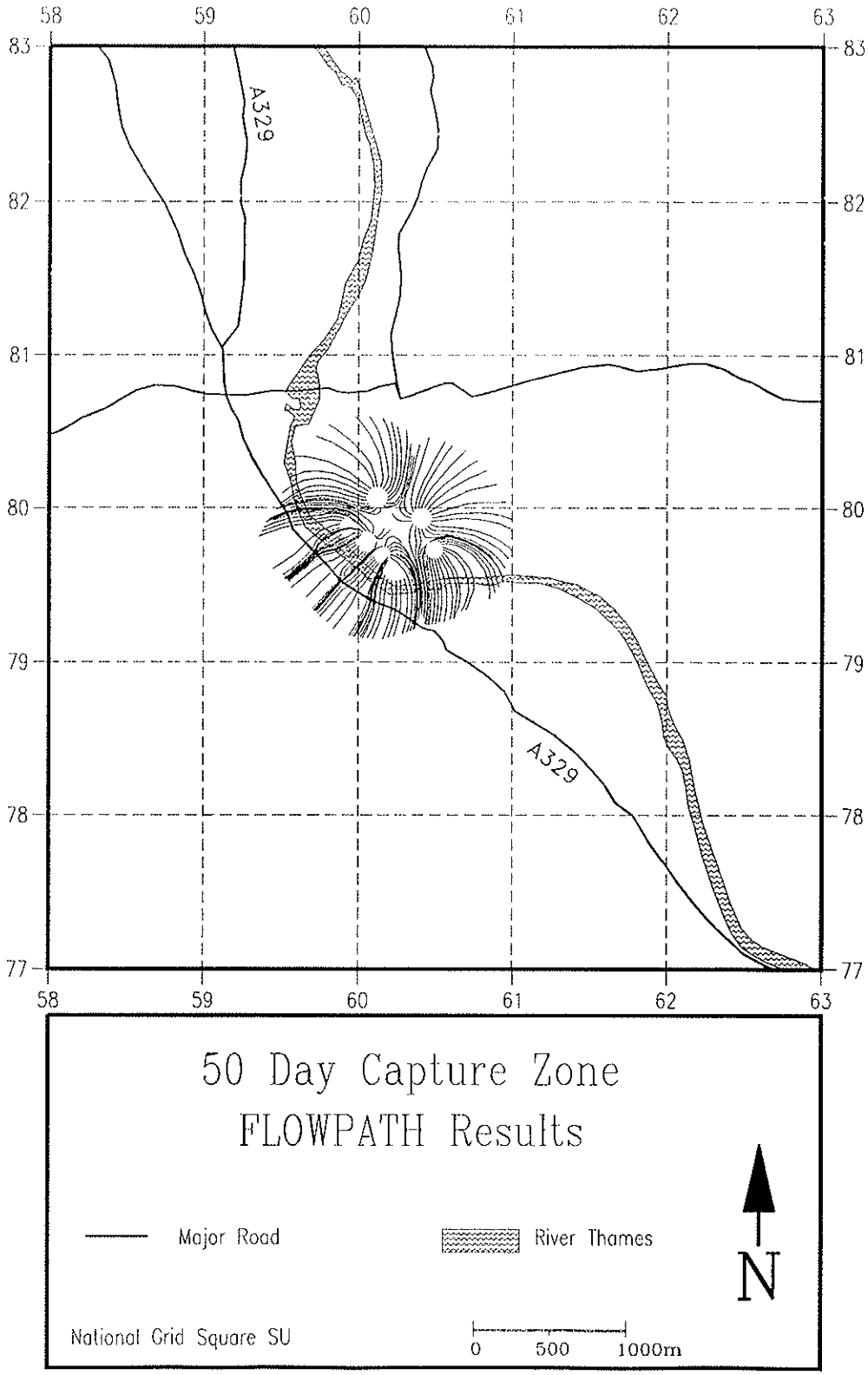
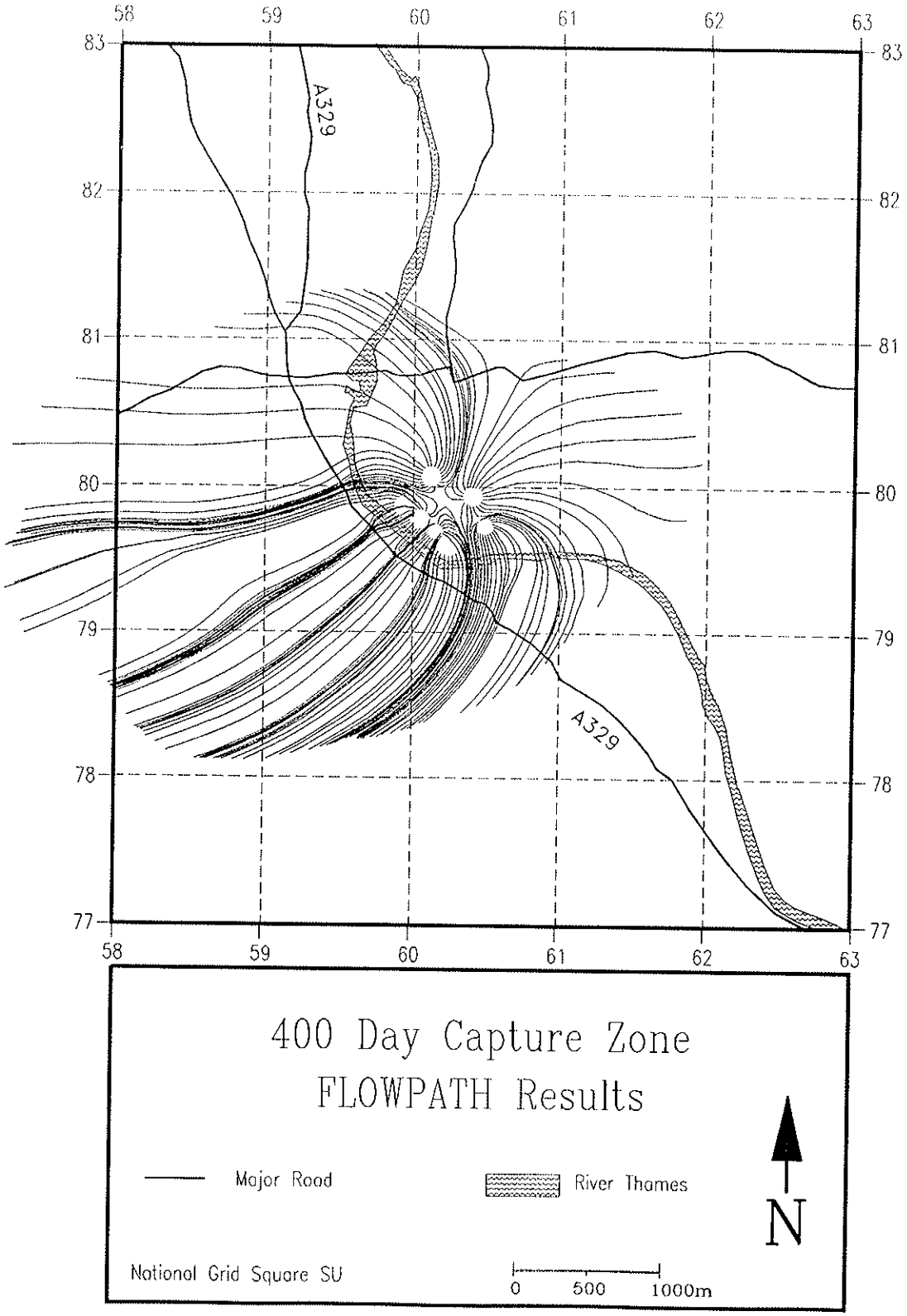

