THE HYDROLOGY OF A MAJOR VALLEY WETLAND AT GOSS MOOR, CORNWALL

C. A. L. ISHEMO

Ph.D. 1999
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THE HYDROLOGY OF A MAJOR VALLEY WETLAND
AT GOSS MOOR, CORNWALL

by

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THE HYDROLOGY OF A MAJOR VALLEY WETLAND
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VOLUME ONE
ABSTRACT

THE HYDROLOGY OF A MAJOR VALLEY WETLAND AT GOSS MOOR, CORNWALL

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This thesis aims to furnish an understanding of the water fluxes and storages occurring at the subcatchment scale in Goss Moor, a large lowland wetland in Cornwall, UK. Goss Moor constitutes approximately 5 km² of poor fen and similar wetland areas sited on clayey alluvial and periglacial deposits in the base of a broad/shallow headwater valley. The bedrock is kaolinised granite and pelite.

The hydrological characterisation was achieved using variables measured directly on site, using spectrally derived stream flow components and using flows output from a calibrated numerical model of transient groundwater flow beneath the wetland. The study demonstrated the use of distributed spectral filtering for source area characterisation and of numerical modelling for investigating the role of groundwater flow in the wetland.

Certain stream flows into and out of the wetland were monitored at an hourly resolution. At each site, slowly- and quickly-varying components of flow were discriminated using a digital filter whose response was based upon an observed summer recession. Quick flows thus defined were found to be conserved during translation from the upstream inputs to the outflow, although in-channel dispersion eliminated their flashiness. Conversely, the slow flow component was found to vary more rapidly at the wetland outflow than at the main stream entry, indicating the dominance of a different source of flow upon exit from the wetland. Overall stream flow gained by 50% in traversing the wetland site.

Evapotranspiration (ET) rates in the wetland and in the outer catchment were estimated using the Penman-Monteith formula with measurements near or within the site. The calculations indicated that evapotranspirative losses would be greater from the wetland than from the remainder of its catchment due to the presence of surface water.

U.S.G.S. MODFLOW was used to model the groundwater flow in the alluvium beneath the wetland. Shallow groundwater levels at 20 piezometer sites within the wetland, together with information on stratigraphy, rainfall and ET, provided boundary and calibration data for the model. The results of in situ slug tests were used to define the aquifer permeability for the model in the transient calibration. Storativity and ET were adjusted to produce a match with the observed summer water table decline. A reduction of ET with falling water table greatly improved the match, and it was postulated that the declining water table had therefore dropped below the zone of greatest evapotranspirative uptake.

By combining the various sources of data, the wetland’s water budget was estimated. The numerical modelling showed that groundwater flow to the river accounted for between only 0% and 3% of the total output from the wetland surface and substrata. ET accounted for 20% and surface runoff for 77-80%. Although wetland surface flow was not measured, the water budget showed that a substantial summer reduction in stored water would result if no peripheral inflows were received onto the wetland surface. In the annual water budget, such peripheral inflows were of a magnitude similar to that of the rainfall input to the wetland. Together, these two inputs traversed the wetland surface to provide the increase in slow flow in the river on its exit from the wetland. The implications of the water budget for the management of the wetland are briefly discussed.
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signed........................................

25/3/2000

date........................................

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CHAPTER 1

INTRODUCTION AND AIMS

1.1 BACKGROUND TO THE STUDY

This thesis describes an investigation into the hydrology of a lowland wetland at Goss Moor which is the largest remaining wetland in Cornwall. The wetland itself is approximately 5 km² in extent, set within a catchment of some 22 km² which forms the headwaters of the river Fal (see Figure 1.1). In 1988 it was declared a Site of Special Scientific Interest and is now leased and managed as a Grade 1 National Nature Reserve. The site exhibits a diverse range of rare plant species in a mosaic of wetland, heathland, peatland and open water (see Plate 1). However, in recent years changes in ecology have been noted and the highly valued open mire and heath communities have been lost to less desirable wet willow wetland.

With these issues in mind, the study aimed to characterise the present-day hydrology of the Goss Moor wetland, determining the balance between water inflows, outflows and storage in relation to the surrounding catchment in order to provide initial indications of where to concentrate watershed improvement schemes. The investigation was conducted at the catchment scale and at a range of temporal resolutions, and monitored rainfall, stream flows, groundwater levels and evapotranspiration. A physically based numerical model of the groundwater domain was calibrated and the relative magnitudes of surface water and groundwater flows were assessed. The propensity of the wetland hydrological system to store water was assessed in relation to that of the surrounding catchment by an examination of the stream flows upstream and downstream of the wetland. A quantitative characterisation of the wetland’s hydrology would help to focus wetland conservation work on the most efficient ways of changing the wetland’s hydrology, thus ensuring a more
effective hydrological amelioration of the wetland degradation. Information from the study could be used to assess the sensitivity of the system to future changes in land management and climate.

The study’s research methods involved a determination of the nature of the wetland’s hydrological system through complementary use of stream flow recession analysis, numerical groundwater modelling and water budget analysis. The characterisation of the geology, topography and drainage network of the catchment was also essential. In terms of scientific research, a primary objective of the study was to demonstrate the effectiveness of these analysis methods when used in combination.

Significantly, the investigation illustrates the innovative value of deterministic modelling in the study of wetland hydrology. Researchers such as Gilvear et al. (1993), Papatolios (1994) and Bradley (1996) have recently begun to explore the possibilities of applying groundwater models to wetlands of various types. The present study extends this exploration further with respect to the estimation of wetland water budgets and the critical evaluation of a groundwater model’s applicability to a wetland system.

1.1.1 Wetlands, Their Function and Their Value

Wetlands were originally extensive in lowland Britain but have been lost as a result of agricultural drainage. In the past wetlands were regarded as waste areas to be exploited (Williams, 1970). Latterly, however, their intrinsic worth as refugia for rare flora and fauna (Wheeler, 1993) and as buffer areas for regulating surface and groundwater flow as well as controlling water quality has been recognised (Heathwaite, 1995). The cultural heritage of wetlands by virtue of the presence of archaeological sites of prehistoric and historic interest is also important. As the intrinsic value and worth of the wetlands is appreciated, managers have recognised the need to base their decisions on sound scientific principles: for example English Nature and the Royal Society for the Protection of Birds have found that for ecological reasons wetland hydrology must be understood. It is also important that
individual wetlands are viewed within their geographical context rather than in isolation from the wider hydrological processes occurring within the basin as a whole.

Wetlands have a number of functions which are linked to their hydrological characteristics (Heathwaite, 1995; O'Brien, 1988). Depending on their position in the catchment, lowland wetlands can have an important influence on runoff by storing flood water. Significance of any flood mitigation depends on the size of wetland, its location relative to the drainage network, as well as the degree of artificial drainage and channelisation which has occurred. Their role in ameliorating water quality is well known: wetlands act as sinks for inorganic and organic elements passing through them. Burt and Haycock (1992) have shown that they have an important role in taking up nitrate and phosphate. Nitrogen, for example, may be taken up by plants, denitrification may occur in waterlogged conditions while phosphate may be held on sediment surfaces.

The research discussed in this thesis considers the example of Goss Moor, which lies in the floodplain of the upper reaches of the River Fal. Set in a broad, shallow depression, the moor demonstrates a number of hydrological features found at similar sites throughout lowland Britain. A number of streams, rising on surrounding land, converge with the nascent river Fal within the confines of the wetland. The floodplain itself consists of clays produced by physical and chemical weathering, overlain by various fine and coarse sediments that were deposited in Quaternary times as part of a braided stream environment. Wetland masses developed in areas of impeded drainage, now leaving traces of peat amid the braided stream sediments. Human activity has resulted in considerable modification of the hydrology of the site. In the recent past the area was mined for alluvial tin, gravel was mined for construction and the river channel was straightened to improve drainage.

Originally, large areas of floodplain in lowland Britain were covered by wetland, but considerable areas have been damaged. Gosselink and Maltby (1990) report that 84 percent of lowland raised bog in Britain was lost between the mid-nineteenth century and 1978. This was caused by afforestation, agricultural reclamation and commercial peat cutting. Drainage and burning have since degraded or damaged most of the remaining lowland bogs:
about 6 percent of them (780 ha) are ecologically intact. The middle decades of the twentieth century saw drastic proportional losses of peat mosses in north-west England (95 percent lost between 1948 and 1975) and of coastal pastures in the Thames estuary (64 percent lost between 1930 and 1980), as reported by Buisson and Bradley (1994). Similar rates of destruction applied to wetlands in other locations. Wetlands remaining in England, made all the more significant in terms of ecological conservation as a result of such loss and degradation, include the Washlands of the Fens in Lincolnshire and Cambridgeshire (Fojt, 1994), parts of the Somerset Levels including the floodplain of the River Parrett (Williams, 1990), areas adjacent to the Severn estuary (Rippon, 1997), Halvergate Marshes in Norfolk and the Isle of Sheppey in Kent.

Water is fundamental in the formation, exploitation and conservation of wetlands (Heathwaite, 1995). Hollis (1994, p.184) wrote simply but emphatically that, “It is hydrology that puts the ‘wet’ in wetlands!”. Similarly, Mitsch and Gosselink (1993, p.68) state that “hydrology is probably the single most important determinant of the establishment and maintenance of specific types of wetlands and wetland processes”.

Wetland hydrology affects species composition and ecological diversity. As explained by Mitsch and Gosselink (1993), many plant species are limited in their tolerance of anoxic soil environments. The resilience of the plant will depend upon season and upon its stage of development, with the result that waterlogging plays a key role in the dynamics of plant establishment, competition and succession. For example, Bornette et al. (1994) show that differences in the level and frequency of periodic flooding by the river Rhône among wetlands established in former channels can account for differences in their species composition. Although the flow regime and inflowing water quality affect the primary productivity, organic accumulation and nutrient cycling of wetlands, the direction of such influence is not consistent between sites. However, the effects are usually noticeable whichever direction they take. Mitsch and Gosselink (1993) relate the outcomes of several studies in the above areas of wetland characterisation. Flow-through conditions usually encourage greater productivity than stagnant conditions, as shown by Steever et al. (1976) in a comparison of saltwater tidal wetlands with varying frequencies of inundation.
Decomposition and organic export rates affect the accumulation of organic matter. Waterlogging may either augment or suppress decomposition (Odum and Heywood, 1978; Chamie and Richardson, 1978). However, the flow-through of water almost always encourages organic export from wetlands with the result that, for example, mangrove swamps export a greater proportion of their net production than do fens. All these processes have their effects on such nutrient cycling as takes place in the wetland (Mitsch and Gosselink, 1993). As a result wetlands are sensitive to small changes in water supply or drainage and therefore any prolonged change in, for instance, water table elevation can disrupt their ecological balance.

Persistence of a wetland depends on the maintenance of the hydrological balance between inputs, outputs and storage, and human activity can dramatically affect the magnitude of each component. The wetland relies on the receipt of sufficient water inputs from precipitation, stream flows and groundwater which are balanced by losses or outflows to streams and groundwater and evapotranspiration demand. Factors such as topography, geometry of the flood plain, soil type and stratigraphy moderate the ability of the wetland to retain and release water.

1.2 THE INFLUENCE OF HYDROLOGY ON WETLAND SURVIVAL

1.2.1 Comparison of Wetland Hydrologies

Tarnocai (1980, p.10) has defined wetlands as

"land that has the water table at, near or above the land surface, or which is saturated for a long enough period to promote wetland or aquatic processes, as indicated by hydric soils, hydrophytic vegetation, and various kinds of biological activity that are adapted to the wet environment."

Wetlands owe their existence to the interaction of climatic and stratigraphic controls, which allow a copious supply of water to a wetland site, with low topographic gradients to promote the storage of water above or near ground level, where it may influence the
vegetation. They are dynamic systems, reacting to changes in these conditions and their associated nutrient levels. The vegetation cover varies according to the temporal regime of soil saturation and nutrient supply, while also progressively modifying these through its own growth and decay. Reduction of the water supply, or the introduction of competing demands for the available water, often leads to eventual death of the wetland. Site hydrology is thus of paramount significance when considering the vulnerability of wetlands.

In the UK, wetlands are often classified in terms of their position (floodplain mires, soligenous mires and other types) or mineral budget (minerotrophic or ombrotrophic mires), as stated by Hughes and Heathwaite (1995). Gilvear et al. (1993) describe the formation of a minerotrophic fen in East Anglia, Badley Moor Fen, through the leakage of groundwater from a regionally confined aquifer. The fen development was possible only through the occurrence of a breach in the confining layer of boulder clay. A similar situation is reported by Wilcox et al. (1986) and Shedlock et al. (1993) for a mounded fen (Cowles Bog) near Lake Michigan, U.S.A. Gilman (1994) considers the hydrology of a small species-rich fen, Cors Erddreiog, in Anglesey, Wales. This wetland is fed by carbonate-rich springs emerging from limestone which outcrops in and around the site, giving a rather basic environment, although the build-up of peat has produced extensive areas with elevations farther above the water table and hence with more acidic conditions.

Examples of ombrotrophic (precipitation-fed) mires are provided by Damman (1986) and Price (1992). Included among such wetlands are blanket bogs, convex raised bogs and plateau bogs. Although groundwater or surface water flows are involved in the initial stages of bog growth, the full development of a bog involves a reversal of the flow regime such that atmospheric precipitation becomes the only input. Consequently, they rely upon a cool, wet, oceanic climate for their complete development, and are restricted to geographic locations such as the northern and western parts of the British Isles and mainland Europe, and the north-western and north-eastern coasts of North America.

The hydrologic functioning of different wetlands also has implications for their vulnerability. The long term viability of wetlands is influenced by the relative magnitudes of groundwater,
surface water and meteoric fluxes through the wetland (Lloyd et al., 1993; Gilvear and McInnes, 1994). Wetland sustainability depends on whether a major component of this water balance has been affected. For instance, wetlands fed by a regional aquifer may be damaged by groundwater abstractions (Bernaldez et al., 1993; Harding, 1993; Suso and Llamas, 1993), while hill slope afforestation may often reduce total runoff (Robinson et al., 1991) and thus may endanger downslope rheotrophic wetlands (Pressey, 1986).

The hydrologic stress on the wetland may be first indicated by subtle changes in vegetation, or may simply take the form of areal shrinkage. In either case, the process of wetland derogation may then accelerate, not only through perpetuation of the original stress, but also through an increased vulnerability to it and to other environmental stresses such as vegetational succession and climatic fluctuation. In this way, hydrologic perturbation may initiate a destructive positive feedback mechanism. An example of such a feedback process is the initial stage in the flow-through succession of mires, presently a widely accepted model for the formation of certain raised bogs (Mitsch and Gosselink, 1993). In some cases, the development of raised bogs from shallow lakes has begun with the deposition of upstream-sourced sediment at the bottom of the lake. Due to this build-up of sediment, marsh vegetation develops and the deposition of peat accelerates due to the retardation of flows and the introduction of denser vegetation onto the peat surface. The process of lake sedimentation thus may initiate positive feedback with bog growth. In a similar way, the construction of a dam may reduce the depth and frequency of floods on a downstream alluvial plain, causing wetland to develop into carr. Processes involved in this may include channel degradation and incision through reduced sediment supply, with a subsequent lowering of the water table, and also the establishment of more evapotranspirative vegetation which lowers the water table even further, potentially reducing waterlogging by rainfall (Ward and Stanford, 1995). Papers by Bakker (1994), Bakker et al. (1994) and Johnson (1997) are further examples of current thinking on the mutual modification of hydrology and vegetation. Such processes, together with the direct effects of the flow regime on the ecosystem (see Section 1.1.1) increase the need for sound empirical knowledge of the hydrology at wetland sites such as Goss Moor in order to help conserve them.
1.2.2 Water Retention in Wetland Survival

Wetlands are, by definition, zones of water retention. Water storage on the wetland surface performs a significant role, both in the wetland's own perpetuation and in the modification of catchment hydrologic response. Depending on their position in the catchment, lowland wetlands can have an important influence on runoff by storing flood water (O'Brien, 1988; Price and Maloney, 1994; Hey et al., 1994), while the retention of a surplus of water protects the wetland surface from desiccation by evapotranspiration. The hydrological vulnerability of wetlands varies according to the local climatic balance and other factors such as size of catchment, efficiency of drainage, and storage in source areas or the wetland itself. Certain wetlands may be highly vulnerable due to the marginal suitability of the land in which they have been established. Groundwater storage often plays an important role in sustaining all types of mires alike.

Wetlands may be groundwater recharge or discharge zones, depending on their position in the landscape and the time of year (Winter, 1988). Siegel and Glaser (1987) investigated the recharge and discharge characteristics in the Glacial Lake Agassiz peatlands, U.S.A., using both potentiometric and hydrochemical methods. Interrelating the hydrology of a raised bog with that of a neighbouring spring fen, they found groundwater discharging at both sites during the winter, but with the raised bog becoming a recharge site in summer. The greater influence of groundwater on the fen raised the pH of its surface waters above 7, which compared with values of less than 4.2 on the bog. Siegel (1988a, 1988b) conducted a similar study of wetlands near Juneau, Alaska, finding simultaneous recharge and discharge within the same patterned fen site, and relatively low mineral content in the associated surface waters. Siegel (1988a, p.432) found that,

"(1) the recharge-discharge function depends more on the hydrogeologic setting than the wetland type classified by vegetation, and (2) the volumes of recharge and discharge are small compared to volumes of ground water in storage and surface runoff in streams."

The above studies by Siegel and by Siegel and Glaser address the question of the influences of both hydrochemical and hydrological regimes on wetland type. It is clear that the
wetlands studied in the above are all different in hydrochemical and vegetational respects. However, while the nature of the wetland flora may be determined geochemically to a greater or a lesser extent, the perpetuation of the wetland's existence is determined hydrologically. In particular, the study by Siegel and Glaser (1987) showed that the maintenance of high water tables can be important in both ombrotrophic and minerotrophic wetlands. The water table, even in a raised bog, can remain high enough that groundwater discharge occurs during the winter. The proximity of the water table to the surface of such an ombrotrophic bog ensures that surface or near-surface water storage is maintained, as, for example, in the small surface water column created during periods of saturated excess overland flow. Thus, given that it is lying on a permeable substrate, the active layer and surface flora of the bog are, even though ombrotrophic, reliant on a high water table.

For minerotrophic wetlands, the maintenance of high groundwater potentials carries further significance in that the communicating aquifer is the required source of water. An aquifer which drains slowly, having a high propensity to store rather than release water, is more likely to provide this continuity. The above conclusion by Siegel (1988a) illustrates the tendency for groundwater bodies discharging to wetlands to have a long water retention time. As pointed out by O'Brien (1988), the size and character of the aquifer, whether superficial or regional, thus influence the wetland's chances of survival.

In the present study, the water table in the Goss Moor wetland is monitored at various locations so as to discover the way in which aquifer water levels vary over the hydrological year. In conjunction with in situ measurements of the wetland aquifer's storage and transmission properties, and with numerical groundwater flow modelling, this allows a rigorous characterisation of the groundwater flow regime and its interaction with the wetland.

Whichever type of wetland is concerned, the dynamics of the groundwater regime beneath the wetland itself and in other associated aquifers may play a major role in its stability. Also important in the hydrology of the wetland is surface water storage on the mire itself. Although wetlands often exhibit flashy hydrological responses (Burt, 1992, 1995; Glenn and
many studies have shown that the detention or depression storage of incoming water can occur in preference to wetland runoff during the summer, when water levels on the wetland surface have been reduced to their annual lowest by evapotranspiration. For instance, Quinton and Roulet (1998) demonstrated the hydrological operation of a wetland in Quebec, Canada. This was a patterned wetland, in which low, narrow ridges of peat separated the intervening pools into clearly defined polygonal patterns. They found two distinct regimes of runoff, corresponding to the time-dependent continuity and discontinuity of the wetland flow path between the catchment outlet and its upland source areas. During phases of discontinuity, the fluctuations in wetland surface water storage exceeded direct precipitation, indicating that the wetland detained lateral inputs from the outer catchment.

Due to a large complement of undrained pools which would release surface runoff only when full, the above patterned wetland would perhaps be an extreme example of wetland surface water storage. However, other examples exhibit similar effects without evidence of such low connectivity and also bring to mind the relevance of examining the character of the stream response at the wetland outlet. Rovansek et al. (1996) demonstrated the storage of runoff in an Alaskan tundra wetland complex. Skaggs et al. (1980, 1991) investigated the drainage of pocosins in North Carolina. Pocosins are “evergreen shrub bogs found on the Atlantic Coastal Plain from Virginia to northern Florida” (Mitsch and Gosselink, 1993, p.55). They occupy broad flat interfluves and are associated with a low drainage network density, although they also fulfil water quality stabilisation functions in estuarine areas (Wilen and Tiner, 1993). The soil is usually deep, waterlogged, nutrient poor peat, although periodic burning also takes place. Skaggs et al. (1980, 1991) showed that the introduction of an artificial drainage network into a North American pocosin produced flashier runoff by shortening the residence time on the wetland surface.

In particular, thick wetland vegetation might be expected to retard surface water flow, as shown by Price and Woo (1988). This latter effect is relevant to Goss Moor, which does not have the distinct pools and ridges of a patterned wetland and yet is covered extensively with the dense hummocky terrain typical of purple moor grass (Molinia caerulea). In Chapter 4 of the present study, the character of the river flow at the outlet of Goss Moor is
analysed and compared with incoming stream flow behaviour, highlighting the combined
effects of wetland runoff contributions and in-stream detention. Through this stream flow
analysis and through the determination of a wetland water budget for Goss Moor, an
indication is gained of the character of runoff from the wetland itself and hence of the
degree of storage occurring within its boundaries.

1.3 WATER BUDGETS AND NUMERICAL MODELS IN HYDROLOGICAL
SCIENCE

The above discussion of water sources, movement and storage in wetland hydrology led to
suggestions of techniques to estimate the storage and exchanges of water in the Goss Moor
wetland and its substrate and thus to establish the major hydrological influences on the
wetland. Many catchment-scale scientific investigations require a similar focus on the
relative involvement of various flow processes in producing certain environments or events.
This objective can usually be achieved by determining the volumes of flow or storage for
which the processes in question are responsible. In the final research declaration, the
respective relative volumetric magnitudes must be stated over an area and a period which is
great enough to make the result generally valid. Given the spatially and temporally non­
uniform way in which many hydrological processes operate within a catchment, the spatio-
temporal distribution of estimation points for such volumes is an important consideration
when trying to obtain representative sampling over the complete region of interest. This has
led to a variety of approaches to the determination of water budgets, often differing in the
degree of involvement of numerical modelling. Physically-based numerical models have
become a well used tool in the science of hydrology (Anderson and Woessner, 1992), and
can greatly facilitate the task of flow estimation.

1.3.1 Preliminary Water Budget Calculations

A time-varying, spatially averaged water budget gives some information on the character of
a catchment. Many papers have been written drawing conclusions on site hydrology from
the analysis of monthly or seasonal water balances. For example, Owen (1995) quantified the water budget of a 92 ha streamside urban peatland over two years. As an initial examination of the site, this study succeeded in defining the main flows and explaining them in terms of site stratigraphy, topography and climate. Groundwater and overland flow patterns were interpolated from sparsely distributed water level measurements made each week. The uncertainty in spatial patterns was undoubtedly high from these data, due to the sparse distribution of the measurement sites. However, the measurements were sufficient to corroborate impressions of general flow contributions, and the weekly sampling period allowed analysis of storage fluctuations in response to rainfall and stream flow events. Nevertheless, as the author states, errors in the water budget were probably too large to permit any more than qualitative conclusions.

Woo and Rowsell (1993) conducted a three year examination of the hydrology of a 3.1 ha prairie slough in Saskatchewan, Canada. Distributed measurements of upland snow ablation, together with rainfall measurements and distributed infiltration and evaporation calculations, formed the basis of spatially averaged calculations of water availability via overland flow to the slough. Groundwater recharge from the slough was calculated separately, based upon slough water level fluctuations. As a result, the variation of slough water balance over the three years was quantified and related to the climatic variations over the same period. Corroboration was also provided for the hypothesis that prairie sloughs are effective for aquifer recharge. As with the study by Owen (1995), the water balance closure error was considerable, pointing to inaccuracies in measurements and calculations. However, this error was not so large as to invalidate the qualitative conclusions regarding the climatic sensitivity and recharge function of the slough.

1.3.2 Distributed Flow Monitoring and Modelling

The acknowledgement of spatial extension in a catchment (as opposed to viewing it as a spatially averaged unit) permits concepts such as the groundwater flow equation, or the Saint Venant equations for surface water flow, to be applied to the catchment as a whole. The adoption of such concepts opens the way to the explanation of hydrological events
under their theoretical frameworks. Such research invariably involves distributed measurements and modelling. One example is the production of stream flow peaks by groundwater ridging. Waddington et al. (1993) inquired into its occurrence in a groundwater-fed streamside wetland, using transects of piezometers and tensiometers in conjunction with chemical hydrograph separation. The measured potentials were used to calculate a soil water flow net and demonstrated that soil water flows were too weak, in this case, to provide peak stream flows.

In a similar fashion, Christensen (1994) used distributed groundwater modelling to substantiate that slow flow in a Danish stream was dependent upon upward leakage from a semi-confined regional aquifer to the phreatic aquifer. A finite difference model for two-dimensional transient Darcian flow, representing the confined aquifer, was coupled to a semi-distributed linear reservoir model representing the phreatic aquifer, with simple treatment of snow and interception storage, evaporation and stream flow generation. The model was calibrated to reproduce observed stream flows and confined groundwater heads, and then was used to predict the effects of groundwater abstraction on the stream flow regime. Although the accuracy of reproduction was not rigorously assessed, it was sufficient for showing that groundwater development would change the water balance considerably. The modelling results also drew attention to the seasonal dependence of the stream flow sensitivity to the perturbations.

There are more cases where scientists have been able to use numerical modelling to good advantage in water budget estimation. These include studies by Shaw et al. (1990), Carter et al. (1994), Hunt et al. (1996), and Siegel (1983). The example given below has particular relevance to the present study, in that the authors' objective was to elucidate the hydrology governing a groundwater-fed peatland.

In the study of a small groundwater-fed mire in East Anglia, U.K., Gilvear et al. (1993) used a three-dimensional steady state groundwater flow model to substantiate their impressions concerning the water flows around the site. Evidence of groundwater supply to the wetland was already provided by the presence of calcium-rich surface waters, tufa
mounds and a rich fen vegetation community on the site. Nevertheless, they used the model to show that the known hydrogeology could support an upward flow system and was therefore a viable explanation of the water budget. In addition, the modelling demonstrated the sensitivity of this flow system to the hydraulic conductances of the surface deposits, lending credence to the opinion that the growth of the tufa mounds might eventually isolate the wetland from the groundwater.

Such work illustrates a certain precept in catchment research. Quantification of the water budget of a site directly corroborates or falsifies impressions gained by any other means (these are often floristic or mineralogical). However, the accuracy of such calculations is usually very limited. This is due to the discrepancy between the point measurements used for estimation and the true spatial average of the quantity. Where groundwater and overland flows are concerned, the spatial variation in the hydraulic properties of the site can be taken into account to improve upon the lumped estimation, via distributed, physically-based modelling (this does not apply when site stratigraphy or topography is known in no more detail than the flow measurements: Chapter 6, Section 6.7.3 addresses this problem).

Distributed models thus provide a way of combining relevant information to improve water budget estimates. Most importantly, the incorporation of such information in the water budget calculation produces a more all-encompassing characterisation of the site, so that the water budget estimates are capable of more definite falsification of competing system conceptualisations. This principle underlies the use of physically-based, distributed models in catchment research and is shown, in the present study, to apply equally well to wetland hydrology.
1.4 SUMMARY

The preceding sections of this chapter established the context of this study. Goss Moor, a lowland wetland site in the headwaters of the river Fal, Cornwall, U.K., is thought to be gradually drying out. If at all possible, this situation should be ameliorated since Goss Moor represents a valuable part of the remnant wetland resource of Britain. The intrinsic worth of wetlands has been well established: they provide flood mitigation, pollutant transformation, archaeological conservation and are now rare ecological habitats. Wetland derogation has proceeded to such an extent that only 4% of lowland mires remain in southern Britain. The present study of the moor's hydrology as it stands today is intended to help English Nature to prevent further desiccation and protect the wetland's endangered ecology. Chapter 1 of this thesis has presented a review of topics relating to wetland hydrological investigation and suggestions for research leading to a better understanding of the present-day hydrological system of Goss Moor.

It has been shown that there is broad variability in the hydrologic function of different wetlands. Consequently, while most wetlands are hydrologically sensitive, the nature of this vulnerability varies from site to site. However, the common factor in the maintenance of essential water levels in mires is that of water retention in the mire itself or in its catchment. This process of storage would be expected to feature prominently in the hydrological regime associated with the wetland.

The above discussion suggests that measurements of components of the water balance, such as groundwater storage, river outflow and evapotranspiration, together with estimation of site characteristics such as hydraulic conductivities and stratigraphy, would at least permit an assessment of the relative importances of groundwater and surface water in maintaining waterlogged conditions on Goss Moor. This in turn would allow the prioritisation of protective land alterations on the wetland, such as tree felling and drainage ditch infilling, and also the prediction of the likely impacts of activities such as road building. A numerically implemented model of the hydrogeology of the catchment would play an essential part in the investigative process, as a means of incorporating extra data into the
estimation of the groundwater budget, thereby improving the force with which alternative conceptualisations of the catchment's hydrology could be rejected. The application of this technique to Goss Moor would extend the current exploration of modelling for wetland hydrological research. Furthermore, the stream flows measured upstream and downstream of the wetland provide an opportunity to assess the wetland's propensity to store water in comparison with that of the surrounding catchment. This should allow further guidance of watershed improvement schemes with the aim of protecting the wetland from further desiccation.

1.5 AIMS AND METHODOLOGY OF THE STUDY

The primary aim of the present study was to broadly categorise the flow and storage of water in the wetland in terms of its basic relationship with the surrounding catchment, and to quantify the relative roles of groundwater and surface water in such wetland flow and storage. The answers were sought for three complementary questions.

A) How much and what type of flow is contributed to the wetland surface and to its substrata?
B) What are the relative water demands from the various drainage processes on the wetland surface and on its substrata?
C) Does the wetland suffer from more rapid depletion than other parts of the catchment? (The answer would highlight whether remediative work should be performed on the wetland as well as on other parts of the catchment.)

In the light of the foregoing discussion, the techniques of stream flow recession analysis, distributed monitoring and modelling and lumped water budget estimation were combined in order to derive complementary observations on the hydrology of the catchment. This involved the study of all components of the site's water balance and the characterisation of the geology, topography and drainage network of the catchment. Four specific objectives were considered.
1 Describe hydrology and relevant site characteristics:
climate, geology, topography and vegetation of the catchment.
detailed stratigraphy of the wetland substrate.
transmission/storage properties of wetland substrate.
water table fluctuation and channel hydrograph behaviour in the wetland.
rainfall and evapotranspiration in the wetland.

2 Assess the variability of flow from the wetland and from upstream slopes:
analysis of stream flow variability upstream and downstream of wetland.
determination of separate flow components distinguished by mode of variation.
conceptualisation of flow component behaviour in hydrophysical terms.
deduction of wetland contribution to flow components.

3 Develop a numerical model of the wetland’s groundwater flow:
physically-based, distributed model of transient groundwater flow.
icorporation of the results of the site characterisation.
calibration with respect to the observed water table fluctuations.
interpretation of calibration measures in terms of wetland-groundwater interaction.
analysis of the sensitivity of groundwater flows to parameter uncertainty.

4 Quantify the overall water budget of the wetland:
precipitation rates.
evapotranspiration rates.
surface channel inflows and outflows.
groundwater inflows and outflows.
changes in groundwater storage.
diffuse surface/near surface inflows from the outside catchment.
1.6 STRUCTURE OF THE THESIS

The methodology outlined above corresponds to the following arrangement of topics in the remainder of this thesis.

CHAPTER 2 Site Description
Entitled “The Goss Moor Catchment”, this chapter reviews previous geological and soil surveys of Goss Moor, examines recent local climatological data, and inspects the physiography and land uses within the catchment. An assessment of catchment structure and probable hydrological character is made on the basis of these data. A brief description of the wetland vegetation is included.

CHAPTER 3 Experimental Design
Entitled “Hydrometric Monitoring Undertaken on Goss Moor”, this chapter describes the synthesis of experimental techniques towards obtaining representative data for budgeting and modelling purposes and stream flow recession analysis. The overall monitoring scheme, lasting one hydrological year and covering most of the wetland area, is introduced. Techniques for determination of rainfall, stream flow, evapotranspiration, groundwater pressures and geohydraulic properties are described and assessed, with some consideration of possible errors in estimation.

CHAPTER 4 Assessment of Observed Variables
Entitled “Variability in Observed Wetland Fluxes and Storage”, this chapter examines the hydrological variables measured in the wetland over a period of approximately one year. Rainfall, stream flow, evapotranspiration and groundwater levels within the wetland are presented and analysed for evidence of water storage in the wetland or its source areas, and related to hydrophysical concepts introduced in Chapter 2. Importantly, the chapter includes a review of previous work on the investigation of source area storage through the analysis of stream flow recessions, explaining why such analysis is capable of comparing storage in various source areas but not of revealing the nature of such storage, whether
groundwater or surface water. It is suggested that the latter objective can be met either by hydrochemical analysis or by hypothesis testing using groundwater modelling.

CHAPTER 5 Estimation of Catchment Water Budget
This chapter, entitled “The Water Budget of the Goss Moor Catchment”, analyses the seasonal and annual balance of inputs, outputs and storage for the entire study catchment. The time-cumulative variations of rainfall, evapotranspiration and river outflow volumes are compared in order to assess the possibility of groundwater inputs or outputs to the catchment as a whole. Seasonal changes in the dominance of the different fluxes are also examined.

CHAPTER 6 Groundwater Modelling
This chapter, entitled “Numerical Modelling of Wetland Groundwater Flow”, deals with the development and calibration of a numerical model of the groundwater flow beneath the wetland. Following the selection of the modelling software, system characteristics such as aquifer and river shape, boundary conditions, rainfall and potential evapotranspiration are incorporated into the model. After establishing the stability of the model with a steady state calibration, the model is calibrated with respect to transient water table behaviour. The wetland-groundwater interactions are considered during the calibration. Provision is made for the analysis of the sensitivity of groundwater flows to changes in uncertain parameters.

CHAPTER 7 Wetland Water Budget
Entitled “The Water Budget of the Goss Moor Wetland”, this chapter compares and combines measured and modelled flows from previous chapters in order to help build a coherent assessment of the hydrology of the wetland. All stream flows and groundwater flows through the wetland boundary are included along with rainfall and evapotranspiration. Storage of water in the aquifer beneath the wetland is compared with storage in the overall catchment. Candidate sources of augmented in-channel slow flows are compared using output from the numerical groundwater model and budgetary evidence of wetland surface water storage. Budget calculations also provide an estimate of the annual input of surficial flows from adjoining hill slopes onto the wetland.
CHAPTER 8  Summary and Conclusion

This chapter reviews the findings of the previous chapters and so evaluates the hydrology of the Goss Moor wetland. The particular approach adopted in the project is examined for its applicability with other sites, and suggestions are made for further research at Goss Moor. Finally, the elicited hydrology is related to possible management treatments for the wetland.
CHAPTER 2
THE GOSS MOOR CATCHMENT

2.1 INTRODUCTION

Goss Moor is the largest mire and heath complex in south-west England and is now designated as a National Nature Reserve with Grade 1 status. However, willow scrub is invading large areas of the reserve, possibly as a result of changes to drainage both within and external to the wetland. Therefore, understanding of the hydrology of the area should be advanced sufficiently to form a basis for management decisions by English Nature. A map of the wetland is shown in Figure 2.1. Several factors combine to give the Goss Moor wetland particular hydrological interest:

1) There is an extensive, detailed literature on the geology of the wetland and a rudimentary inventory of the wetland vegetation and its spatial distribution.

2) Distinct vegetation types exist in clearly demarcated separate sections of the wetland, lending themselves to consideration of their effects upon the local water balances.

3) The wetland is relatively large at about 5 km² and covers numerous sites of ground disturbance by surface mining in past centuries. These mining scars are likely to contribute to the hydrologic character of the wetland.

4) The wetland appears to be gradually drying out, possibly due to canalisation of its river channel in recent times. It thereby provides a case for investigation of the processes involved in wetland desiccation.

These points make the site suitable for both present and future hydrological investigation. In the present study, Goss Moor was examined partly because of the need to understand its hydrology for conservation purposes, and partly because of the availability of background
information. In scientific terms, the wetland provided a case study for the investigation of the effectiveness of several analysis methods in combination: stream flow recession analysis, numerical groundwater modelling and lumped water budgeting.

This chapter will describe in detail the \textit{a priori} information on climate, geology, physiography, vegetation and drainage characteristics of the Goss Moor catchment.

\section*{2.2 CLIMATE}

Goss Moor is situated in the Cornish peninsula of south-west England (Figure 1.1). The peninsula receives precipitation at all times of the year, borne on cyclonic fronts from the Atlantic Ocean. This precipitation, falling almost exclusively as rain due to the mild temperatures, is most frequent and heavy during the months September to March. The period from April to June has the least precipitation.

Mean daily temperatures are lowest in January, at around 6°C, and highest in June, at around 10°C. Monitoring on the wetland was conducted from June 1993 to November 1994. Over this period, the lowest mean hourly temperature recorded by the Meteorological Office at the nearby St. Mawgan airfield was -4.2°C, on 22\textsuperscript{nd} November 1993, while the recorded maximum, at 26.2°C, occurred on 23\textsuperscript{rd} July 1994.

Table 2.1 shows the mean monthly rainfall rates for the years 1975 - 1979, derived from data from the South West Region Environment Agency's recording stations at Roche (4 km east of Goss Moor) and Bugle (7 km east of Goss Moor). This period's mean annual rainfall of 1368 mm compares with a potential evaporation value of 538 mm/annum for Cornwall in the period 1950 - 1964 (Camm, 1981). Combining these figures, the net transfer of water from atmosphere to land is around 830 mm/annum. Potential evaporation is approximately 400 mm in the months April-September and 138 mm during October-March. Corresponding rainfall values are 458 mm and 910 mm respectively, giving a net water gain at the land surface of 58 mm in the spring/summer months and 772 mm in
autumn/winter. Also shown in the table are the monthly rainfall and potential evaporation values measured by the present author and by the Meteorological Office, over the period from June 1993 to May 1994. Rainfall during the study year was generally at the high end of the range of values found during the earlier years. However, of particular note is the low rainfall in August 1993, which when combined with potential evaporation in the same month, gives a net water loss at the ground (-35.0 mm). Groundwater storage would necessarily be depleted, and water tables would fall in such circumstances. The average net gain found for the summer months of the earlier years (Camm's (1981) data), being close to zero at 58.0 mm, also suggests the possibility of net loss during individual months. However, some effect would be expected from including data from 1976, with its summer drought, along with the use of potential evaporation which is an overestimate of actual evaporation during the summer.

In summary, Goss Moor is associated with the general climatic characteristics of north-west Europe, experiencing cyclonic activity for much of the year. The rainfall regime is one of frequent rainfall with low-medium intensities but long durations. In combination with low potential evaporation rates due to mild temperatures and high humidity, this promotes a continuously high terrestrial turnover of water. Overall, the data showed the dominance of rainfall over evaporation, suggesting the climatic suitability of the area for the development of wetlands.

2.3 LAND USE HISTORY

2.3.1 Mineral Exploitation

The river and streams of Goss Moor were worked for tin over many centuries, until the end of the nineteenth century. Camm (1981, p.12) quotes a mining journal of 1858 which stated that "the Goss and Tregoss tin streamworks had long been celebrated for their extent, antiquity and productiveness." There are likely to be long sections of the present and former water courses in the moor which have been the site of excavation of alluvial
material. Such locations must encompass most of the areas of the catchment under which, as required for stream works, a substantial thickness of alluvium may be found. The extent of this alluvium is shown in Figure 2.2 (from Geological Survey of England and Wales Sheet 347, 1982). In the early 20th century, there were several attempts to mine the regolith of the southern section of the moor with mechanised dredges. These schemes were all abandoned in their early stages due to economic problems.

In addition to the tin streaming and dredging, a significant amount of gravel was extracted from Goss Moor. Camm (1981, p.13) states that, "During and after the Second World War, considerable quantities of gravel were graded and removed for road aggregate. The area worked is easily identified as the sites of numerous ponds in the northern central area of the Moor." These ponds are shown in Figure 2.1.

On the fringes of the catchment, both past and present mining are in evidence. To the north, are the abandoned shaft heads of the Castle-an-Dinas and Great Royalton tin mines (Jenkin, 1964), while beyond the south-eastern catchment boundary are the Hensbarrow Downs china clay quarries. These quarries have been in operation since the 19th century. The Goss Moor catchment at its south-eastern extremity encompasses some of the spoil heaps and settling ponds associated with the quarries.

2.3.2 Agriculture

Substantial tracts of the area form part of the Tregothnan Estate of Lord Falmouth, and also come under commoners' rights. The drier land on the slopes, which has been enclosed by walls constructed of granite boulders, is primarily used for stock rearing. The wetland area (shown in Figure 2.1) provides common rough grazing for cattle from adjoining tenant farms. The remainder of the catchment encircles both the wetland and the intrusive ridge of Tregoss Moor at higher elevations and, with the exception of the china clay waste mounds just within the south-eastern periphery, consists of improved agricultural land. This land serves mainly as low intensity pasture.
2.3.3 Wetland Vegetation

The Goss Moor wetland occupies about 5 km$^2$ of the 22 km$^2$ catchment area. As mentioned in Section 2.3.2, most of the remaining area is improved agricultural land suitable for low intensity grazing. The exception to this is the ridge of Tregoss Moor, which supports an extensive community of dry heathland, *Ulex galii* - *Agrostiss curtisi*, H4 in the National Vegetation Classification (Rodwell, 1991). Gorse and grass species appear as community dominants, although some patches of *Calluna vulgaris*-dominated sward also exist.

Goss Moor itself is a complex wet heath, mire and wet woodland site, being the largest lowland wetland in south-west England. The wet heath and mire communities grade into extensive poor fen and swamp communities occupying the lowest parts of the valley basin. Wet heath communities of *Schoenus nigricans* - *Narthecium ossifragum* (M14) occupy the shallow slopes. Intermixed with both H4 and M14 are extensive tracts of M25 *Molinia caerulea* - *Erica tetralix* mire and this tends to develop where the ground is slightly wetter. Waterlogging increases westwards, resulting in widespread open fen and mire communities over most of the valley bottom. Typical of the communities are Poor Fen (M28) *Iris filipendula* - *Juncus acutiflorus* intermixed with M23 *Juncus acutiflorus* - *Galium palustre* subcommunity. At the wettest points, where standing water occurs, Swamp S27 *Carex rostrata* - *Potentilla palustris* often develops as a mat of floating vegetation. Dominating the whole basin is a patchwork of willow fen-carr. This community is entirely W1 *Salix cinerea* - *Galium palustre* woodland which has invaded the wetland over the past 50 years.

2.3.4 Conservation

English Nature hold a long lease for both Goss and Tregoss Moors. The conservation interest of the site is mentioned in Sections 1.1 and 2.3.3. Similar assessments led to its designation as a National Nature Reserve (Category 1). Tregoss Moor, a ridge of land supporting dry heath which extends into the wetland from the high ground in the east, is managed as a relatively undisturbed area following a similar strategy to that of Goss.
2.4 GEOLOGY

2.4.1 Introduction

The current section reviews literature on the geology of the study area, which consist of memoirs and a minerals report from British Geological Survey together with a commercial mineral exploration report and a mining history treatise.

In particular, the commercial report provided most of the information used in this thesis to characterise the shallow geology immediately underlying the wetland. The report details a preliminary evaluation of the alluvial tin reserves of Goss Moor, conducted during 1980-81 by Billiton Exploration (U.K.) Ltd. to determine its potential for further mineral exploitation. This involved drilling exploratory boreholes in a grid of equilateral triangles each 300 m on a side, as shown in Figures 2.2 and 2.3. The area covered by each borehole was thus 77,940 m². A total of 83 cased boreholes were drilled down past the base of the overburden using the shell and auger technique. Depth determinations in the borehole logs were estimated to be accurate to within 0.1 m. The results of this survey, presented by Camm (1981), are used in the present study to characterise the wetland substrate and, along with in situ permeability measurements, to ensure that realistic values of permeability are assumed in the calibration of the numerical model of the wetland's groundwater flow regime.

2.4.2 Solid Geology

Figure 2.2 shows the solid geology of the area. The Goss Moor catchment lies over the north-west corner of the Late Carboniferous St. Austell granite which is the main relief-forming feature of the district. Ussher et al. (1909) state that (p.54) the granite upland "presents no ragged outlines, but consists of a series of low, flat, dome-like elevations separated by broad drainage hollows called moors, as Criggan Moor and Redmoor, and by steep-sided valleys as in the case of the River Fal, the St. Austell River, and the numerous small streams in the eastern part of the granite mass."
The main exposure of the granite, lying to the south of the catchment, dips away to a depth of 600 - 700 m, as estimated by Camm (1981), before resurfacing briefly as Belowda Beacon and Castle Downs 4 km further north. Ussher et al. (1909, p.54) found that

“The subterranean dip of the margin of the granite below the slates is not known with accuracy, but at Carbis, near the Roche Rock, the dip is about 45°, at estimated from the depth at which granite was cut in the shaft at Cornubia Mine. Judging from the width of the metamorphic aureole on the northern side of the mass it is probable that the dip is considerably less than this in some places, and the general circumstances suggest that in this neighbourhood it flattens out in depth.”

As seen in Dines et al. (1956) and Beer et al. (1986), the granite mainly remains well below the surface at Castle Downs, commencing at perhaps 50 m below the hill top. The hill is composed principally of country rock. However, on Belowda Beacon the granite comes to the surface.

The country rock, belonging to the Lower Devonian Meadfoot Beds, is an east-trending sequence of slates and interlaminated siltstones, sandstones and occasional limestones. These beds are known to dip steeply northwards (Dines et al., 1956). Around 4 km north of the Goss Moor catchment’s northern boundary, this sequence gives way to the overlying Staddon Grits which are composed of primarily non-calcareous sediments including gritstones. Beneath Goss Moor, the depression in the granite between Belowda and St. Dennis is overlain by a swathe of the Meadfoot Group. Because of the granite’s vicinity they are weakly metamorphosed, tourmalinized, silicified and spotted, thus earning classification under the general name of ‘pelite’. Except in close proximity to the granite, this metamorphism is sufficiently weak that even the fine laminar structures in the silts have not been destroyed (Ussher et al., 1909). The limestones have been altered to a calc-silicate hornfels facies. These occur in east-trending bands, noted by Ussher et al. (1909, p.8) as “impersistent hard bands approaching calc-flintas”, which exhibit repetition by folding to the north of the study catchment, although only one or two bands pass through the catchment itself - that making up the Tregoss Ridge separating the two main easterly fingers of the alluvial body and that striking westwards across the south-easternmost branches of the
alluvium. Beer et al. (1986, p.3) note that they are “predominantly hard, but brittle and well-jointed,” and therefore they may provide a fracture zone for subsurface flow.

The metamorphic aureole extends as far as 6 km from the granite contact, beyond the boundary of the study catchment. Evidence of earlier igneous activity takes the form of greenstone sills occurring in association with the calc-silicate bands (Ussher et al., 1909). Such greenstones occur north of the eastern end of the St. Austell Granite, and are not observed within the Goss Moor catchment.

The granite and metasediments have been altered to great depths by kaolinisation through the action of low pH, oxygenated hydrothermal and/or meteoric waters penetrating through fractures. Nevertheless, localised areas of the pelite have remained unkaolinised, through a combination of compositional digression and variation in fracture density. These weakly tourmalinized, grey-black, quartz-veined shaley slates are competent enough to have maintained a fairly distinct boundary with the overburden (Camm, 1981). The kaolinised pelites take the form of a soft, highly laminated quartz-veined clay and probably possess very low overall permeability. Factors such as periglacial disturbance and fluvial disturbance have caused a gradual change rather than a distinct boundary between the kaolinised metasediments and the overburden.

Somewhat south of the granite outliers of Belowda Beacon and Castle Downs, a felsitic dyke trending slightly north of east has greisenised the sediments along its walls, to a distance of a metre or so, such that they have become resistant to further alteration (Ussher et al., 1909). Cracks and joints in the dyke, up to several centimetres wide, extend out through this hardened zone to diminish and finally vanish in the neighbouring pelite. Dines et al. (1956) and Beer et al. (1986) describe a north-trending wolfram lode which penetrates down through the hill top at Castle Downs to be subsumed by the later-formed granite. Similar near-surface fracturing might be expected in this lode and its laminar zone of influence. A widespread occurrence of such dykes, fissured in their upper levels, would permit relatively copious groundwater flow through the pelitic strata beneath Goss Moor. However, the catchment contains no other dykes mentioned by the literature, and only one
other visible on the geological map. Thus, it is probable that the kaolinitised pelitic sequence remains a poor transmitter of water beneath the catchment.

Both the felsitic dyke and the wolfram lode have been exploited by mining. In the case of the dyke or elvan, this has been primarily in the nineteenth century. Several mine shafts and adits were developed at the western end of the elvan, one of which extended over three quarters of a mile, that is around 1.2 km, at a depth of 25 fathoms, or 46 m (Jenkin, 1964). Going eastwards along near the foot of Belowda Beacon, further mine passages were developed down to around 20 m below sea level. Records of the geology at this depth were not obtained in the present study, but at the much shallower depth of 46 m below ground, the elvan was traversed by many north-south veins of metal ore (Jenkin, 1964), suggesting considerable fracturing. The direction of the veins is undoubtedly that in which the dyke would most readily have sheared. This suggests that the formation of such veins depended upon the presence of a brittle body within the more plastic pelites, and thus that the fractures do not extend far out into this country rock.

The deep mine shafts developed here may be providing some hydraulic connection between the moor and the deeper groundwater flow system within the Meadfoot Beds, although thermal metamorphism which is stronger with depth, and/or kaolinitisation, may limit the vigour of such a flow system.

The lode on Castle Downs is cut through at eight levels down to a depth of 160 m below the hill top, extending southwards by 400 m and northwards out of the catchment by about 900 m (Beer et al., 1986). The drainage adit of this mine follows the northward direction and, without pumping, would drain down to the fourth level, at a depth of around 90 m. No measurements are available of the discharge from this adit. It might be assumed that the surrounding kaolinitised pelites, which constitute the main part of the hill's mass, do not yield significant flows to the mine. However, in the slopes several hundred metres north of the hill top, a mine developed to 8 fathoms (15 m) in the late 1830s was flooded. Jenkin (1964) writes that "... the water proved so quick that it overpowered their works." Due to the relatively shallow depth of this flooded mine, little can be said about groundwater flows at
lower levels in the hill. However, the flooding provides evidence of a highly transmissive layer of fragmentation existing down to the level of the mine. This earlier mine probably did not extend within the Goss Moor catchment boundary, and so its drainage effect is unlikely to be felt within the catchment. However, drainage of the fragmented layer by the later wolfram mine is likely to have some influence on groundwater flows within the southern slopes of Castle Downs.

The granite of the St. Austell boss comes to the surface at Castle Downs, Belowda Beacon, St. Dennis Crown and all along the southern edge of the Goss Moor basin. Down the slopes of these hills, the upper surfaces of the granite and the adjacent metamorphic rocks are overlain by a thin layer of agricultural topsoil. On other exposures of Palaeozoic intrusions, such as on Dartmoor in Devon, U.K., the upper layers of such pyrogenetic units have been weathered, both physically and chemically (see, e.g., Alexander (1983), and Williams (1983)). Thus, a layer of highly fractured and fissured rock is situated beneath the soil profile in such areas, overlying more coherent granite. From Ussher et al. (1909) it is gathered that the St. Austell granite also exhibits comprehensive weathering. They report (pp. 54-55) that

"The greater portion of the granite is, however, covered by a mantle of decomposed and fragmentary material derived from its breakdown, and this material varies in nature from a fine sandy material, known as ‘growan’, to coarse subangular fragments and blocks of large size which occur to a considerable depth on the slopes and tops of the hills. In the erosion hollows much of the material has been transported from the higher ground, and may be regarded as true alluvium. The enormous remarkably well-rounded, half-buried boulders, ranging in size up to 36 feet in length and in weight from a few tons up to hundreds of tons, which lie scattered over the moors and pasture lands in such countless numbers in the parishes of Luxulian and Lanlivery, appear to have been derived from the abrasion and rounding of the cuboidal rocks into which the granite weathers.”