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Fig 8b


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Zones I and II (Figures 8a and 8b) are seen to be near circular, and in contrast to the results of MODPATH, extend more into the interfluves, on either side of the river.

Conclusions

The present work suggests that the contribution to the Gatehampton source from the River Thames can be substantial; at a pumping rate of 70 Ml/d, it can be as high as 65% of abstraction. The hydrochemical evidence during the 1986-87 Group Pumping Test does, however, suggest little interaction with the river. Lower contributions from the river would require the interfluve chalk to be much more transmissive than assumed, but this is contrary to present knowledge. It is felt that further modelling will not provide more reliable answers without there being more information on the hydraulic properties of the interfluves and the hydrochemistry of the river/aquifer system.

In defining protection zones, the study has demonstrated the importance of a correct representation of the hydrogeological regime. Significant error in this regard may considerably affect the shape and extent of capture zones determined. At Gatehampton, use of a two layered model would seem to be the more appropriate approach.

Acknowledgements

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References

Robinson V.K., and Banks D., 1987, **Hydrogeological Investigation of the Gatehampton Groundwater Source, 1984-1986.** Thames Water Authority, Regulation and Monitoring Division, Internal Report 147. 25pp.

Aspinwall & Company Ltd., 1992, **Definition of Protection Zones.** Report Prepared under Assignment to the National Rivers Authority (NRA).

Hydrogeological Services International Ltd, 1993, **Source Protection Zones - Dorney & Gatehampton.** Report to Thames Water Utilities Ltd.

Younger P.L., 1990, **Stream Aquifer Systems of the Thames Basin: Hydrogeology, Geochemistry and Modelling.** PhD Thesis, University of Newcastle Upon Tyne. 388pp.

Younger P.L., Mackay, R., and Connorton B.J., 1993, Streambed sediment as a Barrier to Groundwater Pollution: Insights from Fieldwork and Modelling in the River Thames Basin. **Journal of the Institution of Water and Environmental Management**, 7, 577-585.

McDonald, M.G., and Harbaugh, A.W., 1988, A Modular Three-Dimensional Finite-Difference Groundwater Flow Model. Techniques of Water Resources Investigations of the United States Geological Survey. USGS open file report 83-875, Book 6.

National Rivers Authority, 1992, **Policy and Practice for the Protection of Groundwater.** London, HMSO.