They go on to state that

"The general form of tors and the loose blocks derived from the granite is determined by the jointing of the region, being the effect of weathering along the fissures, joints, and pseudo-bedding planes of the granite. The jointing is roughly threefold. The principal series of joints or fissures has a bearing about N.E., and corresponds in direction with the lodes of the region. A second series is roughly at right angles to the first, and corresponds with a second series of lodes. The pseudo-bedding is roughly horizontal, or slightly inclined, and the planes vary from a few feet to many feet apart. Another series of joints appears to be present and to correspond with the direction of the principal iron lodes, being about N. 30° W. in the Luxulian district [to the south-east of the Goss Moor catchment]. Small parallel
valleys have been formed along the line of the last series in the region of Lanlivery and Luxulian.”

Since one would expect a fragmentation depth several times greater than the thickness of the largest loose granite boulders in the area, depths of around 30 m would not be implausible. Although Ussher et al. (1909) also note that jointing is rather poorly developed in the granite of St. Dennis Crown, the above information strongly suggests that considerable subsurface water flow occurs down the hill slopes on the southern periphery of the catchment.

2.4.3 Overburden Geology

Goss Moor is the site of Flandrian fluvial debouchement onto the periglacial reworkings of a wave-cut platform formed during the 137 m marine incursion of the Pliocene Period (Camm, 1981). Ussher et al. (1909) note that many other alluvial flats exist at the same elevation in this area of Cornwall. The palaeochannels of the moor have the same gradient as the present river profile, dropping about 6 m per km.

The wetland itself lies directly upon the alluvial deposits distributed along the valley bottom. These deposits consist of coarser grained, higher energy channel sediments interdigitated with lower energy overbanks and Sphagnum spp. rich peats. This suggests a migratory channel, and perhaps a braided stream environment. On the scale of a few metres, the resultant spatial distribution of hydraulic conductivities in the alluvial aquifer is probably highly heterogeneous: lenticular units of starkly contrasting conductivities are located adjacent to one another. Horizontally, the stream deposits exhibit a more gradual transition into the overbanks, due to the migratory nature of the braided stream channels. However, the fluvial sediments found at neighbouring boreholes show no correlation of their positions in the downhole sequence and so even the overbank deposits are less laterally extensive than the 300 m borehole separation. A considerable quantity of peat is intercalated with the silts and clays of the floodplain deposits, with some peat also having developed within abandoned or oversized channels.
The mean depth of alluvium in the Billiton boreholes was 3.8 m, with about 70% of the alluvial overburden being more than 2 m in thickness. Greater depths of alluvium occur occasionally, associated with the higher energy channel areas where fluvial incision into the underlying kaolinised pelite has been greatest (Camm, 1981). Dines et al. (1956) report on earlier activities in which the gravelly fluvial deposits of the moor were extracted for tin, and state that

"The thickness of the deposits ranged from a foot or two to 20 ft. or 30 ft., the irregularities in the shelf or bedrock floor being often steep and trough-like."

Periglacial deposits lying beneath the alluvium accounted for a substantial proportion of the overburden analysed in the Billiton survey (Camm, 1981). Thick deposits of stony clays, characterised as solifluction material, lie over the pelites at the base of slope, while the most abundant non-fluvial overburden type is that of disturbed kaolinised pelite. This latter type takes the form of a creamy-grey/white clay containing pelite-derived fragments. A frost heave mechanism may account for its formation, in which any original pelitic and alluvial structures have become totally disordered. The transition downwards from this periglacial product into well structured kaolinised pelites is usually rather gradual, while upwards, some gradation also occurs with the overlying alluvium.

2.4.4 Geological Summary

The picture of the overall geological stratigraphy of Goss Moor which emerges from the above is as depicted in Figures 2.4 - 2.6. Beneath the centre of the catchment, the granite contact lies at depths of up to 700 m and is overlain by a corresponding thickness of pelites, dipping steeply northwards, at various degrees of kaolinisation. The upper surface of these pelites, along with any early fluvial deposits, has been disturbed by periglacial processes, forming a layer of disordered stony clays. Above this disturbed pelite, the present river has left a sequence of braided channel and flood plain deposits. Of all these strata, the alluvium has the least lateral extent, being confined to the lower parts of the catchment. The kaolinised pelites and their solifluction products extend up the lower slopes of the surrounding granite bosses. Beyond the limits of these units, the exposure of the underlying
granite is manifested in what may be a significantly deep and hydraulically transmissive upper layer of fragmented rock. The kaolinitised basement rocks, taking the form of stratified clays, probably exhibit very low permeability, suggesting an impervious base to the wetland aquifer and other bodies of regolith in the catchment.

Some minor parts of the pelites remain hard and fractured. These are the east-trending calc-silicate bands in the Tregoss Ridge and approximately 1 km south of the ridge. Through their fragmented nature, they may accommodate significant subsurface water flow, at least in their upper sections. While the enhancement of intracatchment groundwater flow within these bodies may thus be probable, intercatchment subsurface flow remains only a remote possibility. The same may be said of the few felsitic dykes and the wolfram lode found within the catchment, and of the deeper mine shafts and passages.

2.5 SOILS

Figure 2.7, based on *Soils in Cornwall II: Sheet SW53 (Hayle)* (Staines, 1979), illustrates the soils found in the Goss Moor catchment. There follows a brief description of the distribution and possible hydrological significance of these soils, based upon an assessment by Heathwaite (1990). Findlay et al. (1984) provide a detailed description of the soil associations mentioned here.

On the slopes just within the southern catchment boundary, an area of the Hexworthy Soil Association is bounded on either side by Manod soils. The Hexworthy Soil Association consists mainly of iron pan stagnopodzols, in which dense, poorly permeable layers of ferric iron precipitates are found a short distance below the ground surface. Due to the metalliferous nature of the parent rock beneath the catchment, there is no lack of availability of iron or aluminium in the soils of the moor for this phenomenon (Staines, 1979). The acid humus at the surface releases organic acids, which form complexes with iron and aluminium and move down through the soil with percolating water, leaving the upper horizon with a bleached appearance, and finally precipitating upon reaching the more basic environment at
lower levels (White, 1987). The poorly permeable ferric iron layer impedes drainage to the water table, creating a perched zone of saturation and causing a high incidence of saturation excess overland flow from such areas.

The areas with Manod Soil Association, found also on the southern slopes of the catchment, consist of quite freely draining brown podzolic soils. Going eastwards from the southern side of Goss Moor, this soil association extends up the south-eastern side of the catchment, merging into the ferric stagnopodzols of the Hafren Soil Association, which occur on the eastern side of the catchment. From here, going northwards and then westwards along the northern periphery of the catchment, Hafren in the east is replaced by the Manod Soil Association in the north-east, which merges again into Hafren in the north-west. The Hafren Soil Association consists mainly of loamy, ferric stagnopodzols which are slowly permeable to water.

The remainder of the high ground around the catchment periphery is accounted for by the Moorgate Soil Association. This covers the igneous crowns of the St. Dennis and Belowda Beacon hills, and is characterised as mainly coarse, loamy, humic brown podzolic soils with good permeability.

Further down slope in the catchment’s southern section, an extensive area of the Yeollandpark Soil Association is also recorded. This association consists mainly of permeable, loamy, groundwater or surface water gley soils. In groundwater gley soils, as explained by White (1987), the reduction of ferric (oxidised) iron to produce blue-grey-white complexes of ferrous iron has arisen through the continuous waterlogging associated with a high groundwater table. The interiors of the soil peds are gleyed, while some reoxidation of iron may occur to give an orange-red mottling in better aerated zones. This contrasts with surface water gleys, in which poor drainage of ponded surface water through the soil produces gleying on ped faces and in the larger pores. In Goss Moor, the gleying of the Yeollandpark soils is likely to be mainly due to a high groundwater table, although some surface water gleying may exist at slightly higher elevations.
On the ridge of Tregoss Moor, the Denbigh 2 Soil Association is recorded. Fine, loamy
typical brown earths, with good/moderate permeability, are the predominant soil types in
this association. This association is also exhibited on the raised ground around Toad Hole
and Enniscaven.

The low-lying, wetland areas of the Goss Moor catchment are covered mainly with the
Laployd Soil Association. These are reasonably permeable humic gley soils, continuously
waterlogged and with a high organic content. The waterlogging here is caused by a high
groundwater table, as in the Yeollandpark soils to the south.

In summary, the overall picture is as follows. The upper slopes of the catchment’s periphery
are characterised by two main types of soil: the freely drained brown, podzolic Manod Soil
Association on the southern, south-eastern and north-eastern sides, and the slowly draining
loamy brown earths of the Hafren Soil Association on the eastern and north-western sides.
Minor areas of Moorgate (freely draining, podzolic) and Hexworthy (badly draining, iron
pan) soils are included. Further in towards the wetland, can be found two more soil types:
in the south, the permeable, loamy groundwater gley Yeollandpark soils; on raised ground,
the moderately permeable loamy brown earths of the Denbigh 2 Soil Association. These
give way to a predominance of the permeable, humic groundwater gleys of the Laployd Soil
Association in the central, low lying expanse of wetland.

2.6 PHYSIOGRAPHY AND SURFACE DRAINAGE

2.6.1 Catchment Boundary and Land Gradients

Figure 2.8 shows the surface features, contours and surface catchment boundary of Goss
Moor. The raised embankment of the former St. Dennis railway line provides a convenient
lower catchment boundary, as it captures all diffuse surface flow from the greatest part of
the wetland, which lies on its eastern side. The river Fal is the major outflow from this
catchment, exiting through a culvert at the northern end of the embankment and flowing
through the western extremities of the wetland, before turning southwards to negotiate its way between the St. Austell granite and its westernmost outlier. The catchment area at the Goss Moor outflow beneath the railway embankment is 22.40 km², containing a wetland area of approximately 5.35 km².

100 m south-west of the outflow of the Fal is the only other outlet from the catchment: a small culvert draining a system of small ponds formed from gravel extraction pits. From this culvert, a channel flows west through the remaining wetland to join the river Fal a few hundred metres downstream of the main outflow.

From the periphery of the catchment, relatively steep slopes, generally between 5% and 15%, slacken off towards the centre of the moor which has, on average, a slight, westward-falling gradient of around 0.8%. The gradient is necessarily more uniform in the flatter, lower parts of the catchment; around the upslope perimeter, wide deviations occur, such as on St. Dennis Crown, with slopes of up to 20%, and just north of the catchment outflow, where the gradient is no greater than in the moor's interior. Catchment perimeter elevations of between 125 m just north of the catchment outflow and 250 m in the Hensbarrow Downs area to the south compare with 123 m in the centre of the moor and 118 m at the outlet. Plate 2 shows the landform of the wetland.

2.6.2 Surface Flow and Channel Network

The majority of channels in the Goss Moor catchment have at one time or another been artificially altered, whether in the course of tin streaming operations over the past centuries, or for the improvement of drainage for agricultural land in more recent times. The drainage density around the edges of the catchment is higher than would be expected naturally, due to a proliferation of field drainage channels. In the centre of the moor, there is much diffuse overland flow. With the exception of the river Fal and a system of ill-defined and unstable miniature channels, this would be the only mode of surface runoff across the areas of wetland, were it not for the introduction of further drainage by man. Nevertheless, what extra drainage has been recently introduced for agricultural purposes has been confined to
the eastern half of the wetland, leaving the western half affected only by discontinuous channel remnants from long-abandoned tin streaming.

Flow from the southern catchment boundary
There is a general emergence of groundwater in springs and seepage zones at the break in slope near the southern periphery of the catchment. From here, the water flows northward in field drains, roadside ditches and streams, into the wetland. Entry into the wetland is marked by a transition from distinct to ill-defined channels, merging with generally sluggish overland flow. The exception to this is the Toad Hole Drain, which receives spring water and runoff from the china clay waste area. This is a deeply incised man-made channel running approximately 1.5 km through the wetland to its confluence with the river Fal. Along the last kilometre of its course, one or two minor rills have been noticed to funnel diffuse flow from the wetland surface into the channel.

In the 1960s, the Central Electricity Generating Board built the St. Dennis substation at the western end of Goss Moor. Three parallel lines of pylons lead out across the southern section of the wetland from the station, serviced by a raised track of granite and clay construction. This track separates the southern periphery of the wetland from the main wetland area, and although a limited number of culverts allow flow beneath it, there is otherwise a general backing up of the diffuse northward flow along its southern side. This is reflected in the higher surface water depths in that section of the wetland. Water channelled through the culverts dissipates once again into diffuse flow on the northern side of the track.

Flow from the northern and north-eastern catchment boundary
The agricultural land on the northern periphery of the catchment is drained by field drains and drainage ditches which then run down to meet the main east-west-running road (the A30) which separates the northern section from the remainder of Goss Moor. A small number of culverts, often inactive, allow the flow to continue southwards beneath the road. Below the base of the northern slopes, there is an extensive area of wetland. The ditches carrying water from the higher land on Belowda Beacon and Castle Downs, upon entering
these lower elevations below the base of slope, also receive substantial contributions from the subdued overland flow and soil water flow of the wetlands.

The A30 is bordered on either side by drainage ditches. However, these were found not to be carrying significant amounts of water and sometimes exhibited reversals in bed slope which inhibited the flow of water.

At the eastern end of the catchment, culverts beneath the main road allow northward drainage of the Tregoss ridge. This drainage joins channels flowing westward through the north-eastern limb of the wetland, eventually recrossing the road southwards beyond the western tip of the ridge. These features can be seen in Figure 2.1.

**The river Fal and tributaries**

The river Fal originates in the south-eastern corner of the catchment, among the china clay waste mounds. In its upstream reaches it is joined by many small field drains and ditches, together with some contribution from diffuse surface flow in the easternmost parts of the wetland. On entering the central area of the wetland at Tregoss bridge, it has accumulated an average annual flow of around 0.2 m³s⁻¹. The 1.6 km stretch of river downstream from here has been canalised to give a trapezoidal channel incised to depths of between 1 and 2 m, the river profile having an overall gradient of around 0.6%. This reach of the river is joined by two tributaries from the north, draining the southern and western sides of the Tregoss ridge. The Toad Hole Drain converges from the south, at the downstream end of this reach. Additions to the river by diffuse surface flow in this area are uncertain, although some small temporary rivulets have been observed discharging into the streams during periods of prolonged rain.

In the northern central area of the moor, the river flows through a system of ponds created by gravel extraction operations in the mid-twentieth century. These ponds act as substantial storage reservoirs, receiving water from the river and from the direction of the main road to the north. The water flow from the culverts beneath the main road runs southwards from the highway in the form of a finely interconnected network of sometimes ill-defined ditches.
and diffuse overbank flow, joining these ponds and the downstream stretches of the river in a correspondingly diffuse manner.

**Diffuse overland flow in the centre of the moor**

Much of the central area of Goss Moor is covered by surface water. Some sluggish overland flow results from this and moves in a westerly direction, in accordance with the ground surface gradient, towards the disused railway embankment at the western end of the moor. This embankment causes a backing-up of the water along its eastern side, and a redirection northwards towards the pond system just south of the catchment outlet. These pits receive overland flow also on their eastern side, and discharge through the small culvert in the embankment which serves as the secondary catchment outflow, the water joining the channel of the river Fal a short distance downstream.

**Overall hydrological character of the catchment**

The storm response of the upper slopes of the catchment periphery is influenced by two factors: the steep gradients and the well permeable soil. Firstly, the steepness of the ground surface encourages swift runoff to the ditches and drains, and lessens the available time during which surface and soil water might percolate downwards to replenish groundwater storage. The effect of this is to allow a flashy response of the drainage ditches and quick delivery of storm water to the lower parts of the catchment.

In the second instance, the characteristically open pore structure and high permeability of the brown earths and brown podzols in these upper areas of the catchment may contribute to the rapidity of transfer to ditches via subsurface lateral flow. Nevertheless, this efficiency reduces the possibility of saturated excess surface runoff which, with wet antecedent conditions, is the most rapid runoff process. Since downward percolation is easy in such soils, provided that the underlying regolith is reasonably permeable, significant amounts of runoff might be expected to reach the base of the permeable layer, forming a saturated groundwater zone. However, given that this hill slope stratum is particularly thin compared to its range in altitude, such a groundwater body would drain very rapidly downslope and therefore be very limited in its longevity.
Nevertheless, at times when they are regularly replenished, the physically weathered layer of granite on the exposed surfaces of the St. Austell boss and the similarly degenerate exposure of the adjacent metamorphic rocks, may embody a substantial part of the groundwater reservoir in the catchment, and are responsible for the springs, found at the breaks in slope, around the edges of the wetland. Alexander (1983) and Williams (1983) discuss the extent of such weathered layers, and its implications regarding catchment hydrological response and the chemical composition of stream water, for similar geology in the Narrator catchment, Dartmoor, 50 miles to the east.

At the base of the peripheral slopes of the Goss Moor catchment, these aquifers form a fairly disordered contact with both the stony clays of the disturbed kaolinised pelites and the silts, sands and clays of the flood plain alluvium. Aside from the softening of gradients at the bases of slope, the gradation of the high conductivity weathered layer into finer-grained sediments on the edge of the moor and the consequent reduction in transmissivity is a principal cause of the emergence of the hillslope water as springs in these areas. However, some direct transfer of water from the hillslope aquifers to the alluvial deposits cannot be ruled out.

As the gradients flatten out below the break in slope, the peripheral dry land of the catchment gives way to wetland, fed by rain water and by the springs emerging from the hillslope aquifers. With regard to rainfall response, the creation of extensive saturated areas by groundwater discharge is thought to be the most important factor at these elevations, causing surface runoff by saturation excess.

Diffuse surface flow at these elevations, and lower in the catchment, is slow due to the gentle gradients and high flow resistance offered by the wetland vegetation, with the result that the response to rainfall events is rather drawn out. Considerable depths of surface water, from 0.1 to 0.4 m, are encountered from here on into the centre of the moor, indicating the strong propensity for storage on the ground surface and in depressions and hollows. Since the downstream reaches of the river Fal receive a substantial contribution of
their flow from rain falling on this area, a subdued response can be expected at the
catchment outflow. However, the most important observation regarding the hydrological
character of the wetland is not its effect on the response at the catchment outlet, but rather,
its reluctance to release water from its own reserves. This, in addition to the steady nature
of the input from the peripheral hillslope aquifers, is a major factor in the maintenance of the
wetland.

The alluvial silts, sands, clays and gravels at the centre of the moor, discussed in section
2.4.3, form a dish-like aquifer as shown in Figure 2.5, which is recharged by rainfall and
possibly by peripheral surficial inflows. This aquifer discharges to the river Fal by direct
seepage. If the efficiency of drainage of the surrounding land were to be improved, by the
introduction of new drainage ditches or the deepening of those already in existence, the
availability of recharge water for this aquifer might be substantially reduced, accompanied
by a conspicuous alteration in the wetland's water balance. The introduction of more
drainage channels in the central, wetland area of the catchment, by acting as a sink for
groundwater flow in the alluvium, might produce a similar disturbance of the wetland water
balance. The primary aim of the present study was to assess the hydrology of the wetland
and thereby facilitate further evaluation of its sensitivity to such disturbances.

2.7 SUMMARY

For conservation purposes, it was required to understand the hydrology of Goss Moor.
Furthermore, good availability of background information and the current transience of its
hydrological system made the site suitable for research. The present chapter established the
a priori knowledge on the site.

The 5 km² wetland occupies a 22 km² headwater valley catchment surrounded by granitic
uplands, in Cornwall. At this location in north-west Europe, cyclonic rainfall dominates
over potential evaporation, making the area climatically suitable for the formation of
wetlands. The river Fal rises in the catchment and exits through the wetland.
The bedrock has been extensively kaolinised, making it impermeable and suggesting that the groundwater basin is isolated from deeper groundwater systems, although the presence of geological fractures in hardened, unkaolinised zones and consequent exchange of groundwater with neighbouring areas cannot be ruled out. The granite slopes down from the surrounding uplands towards the alluvial flat of the moor are likely to be weathered to depths of some tens of metres, permitting water storage and transmission. An overburden of badly sorted, clayey periglacial deposits and alluvial sediments mixed with remnants of peat forms the unconsolidated substrate of the wetland.

Currently, peat accumulation continues at scattered locations in the central complex of wet heath, mire and wet woodland. The soil in this area is characterised as a groundwater gley, testifying to its permanent saturation. Farther out from the central wetland, the hillsides carry a combination of freely-moderately permeable brown earths and slowly draining loamy brown earths, with minor areas of freely draining podzols and impervious iron pans. Peat was present in only minor amounts in the 83 borehole logs previously recorded on the moor and was not registered in the Soil Survey description of the area. Therefore its effect upon the groundwater flow regime was likely to be insignificant although its chemical effects might be greater. The hill slopes support low intensity pasture, while the wetland margins provide common rough grazing which has decreased in intensity over recent decades.

Ground slopes in the wetland, at the centre of the catchment, are uniformly gentle at around 1% and so encourage retardation of surface water flow. This retardation is augmented by pool storage in pits left by previous tin streaming operations. Near the catchment boundary, the hillside gradients vary between 5% and 15% and so surface or near-surface runoff is rapid, providing overland flow or spring seepage onto the wetland periphery. Further rapid flow comes from the hillsides via agricultural drainage ditches. Many of these ditches join together to form the streams entering the central wetland area, although some also empty onto the wetland surface. Some rerouting of these converging wetland surface flows by road embankments occurs to the north and south of the main wetland area. However, the main obstacle to surface water flow is at the downslope end of the wetland, where a disused
railway embankment stops the further progress of all surface water except that of the river Fal. In the middle of the wetland, the main river channel joins several large pools left by gravel extraction operations. Upon leaving these ponds, the water course continues for approximately 1 km before leaving the catchment through a culvert beneath the disused embankment.

The site description given in the present chapter is particularly significant when considering the extent and design of the hydrological measurement programme on Goss Moor. Chapter 3 will use this information, combined with the project's objectives set out in Chapter 1, to delineate this programme.
CHAPTER 3

HYDROMETRIC MONITORING UNDERTAKEN ON GOSS MOOR

3.1 INTRODUCTION

As stated in Chapter 1, the aim of the present study was to identify the hydrological processes affecting the Goss Moor wetland. This involved quantifying the flows between the wetland and its surroundings together with the storage volumes in the wetland itself, and therefore required monitoring of rainfall, stream flows, evapotranspiration and groundwater levels/potentials in the wetland and on its boundaries. Aside from direct incorporation into a water budget, the stream flows would be used to assess storage characteristics in the wetland and its source areas. Appropriate accuracy of measurement was therefore an important objective. Rainfall, evapotranspiration and groundwater potentials would also be used in the numerical modelling of groundwater flows beneath the wetland. It was therefore important to choose suitable locations for the water table measurements. Also in anticipation of the requirements of the numerical model, slug tests were carried out in several locations to determine the range of permeabilities in the sediments beneath the wetland.

The installation of measuring equipment on Goss Moor began in March 1993 and continued until October 1993, by which time all components of the monitoring programme were in operation. Measurement of the catchment's river outflow began almost immediately, shortly after installation of the first river stage recorder in March 1993, while most other regular hydrometric measurements began during October 1993. The regular monitoring was concluded at the end of August 1994. The periods of measurement are given in Table 3.1.

As shown on the map of instrument locations in Figure 3.1, the hydrometric measurements were
distributed within and around the periphery of the wetland itself, rather than over the catchment as a whole. This placement allowed some quantification of the surface water and atmospheric inputs and outputs to and from the wetland itself, of variations in water storage in groundwater and of the distribution of groundwater potentials over the moor. By interrelating the short term and seasonal variations in the input variables and state variables quantified through this monitoring, the nature of the wetland’s water balance could be determined. The data would also be used in the validation of a physically-based numerical model of the wetland’s groundwater flow domain, as described in Chapter 6.

The nature of the instruments used and the periods of sampling at each instrument are given in Table 3.1. Figure 3.1 shows the locations of the instruments.

3.2 STREAM AND RIVER FLOW MEASUREMENT

3.2.1 Aims of the Stream Flow Measurement

The flow in the channel network of the Goss Moor catchment constitutes both major inputs to and, in the case of the river outflow, a major output from the wetland area. Hence, with the aim of quantifying the water budget of the wetland, high priority was given to the measurement of flow in the river and streams of Goss Moor. This measurement was achieved through the use of stage recorders and current meters. The frequency and duration of these measurements was sufficient to allow estimation of the mean magnitudes, over the durations of the dry and wet seasons of the year, of the surface channel flow components of the wetland water budget.

In the case of the river stage recorders, the high frequency of measurement ensured the possibility of a detailed analysis of the hydrographs at certain points in the channel network, permitting further quantification of the catchment and wetland water balance.
3.2.2 Locations and Frequency of Flow Monitoring

With the exception of a small outlet from a system of ponds at the western end of the catchment, surface water leaves the catchment via the river Fal. The flow of the river at this point (site C6 as shown in Figure 3.1) was monitored with a continuously recording stage gauge, thus capturing the surface outflow component of the wetland water balance. The high frequency of measurement (effectively 1 hr⁻¹ after data retrieval) given by this type of gauge also ensured that detailed analysis of the wetland/catchment outflow hydrograph could be undertaken.

Stage recorders were installed in two other, additional locations in the wetland area, known as sites C5 and C1. Site C1, at the point of entry of the river Fal into the wetland, is the greatest single surface channel input. As shown in Figure 3.2, a subcatchment of approximately 5.20 km², extending out to the overall boundary of the Goss Moor catchment, drains to this point. The hydrograph characteristics here are representative of the hydrologic response characteristics of the outer parts of the catchment, and provide a comparison with the wetland-affected response of the overall catchment outflow. In addition, the average flows at this point provided an indication of the likely seasonal flow levels coming from any ungauged subcatchments of Goss Moor which have similar topography, soils and geologic stratigraphy. Further quantification of the surface inflow component of the wetland water balance was thus achieved.

Site C5 is the major point of transfer of surface water from the northern wetland section to the main area of wetland. Although the catchment area of this gauging point, at 6.32 km², exceeds that of site C1, the presence of upstream distributaries undoubtedly reduces the effective catchment area. The flow at this point constitutes the second greatest single channel input to the main wetland area, and the major outflow of surface water from the northern section of the wetland. In a manner similar to that used on sites C6 and C1, the flows here were inspected to give the response characteristics of the northern subcatchment and to provide further seasonal quantification of the surface water input to the main wetland area.

As part of a separate research programme, flow measurements were performed by current meter in three tributaries of the river Fal, within the wetland boundaries. These measurements were
performed on an occasional basis. Their locations (points C2, C3 and C4) are shown in Figure 3.1.

3.2.3 Use of Stage Recorders for Channel Flow Measurement

The volumetric flow of water past a point in a river can be related to the water level, or stage, at that point. Such a relation, known as the stage-discharge relation, varies with the position of measurement along the channel according to the uniformity or non-uniformity of the channel geometry, roughness and slope. As a consequence of this relation, by measuring stage, the discharge may be estimated and a flow record for the river may be obtained.

For the present study, Ott R16 drum chart stage recorders were used. The construction and mode of operation of such recorders are, briefly, as follows.

A stilling well, as shown in Figure 3.3, is secured to the bank with its inlet facing out into the river water and slightly downstream to ensure no blockage by water-borne debris and no velocity head capture by the internal water column. The recorder lies on top of the stilling well. Through a float, pulley and counterweight mechanism, the movement of the pen over the drum-mounted chart corresponds to the variation in water level in the stilling well. The uniform rotation of the drum, powered by a spring-wound motor, marks the passage of time. A stage board is situated next to the stilling well to provide a reference against which to register the displacement of the pen on the stage chart and is read regularly as a check for pen slippage.

The stage-discharge relations for the gauging stations on Goss Moor were developed from measurements of flow for a range of stages at each site. A current meter was used, along with the velocity-area method, as discussed by Rantz (1982a), Dingman (1984) and Shaw (1994), in order to determine the discharge in each case. Stage is read from the stage board next to the stilling well.

The effects of random errors in the obtained values of stage and discharge, caused by rounding of the stage board reading, by turbulence around the current meter, by inaccurate placement and
orientation of the current meter and by discretisation of the flow area into finite columns (see Herschy (1971), Rantz (1982a) and Shaw (1994)), can be mitigated by the repetition of measurements on as many different occasions as possible. This approach has the added benefit of providing points over a longer stretch of the stage-discharge curve, thus reducing the need for extrapolation and allowing better estimation of the parameters in the rating equation. However, due to the limited time available for the present study (stage-discharge relations at most gauging stations are developed over many years of monitoring, whereas the monitoring on Goss Moor had, at the time of analysis, been going for only two years), the number of stage-discharge measurements obtained for the Goss Moor gauging stations was small, and thus the estimation of their stage-discharge relations was subject to a wide margin of error. Due to the small sample size, no estimate was possible for the error or bias in these relations. Figures A1 - A3, Appendix A, show the derived rating curves.

The stage-discharge equations thus derived are variants of the Manning formula,

$$Q = \frac{1}{n} R^{\frac{2}{3}} \cdot A \cdot S_0^{\frac{1}{2}}$$

where

- $Q$ is discharge ($m^3s^{-1}$),
- $R$ is the hydraulic radius of the stream ($m$),
- $A$ is the cross-sectional area of flow ($m^2$),
- $S_0$ is the channel bed slope, assumed to represent the frictional energy slope (dimensionless), and
- $n$ is the Manning roughness coefficient ($m^{1/3}s^{-1}$).

Manning's formula is widely regarded as being the best available approximation for the relation, in steady, uniform flow, between velocity, energy slope and stage in terms of the ease of practical assessment of the necessary parameters in the field. Good documentation of typical values of $n$ exists (Chow, 1959). Errors arising from its use occur when $n$ is held constant despite changing depth of flow. This neglects the associated change in relative roughness of the channel bed. However, Manning's $n$ is relatively insensitive to such changes in conditions, leaving flow estimates essentially unaffected. This provides justification for the adoption of a constant value
over the full range of observed stages at a station. The effects of viscosity at the channel bed are also neglected, but such effects are worth considering only for slow flows over fine-grained bed sediments (Dingman, 1984; Featherstone and Nalluri, 1988).

The Manning equation and its equivalents are strictly applicable only to flow conditions in which the gravitational and frictional forces are dominant and balance each other to produce a steady state. A full examination of the forces involved in stream flow reveals that this is not always the case and that such conditions are met only in steady, uniform flow. In addition to gravity and friction, a longitudinal pressure gradient along the stream axis is generally present in non-uniform flow. Imbalance in the overall force on the water in a stream is not uncommon, with the result that acceleration/inertial effects also occur. These governing factors of the motion of stream water are expressed in the Saint Venant momentum equation for 1-dimensional stream flow (Chow, 1959; Strelkoff, 1969; Dingman, 1984; Chow et al., 1988). As a consequence of these factors, the relation between stage and discharge in a particular stretch of river cannot strictly be described with a single algebraic expression, but is dependent upon the changing slope of the water surface as a flood wave passes and upon antecedent conditions in the channel. The rating curve is event-specific and takes the form of a loop with discharges higher on the rising limb of the flood hydrograph than on the falling limb and with subsidiary loops depending on the event history and depending on the presence of different water level controls such as channel constrictions, reservoirs and channel junctions (Rantz, 1982b; Chow et al., 1988).

Justification of the use of Manning-type formulae for the gauging stations on Goss Moor was provided by the observation that the response of the catchment to a spell of rain was of a gradual nature, producing only slow variation in river flow. This is due partly to the fact that rain in the area is usually of low intensity and long duration, thus loading the hydrological system only gently, and partly to the apparently large storage effect of the catchment itself. Because of the absence of abrupt changes in river flow, flood waves from the moor possess little change in water surface slope, and flow velocities vary little, with negligible involvement of inertia. Thus, the looping of the rating curve for each gauging station was assumed negligible, and a single-valued stage-discharge relation was considered to be an adequate approximation. These assumptions are strongest in the case of stations C5 and C6, where the immediate upstream
reaches slope at less than 1%, and weakest for station C1, where the source area consists of a 
predominance of steeper slopes with a greater proportion of surface runoff (see Figure 3.2). The 
rating curves at sites C6, C5 and C1 are shown and discussed further in Appendix A.

Besides the gentle response characteristics of the stream flow leaving the moor, it is worth noting 
that the paucity and inaccuracy of stage-discharge data also affects the benefit derived from 
attempting to evaluate looped rating curves for the gauging stations. With only a few, scattered 
data points, it is impossible to discern any looped structure in the rating curve. Thus, any 
adopted form of loop would constitute an uninformed guess on the character of the gauged stream.

Stage data were read directly from the collected stage charts onto computer using a digitiser. 
The random deviations in discharge caused by errors in this process of data retrieval, were 
estimated in Appendix A. These digitisation-incurred errors were not negligible and varied from 
0.003 to 0.04 m$^3$s$^{-1}$ depending on the location and on the stream flow at the time for which 
estimation was required. Such values were between 2% and 6% (extreme value: 15%) of the 
total flow in the channel. They were likely to be smaller than the error or bias of the station's 
rating formula. This led to the conclusion that the biases in flow estimation due to inaccurate 
rating curves may have been significant as a proportion of the total flow.

3.3 MEASUREMENT OF RAINFALL AND EVAPOTRANSPIRATION

3.3.1 Introduction

A detailed understanding of wetland hydrology at Goss Moor depended upon the availability of 
high resolution data on three quantities: rainfall, stream flow and evapotranspiration. In this 
thesis, the methods used in the acquisition of detailed river outflow data on Goss Moor are 
described in Section 3.2, while the current section is concerned with describing the measurement 
of the remaining two necessary quantities.
3.3.2 Measurement of Rainfall

A single tipping-bucket rain gauge giving 1 tip for 0.5 mm was installed at site M on the Goss Moor wetland in order to measure rainfall intensity. An electronic data logger recorded the cumulative number of tips occurring every 3 minutes. In the present study, these 3-minute totals were aggregated up to daily values (for the analysis of seasonal rainfall trends and the comparison of rates of rainfall variation with rates of variation of other variables in Chapter 4) and to weekly values (for input to the transient groundwater flow model described in Chapter 6).

Rainfall data were obtained in this way over the periods 6/7/93-16/7/93, 30/7/93-7/8/93, 2/9/93-7/9/93, 24/9/93-10/1/94, 12/1/94-26/1/94 and 2/2/94-16/5/94. Additional daily data for the periods not included in the Goss Moor coverage, mainly in June-September 1993 and June-November 1994, were obtained from the Meteorological Office station at St. Mawgan airbase, approximately 9 km north-west of Goss Moor and, as a back-up, from the Environment Agency South Western Region monitoring station at Roche, less than 4 km east of the moor.

Two problems affect the accuracy and applicability of rainfall data. These are:

1. the degree of equivalence between the amount of rain caught by the gauge and the amount falling on an equal area of horizontal terrain, and
2. the representativeness offered by the rainfall rate gauged at a few isolated and displaced points with respect to the average rate over the study catchment.

The first problem, that of the equivalence between gauge catch and ground interception, is affected by the location of the gauge. Since rain drops falling to the ground can be deflected by air movement, it has been found that the nearby presence of bluff structures such as trees and buildings can cause enough distortion of the wind field to produce a bias in the amount of rain caught by the gauge. This bias may be either positive or negative. In addressing this problem, the Great Britain Meteorological Office (1975) suggest that "nearby obstacles should not subtend an angle of greater than 30 degrees to the [horizontal] gauge orifice" (Sumner, 1988, p.287). The rain gauge on Goss Moor was situated (see Figure 3.1) on a flat area of wet heath, far enough away from any trees or buildings to satisfy the above guideline.
In a similar manner, the structure of the gauge itself may cause enough air turbulence to deflect rain drops falling near the gauge funnel. The gauge on Goss Moor was simply placed on the ground surface, presenting a 0.3 m high obstacle to wind movement. In a study of the effects of exposure on the catch of rain gauges, Green (1969) finds that the catch obtained with such placement on short grazed grass was up to 8% less than the true ground interception, given similar wind speeds to those found on Goss Moor. Hence, although the situation on Goss Moor differed in that 0.5 m high *Juncus* reed was prevalent within a few metres of the gauge, a possible shortfall of a few percent in the gauge catch could be expected. Similar conclusions with respect to unsheltered gauges are reached by Andersson (1963) and by Robinson and Rodda (1969).

However, such discrepancies are unimportant in comparison with the problem of areal unrepresentativeness from the point sample (problem 2). This problem is caused by the spatial variability of rainfall during the storm. During, for example, the approach of a cyclonic warm front, a band of quite uniform, low-moderate intensity precipitation covering an area some hundreds of square kilometres may pass over (Browning et al., 1973; Sumner, 1988). The spatial and temporal variations of rainfall intensity during the passage of such a feature are gradual, with the result that simultaneous measurements of intensity separated by a few kilometres exhibit little discrepancy. However, rainfall in the mid-latitudes also occurs in discrete, higher-intensity cells covering areas of 5 to 20 km² and separated by distances comparable to their size (Austin and Houze, 1972; Harrold, 1973). These may be generated in cyclonic fronts (see Browning et al., 1973, and Harrold, 1973) and also, particularly, as a result of small-scale convection over land or ocean.

This discrete spatial organisation on the sub-meso scale gives rise to significant variation in simultaneous rainfall, even within catchments as small as Goss Moor. Berndtsson et al. (1994) report a cell-shaped rainfall structure, covering an area of a few square kilometres, detected by a dense rain gauge network with a temporal resolution of 1 minute, in Lund (Sweden). The spatio-temporal distribution of rainfall can be quantified using a technique known as "correlation analysis", in which the measured rainfall rates at different locations and time lags are cross-correlated to show the spatial scale, velocity and direction of movement of the rainfall pattern.
Felgate and Read (1975), Marshall (1980) and Shaw (1983) employ techniques such as this in the analysis of 2-minute data from convective storms, finding cell sizes similar to those found by Berndtsson et al. The single case of frontal rainfall analysed by Shaw (1983) exhibits greater areal uniformity than the convective cases, and yet the correlations within the 2-minute rainfall fields still fall off to values of around 0.5 at separations of 3 km. This illustrates the non-synchronous nature of short-interval activity, even for frontal rainfall measured at a small scale.

This non-synchronous behaviour becomes less apparent as time averaging is introduced, such as when considering daily rainfall accumulations instead of hourly totals. For instance, the correlation coefficient of daily rainfall depths was found to remain above 0.7 at 6 km for short-lived, slowly moving convective rain cells in northern Tunisia (Berndtsson and Niemczynowicz, 1986). The frontal rainfall affecting Goss Moor probably exhibited even greater aggregation than this over an interval of one day. Whereas the rain cells in the above Tunisian study were relatively stationary, many frontal cells may move in succession over Goss Moor in one day. At the average wind speed of 500 km/day, the 9 km between St. Mawgan and Goss Moor is traversed over 50 times in one day by the clouds in a typical frontal system, assuming that low level wind speed = rain cell travel speed. This suggests that the distance from Goss Moor to the St. Mawgan rain gauge or to the nearer Roche site was unlikely to have caused significant loss of correlation between the required daily totals of the local rain gauge catches.

However, this did not preclude the possibility of an increasing or decreasing trend in rainfall totals along the direction of the storm tracks due to orographic effects. Together with the possibility of bias in the rain gauge response at one or more of the monitoring sites, this suggested the need to examine the relation between those data recorded at Goss Moor and those obtained from the surrogate monitoring sites.

In Figures 3.4 and 3.5, daily rainfall totals for the Goss Moor site are plotted against corresponding data at St. Mawgan and at Roche for the amalgamated periods 6/7/93-16/7/93, 30/7/93-7/8/93, 2/9/93-7/9/93, 24/9/93-10/1/94, 12/1/94-26/1/94 and 2/2/94-16/5/94 during which Goss Moor rainfall was available. The coefficient of determination, \( R^2 \), was 0.52 for St. Mawgan and 0.52 for Roche. Both values correspond to a Pearson correlation coefficient of
Thus, Goss Moor daily rainfall totals were correlated to both St. Mawgan and Roche daily rainfall totals to the degree suggested by the findings of Berndtsson and Niemczynowicz, 1986. Goss Moor rainfall in metres was found to be approximately 0.014 m + 1.01 × St. Mawgan rainfall (m), and a value of 1.00 fell within the (two-tailed) 95% confidence interval for the gradient. On the other hand Goss Moor rainfall was 0.0015 + 0.70 × Roche rainfall, the gradient being much lower than 1.00, even at the (two-tailed) 95% confidence level. Therefore, annual averages of the St. Mawgan data, such as over the water budget period of the present study, 1/9/93-31/8/94, were likely to reflect accurately the annual rainfall input to the wetland, whereas the annual average of the Roche data would overestimate input at Goss Moor and would have to be scaled by a corrective factor of 0.70 before adoption for the wetland. The correspondence between rainfall at Roche and Goss Moor would seem to have suffered from the above-mentioned orographic effects or instrument bias. However, only St. Mawgan rainfall data was used for the water budget. No scale factor was introduced before the adoption of this data.

At the weekly resolution necessary for input to the groundwater simulations in Chapter 6, the errors due to the distance of the St. Mawgan monitoring site from the study area were assumed to be acceptable.

3.3.3 Measurement and Calculation of Evapotranspiration

**Background: the thermodynamics of evaporation:**

Evaporation or condensation at a liquid-gas interface is driven by the tendency for the liquid-gas system to increase its entropy (McClelland, 1973; Atkins, 1988). In a fully-mixed, fixed mass, isothermal system at constant pressure, initially deficient in vapour, evaporation will occur until the increase in entropy associated with the conversion of liquid to gas has been balanced by the mechanical work done by the expansion of the gas. When this balance occurs, the system has reached saturation point.

In normal atmospheric conditions, such an equilibrium is rarely reached over any appreciable depth of atmosphere, due to continual changes in temperature and pressure, removal or addition of vapour by wind and the input of solar heat energy. Rather, a local equilibrium is assumed to
exist continuously in a very thin layer of saturation at the phase interface (Merlivat and Coantic, 1975). This zone of saturation fluctuates in thickness, and condensation or evaporation continuously occurs, as the layer responds to externally-subjected perturbations in vapour mass, temperature and pressure, by continuously augmenting entropy.

**Aerodynamic and calorimetric measurement of evapotranspiration (ET):**

The most common situation in the hydrological cycle is for evaporation, rather than condensation, to occur at a water surface. Thus, convective and dispersive aerodynamic motions transport vapour away from the zone of saturation, which is replenished by evaporation at the liquid water surface, a process enhanced by solar heating of the liquid. The determination of the evaporative flux from a water surface can thus be attempted by measurement of the vertical flux of water vapour in the atmosphere just above the surface, and also by measurement of the energy balance of the water body to obtain the rate of energy loss through evaporation, which is related to the evaporative water flux through the latent heat of vaporisation.

**The aerodynamic approach:**

A common way of considering the aerodynamic transport of momentum, heat or vapour is through the dimensional analysis of the quantities involved to derive semi-empirical relations which can readily be solved given appropriate measurements. In accordance with the mixing length theory of Prandtl (1932), the wind velocity in the lower part of the atmospheric boundary layer increases logarithmically with height above the ground level (Thom, 1975; Brutsaert, 1982). Above a vegetation canopy, this logarithmic increase begins at the level $d + z_0$ above the ground, where:

- $d$ is the zero plane displacement (m), and
- $z_0$ is the roughness length of the canopy (m).

Thus, the wind velocity above the canopy, $u(z)$ (m), is given by

$$u(z) = \frac{u_*}{k} \cdot \ln\left(\frac{z-d}{z_0}\right)$$  \hspace{1cm} (3.2)
In the above,

- $z$ is the height above ground (m),
- $u^*$ is the "eddy velocity" (m s$^{-1}$), and
- $k$ is von Karman's constant (= 0.4, dimensionless).

This logarithmic wind profile strictly applies only to conditions of neutral stability in the boundary layer.

Equivalently, there is a downward diffusive flux of momentum, $\tau$ (kg m$^{-1}$ s$^{-2}$), given by

$$\tau = \rho \cdot k \cdot u^* \cdot (z - d) \cdot \frac{\partial u}{\partial z} = \rho \cdot u^2$$  \hspace{1cm} (3.3)$$

where:

- $\rho$ is the mass density of the air (kg m$^{-3}$).

Hence, a uniform momentum flux exists in this part of the boundary layer. By the "similarity hypothesis", other entrained quantities such as heat and water vapour are subject to the same diffusive fluxes. Using Fick's law for any one of these quantities, such a flux can be related to a gradient in the concentration of the quantity, and hence to a difference in concentration between levels above the canopy. This permits the evapotranspiration flux above a canopy to be determined using measurements of humidity at two or more levels, and using resistance terms which are dependent on wind speed and plant physiology.

**The energy balance approach:**

Thom (1975) and Brutsaert (1982) detail the evaluation of the net energy fluxes through a plant community. The total vertical heat flux in W m$^{-2}$,

$$H = C + \lambda \cdot E$$  \hspace{1cm} (3.4)$$

where
C is the sensible heat flux (W m\(^2\)),
E is the evaporative moisture flux (kg m\(^2\) s\(^{-1}\)), and
\(\lambda\) is the latent heat of evaporation of water (J kg\(^{-1}\)).

This flux is part of an overall energy balance:

\[
H = R_n - D - G - J - A
\]  

(3.5)

where

- \(R_n\) is the net radiative input to the community (W m\(^2\)),
- \(D\) is the net horizontal diverging rate of sensible and latent heat (W m\(^2\)),
- \(G\) is the heat flux into the ground (W m\(^2\)),
- \(J\) is the heat flux absorbed into physical storage (W m\(^2\)), and
- \(A\) is the energy flux absorbed into biochemical storage (W m\(^2\)).

After accounting for all other components of the energy balance, the residual energy flux is often partitioned between \(E\) and \(C\) using the Bowen ratio, \(B_o\) (dimensionless), defined as the height-invariant ratio between \(E\) and \(C\). The assumption of height-invariance in \(B_o\) relies upon the similarity hypothesis for the vertical turbulent transport of heat and water vapour, and upon the assumption of similar concentration profiles. This assumption should be treated with caution, since the profiles above the canopy may be affected by dissimilar vertical distributions of sources or sinks within the plant community, although Cellier and Brunet (1992), among others, find it to be valid in certain circumstances. Hydrological studies using the energy balance methods include Smid (1975) and Lindroth et al. (1994).

In the present study, it was assumed that \(H = R_n\) due to the relatively small magnitudes of \(D, G, J\) and \(A\). The discrepancy arising from this assumption is likely to be outweighed by errors in measurement of \(R_n\) and other quantities.

The Penman-Monteith equation:

Penman (1948) considers potential evaporation from an open water surface. By combining
aerodynamic and calorimetric approaches, he produces an expression for $E$ in which the
measurement of temperature, humidity and wind speed at two levels is no longer necessary.
Implicit in the Penman equation is a bulk aerodynamic resistance to the transfer of water vapour
from the evaporating surface through the Prandtl layer. Since the equation applies to open water
bodies, no account is made for extra resistances such as that encountered in evaporation from
vegetation.

However, Monteith (1964) introduces a bulk physiological resistance term for the diffusion of
water vapour from the sub-stomatal cavities to the aerodynamic surfaces of a vegetation stand.
The resulting Penman-Monteith equation is applicable to evapotranspiration. A somewhat more
rigorous variant of this equation, derived by Thom (1972), introduces a more accurate
physiological resistance term, known as the "bulk stomatal resistance", which is now widely used
by workers in hydrology (for example: Great Britain Meteorological Office, 1983; Lafleur and
Rouse, 1990; Wigmosta et al., 1994; Lindroth et al., 1994). Thom (1972) concludes that the
evapotranspirative water flux over a vegetation stand can be expressed as:

$$\lambda \cdot E = \frac{\Delta \cdot H + \frac{\rho \cdot C_p}{r_{av}}(e^*_a - e_a)}{\Delta + \gamma \cdot \left(1 + \frac{r_{st}}{r_{av}}\right)}$$

(3.6)

where

- $\Delta$ is the rate of change of saturation vapour pressure with respect to temperature at
  the ambient air temperature (Pa K$^{-1}$),
- $C_p$ is the specific heat capacity of air at constant pressure (J kg$^{-1}$ K$^{-1}$),
- $r_{av}$ is the bulk aerodynamic resistance to the turbulent transfer of water vapour (s m$^{-1}$),
- $r_{st}$ is the bulk stomatal resistance to the diffusion of water vapour (s m$^{-1}$),
- $e^*_a$ is the saturation vapour pressure at the ambient air temperature (Pa),
- $e_a$ is the actual vapour pressure in the air (Pa), and
- $\gamma$ is the psychrometric constant (Pa K$^{-1}$).
Further expressions are required for $\Delta$, in terms of the ambient air temperature (for example, Richards, 1971) and for $r_{av}$, in terms of the wind speed, the roughness length of the plant canopy and the zero plane displacement. Conditions of either static stability, neutrally static stability or instability in the atmospheric boundary layer may be assumed in the formulation of $r_{av}$. Penman (1948) developed an empirical relation between evaporation and wind speed for unstable conditions. Whilst giving satisfactory accuracy when employed in the correct context (Slatyer and McIlroy, 1961; Thom and Oliver, 1977), this relation and its equivalent expression for $r_{av}$ apply only to open water or short grass surfaces. The next subsection of this thesis briefly mentions the approach taken in the present study to extend the applicability of $r_{av}$ to other surfaces.

For transpiring vegetation, factors such as the stage in the growth cycle and the availability of water to the roots affect the status of its stomata. Such factors can be accounted for by adjustment of $r_{st}$ in the Penman-Monteith equation. Thus, although developed from Penman’s initial consideration of potential evaporation, the evapotranspiration equation addresses actual, rather than potential, conditions.

An automatic weather station for the determination of ET by the Penman-Monteith method was installed at site M on Goss Moor. This was situated over a stand of *Juncus* rushes about 0.5 m in height, and remained there for the entirety of the measurement programme. As shown on Table 3.1, the apparatus consisted of a net radiometer, dry and wet bulb thermistors and an anemometer, monitored by an electronic data logger. Data from an accompanying rain gauge were also used in analysing the data from the other instruments.

**Analysis of the evapotranspiration data:**

As mentioned above, the Penman-Monteith equation can be used alone with appropriately specified stomatal resistances in order to calculate actual evapotranspiration. However, the variation of $r_{st}$ with plant moisture stress is not well documented and so this method of calculation of actual ET would necessitate experimental work on plant responses to environmental changes which would be beyond the scope of the present study. Instead of adjusting $r_{st}$ according to both soil moisture conditions and season, a more usual approach is to
apply only seasonal adjustments so that potential ET is calculated initially, and then to derive actual evapotranspiration from this on a daily basis using an empirical relation accounting for rainfall and soil moisture status. This procedure was followed in the present study. The current subsection describes the treatment of wind drying, intercepted rain, spatial variations in plant stand characteristics, understorey evapotranspiration and shading, all of which involved modifications to the calculation of potential ET rates before addressing the effect of soil moisture status in the calculation of actual from potential ET. The final part of this subsection explains the use of the Penman-Grindley formula in the latter calculation.

The aerodynamic resistance to vapour transfer from the evaporative surface was determined empirically by Penman (1948) for an unstable boundary layer over an open water or short grass surface. Thom and Oliver (1977) modified Penman’s empirical wind function in order to apply it to evaporating surfaces with different roughnesses. Penman’s wind function addresses conditions of partially forced convection (that is, statically unstable conditions) in which vertical transport by turbulence is augmented by buoyancy and the wind speed profile is not precisely logarithmic. To account for such conditions, the equivalent theoretical equations also augment vertical transport for any given wind speed gradient. This is done non-linearly with respect to the turbulence-derived flux since greater turbulent displacements receive disproportionately larger buoyant enhancements. As noted by Thom and Oliver (1977), increasing roughness lengths tend to reduce the vertical temperature gradient and thus decrease the role of buoyancy in vertical transport at a given measured wind speed. Thus, while the turbulence structure in the boundary layer then allows greater vertical transfer of momentum and moisture, the non-linear augmentation of the vertical fluxes by buoyancy is reduced. Thom and Oliver (1977) found theoretically that buoyancy thus acquires an approximately linear role in the augmentation of vertical fluxes resulting from a growing roughness length. They concluded that the dependence of the net aerodynamic resistance upon the roughness length is approximately the same for partially forced convection (statically unstable flows) as it is for fully forced convection ( neutrally stable flows). Consequently,

\[
\frac{r_{a,z_0}}{r_{a,z_0}} = \frac{r_{oN,z_0}}{r_{oN,z_0}}
\]  

(3.7)
where

\[ r_{a,z0} \]

is the aerodynamic resistance (s m\(^{-1}\)) to vapour transfer under unstable flows, with roughness length \( z_{01} \) (m),

\[ r_{a,z2} \]

is the aerodynamic resistance (s m\(^{-1}\)) to vapour transfer under unstable flows, with roughness length \( z_{02} \) (m),

\[ r_{a,n,z0} \]

is the aerodynamic resistance (s m\(^{-1}\)) to vapour transfer under neutrally stable flows, with roughness length \( z_{01} \) (m), and

\[ r_{a,n,z2} \]

is the aerodynamic resistance (s m\(^{-1}\)) to vapour transfer under neutrally stable flows, with roughness length \( z_{02} \) (m).

Given ground-to-atmosphere sensible heat fluxes within the range 20% to 200% of the average summer value of 50 W m\(^{-2}\), this approximation was found to be valid for wind speeds above 0.5 m s\(^{-1}\) (Thom and Oliver, 1977). Setting \( z_{01} \) equal to Penman's roughness length, then the aerodynamic resistance for \( z_{02} \) in unstable conditions could be obtained by introducing Penman's resistance into Equation 3.7 and rearranging, giving

\[
\frac{4.72 \cdot \left( \ln \left( \frac{z}{z_0} \right) \right)^2}{\left( 1 + 0.54 \cdot u \right)}
\]  

(3.8)

This expression was used in the calculations of evapotranspiration in the present study.

The Penman-Monteith equation was used to calculate daily potential evapotranspiration totals from daily averages of the measured quantities. The FORTRAN 77 program used for these calculations is listed in Appendix B. This program alters the effective bulk stomatal resistance depending upon the occurrence of rainfall on the day in question: a value of zero is assumed for days of rain, since, as suggested by Thom and Oliver (1977), leaf wetting by rain produces a zone of saturation on the leaf surfaces rather than in the stomatal pores. Thus the governing resistance becomes the vegetation canopy's aerodynamic resistance alone. Lindroth et al. (1994) adopt the same approach in modelling the water-use efficiency of willow. They state (p.6) that, "Transpiration and evaporation of intercepted water were assumed to be mutually exclusive," although they qualify this with the statement (p.6), "In reality, intercepted water does not inhibit transpiration (Larsson, 1981), but with a model time step of a full daytime period, it was an acceptable approximation." Wigmosta et al. (1994) adopt a similar approach.

For the purposes of calculating evapotranspiration, the land cover on Goss Moor was divided
into 4 classes:

i) open water,
ii) willow scrub,
iii) wet heath, and
iv) pasture.

These classes were based upon a Geographical Information System database of the wetland vegetation and are related to the assessment of site vegetation in Section 2.3.3. Since the measurements were taken in only one location on the moor, and over only one stand of vegetation, some method was required to synthesise simultaneous evapotranspiration rates for land cover types other than the *Juncus* rush of the weather station site. The approach taken here was to assume that the profiles of wind velocity, temperature and humidity remained identical over all surfaces. Hence the wind speeds, temperatures and humidities obtained at a level of 2 m above ground, over the 0.5 m high *Juncus* rush (wet heath) at the measurement site were assumed to apply also to the same level over open water and pasture. The same wind speeds were taken to be present at about 1.5 m above the 4 m canopy of willow scrub. Net radiation was taken to be similarly invariant. However, a different roughness length and stomatal resistance was assigned to each of the vegetation types, according to findings in the literature. The values and references used are listed in Table 3.2.

Limitations in the above approach could be expected due to the dependence of net radiation, temperature, humidity and vertical wind speed profile on the characteristics of each vegetation stand. Ideally, these variables should have been measured over each vegetation type, but this would have required more equipment than was available in the present study. A survey of the literature reveals that observational studies of evapotranspiration from individual stands in a patterned canopy are rare, leaving a lacuna in the research to date. Recent modelling studies vary in their treatment of spatial variation of such quantities, again due to a lack of data. For example, the sophisticated distributed hydrology-vegetation model of Wigmosta *et al.* (1994) was tested with a uniform wind speed over a topographically and vegetationally complex catchment. Compensation may have been made for the constant wind speed by adjusting the height above zero plane displacement to which it applied. Actual vapour pressure was adjusted.
for elevation above sea level, but not for the local canopy or soil moisture. An investigation by Band (1993) into the effect of grid scale on modelled hydrological and carbon budgets, used Thematic Mapper imagery to represent leaf area indices at a sub-stand scale on a detailed representation of catchment topography, but was unable to incorporate distributed wind speed or temperature into the high resolution model. Net radiation and vapour pressure deficit were, however, varied between individual hill slopes, at a scale greater than that of the vegetation pattern.

Closer investigation at Goss Moor is beyond the scope of the present study. However, some indication of the nature of the problem is available from the literature. With respect to variations in net radiation, Kessler and Jaeger (1999) compared radiation fluxes above a pine forest and a grass surface in Germany. Although the two vegetation stands were not at precisely the same latitude, compensation was made for this by normalising the fluxes with respect to the locally measured global radiation (defined as the direct and scattered solar radiation in cloudless or cloudy sky, and dependent on latitude - see Monteith and Unsworth, 1990). The short wave albedo and the normalised upward long-wave emission were significantly smaller over the pine forest than over the grass surface. As a result, the normalised net radiation for the pine forest was approximately 1.5 times that for the grass surface, thus allowing greater latent and sensible heat fluxes. Similar differences might be found between the willow and wet heath areas on Goss Moor and might possibly augment the higher evapotranspiration rates of the willow already caused by its lower aerodynamic and bulk stomatal resistances. However, the presence of surface water beneath the canopies on Goss Moor complicates the comparison because of its low albedo and high specific heat capacity (Boudreau and Rouse, 1995).

Variations in temperature and humidity between neighbouring stands of contrasting vegetation are poorly represented in the literature. It can be said that both variables will be affected by the above differences in net radiation and by the differing canopy heat capacities (Kessler and Jaeger, 1999). The greater heat capacity of the willow canopy is likely to depress its daytime temperature, but increase its night-time temperature in comparison with that of the wet heath. The effect of this might be, on the one hand, to reduce willow evapotranspiration through a lower vapour pressure deficit during the period when the stomata are open. However, the
nonlinearity of black body radiation implies that, because temperatures in both canopies are lowest at night, the long-wave radiation losses of the willow canopy are generally smaller than those of the heath (e.g., Kessler and Jaeger, 1999). This again would leave more energy available for ET or sensible heat loss in the willow canopy.

The temperature and humidity are also strongly affected by advection, and thus by inter-canopy variation in wind speed. Moreover, increasing wind speeds augment evaporation (see Equation 3.8). As mentioned above, wind speed was measured at only one location on Goss Moor - 1.5 m above the canopy of a stand of 0.5 m high *Juncus* rush. Since it was necessary to use the same data in willow areas, it was ensured that this wind speed was assigned to the same height with respect to the top of the willow canopy.

In reality, the wind speed and associated turbulent vertical fluxes of moisture and heat do not follow such simple rules. In particular, the wind field downstream of a discontinuity in roughness, such as the edge of a stand of willow, may take some hundreds of metres to reach a new equilibrium. The vertical wind speed gradient and eddy velocity must both adapt to the new surface conditions (Brutsaert, 1982). This is the subject of much ongoing theoretical and practical research for the purpose of estimating the appropriate distance or "fetch" required between evapotranspiration measuring instruments and the upstream edge of the canopy (Horst and Weil, 1994, 1995; Hsieh *et al*., 1997). van Breugel *et al.* (1999) analysed vertical wind speed gradients at up to 60 m above ground in a mixed forest in the Netherlands. Two measurement towers were used, both approximately 150 m from the edge of the forest. The vertical wind speed gradient normalised with respect to wind speed was analysed as an indicator of local turbulence, for different wind directions. It was found to decrease when the wind came from the direction of areas with a higher forest canopy, and increased in wind from the forest edge, indicating that adjustments of the wind field were still occurring at this height.

At lower heights, the wind field reaches equilibrium within a shorter fetch. Consequently, fetch requirements are often expressed in terms of a minimum fetch:height ratio for which values of between 16 and 200 have been quoted (Blanken and Rouse, 1995; van Breugel *et al*., 1999). Taking a moderate value of 100 for this ratio, the required fetch for the 2 m wind speed
measurements on Goss Moor would be 200 m. From Figure 3.1, it can be seen that this was unlikely to have been satisfied in the predominant south-westerly wind. The wind profile over the Juncus stand in which the weather station was situated probably remained somewhat similar to that over the neighbouring willow canopy. Evapotranspiration from the wet heath at the measurement site was probably thus slightly reduced in comparison with that from more open areas of wet heath. Humidity and temperature were measured at about 1.5 m above the ground, and so were probably less affected by the same sheltering effects. Finally, such considerations of local advection near changes in vegetation suggest that the spatial variations in wind speed, temperature and humidity over Goss Moor do not follow precisely the boundaries of the vegetation zones, leading to further uncertainty in the calculated evapotranspiration. This is one reason why changes between the zoned evapotranspiration rates may not have such distinct boundaries in reality as in the model adopted in the present study (see Figure 6.18). In conclusion, the overall bias in ET estimates for each land cover type due to adoption of a spatially uniform micrometeorology is affected by many factors, and cannot be quantified in the present study. However, the most accurate parameterisation was likely to be that over the wet heath areas of the wetland, where the monitoring station was located.

Lindroth et al. (1994) consider the evaporation of water from the soil beneath the tree canopy. This is assumed to occur at the potential rate, an assumption appropriate for the greater part of Goss Moor, where the ground beneath the willow carr and the wet heath is often waterlogged; a sheltered water surface usually exists below the canopy. Evaporation from such a water surface is distinguished from open water evaporation by the effects of the sheltering on both wind and insolation. In the present study, the wind speed beneath the canopy was taken to be zero. Using this assumption, conditions of equilibrium evaporation can be inferred, since without wind, the air above the water surface may eventually reach saturation. Penman's equation for open water evaporation then reduces to a form now known as the Priestley-Taylor equation (Brutsaert, 1982). This was used in the Goss Moor analysis. Evaporation thus determined depends upon only the available energy and the temperature of the system.

Research on sheltering from sunlight by plant canopies has been reviewed by Ross (1975). As sunlight filters through the canopy towards the ground, its intensity is progressively attenuated
according to an exponential relationship with the number of layers of foliage encountered. This
results in an exponential fall-off with both depth and leaf density. The present analysis used an
exponential decay function given by Lindroth et al., depending on total leaf area index and bark
area index, to describe the net radiation for the water surface beneath the plants of Goss Moor.
As a result, the values of available energy used in the calculation of evaporation from these water
bodies were appropriately reduced. Wigmosta et al. (1994) develop a radiation budget for a
more comprehensive vegetation water use model, according to similar physical principles.

Thus, the potential evapotranspirative fluxes from the different land cover types on the moor
were calculated as follows. For any open water bodies, the original formulation of Penman's
equation was used. For both willow scrub and wet heath, the Penman-Monteith equation (with
a modified aerodynamic resistance) was used, along with the Priestley-Taylor formula (with
exponentially reduced net radiation) to account for evaporation from the underlying water
surface. For pasture, the Penman-Monteith equation was used alone. The FORTRAN 77
program code in Appendix B illustrates this scheme. Table 3.2 presents the values of parameters
contributing to the distinction between different vegetation types in this calculation. The
assumed seasonal variations in the leaf area indices and bark area indices are included.

Actual ET was assumed equal to potential ET for the wetland areas in the centre of the
catchment, since these areas were assumed to suffer no soil moisture deficit. However, areas of
pasture were generally on steeper slopes than the wetland and were relatively well drained,
suggesting that deficits in soil moisture would be possible, causing reductions of actual
evapotranspiration from the potential rate. In order to account for this, a soil moisture
accounting procedure was applied to the pasture areas. This procedure was combined with an
empirical relation between the soil moisture deficit, the rainfall and the fractional attenuation of
the evapotranspiration rate, in order to estimate actual evapotranspiration from the pasture.

The particular soil moisture accounting technique used in the present study, known as the
Penman-Grindley model, is described in Lerner et al. (1990), who draw attention to the
inherently simplistic and empirical nature of such methods and their neglect of processes such as
macropore flow and the spatially varying evolution of moisture content within the soil column.
Nevertheless, the general approach of soil moisture accounting is widely employed in hydrological and climatological simulation and prediction. For example, Houston (1982) successfully used the Penman-Grindley model to estimate monthly groundwater recharge in a semi-arid area of Zambia. Rushton and Ward (1979) used a modification of the Penman-Grindley method to estimate groundwater recharge with daily time steps in Lincolnshire. While they found no need to change the direct treatment of actual evapotranspiration, they suggested modifications to the model's criteria for recharge which would indirectly affect the estimated ET losses. This problem is considered below.

In the present study, the soil moisture accounting calculations proceeded in daily time steps using the Penman-Grindley formula:

\[ psmd_{i+1} = smd_i + ae_i - p_i \]

\[ r_i = -psmd_{i+1} \text{ when } psmd_{i+1} < 0 \]  \hspace{1cm} (3.9)

\[ smd_{i+1} = psmd_{i+1} + r_i \]

Actual ET was derived from potential ET as follows:

\[ ae_i = pe_i \text{ when } smd_i < C \text{ or } p_i \geq pe_i \]

\[ ae_i = F \cdot pe_i + (1 - F) \cdot p_i \text{ when } D > smd_i \geq C \text{ and } p_i < pe_i \]

\[ ae_i = p_i \text{ when } smd_i \geq D \text{ and } p_i < pe_i \]  \hspace{1cm} (3.10)

where

- \( smd_i \) is the soil moisture deficit below field capacity at start of day \( i \) (m),
- \( ae_i \) is the actual evapotranspiration during day \( i \) (m),
- \( pe_i \) is the potential evapotranspiration during day \( i \) (m),
- \( p_i \) is the precipitation during day \( i \) (m),
- \( r_i \) is the recharge during day \( i \) (m),
- \( psmd_i \) is an intermediate variable (m),
- \( C \) is the root constant (m),
\( \text{D} \) is the wilting point (m), and \\
\( \text{F} \) is an empirical constant between 0 and 1 (dimensionless).

Certain features of this procedure are noteworthy. Firstly, hillslope runoff is not explicitly represented in the formulation as used here, although \( \text{r}_{\text{i}} \) might be considered to include the runoff (Rushton and Ward, 1979). In this way, runoff would be considered to be non-Hortonian, that is, not governed by any threshold in infiltration rate. The absence of a scaling factor between \( \text{r}_{\text{i}} \) and \( \text{psmd}_{\text{i}} \) limits the applicability of the formula in this respect.

Secondly, the treatment of \( \text{ae}_{\text{i}} \) for \( \text{D} > \text{smd}_{\text{i}} \geq \text{C} \) and \( \text{p}_{\text{i}} < \text{pe}_{\text{i}} \) implies that:

- \( \text{ae}_{\text{i}} \) never rises above \( \text{pe}_{\text{i}} \)
- \( \text{ae}_{\text{i}} \) never falls below \( \text{p}_{\text{i}} \) and thus the vegetation will take up all available rain water in addition to some of the resident soil moisture when the soil moisture is insufficient to supply potential evapotranspiration and the rain constitutes insufficient supply for potential evapotranspiration.

The above behaviour of \( \text{ae}_{\text{i}} \) implies that the Penman-Grindley model does not consider the possibility that sufficient supply for potential evapotranspiration may be constituted by rain together with stored soil moisture. This is likely to be important during times when the soil moisture deficit is in transition between values below and above the root constant \( \text{C} \).

Thirdly, the percolation or groundwater recharge, \( \text{r}_{\text{i}} \) becomes non-zero only when \( \text{psmd}_{\text{i}} \) is less than zero, that is when the soil water store is at or above field capacity. This has been shown to cause significant underestimation of groundwater recharge (Rushton and Ward, 1979) and consequently, can be expected to contribute to an overestimation of actual evapotranspiration and the amount of time for which evaporation occurs at the potential rate. Such ET overestimation was also found by Karongo and Sharma (1997) in four humid catchments in Kenya. If accompanied by underestimation of recharge, this may sometimes be corrected by reducing the specified value of \( \text{C} \), the root constant. However, Rushton and Ward (1979) found that the optimum value of \( \text{C} \) increased with rainfall whereas it should be independent (being an ET-governing parameter), suggesting that the direct handling of recharge, rather than the treatment of ET, should be altered.
Fourthly, \( r_j \) is assumed never to be limited by a high and perhaps rising water table. If such limits on \( r_j \) were to be represented and the water table were sufficiently high, then a soil moisture distribution would be implied which was either in equilibrium or exhibiting upward head gradients. The soil water store would then be at or above field capacity. Potential ET would then apply. In neglecting this possibility, the Penman-Grindley model underestimates actual ET for high water tables, corresponding to part of the time when \( \text{psmd}_i \) is small or negative. However, this error may on occasion offset the error of recharge underestimation mentioned above. Furthermore, a value of \( r_i \) which is greater than the groundwater recharge may be appropriate in those instances when it is required also to account for runoff.

Assessing the combined effects of all the above model characteristics on the accuracy of the evapotranspiration estimate, it would appear that a great deal of interaction is possible between the various potential misrepresentations. This highlights the empirical, quasi-physical nature of the model and also indicates that much freedom may be found in the calibration of its parameters. Other, slightly different quasi-physical techniques such as that of Thornthwaite and Mather (1955), and Kachroo (1992) would also suffer from such uncertainties.

The Penman-Grindley procedure, as formulated above, was applied over the total study period, beginning at 1/6/93. In this way, there was sufficient time interval between the beginning of the simulation and the beginning of the water budget period to remove the effect of the initial condition of soil moisture (which had been arbitrarily set to zero). A problem for the application of the Penman-Grindley procedure in the present study was the lack of water table and directly measured ET data with which to calibrate the model parameters on the pasture slopes. Therefore, as suggested by Lerner et al. (1990), the empirical constant \( F \) was given a value of 0.10 and the root constant \( C \) a value of 0.076 m, suitable for permanent grass. Wilting point \( D \) was set at 0.095 m using figures provided by MAFF (1988). The final result of the soil moisture accounting procedure was that rainfall remained too frequent throughout the year to allow any appreciable deficit of soil moisture in the pasture. Potential rates of evapotranspiration therefore applied over the entire catchment for the duration of the study period.
Data input and results:
A complete set of meteorological data were obtained on the wetland for a continuous period from 12th January 1994 to 16th May 1994. In addition, other periods for which the data set was incomplete or absent were filled in with daily meteorological data from the nearby R.A.F. St. Mawgan airfield, to provide coverage from 1st June 1993 to 30th November 1994.

The missing variables were synthesised from the St. Mawgan data using linear regression. The St. Mawgan data used for this purpose were:
• mean daily temperature,
• daily hours of sunshine, and
• mean daily wind speed.

Daily rainfall totals at St. Mawgan were also used to fill the gaps in the on-site data. The correlation between these data and the available Goss Moor data is assessed in Section 3.3.2.

The other St. Mawgan data were used in the synthesis of the following variables.

• Net radiation was first estimated using astronomical information and the daily hours of sunshine recorded at St. Mawgan. The astronomical calculations can be seen in FORTRAN program PENMONT5.FOR, listed in Appendix B. Following this, the Goss Moor measurements were regressed on the St. Mawgan estimates. Figure 3.6 shows this regression (coefficient of determination, $R^2 = 0.90$). The regression equation was then used to convert St. Mawgan estimates into Goss Moor estimates for periods when net radiation measurements were unavailable on Goss Moor.

• Daily mean wind speed measured on Goss Moor was regressed on that measured at St. Mawgan. Figure 3.7 shows this regression ($R^2 = 0.92$). The regression equation was then used to estimate Goss Moor wind speed from St. Mawgan wind speed when the former was not measured.

• Mean air temperature measured at St. Mawgan was substituted directly for daily Goss Moor air temperature when the latter was unavailable. Figure 3.8 shows the relation between temperature at the two locations via a linear regression line ($R^2 = 0.87$). Although the gradient of this line was not significantly different from 1 at the 95% level, inspection of the data suggests that, over the interval 270 Kelvin - 280 Kelvin, St. Mawgan temperatures are,
on average, around 1 Kelvin higher than those on Goss Moor. The spread of the data (standard error = 0.97 Kelvin) allows for errors as low as 0 Kelvin and as high as 2 Kelvin. In summary, some overestimation of evapotranspiration resulted from this approach.

- Actual vapour pressure (avp) was regressed on saturation vapour pressure (svp), both variables having been calculated from such wet/dry thermistor readings as were available on Goss Moor. The methods of calculation are shown in FORTRAN program PENMONT5.FOR, listed in Appendix B. This regression equation, shown in Figure 3.9 (R² = 0.96), was required in order to estimate actual vapour pressure when wet/dry thermistor readings were not available on Goss Moor. Before the estimation of avp from svp, svp was first calculated using the estimate of Goss Moor air temperature derived from St. Mawgan air temperature as above.

Analysis of the resulting evapotranspiration estimates begins in Chapter 4.

3.4 MONITORING OF GROUNDWATER POTENTIALS

3.4.1 Aims of the Groundwater Monitoring

These were:

1. To assess the fluctuations and seasonal variations of the water table in comparison with changes in rainfall and evapotranspiration, and thus to indicate the nature of interactions between the wetland and the groundwater.

2. To provide spatially distributed and time-dependent potentials with which to calibrate the groundwater flow model and arrive at estimates of groundwater flows through the wetland aquifer.

3.4.2 Methods Employed in the Groundwater Monitoring

Piezometers were installed at 20 sites in the wetland, as shown in Figure 3.1. Their siting and
installation are discussed below.

In addition to the piezometers, stage boards, each consisting of a vertical board marked every 0.01 m along its 1 m length, were installed in a few of the deeper pools in the wetland. As mentioned in Chapter 2, these pools were originally pits dug for the extraction of gravel at earlier stages of the twentieth century. Four such pools were chosen for observation, marked on Figure 3.1 as sites S1, S2, S3 and S4.

Stage boards S1 and S2 were located in a closely spaced group of pools near the outlet of the catchment. Assuming that they are man-made and that no lining material was installed, they are unlikely to be perched. The elevations of the different pools form a stepped sequence, increasing eastwards away from the catchment outlet (see Figure 3.1). Surface water trickles from the highest to the lowest in a cascade, finally running into a small culvert and exiting the catchment. In periods of high water tables, therefore, these pools may act as groundwater sinks or discharge points, albeit for a restricted area due to the low surrounding permeabilities.

Stage board S3, on the southern side of the wetland, occupies a closed pool with no surface water connections. Since the pool is man-made, and yet unlined, the variations in the water level at S3 could therefore be expected to follow those of the surrounding piezometric surface. S3 thus provided further observations on the water table. In contrast, the pool occupied by stage board S4 is part of the River Fal flow system, thus providing an extended zone of interaction between the river and the groundwater domain.

A major factor in the positioning of the piezometers on Goss Moor was the need to assign boundary conditions to the wetland groundwater model. The two alternative boundary conditions relating to the siting of peripheral observation wells are the constant head condition and the constant flux condition. Because of uncertainties in hydraulic conductivity and aquifer thickness, flux conditions are usually subject to large errors of determination, and are therefore usually neglected in favour of simple head measurements. Modelling problems for which the primary question is the distribution of groundwater heads, rather than the flux distribution, also favour constant head boundary conditions. However, the strong influence of fixed head
boundaries upon nearby heads means that uncertainties in boundary values are also closely expressed in the interior of the model domain. A reasonable degree of accuracy in evaluating such boundaries is therefore demanded. In these terms, the Goss Moor problem was no exception, and so a large proportion of the monitoring effort was directed towards measuring groundwater heads at the periphery of the wetland. Thus, of the 20 groundwater monitoring sites in the wetland, 12 were used to provide boundary data. These sites were P1 at the western end of the wetland, P5, P7, P9 and P4 at the southern edge, P17, P18 and P19 at the northern edge, and P15 and P20 at the eastern end of the moor (see Figure 3.1).

In previous surveys of the catchment, a spring line was observed at the break of slope above the southern edge of the wetland. While supplying stream water to the wetland, this discharge zone also suggests the possibility of upward groundwater discharge in nearby downslope regions. Consequently, some provision was made to measure vertical potential gradients in this region of the wetland, by the installation of differentially penetrating piezometer pairs. These pairs of observation wells, at sites P5, P7, P8, P9 and P10, were located on the edge of the wetland, and so most of them also served to provide boundary data.

For those piezometers not employed in supplying boundary data, an even distribution over the remaining wetland area was attempted, subject to accessibility which was limited by the boggy nature of the ground. In order to avoid redundancy caused by proximity of other calibration points, efforts were made towards siting away from the river and stream channels, thus maximising the degree of areal representativeness of the data, and hence increasing the accuracy with which the groundwater flow model could be calibrated. Observation wells sited in this manner were P2, P3, P10, P11, P12, P13 and P14. Unfortunately, the necessity of using riverside tracks for access meant that certain of these sites were closer to the river than they otherwise would have been.

As seen in Figure 3.10, each piezometer consisted of a plastic pipe, with a perforated section about 21 cm in length at the capped end. The top of each piezometer was kept covered to prevent ingress of rain. The boreholes into which the piezometers were inserted were only just wide enough to accept them, due to the unavailability of larger drilling heads. The lack of any
appreciably large gap between piezometer tube and borehole wall then precluded the introduction of any filter pack or backfilling materials, which should have been included in order to reduce the entry of silt into the piezometer bore and to prevent bypass flow occurring along the outside of the piezometer tube (Morrison, 1983). However, an attempt was made to seal the top of the annular gap with earth. This sealing appeared to be effective and the remaining unfilled gap to be insignificant since there was no early recovery due to bypass flow in any of the slug tests reported in Section 3.5.4.

The boreholes were drilled either by hand auger, by motor-driven auger or with a non-rotary percussion drill with a cylindrical cutting head. Some smearing of the wall of the hole could be expected from all three of these methods. Ideally, a well development technique such as “surging” would have been employed after installation. For surging, a close-fitting surge block or piston is attached to a rod and forced up and down within the water column so as to force water back and forth through the screen and dislodge the finer particles in the well skin (Wilson, 1995). However, since no such measures were taken in the present study, significant inaccuracies were possible in subsequent measurements of both water table elevation and alluvial hydraulic conductivity. Some attention is given, in Section 4.6.3, to the consequences for the accuracy of the water table measurements, while the effect upon the accuracy of the slug tests is considered in Section 3.5.4.

At each site, the borehole was drilled to a depth such that a 20-50 cm section of the piezometer tube would be left protruding above ground when fully inserted, to allow for groundwater potentials above ground level. Most of the wells penetrated to a depth of 1 m below the surface, with a further seven penetrating to 3.5 m. Of the latter, five were placed at the same site as a 1 m well, thus forming a differential pair for the purpose of measuring vertical potential gradients in a region of possible upward groundwater flow. These pairs were at sites P5, P7, P8, P9 and P10, as mentioned above. Further information is given in Figure 3.1 and Table 3.1.

The water levels in the wells were monitored weekly with a portable electric dip probe. The resulting data are analysed in Section 4.6.3.
3.5 ESTIMATION OF HYDRAULIC PARAMETERS OF SUBSOIL

3.5.1 Introduction - Suitability of the Slug Test

As indicated in Chapter 1 of this thesis, the hydrologic functioning of a groundwater wetland is greatly influenced by the hydraulic properties of the contributing aquifer. The competence of a wetland in resisting periods of hydrologic stress is directly linked with the predisposition of the subsoil towards storage of water over long periods. The residence time of water in the subsoil is longer in groundwater systems with poorly conducting media and low potential gradients. A high specific storage or specific yield is also a factor in encouraging the longer residence of water. Measurement of the hydraulic conductivity and estimation of the storage coefficients of the wetland's contributing aquifer is thus essential to characterise the wetland in terms of its sensitivity to environmental change. The values of these parameters are best measured over as wide an area of the wetland as possible and, depending on the available information on the underlying stratigraphy, appropriate parameter ranges may be allocated to the particular types of stratigraphic unit which are present. These measurements also contribute to the overall characterisation necessary to allow modelling of the groundwater system.

Currently employed methods of estimating the conductivity and specific storage of water-bearing strata include steady state pumping tests, step-drawdown pumping tests, ring permeameter tests (for surficial deposits), and slug tests. Among those techniques applicable to subsurface deposits, the slug test involves the least commitment of time, manpower and equipment, and the least disruption to the water balance of the system. Although in some cases suffering from more problems of uncertainty in analysis, slug tests are also applicable to the measurement of much lower hydraulic conductivities than are pumping tests since they do not necessitate the maintenance of large fluxes of water from the medium around the borehole. They are therefore more suitable for use in the clay-rich material of the Goss Moor alluvial aquifer. In the present study, slug tests were employed as the sole means of estimating the hydraulic conductivity of the alluvium.
3.5.2 Performance of Slug Tests

Slug tests are most commonly carried out in piezometers. Figure 3.10 shows the configuration of a piezometer slug test in an unconfined stratum. The most important aspects of the test are the sudden removal of a "slug" of water from the piezometer bore and the monitoring of the subsequent recovery of the water column back to its original level. The speed of this recovery is analysed to yield the hydraulic conductivity and, in some cases, the specific storage of the subsoil. In a conventionally conducted piezometer slug test, the water level in the hole will always stay above the screened interval of casing, ensuring that the effective screen length remains constant throughout the test. This is not the case for tests carried out in fully screened wells or uncased auger holes.

Successful slug tests were performed on 8 of the 26 piezometers on Goss Moor. The water column in each piezometer was pumped down to the base of the tube, and this constituted the slug removal for each test. Hence the tests were conducted with initially varying effective screen lengths which became constant as the water level in each piezometer rose above the top of the screen. This unconventional procedure resulted from the need to remove water from the piezometer tube as quickly as possible in order to approximate an instantaneous withdrawal. During the approach of the water column to the upper end of the screen, the remaining interval of screen above the water column would act as a seepage face, on which the hydraulic head expressed the decrease in elevation down towards the water surface. Most models used in the analysis of slug recovery assume a constant screen length along which the hydraulic head is uniformly equal to that in the water column. The experimental arrangement used in the present study replaced the uniform head with a linearly increasing head along the upper part of the screen. This would result in slightly less flow into the piezometer than expected from the standard conceptualisation of the flow system, and consequently in an apparent initial delay in the water column recovery. The effect of this digression on the subsequent estimation of hydraulic conductivity and storage coefficient is discussed in Section 3.5.4.
3.5.3 Analysis of Slug Tests

The rate of recovery of the water column varies according to the stratigraphy of the subsoil, the siting of the piezometer and the presence or absence of anisotropy in the conducting medium. The two configurations thought to be applicable in the present study are shown in Figures 3.10 and 3.11. As explained in Chapter 2, the surficial deposits of Goss Moor consist of unconsolidated, silty, clayey sediments intercalated with units of gravel and sand 0.1 - 0.5 m thick. This interleaving of sediments may give parts of the aquifer a locally confined and/or anisotropic nature. Such conditions may be represented by the configuration shown in Figure 3.11, where the screen is assumed to penetrate fully a permeable stratum which is confined over the area of influence of the slug test. For the analysis of this type of configuration, an analytical solution of the flow problem given by Cooper et al. (1967) may be used. The alternative configuration, shown in Figure 3.10, is assumed for all cases not involving anisotropy or confined conditions. Tests in these situations may be analysed with a semi-empirical method given by Bouwer and Rice (1976).

The slug test solution of Cooper et al. (1967):

The method of Cooper, Bredehoeft and Papadopulos (henceforth referred to as the CBP method) gives estimates of hydraulic conductivity in the horizontal plane, and order-of-magnitude estimates of specific storage. The flow problem for which it is formulated is illustrated in Figure 3.11. The following assumptions are made.

1. The aquifer to be tested is confined.
2. The aquifer is composed of a homogeneous material which is either isotropic or anisotropic with a preference for flow in the horizontal plane.
3. The piezometer is screened through the total thickness of the aquifer.
4. The piezometric surface is not drawn down below the upper boundary of the aquifer.
5. The head in the aquifer is initially uniform and steady.
6. The withdrawal/injection of the slug is instantaneous.
7. There are no well losses.
8. The column of water in the well exhibits no inertia.
Based upon these assumptions, differential equations and boundary conditions are formulated for the transient, radial flow problem. A Laplace transformation of the problem is solved with power series, following which the inverse transformation yields an expression for the variation of the head in the aquifer with time and distance from the piezometer. Substitution of a distance equal to the screen radius gives the evolution of the head in the piezometer itself. A family of type curves, with which to evaluate the required parameters from field data, can then be defined in terms of the non-dimensionalised drawdown, time and storage coefficient. See Cooper et al., 1967, and Papadopulos et al., 1973.

Applicability of the CBP model:

The applicability of any method of analysis is limited to those uses in which its basic assumptions are not violated. In the present study, assumptions (4) to (6) were initially taken to be acceptable, given proper piezometer installation and careful operation of the tests in the field.

Violation of assumption (8) is characterised by oscillation of the water column in the piezometer, requiring that the test aquifer has a very high permeability and that the well contains a very massive water column. In other words, the system of aquifer and water column is underdamped (see van der Kamp, 1976). In assessing whether this occurred during any of the slug tests on Goss Moor, the recovery of the water level during each test was examined for oscillations. As expected, no oscillations were found and assumption (8) was thus validated.

The CBP model can be used only for cases with the appropriate hydrostratigraphy and siting of the piezometer, as dictated by assumptions (1), (2) and (3). The presence of strong anisotropy with a preference for horizontal flow (assumption (2) ) would justify the use of this radial flow model. Assumptions (1) and (3) could then be ignored. Hence, anisotropy becomes a key factor in the suitability of a location for analysis with the CBP model, and must always be borne in mind.

In the present study, borehole logs were taken for the purpose of defining the stratigraphy around the piezometer. Also, an attempt was initially made to examine the borehole products for
anisotropy, but soon proved to be unsuccessful as the samples were too disordered. With the consequent lack of information on the directionality of the aquifer material, the selection of cases appropriate for the CBP model could not be made on the basis of the bore-hole logs alone. Rather, information from both bore-hole logs and estimates of specific storage was used, as described in Section 3.5.4.

Assumption (7) requires that the piezometer should not possess a low-conductivity skin. This requirement was unlikely to have been satisfied in the present study since no well development was performed after installation of the piezometers (Section 3.4.2). The implications for the permeability estimates are discussed in Section 3.5.4.

The slug test model of Bouwer and Rice (1976):

The Bouwer and Rice method (hereafter the B-R method) gives estimates of isotropic hydraulic conductivity only. Figure 3.10 shows the flow problem for which it is formulated. Formulation of the model involves the following assumptions.

1. The aquifer to be tested is either unconfined or confined with leaky upper boundary.
2. The aquifer is composed of a homogeneous, isotropic material.
3. Drawdown of the water table is negligible.
4. Flow in the capillary fringe can be ignored.
5. The head in the aquifer is initially uniform and steady.
6. The withdrawal/injection of the slug is instantaneous.
7. There are no well losses.
8. The column of water in the well exhibits no inertia.
9. The aquifer exhibits no piezometric storage.

Unlike the CBP solution, the B-R model does not involve the assumption of full penetration. Hence it is widely used for the assessment of data from partially penetrating slug tests in isotropic media.

Derivation of the B-R model utilises the Thiem equation for steady flow from a confined aquifer
into a fully penetrating well. This equation (for which a derivation is available in Todd, 1959) expresses the flow in terms of geometrical factors, and utilises the concept of an "effective radius", "dissipation radius" or "radius of influence", $R_e$, beyond which the piezometric surface is unaffected by the well. By relating this expression to the rate of rise of the water level in a recovering well, Bouwer and Rice arrived at a formula for the time dependence of the water level in which the flow need not be known. In order to use the formula, however, the effective radius must be determined. Bouwer and Rice conducted electrical analogue experiments to obtain a range of values for this radius. Optimisation of these values would provide some correction for the use of the Thiem equation in unconfined, partially penetrating conditions, and the effective radius was finally given as a function of the geometrical configuration of the test, taking into account such factors as the radius and the degree of penetration of the piezometer and the proximity of the screen to the water table.

**Applicability of the B-R model:**

Examination of the assumptions used in formulating the B-R model reveals that they are similar, although not identical to those of the CBP model, which were examined above. Assumptions (5) to (8) are shared exactly with the CBP model, having a similar effect on the applicability of the B-R method.

Assumption (4) of the B-R model is linked with assumption (3), since the drawing down of the water table will create a new zone of partial saturation (referred to by Bouwer and Rice as a "capillary fringe") from which water drains gradually. This process, in which water is withheld from the saturated zone and hence is temporarily unavailable for replenishment of the water level in the well, is known as "delayed yield". It was first introduced by Boulton (1954). Through assumptions (3) and (4), the B-R method neglects the possibility that this might happen.

The use of the concept of an effective radius allows further understanding of the limitations of the B-R model. In a slug test, this radius must firstly expand as the change in pressure propagates outwards, and then contract as the pressure disturbance is dissipated. This is illustrated by Brown *et al.* (1995). The speed at which these changes occur depends upon the specific storage and hydraulic conductivity of the test medium.
In developing their slug test model, Bouwer and Rice (1976) defined a single dissipation radius for the whole duration of the test, making use of assumption (9) that piezometric storage in the aquifer is negligible. In truth, $R_e$ is reduced by finite aquifer storage, rendering the Bouwer and Rice value something of an overestimate. This leads to an overestimate of the hydraulic conductivity. Storage of water in the piezometer bore has a similar effect. Hyder and Butler (1995) demonstrate numerically the progressive overestimation of permeability as aquifer storage increases, adding that the effect is particularly marked for low-permeability, clay-rich formations.

However, Hyder and Butler (1995) also neglect the possibility of delayed yield. Narasimhan and Zhu (1993) demonstrate the effects of this process for the initial stages of continuous pumping tests. Delayed drainage of the newly created vadose cone temporarily reduces the apparent specific yield, expanding the physically realised radius of influence. The effect on slug tests is similar, and reduces the overestimation of permeability otherwise inherent in the B-R method.

Despite the above problems, the accuracy of permeability estimates with the B-R model is not strongly affected by changes in the specific storage of the medium. This insensitivity, noted by Brown et al. (1995), arises because $R_e$ appears only as the argument of a logarithmic term in the B-R formula.

As with the CBP model, assumptions (1) and (2) restrict the use of the method to certain hydrostratigraphic conditions and sitings of the piezometer. In the context of the present investigation, condition (1) was satisfied at most of the piezometer sites. However, as explained above, it remained unclear whether anisotropic conditions (violating assumption (2)) existed at any of the slug test locations, because of uncertainties in interpreting samples of the drill-hole material. The presence of anisotropy at any given test location was thus inferred from the estimates of specific storage obtained with the CBP model, as described in Section 3.5.4, and the suitability of the B-R method decided accordingly.

The Bouwer and Rice technique shares its overall methodology with that of Hvorslev (1951), in that both methods employ the analogy of a radial permeameter. Bouwer and Rice (1976) and
Brown et al. (1995) found that these two formulations yield comparable estimates of permeability, although the B-R results appear to be slightly more accurate (with errors of around 20% relative to the true value). This advantage in accuracy is due to a better representation of the flow field by the experimentally determined geometrical factors of Bouwer and Rice than by the analytically determined “shape factors” of Hvorslev. For the present study, the B-R model was used.

3.5.4 Results of Slug Test Analysis

The data from slug tests conducted on Goss Moor were analysed with both the Bouwer and Rice method and the Cooper et al. method. Use was made of a computer software package known as AQTESOLV (Geraghty and Miller Inc., 1991), which can implement both techniques. For cross-checking, the B-R method was also implemented in spreadsheet form. Agreement between the two implementations was generally very good.

Application of the Cooper et al. model:
The fitting of the CBP model to the slug test data can be seen in Figures 3.12 to 3.19, which show the estimated transmissivities, in m² s⁻¹, and storage coefficients (dimensionless). Table 3.3 shows the equivalent hydraulic conductivities and specific storages, assuming a conducting layer thickness equal to the screen length (i.e., 0.21 m). As expected given the clayey nature of the Goss Moor alluvium, the estimated permeabilities were rather low (see Table 3.3 for permeabilities in m/day). Since the CBP model assumes horizontal flow, these were estimates of horizontal permeability.

With a couple of exceptions, the slug test analyses underestimated the values of specific storage by a factor of between 10³ and 10⁵ when compared with the expected range of values, as seen in Table 3.4. Expected values in Table 3.4 were calculated, using Jacob's expression, from ranges of vertical compressibility given by Domenico and Schwartz (1990) and ranges of porosity given by Freeze and Cherry (1979). The sites at which the slug test estimate was compatible (to within one order of magnitude) with the expected range were P05d, P11s and P12s. The underestimates obtained at the five remaining sites may have indicated either the influence of a
retarding well skin or the involvement of non-horizontal flow in the recovery of the water column and therefore cast doubt on the applicability of the CBP model to the situations concerned. Moreover, the obtaining of an appropriate value of specific storage was in itself no guarantee of the suitability of the model: the degree and nature of deviations of the modelled from the observed water level response should also be taken into account, as discussed below. Despite these potential problems with applicability, expected differences in site characteristics seemed to be discernible through the specific storage values given by the CBP model. The variation in specific storage appeared to follow the trend in the gravel content of the sediments surrounding the piezometers: those regions with more gravel exhibited lower specific storages, whereas those slug tests performed in purer clay gave results in accordance with a more compressible medium.

Viewed overall, the CBP model appeared graphically to fit the experimental data reasonably well. However, in general, apparent goodness of fit of any model to the data is no absolute guarantee, in itself, of the appropriateness of the model. Karasaki et al. (1988) illustrate this by the similarity between type curves obtained for many different test configurations ranging from radial flow to two-layered aquifers, linear flow and spherical flow. They conclude (p.123) that "analyses of slug tests suffer problems of nonuniqueness, more than other well tests." For the Goss Moor slug tests, this implied that the assumptions of confined, anisotropic and skin-free conditions might not be appropriate in every case despite broad correspondence of the modelled with the observed responses.

Type curves presented by Hyder and Butler (1995), for recovery in a well partially penetrating an unconfined aquifer, demonstrate the broad similarity of the response under these conditions to that under radial flow conditions. Nevertheless, this paper notes that a more spherical flow regime produces a type curve which, when compared with radial flow type curves, exhibits retardation of recovery in the initial stages, followed by somewhat faster recovery after a while. This pattern is also followed when a retarding skin, one or two orders of magnitude less conductive than the test material, is present (Hyder et al., 1994). Steepening of the type curve is also effected by a decrease in the specific storage of the medium, with the consequence that the best fit for the radial flow CBP model to data from a partially penetrating test, or from a test with
a retarding skin, is obtained by the specification of low specific storages. This can be seen in
Table 3.4: the specific storages of $5 \times 10^{-8}$ estimated at piezometers P04s and P10d compared
unfavourably with the values of around $10^{-3}$ applying to clayey unconsolidated sediments. Such
underestimation of the specific storage allowed a close fit to the data (Figures 3.12 and 3.16). In
contrast, the best fit curve for piezometer P05d (Figure 3.13) gave a plausible value of specific
storage but did not closely fit the measured data which exhibited a trend characteristic of the
aforementioned spherical flow / undeveloped well / low specific storage regimes. The CBP
model suffered similar problems when applied to the slug test responses from piezometers P09d
and P16s.

Application of the Bouwer and Rice model:

While the Cooper et al. model was used successfully for the analysis of a few of the slug tests in
the present study, it was thought possibly to be inappropriate in relation to piezometers P04s,
P05d, P09d, P10d and P16s. For these five piezometers, the method of Bouwer and Rice was
also used, since the behaviour of the data may have been indicative of unconfined, isotropic
conditions as discussed above.

Figures 3.20 to 3.24 show the fitting of the B-R model to the data from the relevant sites. Table
3.3 gives the estimated conductivities in m/day. As expected, the estimates of isotropic
permeabilities obtained with the Bouwer and Rice method were lower than those of horizontal
conductivity given by the Cooper et al. model. Assuming no low-conductivity skin and no
anisotropy, these were the lower bounds for the permeability of the aquifer at these locations.

Adjusting for the effect of a retarding well skin:

The assumption that a retarding skin, 10 to 100 times less conductive than the aquifer material,
has been present during the slug test, provides the upper bound of (horizontal) permeability for
those formations not containing any clay at the level of the piezometer screen. As noted by
Faust and Mercer (1984) and Hyder et al. (1994), the presence of a retarding well skin can
produce severe underestimates of aquifer permeability: the evaluated permeabilities may be more
representative of the skin than the formation. In their modelling assessment of the performance
of the Cooper et al. method and the method of Hvorslev (1951), Hyder et al. (1994) found that
skin conductivities from 10 to 100 times less permeable than the formation gave estimates between 2 and 10 times less than the true conductivity.

Koppi and Geering (1986) tested the effectiveness of a technique to remove the smearing from an “alluvial Prairie Soil” surface with epoxy resin. The resin was spread onto the flat surface, allowed to set and then removed, ripping away the soil membrane. Infiltration rates of water through surfaces prepared in this manner were found to be between 11 and 19 times those for smeared surfaces. However, this improvement was found to depend on the type of soil tested: for a “Red Podzolic” soil, the prepared surface was only 1 - 7 times as permeable as the smeared surface. In a similar study with a strongly structured clayey soil (a silt loam), Campbell and Fritton (1994) were able to remove borehole skins by levering with an ice pick and found that after this treatment, the permeabilities given by constant head Guelph permeameter tests were an average of 14 times higher. Such results may be considered in an attempt to compensate for the probable presence of well skins in the present study.

Since the Goss Moor alluvium contains a high proportion of clay layers which contrast with the sand/gravel layers, it was plausible that the smearing of the sediments over the wall of each borehole during drilling would produce a very effective well skin, readily retarding flow by factors greater than those observed in the above studies. However, those sites in which the layer next to the piezometer screen was clayey would be relatively unaffected by a clay skin, removing from consideration all piezometers in the present study apart from P09d and P16s. For these two piezometers, it was further assumed that the problems in model fit were caused entirely by well skins rather than by non-horizontal flow. Thus, the final estimates of \( K_h \) for piezometers P09d and P16s were obtained by adjusting the CBP-derived value upwards by a factor of ten, as shown in Table 3.3.

**Vertical hydraulic conductivities:**

Vertical conductivities could not be evaluated with any accuracy in the present study, since no techniques appropriate to their measurement were used. One of the formulae given by Hvorslev (1951) provides estimation of vertical permeability for a slug test configuration in which the piezometer tube is unscreened along its length but screened across its bottom end. However,
this formula could not be used since all piezometers were screened along their vertical limbs but sealed across their bottom ends. Had undisturbed samples of the unconsolidated material been taken, directional measurements of permeability could have been made with a laboratory permeameter. However, no such samples were taken.

Since such techniques were not used in the Goss Moor programme, the vertical conductivities could only be guessed. At locations where the model of Cooper et al. (1967) is applicable to the data, the anisotropy ratio, defined as the ratio of vertical to horizontal conductivity, is likely to be of the order of $10^{-1}$ or less (Hyder et al., 1994; Hyder and Butler, 1995.). Where the Bouwer and Rice model is applicable, the material is assumed isotropic. Plausible vertical hydraulic conductivities at the slug test locations are therefore given in Table 3.3.

3.6 SUMMARY

With the aim of quantifying water flows and storage associated with the wetland, rainfall, stream flows, evapotranspiration (ET) rates and water table elevations in the wetland were monitored over a total duration of 19 months. The different periods of monitoring and the sampling intervals are listed in Table 3.1. Instrument locations are shown in Figure 3.1.

Stream flows were monitored using stage recorders at the catchment outflow, at the main inflow to the wetland and at one secondary wetland inflow. Due to a lack of stage-discharge data, the overall errors in the stream flow estimates could not be assessed. However, these overall errors would be greater than the errors due to inaccuracies in digitisation of the stage charts. Thus, errors ranging from 6% to greater than 15% might be expected.

Daily rainfall was monitored at one location on the wetland for 8 months of the study period. A further 10 months of daily rainfall totals were obtained from nearby (i.e. less than 10 km distant) monitoring stations. Despite the frontal nature of rainfall in the region (which would give a highly uniform spatial distribution of rainfall intensity, orographic processes and instrument bias would potentially cause differences between the daily totals at the different locations. Such
differences were found between rainfall totals at Roche and Goss Moor, but were minimal between St. Mawgan and Goss Moor. Roche rainfall totals were not used to quantify the Goss Moor water budget.

ET variables were measured on the wetland for 4 months with a further 13 months of supplementary data from nearby monitoring stations. The Penman-Monteith equation was used to calculate daily potential ET from 4 land cover types known to exist in the catchment: Willow Scrub, Wet Heath, Pasture and Open Water. For Pasture, actual ET was calculated using the Penman-Grindley formula and was found to be equal to the potential ET, due to high rainfall. Therefore, actual ET in the 3 remaining (wetland) land cover types was also equal to potential ET. A short discussion illustrated the inherent simplifications involved in the parameterisation of several highly variable meteorological quantities.

Water table elevations at 20 sites and pool surface elevations at 4 sites were monitored weekly for 12 months with stage boards and with piezometers of 1 m and 4 m depth. The water table data would be used to parameterise a numerical model of groundwater flow beneath the wetland, described in Chapter 6, and so the piezometer locations were distributed partly along the boundary of the wetland and partly for uniform areal coverage within the wetland.

Recovery of the piezometer water column after sudden bailing, recorded at 8 of the piezometer sites, was analysed using the methods of Cooper et al. (1967) and Bouwer and Rice (1976), revealing wetland sediment permeabilities from $4 \times 10^4$ m/day (clay) to 13 m/day (silty, sandy gravel).
4.1 AIMS AND INTRODUCTION

Chapter 2 of this thesis described the climate, geology, geomorphology and drainage structure of the Goss Moor catchment in relation to their possible effects on the hydrological regime of the central wetland. While these factors were seen to be important in their influence, a need to examine the Goss Moor wetland's hydrological behaviour led to the concentration of monitoring activity on the measurement of hydraulic conductivity and of state variables such as groundwater potential and stream flow in the wetland itself, as described in Chapter 3. The current chapter presents and analyses these measurements in terms of hydrological processes, in a first attempt to satisfy the above requirement.

Presented here are the measured variations in rainfall, stream flow, evapotranspiration and groundwater levels in the wetland over a period of approximately one year. A major theme running through the analysis of these variables is the assessment of evidence concerning the storage of water in the wetland or its source areas. Thus, the loss of water by evapotranspiration (ET) from within the wetland is compared with that from the outer catchment, as permitted by the parameterisation of plant properties provided by the Penman-Monteith approach to ET estimation. The concept of quick and slow components in stream flow is used to compare the degrees of storage in the flow paths supplying the different points in the channel network. Variations in groundwater head at the various piezometer sites reveal the influence of the river and the ground surface on the wetland groundwater column. For climatic context, the fluctuations of the wetland's rainfall input over the study year are considered before examining the modes of variation of the stream.
flows and groundwater levels. With the help of hydrophysical concepts introduced in Chapter 2, the results of the analyses are used to indicate three possible sources of water storage and flow retardation, all within the wetland itself.

4.2 RAINFALL

A fundamental consideration in the undertaking of any catchment study of limited duration is the relation of the rainfall input during the study year to its long term average. This provides qualitative guidance in the interpretation of the current findings within a longer time scale and is vital with regard to wetland management. Consideration of potential evaporation in the same way would also be desirable, but long term data for this quantity are unavailable.

In 1993, 1.673 m of rain was recorded at the Environment Agency's (E.A.) nearby Roche Weather Station (note that the depth is expressed here in metres rather than in millimetres for consistency with the water budget calculations in Chapters 5 and 7). The rainfall total for the following year, 1994, at the same station was 1.788 m. These amounts correspond to 122% and 130%, respectively, of the site's long term average annual rainfall of 1.371 m (data supplied by E.A. South Western Region, Cornwall Area). Table 4.1 shows the monthly data and long term averages at Roche during the water budget period of the present study (September 1993 - August 1994). Monthly rain during this period was consistently greater than the long term average, with the exception of October, June and July 1993-94. The total rainfall for these 12 months was 1.809 m, 132% of the long term average. This implies that Goss Moor may have been wetter than usual during the study period, increasing the apparent sensitivity to desiccation since high water tables could have been encouraging greater surface runoff and possibly greater flashiness in the stream response. The water table fluctuation between winter and summer may also have been greater than usual, due to higher than normal winter rainfall followed by lower than normal rainfall in June and July.
The daily rainfall over the water budget period of the present study is shown together with an approximate antecedent precipitation index (API) in Figure 4.1. This API is notionally the same as that used by Fedora and Beschta (1989) in the simulation of storm runoff. The API is given an arbitrary initial value and then is calculated in subsequent time steps according to:

\[ API_t = \lambda \cdot API_{t-\Delta t} + P_t \]  \hspace{1cm} (4.1)

where

- \( t \) is total time elapsed (days),
- \( \Delta t \) is the constant time step (days),
- \( \lambda \) is a constant scaling factor (dimensionless), and
- \( P_t \) is the amount of rain (m) falling in an interval \( \Delta t \) beginning at time \( t \).

They used the stream flow recession constant for daily time steps, equivalent to \( e^{24s} \) in Equation 4.6, page 104, as the scaling factor \( \lambda \) for the wetness of the catchment. The same constant, calculated from the slow flow of Goss Moor at gauging station C6 where \( s = 0.001793 \text{ hours}^{-1} \) (see Figure 4.20), was 0.958 at the 1-day time scale. Since the API evolves from an arbitrary initial value, sufficient run-in time must be allowed for the effect of this value to decay away. The period of interest and the focus for water balance and modelling considerations in the present study was 1/9/93 - 31/8/94. In comparison, API calculations began at the earliest date for which rainfall data had been procured, this being 1/6/93. Thus, a run-in period of 3 months was allowed before the period of interest. The resultant API was scaled to form an envelope around the bulk of the rainfall values in Figure 4.1.

The maintenance of wetness in a humid-temperate catchment such as Goss Moor depends not only upon the water storage and transmission properties of the catchment itself, but also upon the constancy of meteoric input from the local, current regime of rainfall. The API is intended to show the decay of the influence of periods of rain upon the catchment’s wetness, and so can be used for a qualitative comparison of the sustaining effect of rainfall at different times in the catchment’s long term or short term history. In particular, it assists...
in differentiating the relative significance of rainfall spells of different intensities and durations. This may help in the interpretation of other observations of catchment behaviour. In a sense, the API is a function fitted to represent catchment soil moisture conditions. However, it is also an approximation to the outflow response of the whole catchment, encompassing the effects of channel routing, soil storage, groundwater storage, surface water storage and evapotranspiration, and so it is being examined in the present section as a precursor to inspecting the true catchment outflows. The main shortcoming of the linear API is that, being derived using the recession constant, it reflects only the later stages of catchment response and is much less representative of runoff immediately after storms. A non-linear API, using parameters fitted to include early stages of the stream flow recession, would be a quick way of giving a closer reflection of the decay of catchment wetness after rainfall.

In the present study, the comparison of the linear API in different time periods is restricted to the chosen 12 month water budget period, which begins on 1/9/93 (see Chapter 5). It can be tentatively concluded from the graph that November, July and August were the driest periods during the water budget year. More rapid response than accounted for by the API in the earlier stages of the recession may shift these periods closer to the time at which rainfall ceased, that is, up to one month earlier. Lack of rainfall in June and July was responsible for the reduced summer proportion of quick flow discussed in Section 4.5.4.

The monthly numbers of rain days shown in Figure 4.2 follow the pattern of monthly rainfall totals, again showing shortfall in October, November, June and July. Between December and March, the large proportion of days with rain undoubtedly contributed to the maintenance of catchment wetness and therefore to high proportions of quick flow.
4.3 EVAPOTRANSPIRATION

Evapotranspiration from the different parts of the Goss Moor catchment was calculated on a daily basis using the Penman-Monteith equation with a combination of meteorological data obtained from an automatic weather station situated on site and from the weather station at the nearby St. Mawgan air force base. Separate values of evapotranspiration (ET) for each of the various types of land coverage in the catchment were obtained by the adoption of different parameter values for each distinct type of area, as described in Section 3.3.3. Also in Section 3.3.3, rough calculations of soil moisture deficit and actual evaporation using the Penman-Grindley approach (Lerner et al., 1990) showed that even the non-wetland parts of the catchment retained sufficient soil moisture to maintain ET at the potential rate throughout the study period. Problems associated with the assignment of uniform net radiation, wind speeds, temperature and humidity over the catchment were discussed. It was concluded that the true variations in such quantities were likely to be highly complex. The direction of bias in the evapotranspiration estimates arising from assumed uniformity in these quantities was therefore unknown, but the estimates were likely to be most accurate for the vegetation type over which the monitoring station was located, i.e. wet heath.

A representation of the land cover pattern in the Goss Moor catchment, illustrated in Figure 2.1, was obtained from a data set for geographic information systems (GIS) provided by English Nature. The data set contained four specified land types: willow scrub, wet heath, open water and (by elimination) pasture. A very small proportion of the pasture on the southern fringe of the catchment is in fact covered with mica waste from nearby kaolin mining operations. The respective areas of these land types in the catchment are shown in Table 4.2. The wetland contains all of the willow scrub, wet heath and open water. All areas in the outer catchment are pasture, excepting less than 1 km² of dry heath and the restricted mine waste area mentioned above.

The variation of the daily evapotranspiration rate (in m³m⁻²d⁻¹) over the study period is shown for each different land cover type in Figure 4.3. These rates vary over the year in a
roughly sinusoidal manner according to the seasonal variation of net radiation, reaching their minima during the months October to February and their maxima during April to August. In concert with seasonal variations of rainfall, this contributes to the winter moisture surplus and the summer moisture deficit. Figure 4.3 also demonstrates that the wetland land types willow scrub, wet heath and open water lose water at a greater rate than the surrounding pasture.

Figure 4.4 shows the mean evapotranspiration rates for each season. In general, the flux rates were greatest for willow, with wet heath, open water and then pasture in descending order. Unlike the willow and heathland grasses, the open water and pasture do not undergo seasonal changes in leaf area index. However, this difference did not result in any summer enhancement of wetland vegetation ET over that of the other land cover types. Indeed, evaporation from open water gained relative strength during the summer. This was due to the shading of sub-canopy evaporative surfaces in the willow carr which reduced much of the evaporative gain otherwise made in such areas by the denser foliage. Figure 4.5 illustrates this by comparing sub-canopy evapotranspiration with canopy evapotranspiration rates in the willow carr. Both fluxes have been normalised with respect to the simultaneous open water evaporation rate, removing the effects of variations in net radiation, humidity and temperature imposed by external atmospheric processes. The resulting functions show the step-like manner in which the leaf area index was varied for the vegetation model. Evapotranspiration rates from the canopy followed monthly changes in the leaf area index (see Table 3.2). Sub-canopy losses, in turn, decreased upon increasing leaf area index with the result that the two sources of evapotranspiration appeared to be in anti-phase. This relationship was most evident during summer and early autumn when wind speeds were low, since high wind speeds drove up the evapotranspiration from the canopy even while leaf coverage was minimal. Increased wind speeds did not affect the sub-canopy evapotranspiration because of sheltering (the Priestley-Taylor equation for equilibrium evaporation was used beneath the canopy - Section 3.3.3). Nevertheless, even in windy weather with low leaf area indices (for example, between November 1993 and March 1994), the calculations programmed in Appendix B partitioned available solar energy such that rising evapotranspiration from the canopy reduced the evapotranspiration from below.
The energy budgeting used to obtain this result did not consider the physical processes involved and so these remained unidentified. However, the validity of the higher rates of canopy evapotranspiration during times of near-zero leaf coverage was questionable and was probably due to the replacement of zero values with values of 0.1 for the leaf area indices. Therefore the winter evapotranspiration rates obtained with these calculations should be treated with caution.

The ET losses from the wetland and from the rest of the catchment are shown in Figure 4.6. The wetland contributes 23% of the annual ET losses from the catchment (see Table 4.2), despite constituting only 17% of the catchment area. Hence the wetland exhibits hydrometeorological conditions different from those of the rest of the catchment. Taking into account the minor proportion of wetland in the catchment area, the contribution of the wetland to the catchment ET flux volume was little affected by any seasonal variations in relative ET rates. This wetland contribution remained steady at around 23% (see Table 4.2).

The relatively high rates of water loss from willow scrub and wet heath seen in Figures 4.3 and 4.4 may be explained via the moisture fluxes coming from two different levels in the vegetation stand: the plant canopy and the ground surface:

i) **Plant canopy.** Three main factors affect the canopy transpiration as calculated with the Penman-Monteith equation (see Section 3.3.3 and Appendix B). Firstly, the bluff body obstruction of air movement by the plant stand affects the diffusion of water from the vegetation, transferring momentum towards and moisture away from the leaf surfaces. The willow canopy, presenting the greatest bluff body area to wind motion, thus transpires most readily in response to stronger winds. As mentioned above, enhancement of ET through this mechanism appeared to occur mainly during autumn, winter and spring. The winter enhancement, continuing despite an absence of leaves, was attributable to evaporation of intercepted water from branches, although it was possibly overestimated due to the specification of a seasonally invariant roughness length and zero plane displacement for the canopy. Secondly, the total area of transpiring leaves varies between the plant types and
also according to season. As shown in Table 3.2, the leaf area index (L.A.I.) of willow varies from 0.0 in the months November - April up to 6.0 during August (Lindroth et al., 1994). While pasture maintains a constant L.A.I. of 1.0, the moorland grasses of the wet heath develop this amount of living foliage only during July and August (Ripley and 'Redmann, 1975). Thirdly, the stomatal resistance to water vapour diffusion is lower for the willow than for the moor grasses or pasture, as seen in Table 3.2. Combined with the differences in leaf area index, this identifies the willow as the greatest transpirer during the summer.

ii) Ground surface. While the above consideration of the plant canopies shows that the willow is the strongest transpirer during the summer, this effect is less important than the presence of an extra source of evaporation in the wetland, namely the surface water beneath the vegetation stands. This can be seen from Figure 4.7, in which the water losses from the open water and from the pasture are compared with the sub-canopy evaporation from beneath the willow and the wet heath grasses. The two wetland vegetation types, willow carr and wet heath, conceal an evaporative surface with a flux similar to that found above the other land cover types, although slightly reduced by shading and by sheltering from wind. The addition of transpiration to this evaporation raises the total flux from the wetland above that of the surrounding farmland.

4.4 STREAM FLOW STATISTICS AND COMPARATIVE SPECTRAL CHARACTERISTICS

4.4.1 Daily and Hourly Stream Flow Time Series

Figure 4.8 shows the daily average stream flows at the three monitoring locations in the wetland over the study year. For comparative purposes, the daily rainfall rates are shown along the top of the graph. The main period of high flows was from December to April, with a short revival of flows in late May and early June. C5 and C1 are situated on tributaries of the channel upstream of C6, as seen in Figure 3.1. Both the outflow, C6, and
the flow from the northern part of the wetland, C5, appear to decay faster than the inflow at
C1 during the early summer. This leaves the slow flow at the outlet equal to that at the inlet
later in the summer. The sum of the two contributory flows is compared with the outflow
in Figure 4.9, showing again that outflow during July and August can be accounted for
almost entirely by the two measured tributary flows. At other times of the year, flows at C6
may be up to three times the sum of those at C5 and C1. The contribution of these
tributaries to the study year's accumulated outflows is shown in Figure 4.10. As shown in
Table 4.3, they account for 44.4% of the total flow at C6. The remaining 55.6% is
provided by rainfall within the main part of the wetland itself, by seepage zones at the
wetland periphery and by ungauged tributaries. Of such ungauged channels, the most
important accompany drain C5 in transferring flow from the northern section of Goss Moor
beneath the barrier of the dividing trunk road to the main wetland area. Certain of these are
distributaries of the drainage channel supplying C5, which is situated on the largest of such
distributaries. A rough estimate of the total flow from the northern area can be obtained by
doubling the flow measured at C5. This estimate is based upon a short inspection of other
culverts crossing the main road before the installation of the stage recorder at site C5. It
was found that only one other culvert was carrying an amount of water comparable to that
passing site C5. The expedient of doubling flows recorded at C5 takes into account a
further 11% of the contributions to the volume passing C6. However, significant error
could be expected from such a rough estimate, giving rise to errors of similar magnitude in
certain water budget quantities calculated in Chapters 5 and 7.

Differences in stream flow characteristics between the three monitoring sites are not
discernible from the graph of daily average flows in Figure 4.8. However, Figure 4.11
gives an example of the marked difference in hourly response at the upstream gauging
station C1 from that at the other two sites. Stormflow at C1 increases and decreases more
sharply and is briefer than that at the other locations. Flow at C5 and C6 reaches its peak
approximately 14 hours after the peak rainfall, whereas C1 flows take only 9 hours to reach
their peak.
4.4.2 Stream Flow Duration Curves

The duration curves for daily average flows at C5, C6 and C1 over the study year are plotted in Figure 4.12. The shape of each curve depends upon the statistical distribution of flows at the site. If the daily average flows were lognormally distributed, then the duration curve for each site would be a straight line. The flattening-out of the lines at low flows seen in Figure 4.12 indicates that flows in the river are unlikely to fall any lower. The fact that the catchment outlet appears to have low flows which are lower than those further upstream at C1 would normally indicate losses occurring from the intervening reach. However, the short distance (3.7 km) and travel time of less than five hours, derived from the times to peak presented in Section 4.4.1, between the two sites gives little chance for either evaporation or infiltration. It is possible that the apparent superiority of upstream slow flows is due to a mechanical fault in the stage recorder at low stages, as suggested by the glitchy nature of upstream slow flow shown in Figure 4.11.

All three lines have similar shapes at probabilities of exceedance greater than 50%. However, for site C5, reduction of the probability of exceedance below this value brings about a flattening of the line. In other words, the higher flows at this site are less pronounced in relation to the lower flows than would be expected from a lognormal distribution, and can be said to be curtailed by upstream storage. By comparison, sites C6 and C1 behave similarly to each other and according to the lognormal model except at the very lowest probabilities of exceedance. Here, the line for C1 steepens, indicating more pronounced high flows caused by a lack of upstream storage. This assessment of the duration curves of the daily flows is consistent with the general flow characteristics seen at hourly resolution in Figure 4.11.

Recent literature includes a small number of flow duration curves for wetland catchments. Devito et al. (1996) studied two wetlands lying in small catchments above impermeable bedrock. The catchments differed in the depth of surficial till deposits available for groundwater storage. Thin deposits resulted in a cessation of summer groundwater flow to the first wetland and its stream outlet while the groundwater supply was maintained for the
second wetland which was bordered by thicker sediments. For example, at 70% exceedance, normalised low flows from the catchment with thicker sediments were an order of magnitude higher than those from the catchment with less overburden, illustrating the part played by groundwater in slow flow production in these two catchments. Burt (1992; 1995) illustrated similar differences in the characteristics of runoff from 3 British headwater catchments whose substrates had been characterised variously as impermeable clay (with clay soils), permeable limestone and peat. The corresponding flow duration curves are shown in Figure 4.13 along with the Goss Moor outflow duration curve and those from Devito et al. (1996). Ready percolation and storage of meteoric inputs, with gradual release from the limestone water body, set the flow regime from the limestone catchment apart from the clay and wetland regimes. However, Goss Moor and the wetland catchment with appreciably deep till deposits sustained greater slow flow than the other wetland catchments, with middle-magnitude flows comparable to those from the limestone catchment. Thus, Goss Moor exhibited a less flashy response than a significant range of other headwater wetland catchments. Since Goss Moor is a larger than average wetland, this may have been due to its longer length of dispersive flow path. Equally well, greater depths of available storage in the Goss Moor catchment would account for its steadier flow regime. Further analysis and modelling in the present study will be used to refine and explain this initial assessment of the wetland's hydrology. The following section describes further characterisation by spectral analysis.

4.4.3 Spectral Characteristics of Daily Stream Flows

The flow duration characteristics discussed above give information on the statistical distribution of flow magnitudes over the study year. Apart from the statistics of flow magnitudes, the statistics of the rates of variation of flows are also useful in assessing the character of hydrological systems. Such information is provided by spectral analysis.

The spectral analysis of river flows has been used by some researchers in order to make basic observations on the characteristics of catchments. For example, Hino and Hasebe (1981) utilise changes in the coherence and phase between daily rainfall and stream flow to
estimate the cutoff frequency for a flow separation filter. Papps (1990) uses the localised flow variance of flow data as a detection criterion for quick flow in the Hurunui River, New Zealand, while Angelini (1997) uses plots of coherence, phase, cross amplitude and gain, along with the stream flow spectral densities and autocorrelations, to show the substantial integrating effects of some karst aquifers in central Italy.

The spectral sampling which produced the spectral plots in the present study was performed using SPSS, a statistical analysis software package. A Parzen window 21 days wide was used for smoothing, and the period analysed was the water budget period, 1/9/93 - 31/8/94, wrapped around upon itself to reduce loss of data. The sampling and the discussion below were limited to frequencies below 0.5 cycles per day, the Nyquist frequency for the daily data set.

Evidence of the most prevalent modes of variation of flow down to a period of two days is provided by the plot of spectral densities of daily data in Figure 4.14. In this graph, the spectral power density of each data set has been normalised relative to the total variance in the series. Evapotranspiration data is included along with rain in order to show the degree of influence of hydrometeorological variations. Most noticeably, the degree of periodicity at all frequencies above 0.03 cycles per day is greatest for rain, followed by flows at C1, C5 and then C6. Also, the spectral power of flows at C5 and C6 falls off with increasing frequency at a greater rate than at C1. This is evidence of storage in the source areas for C5 and C6, since such storage eliminates rapid variations in flow. The areally averaged evapotranspiration also exhibits a marked preference for slow, seasonal variations, despite the great variability from day to day in the individual land classes (see Figure 4.3). For rainfall, fast fluctuations account for a much greater proportion of the total variance, resulting in a flatter spectral curve.

Variations in phase between rain and other processes at various frequencies are shown in Figure 4.15. Changes in evapotranspiration maintain a phase of ± π over most of the frequency band, indicating that ET is decreasing when rain is increasing, and vice versa. The phase between variations in stream flow and rain increases steadily with increasing
frequency for C5 and C6, whereas it is approximately constant for the upstream site, C1.

The rate of change of phase with frequency is proportional to the delay between variations in rain and variations in flow. From this phase graph, it appears that variations in rain precede flow variations at C6 by 1.48 days, at C5 by 0.39 days and at C1 by zero days. Since the data was here analysed at a daily resolution, these values are best expressed in whole days, obtaining delays of either one or two days at C6, either zero or one day at C5 and zero days at C1. It may be noted that the two estimates of delay at C6 from the phase plot do not encompass the 14 hour time to peak estimated at the start of Section 4.4.1.

This discrepancy results from the shape of the hydrograph at C6. As seen from the graph of hourly flows in Figure 4.11, the flow at C6 rises to its peak more quickly than it falls away. Hence there is a greater mass beneath the flow curve in the day or so following the peak flow than in the day preceding it, shifting the daily average flows away to the right of the hourly peak and increasing the perceived delay in response.

The delays in the response of stream flow to rainfall cannot be ascribed purely to any one type of flow path. Rather, several parallel flow paths acting within each subcatchment each exhibit their own characteristic delay, the relation of which to the overall hydrograph translation is dependent upon the relative sizes of the respective flow peaks. The possible flow paths all involve the interception and release of rain water by vegetation. Subsequently, the paths diverge into overland flow, throughflow and groundwater flow, whose relative contributions to the hydrology of Goss Moor this study purposed to evaluate. Finally, channel routing may be responsible for a small proportion of the hydrograph translation.

The frequency dependence of the coherency between rain and other processes is shown in Figure 4.16. The contrast in response characteristics between the upstream site, C1, and the other two sites is again highlighted by differences in this property. Slow variations in rainfall produce a stronger response in stream flow at C5 and C6 than do faster variations. This is due to the removal of fast variations by storage in each source area. Variations in flow at C1 behave in the opposite manner, following the faster components of stimulation.
by rainfall and therefore lacking the storage necessary to reflect gradual changes in rainfall regime.

Variations in ET are shown to be independent of variations in rain by virtue of the low level of coherency. However, a small increase in coherency between ET and rain in the range 0.3 - 0.45 cycles per day perhaps indicates that this high frequency of rainfall variations is associated with the strongest variations in sunshine or cloud cover and possibly wind speeds. The restricted section of the flow - ET phase spectrum which shows smoothly varying phase, and the slight upturn in flow - ET coherency (Figures 4.17 and 4.18) in the same frequency range are due to the shared correlation of ET and flow with rainfall. High coherencies of flows with ET and of ET with rainfall at very low frequencies demonstrate the reduction of flow by stronger ET and the reduction of ET by the conditions associated with stronger rain, respectively.

4.5 STREAM FLOW SEPARATION AT THE THREE MAIN STREAM GAUGING SITES (C6, C5 AND C1)

4.5.1 Aims

Chapter 1 established the overall aim of the present study in terms of three complementary questions.

A) How much and what type of flow is contributed to the wetland surface and to its substrata?
B) What are the relative water demands from the various drainage processes on the wetland surface and on its substrata?
C) Does the wetland suffer from more rapid depletion than other parts of the catchment? (The answer would highlight whether remediative work should be performed on the wetland as well as on other parts of the catchment.)
The stream flow recession analysis described in the current section aims to provide the basis for answering the third of the questions above. By deriving slowly varying and quickly varying flows both upstream and downstream of the wetland, a comparison can be made between storage depletion in the upstream and downstream source areas. Firstly, a comparison is made at the end of Section 4.5 in terms of hydrophysical concepts. Secondly, a quantitative comparison with other components of the water budget is undertaken in Chapter 7.

The foregoing section established the fact that the flows at the three stream gauging sites in Goss Moor exhibit different hydrograph shapes and response times. Dissimilarity could also be seen in the duration curves and the spectral distributions of variations in daily mean flows. These observations suggest that the proportions of the total flow in the channel attributable to slow flow and quick flow, and therefore to slow and fast flow paths in the source area, would differ between the sites.

The current section estimates these proportions for each of the three sites and to discuss the uncertainty in the proportions so obtained. A review of methods of stream flow recession analysis provides explanation for the method adopted in the present study and elucidates the applicability of such methods to the investigation of flow paths other than groundwater. It is shown that, while the technique is applicable to the evaluation and comparison of storage in source areas irrespective of the surface/groundwater nature of such storage, the determination of the nature of the storage requires additional investigation either by hydrochemical analysis or by the testing of competing hypotheses using numerical modelling. The possible alternative flow paths in the Goss Moor catchment, whether over land, through the unsaturated soil zone or through the saturated soil zone, are considered in relation to the two derived flow components. In the subsequent chapters of this thesis, evidence distinguishing between these flow paths is considered with the hypothesis testing method.
4.5.2 Use of a Digital Filter for Stream Flow Separation

In this study, slow flows are estimated using the non-oscillatory spectral filter put forward by Hino and Hasebe (1984). This is a low-pass filter that is functionally equivalent to a mass-spring-dashpot system. The output from this system in response to an instantaneous impulse of unit strength is given by $w(\tau) \text{ (hours$^{-1}$)}$, the impulse response. The system's response, $q_B(t) \text{ (m$^3$s$^{-1}$)}$ to extended stimuli $q(t) \text{ (m$^3$s$^{-1}$)}$ is obtained by convolution of the impulse response with the stimuli:

$$q_B(t) = \int_{0}^{\infty} w(\tau) \cdot q(t - \tau) \, d\tau$$

(4.2)

where

t is time elapsed (hours) and
\( \tau \) is a dummy time variable (hours).

In discretised form,

$$q_B(t) = \sum_{k=0}^{\infty} w(k \cdot \Delta \tau) \cdot q(t - k \cdot \Delta \tau)$$

(4.3)

where

\( \Delta \tau \) is sampling interval (hours).

In general, $w(\tau)$ may be oscillatory, which is undesirable for a hydrologic data filter. However, when the level of damping is great enough, giving \( c^2 > 4k \), the impulse response takes the form

$$w(\tau) = e^{-\frac{1}{2}c \cdot \tau} \cdot \left[ \frac{1}{4} c^2 - k \right]^{-\frac{1}{2}} \cdot \sinh \left( \left[ \frac{1}{4} c^2 - k \right]^{\frac{1}{2}} \cdot \tau \right)$$

(4.4)

where

c \text{ (hours$^{-1}$)} is analogous to the damping constant, and
k \text{ (hours$^{-2}$)} is analogous to the spring force constant.

This function is non-oscillatory. As applied in the present study to river flows at gauging station C6, it takes the form shown in Figure 4.19. The convolution of this function with
the flow record is a form of weighted time-averaging, smoothing off the flow variations and
thus attenuating the higher frequency Fourier components.

Hino and Hasebe (1984) determine the two parameters c and k from semilogarithmic plots
of slow flow recessions at the monitoring station. They note that after a long intermission,
the impulse response \( w(\tau) \) above may be approximated by

\[
\frac{k \cdot e^{-\frac{\tau}{c}}}{2\left[\frac{1}{4}c^2 - k\right]^\frac{1}{2}}
\]  

(4.5)

They equate this with the exponential slow flow recession such as that shown in
semilogarithmic form in Figure 4.20. The slope of the straight line segment at the end of
the recession gives the decay constant \( s \) (hours\(^{-1}\)) in the descriptive equation

\[
q(\tau) = A \cdot e^{-s \cdot \tau}
\]  

(4.6)

where

\( A \) is a constant magnitude of flow (m\(^3\)s\(^{-1}\)).

Using the condition of overdamping and pre-setting a value for the dimensionless damping
factor \( \delta = \frac{c}{\sqrt{k}} \), > 2, then

\[
c = \delta^2 \cdot s \quad \text{and} \quad k = (\delta \cdot s)^2
\]  

(4.7)

The convoluted flows, \( q_B(t) \), are taken to be the slow flow, while the residual flow in m\(^3\)s\(^{-1}\),

\[
q_R(t) = q(t) - q_B(t)
\]  

(4.8)

is adopted as the faster runoff. It was found necessary in the present study to experiment
with the values of \( \delta \) and to scale the slow flow to ensure that the residuals did not go
negative. That is, the filtering operation actually implemented was modified from Equation
4.3:
\[ q_B(t) = \alpha \cdot \sum_{k=0}^{\infty} w(k \cdot \Delta \tau) \cdot q(t-k \cdot \Delta \tau) \]  \hspace{1cm} (4.9)

where

\( \alpha \) is a positive scaling factor (dimensionless), chosen so as to satisfy the condition that the residual output \( q_R(t) \) should not be negative.

Since the damping factor \( \delta \) was not directly governed by any hydrological constraints except that of non-oscillation of the filter, it could be adjusted to improve the fit of the output slow flow to the original flow recession. This adjustment affected, and was affected by, the above scaling and thus \( \delta \) and \( \alpha \) were interdependent. Eventually, not only \( \delta \) and \( \alpha \) were adjusted for this purpose, but also the "cutoff frequency" or decay constant \( s \). The rationale for this is given in Section 4.5.3 when discussing the amplitude spectrum of the stream flow at the Goss Moor catchment outflow in relation to the amplitude spectrum of an exponential decay and the gain spectrum of the filter, and reflects the fact that the stream flow recession constant can only be used as an initial estimate of the appropriate "cutoff frequency".

The problems encountered and the results of this flow separation for the Goss Moor data are discussed in the next section.

4.5.3 Alternative Approaches and Physical Interpretation

In the present study, slow flow separation was achieved using the above non-oscillatory spectral filter applied to the observed total flow at each stream gauging site. This approach followed from many theoretical and practical analyses of the drainage behaviour of aquifers. Tallaksen (1995) has reviewed slow flow recession studies, classifying them according to their methods of deriving the recession curve. Three methods are mentioned:

1) Derivation from intrinsic flow equations (with simplifying assumptions)
2) Modelling recession as reservoir outflow
3) Trial of empirical formulae (curve-fitting).
Category (3) methods are not discussed here. The following is a discussion of the remaining two categories. It will be seen that slow flow analysis suffers from the complexity of theoretical groundwater drainage, which may at times mask the complexity of the assemblage of source units within the catchment.

The mathematical studies in Category (1) follow after the work of Boussinesq (1877), who derived the basic differential equation governing two-dimensional flow in an unconfined aquifer. This equation is non-linear since the thickness of the flow domain is a function of the groundwater head. The linearised form of the equation states that

\[ S \cdot \frac{\partial h}{\partial t} = \nabla \cdot [T \cdot \nabla h] + N \]  

(4.10)

where

- \( t \) is elapsed time (days),
- \( h \) is the groundwater head (m),
- \( S \) is the specific yield of the aquifer material (dimensionless),
- \( T \) is the transmissivity of the aquifer (m\(^2\)/day) and
- \( N \) is the recharge rate less abstraction rate (m/day).

This equation is applicable only when the groundwater flow is approximately horizontal.

The recession of a water table during a dry spell is approximated by applying Equation 4.10 to an initially saturated rectangular aquifer with three impermeable sides, draining to a fully penetrating stream along the fourth side. Brutsaert and Nieber (1977) give the resulting flow per unit length of channel in a form similar to:

\[ q_B = C \cdot \sum_{n=1,3,5,\ldots} \left[ \frac{(n-\pi)^2 \cdot T \cdot t}{4S} \right] \quad (t > 0) \]  

(4.11)

where \( C \) (m\(^2\)/day) is proportional to the transmissivity, the inverse of the aquifer width and the initial head difference between aquifer and stream. Thus, the slow flow recession in this idealised situation is a superposition of many exponential decays, with the slowest decay becoming dominant after some time (Nutbrown and Downing, 1976). In many slow flow studies, the smallest decay constant, obtained by plotting the recession semilogarithmically
as in Figure 4.20, is taken to be fully representative of the groundwater recession. For example, Padilla et al. (1994) investigated the relative importance of slow flow and quick flow from four European karst springs. They successfully used the above technique to show that the relative contribution by slow flow, and hence by the saturated rock matrix, is reduced by greater connectivity in the karst's channel network. Mau and Winter (1997) used an Institute of Hydrology (1980a,b) procedure based partially upon the same approach, in a comparison of slow flow and recharge in two small montane catchments.

The use of a single exponential decay to represent the slow flow recession corresponds to the simplest of the reservoir recession models (Category 2). These models involve linking several conceptual reservoirs together in a combination of series and parallel arrangements (Chow et al., 1988). Numerous lumped-coverage rainfall-runoff models have utilised this system of conceptualisation, including the Dawdy-O'Donnell Model, shown in Figure 4.21, and the Stanford Watershed Model (Fleming, 1975). It is common to find conditional routing of flows along alternative pathways in such models, as thresholds of current or antecedent rainfall or flow are used to induce different response characteristics. This approach arises through the need to represent different non-simultaneous modes of response, such as snow storage and ablation, interception by vegetation, or depression storage. Such devices are not discussed in the present study since the ability to detect thresholds in the hydrograph is still a matter for debate, thus making the inverse estimation of the parameters from hydrograph analysis rather difficult.

Each reservoir in the model is subject to the continuity equation

\[
\frac{dV(t)}{dt} = I(t) - Q(t) \quad (4.12)
\]

where

- \( V \) is the volume of stored water (m\(^3\)),
- \( I \) is the flow into the reservoir (m\(^3\)/day) and
- \( Q \) is the flow out of the reservoir (m\(^3\)/day),
and behaves according to a specified relation between its stored volume and its discharge. In the case of a linear reservoir, this relation takes the form
\[ V(t) = \kappa \cdot Q(t) \]  
(4.13)
\( \kappa \) being known as the reservoir constant (days).

A collection of such reservoirs in parallel corresponds to the selection of the appropriate decay terms from the series in Equation 4.11. In order to correspond exactly with the truncated series, the reservoirs must all be subject to the same input. However, in general, the rainfall input may be distributed unequally between the parallel reservoirs.

A collection of \( n \) linear reservoirs in series has the global response \( Q(t) \) to rainfall input \( I(t) \), where
\[ \prod_{i=1}^{n} \left( 1 + \kappa_i \cdot \frac{d}{dt} \right) Q(t) = I(t) \]  
(4.14)
(Spolia and Chander, 1974). This equation can be solved using the Laplace transformation to give the impulse response of the system, in days\(^{-1}\):
\[ u(t) = \sum_{i=1}^{n} \left( \kappa_i^{n-2} \cdot e^{-\frac{t}{\kappa_i}} \cdot \left( \prod_{j=1}^{n} \left( \kappa_i - \kappa_j \right) \right)^{-1} \right) \]  
(4.15)
for \( j \neq i \).

Spolia and Chander (1974) have shown how the above type of linear reservoir model in continuous time can be related to an autoregressive moving average (ARMA) model in discretised time.

Although similar in appearance to Equation 4.11, Equation 4.15 differs firstly through the inequality of the factors in the summation terms, and secondly in that one or more of these factors may be negative. This delays the impulse response, producing a rising as well as a falling limb. As with the other models of slow flow from a linearised aquifer, this linear reservoir cascade model gives a single exponential decay term after an extended wait.

A major difference between reservoir models, such as the above, and recent theoretically-derived recession curves, is in the initial time from which the hydrograph is considered. While the theoretical models currently address only the hydrograph recession, assuming that
the time of concentration has passed and thus ignoring the rainfall characteristics, the reservoir outputs are expressible in terms of the time since the last rainfall event and its intensity. Hence, the output of a reservoir-based model, as long as it possesses more than one reservoir in series, may represent the whole hydrograph and can be used in the determination of unit hydrographs. Such an assemblage of reservoirs in series could also be fitted to each spectrally filtered flow component produced in the present study to produce separate unit hydrographs for the slow flow and quick flow.

The foregoing discussion indicates that Boussinesq’s equation plays a major part in the rationale for loglinear analysis of slow flow recessions. This analysis applies only to the tail end of the Boussinesq response. The presence of faster-decaying terms in Equations 4.11 and 4.15 invalidates such a simple analysis for the earlier part of the drainage curve (Singh and Stall, 1971; Brutsaert and Nieber, 1977). Nevertheless, numerous researchers have extended the application of the loglinear analysis to earlier stages in the recession. In certain catchments and for certain kinds of rainfall event, the storm hydrograph recession appears divisible into more than one interval of exponential decay, each with its own distinct decay constant. The earlier loglinear periods are taken to be due to the final drainage of less retentive water storage units such as more permeable aquifers, soils or entrainable ice.

For example, Mulholland (1993) interrelates some loglinear sections of hydrograph recession in a small, forested, experimental catchment so as to postulate drainage from three water stores: the unsaturated soil zone, a saturated soil zone and a saturated “bedrock” zone. Hydrograph separation is also suggested by an end-member analysis of stream Ca$^{2+}$ and SO$_4^{2-}$ concentrations. However, the data presented show no synchronisation between the chemically derived and the hydrometrically derived hydrograph components. Moore (1997) adopts a model with two linear reservoirs in series for the fitting of a non-linear hydrograph recession. This model produces a better fit to the observed hydrograph than do alternative single non-linear reservoir models, and also provides wider scope for physical interpretation: the researcher relates the two reservoirs conceptually to a downslope, near-saturated, streamside soil zone and an upslope, unsaturated soil zone. Jakeman et al. (1990) fit alternative z-transfer, or ARMA, functions to the total hydrograph (including the
rising limb) of a small Welsh catchment. They relate the different components of the optimum transfer function, describing a parallel pair of linear reservoirs, to fast and slow responses in the hydrograph. With regard to this study, Young and Beven (1991) state that the exact correspondence of these conceptual reservoirs with true hydrograph components is limited by even small uncertainties in the values of the ARMA coefficients. Furthermore, the observed hydrograph peaked after a delay of only one time step due to the small size of the catchment. The number of alternative sets of parameters, corresponding to alternative reservoir assemblages, would be far greater for larger catchments in which the rising limb of the storm response was more gradual and necessitated a series configuration. Given that some uncertainty in parameter values will always remain after model optimisation, this fuels doubts about whether observed hydrographs can reliably be separated into different flow components on the basis of the parameters in the best fit transfer function.

Further doubt is cast on the physical interpretation of multiple linear reservoirs by Eisenlohr et al. (1997), who address the hydrograph analysis of numerically simulated karst aquifers exhibiting various degrees of heterogeneity. They find it possible to infer from the hydrograph more types of aquifer material than are really present in the numerical models. Furthermore, although the slow flow recession obtained in their hydrograph analysis coincides with the depletion of the lower-conductivity zones of the aquifer, the decay constant for this recession does not depend solely on the hydraulic properties of these zones, but also on the domain’s geometry and the properties and distribution of other zones in the aquifer, thus reflecting the global configuration of the karst. This renders unfeasible any inter-basin comparison of porous media properties through differences in recession constants, and also highlights the possibility that in any particular catchment, some faster parts of the recession may perchance correspond to the drainage of more storative units as a result of their relation to the stream channel network and/or position within the catchment’s stratigraphy. Nutbrown and Downing (1976) encapsulate doubts on simple hydrograph separations by demonstrating numerically that even a structurally simple aquifer exhibits apparent alterations in the speed of recession, due to a gradually changing emphasis in the superposed exponential decay terms making up the solution to the linearised Boussinesq equation.
Given that all but the very tail end of the hydrograph recession poses such difficulties in interpretation, the method of digital filtering proposed by Hino and Hasebe (1984) can be seen to offer a way of circumventing many problems by shifting attention from the time domain to the frequency domain. As illustrated in Section 4.4.3, the spectral analysis of flow time series can provide information on the modes of variation of flow in the hydrological system. The use of spectral criteria is thus a natural technique in the determination of fast and slow flow components.

The amplitude spectrum of an exponential decay is shown in Figure 4.22. This particular example uses the decay constant determined from the semilogarithmic plot of the flow recession at the Goss Moor catchment outflow (site C6). The figure also shows the amplitude spectrum of the complete outflow and the gain spectrum of the digital filter used in the present study to separate slow flow at this site. Within the relevant frequency band 0 - 0.08 hours⁻¹ (equivalent to periods greater than 12 hours), less emphasis of low frequencies is found within the spectrum of the complete flow than in the spectrum of the exponential decay. The filter passes not only the slowest recession component, from which its cutoff frequency was initially derived, but also a limited range of the faster components in the hydrograph, thus taking into account higher terms in the Boussinesq series. The cutoff frequency finally used in this filter was revised upwards from 0.001793 hours⁻¹ to 0.0035 hours⁻¹ during the process of fitting the derived slow flow to the observed hydrograph. This is possibly due to the presence of quicker terms in the slow flow.

Although the above discussion is expressed in terms of the solution to the Boussinesq equation, the general conclusion that there may be a need for accepting slightly higher elements of the hydrograph spectrum with the tail end component is not limited to a groundwater-produced slow flow, but may also apply to other slow flow-producing mechanisms. Consequently, the physical identity of the separated slow flow is still very much in doubt. These observations may be applied to the present study site: while the approach of hydrometric slow flow separation is an undeniably invaluable tool in
ascertaining the degree of rainfall integration or storage in the catchment, it still leaves open the choice of source flow paths for the quick and slow flow components.

Additionally, the technique suffers from difficulties with non-uniform catchments, source areas and times of travel over the stream network. Jakeman et al. (1990) mention that the treatment of a catchment in a lumped fashion, as done in hydrograph analysis, effectively assumes uniform infiltration capacity, rainfall and rainfall intensity. However, as long as the spatial pattern of the infiltration capacity is random, then the infiltration characteristics of the various parts of this pattern in the catchment are not likely to have distinguishable individual effects on the slow flow hydrograph. Fitzjohn et al. (1998) have shown that the variations in hydraulic properties between neighbouring areas of soil tend to cancel each other out in the transmission or infiltration of runoff. However, in the case of large scale, spatially systematic variations in soil properties, the response of each distinct zone is less affected by the responses of neighbouring zones. If such zones are large enough to capture a significant fraction of the catchment’s rainfall, the potential for a multicomponent hydrograph may be enhanced. In catchments where such zonation is deemed likely, this increases the number of possible explanations for a given form of recession, adding to the problem of attributability of quick and slow flow components.

Since flow recession analysis cannot identify the origin of the slowly varying and quickly varying components of the stream flows, an additional technique must be applied for this purpose. Two methods considered here for this purpose are hydrochemical analysis and hypothesis testing using numerical modelling. Hydrochemical analysis for the determination of contributing flow paths involves the measurement of naturally occurring conservative solute concentrations in the stream and in the various possible source bodies at various times in the stream hydrograph. As long as the flow paths exhibit sufficient differences in the concentrations of the analysed solutes, concentrations of these solutes in the stream water will indicate their relative contributions. This technique has been used many times in the past to determine flow paths to water courses before, during and after rain events.
For example, given a conservative solute which has concentrations $C_{\text{old}} \text{ kg/m}^3$ in “old” water, $C_{\text{new}} \text{ kg/m}^3$ in “new” water and $C_{\text{stream}} \text{ kg/m}^3$ in the stream, the stream flow is made up of “old” and “new” water in the proportions dictated by the following equations.

\[ Q_{\text{stream}} = Q_{\text{old}} + Q_{\text{new}} \]  
\hspace{1cm} (4.16)

\[ C_{\text{stream}} \cdot Q_{\text{stream}} = C_{\text{old}} \cdot Q_{\text{old}} + C_{\text{new}} \cdot Q_{\text{new}} \]  
\hspace{1cm} (4.17)

\[ \Rightarrow \frac{Q_{\text{old}}}{Q_{\text{stream}}} = \frac{C_{\text{stream}} - C_{\text{new}}}{C_{\text{old}} - C_{\text{new}}} \]  
\hspace{1cm} (4.18)

where

- $Q_{\text{stream}}$ is the total flow in the stream (m$^3$s$^{-1}$),
- $Q_{\text{old}}$ is the flow of “old” water to the stream (m$^3$s$^{-1}$), and
- $Q_{\text{new}}$ is the flow of “new” water to the stream (m$^3$s$^{-1}$).

The solute concentration in each flow path is typically sampled at regular time intervals on a sub-event time scale to allow calculation of the variation of the different flow contributions throughout a flow event. Sampling must be performed very close to the stream in order to eliminate differences in travel times. In general, the number of different solutes used must be one less than the number of flow pathways to be inferred.

Example studies include those of Dewalle \textit{et al.} (1988), Ogunkoya and Jenkins (1993), O’Brien and Hendershot (1993) and Hinton \textit{et al.} (1994). Dewalle \textit{et al.} used $^{18}$O concentrations to separate groundwater and soil water contributions, also compensating for direct precipitation into the channel. Hinton \textit{et al.}, using $^{18}$O and SiO$_2$ concentrations in a three-pathway equation, found that groundwater contributions in a glacial till headwater catchment were significant despite till conductivities from 1 m/day down to $10^{-4}$ m/day. A comparison of the timing of the groundwater contribution with that of the stream flow event, along with piezometric observations, led the investigators to conclude that groundwater was flushed into the stream during a rain storm after building up in the upper soil horizons prior to the event. O’Brien and Hendershot addressed a similar situation in a...
first-order catchment with humic soils underlain by thin glacial tills. They drew attention to the fact that certain solutes which may be conservative in one groundwater regime may be reactive in another. Innovatively, they analysed both conservative and non-conservative solutes (in this case, a range of solutes of apparently various degrees of reactivity, including acid extractable aluminium, monomeric aluminium, hydronium, fluoride, magnesium and dissolved silicon) to determine the amount of flow which groundwater had contributed to soil water before entering the stream. Ogunkoya and Jenkins assessed uncertainties in estimated flows from groundwater, soil water and incident precipitation occurring due to spatial variations in end-member chemistry and due to differing transit times of the solutions. They reported errors of between 11% and 50% for each component of a particular storm response.

The above demonstrates some of the capabilities and limitations of hydrochemical analysis applied to stream flow separation. This type of analysis is aimed at identifying the source bodies rather than their rates of depletion, and so will not address the evidence regarding their propensity to store rather than release water which is required in the present study. However, the identification of the source body of the slowly varying flow, whether groundwater or surface water, was also necessary for a principal aim of the study, that of focusing land management measures on the most appropriate hydrological subsystem for the protection of the wetland. For this purpose, hydrochemically based stream flow separation was an alternative to the technique of hypothesis testing by numerical modelling of groundwater flows.

Numerical groundwater modelling may provide estimates of the groundwater flows to the stream by virtue of its solution of the groundwater flow equation given boundary conditions representing the alluvial aquifer and stream system. The present study used this technique in preference to hydrochemical stream flow separation for two reasons. Firstly, the modelling effort would involve the consideration of information on many aspects of the wetland's hydrological system such as evapotranspiration, sediment conductivity, bedrock depth, stream channel geometry and drainage at the ground surface, and thus would provide a framework for the assessment of the effects that these factors had on the wetland's
hydrology. Secondly, the groundwater model so developed could be used at a later date to investigate the effects of changing such characteristics on the water table underlying the wetland. This would facilitate the planning of future watershed improvements. Information on the applicability of numerical groundwater modelling in wetland situations is given in Section 6.2.

4.5.4 Results of the Stream Flow Separation

The determination of which flow paths were responsible for the fast and slow components of the river flow would indicate the nature of water flow through the Goss Moor wetland. Later stages of this thesis describe the numerical modelling and water budget analysis which was undertaken with the aim of identifying or eliminating the general candidate flow paths. The current section assesses the results of the stream flow separation at the three stream gauging sites in terms of the potential errors incurred during implementation and gives alternative, qualitative interpretation of the nature of the separated flows in the context of the hydrogeomorphology of Goss Moor.

Consistency of results in the stream flow separation

Before allocating filtered river flow components to any particular flow paths, it is best to confirm that the relative sizes of these components are consistent with the known hydrograph characteristics. Possible measurement errors must also be taken into account. In view of this, the characteristics of the separated stream flows are inspected here. An initial exploration of the possible interpretations can then be made, as proposed in Section 4.5.1.

As mentioned in Section 4.5.2, the “cutoff frequency”, or decay constant, of the low-pass filter for the hourly catchment outflow at site C6 was initially taken from the slope of the loglinear recession. The damping factor $\delta$ was initially set at 2.5 as used by Hino and Hasebe (1984). The two parameters were then adjusted, along with a linear scaling factor (see Section 4.5.2) to ensure that the calculated slow flow exceeded the total flow in as few instances as possible during the whole flow record, whilst also maintaining the closest
possible fit in the recession period. Slow flow was then subtracted from total flow to obtain quick flow. The same procedure was undertaken for the separation of hourly flows at site C5, the intra-wetland site, and CI, the site of inflow to the wetland. For proper operation of the filter over the period of interest, the total length of the input flow data sequence had to include a run-up period equal to the length of the filter, as shown in Table 4.4. Flows recorded at site CI were not available for a sufficiently long time before the water budget period to allow correct separation of the flow components before 15:00 hours, 25/10/99. This resulted in the gap of "missing data" shown in Figure 4.29.

The parameter values found by the above method are shown in Table 4.4. Plots of the daily means of the separated stream flows over the study year, and at hourly resolution for the considered winter and summer periods, are shown in Figures 4.23 - 4.31. For both C5 and C6, the cutoff frequency used in the filter was higher than the recession constant of the initially analysed recession period. This reflects the fact that the recession constant must be used only as an initial estimate of the cutoff frequency, there being no direct physical correspondence between the two constants. The recession constant is the frequency which is characteristic of the later stages of the recession alone, and does not account for earlier stages of the recession which exhibit higher characteristic frequencies (as discussed in the previous section). Furthermore, the damped-oscillator filter is not physically related to the stream flows, and so the form of its gain spectrum is not necessarily similar to that of the flows. Compensation for this involves adjustment of the filter's parameters.

At site CI, the cutoff frequency decreased below the recession constant. This was due to the initial choice of recession period (8/6/94 - 17/6/94), which can be seen in Figure 4.29. Attempts to fit the filtered flows using cutoff frequencies as high as or higher than the recession constant from this period resulted in gross exceedance of the observed flow by the filtered flow for most of the year. This indicated that the recession in question contained the same spectral components as the higher, peakier flows at the site, which were not gradually varying. The filter parameters were therefore adjusted to give the highest possible proportion of gradually varying or slow flows, which were found to be expressed in recession periods such as that at the end of April and that during July 1994.
The exceedance of observed flows by the filtered slow flow was kept to a minimum as far as possible during the scaling of the filter, as considered in Section 4.5.2. Some exceedance nevertheless was allowed to occur occasionally at sites C5 and C6, although ideally it should not have occurred. This was due to the importance of maintaining a correspondence between the filtered flows and the observed recessions in as many cases as possible since there would otherwise be no criterion for the determination of the gradually varying proportion of total flow.

Low flows at site C5 were reduced by the installation of gabions for channel bed protection shortly after the installation of the stage recorder. At times of low flow, water would flow down a nearby upstream distributary in preference to the main drainage channel, due to obstruction of flow in the main channel by the gabions. This led to the very low summer flows and the abrupt cut-off of storm event recessions in summer and early autumn. Some compensation for this was achieved through the above-mentioned exceedance of observed low flows by the filtered slow flow at certain times. Nevertheless, as shown in Table 4.4, C5 had the lowest proportion of slow flow among the three stream monitoring sites, despite exhibiting a far less flashy character than C1. The estimate of slow flow for C5 was therefore taken to be very conservative, although the discrepancy may also have been due in part to the overestimation of C1 slow flows through a malfunction in the stage recorder, as mentioned in Section 4.4.2.

Also shown in Table 4.4 are the annual and seasonal mean percentages of the total flow accounted for by slow flow and by quick flow. These results are discussed in terms of hydrophysical processes in the next subsections. Their significance for the water budget of the catchment and the wetland is considered in Chapter 7.

Notwithstanding the above problems at C5 and C1, there was no reason to suspect the same complications in the flow separation at site C6, the catchment outflow, and so the values shown in Table 4.4 for C6 were taken to be valid. Therefore, a value of 51.2% could be adopted as the annual proportion of river water from the catchment which had at some
stage during its transit been incorporated in a reservoir with significant residence time. The equivalent figure for site C1, 40.7%, might be taken as an upper limit in view of the possible overestimation of C1 slow flows. The true figure would therefore be smaller and so the subcatchment at C1 might be seen to be biased in favour of quick flow when compared with the overall catchment. The value of 23.1% for C5 was assumed to be in need of an upward revision in light of the aforementioned low flow problems.

Limitations of the stream flow separation procedure

Besides the above problems of site alteration and equipment malfunction, the stream flow separation suffered two omissions from its procedure. These are mentioned here for completeness although their effects are probably very much outweighed by the practical problems described above.

1) Seasonality in the Slow flow Parameters

Earlier in this chapter, researchers were cited (Nutbrown and Downing, 1976; Eisenlohr et al., 1997) who revealed that changes in the semilogarithmic slope of the stream flow recession cannot necessarily be attributed to the depletion of aquifer units with different properties. Nevertheless, the properties and configuration of the draining storage zone affect the form of the stream flow recession, albeit perhaps not in a way amenable to loglinear analysis. Consequently, the slow flow may recede differently at different times of the year due to summer contraction of the draining reservoir and the consequent migration of its centre of mass through areas with different hydraulic characteristics. Such seasonality of the slow flow recession was not investigated for Goss Moor, the digital filter being fitted for an overall optimum fit, rather than locally optimum fits, of the output slow flow to the observed flow recessions during the study year. However, this approach minimised the error in annual average slow flow. Regarding the seasonal accuracy of the slow flow, the random, small-scale nature of the heterogeneity in the storage body may have provided some mitigation of the potential errors.
2) Variations in Evapotranspiration

Evapotranspiration may cause particularly marked seasonality in the form of slow flow recessions. Daniel (1976) showed that the semilogarithmically plotted recession line exhibits downward curvature under the influence of ET. Although the recession analysed in the present study showed no such behaviour, it is known that the catchment ET rates vary seasonally by as much as a factor of two. Hence, the degree of involvement of ET in each recession may vary according to season, making the annually optimum derived slow flow misrepresentative of those recessions in which ET played a greater or a lesser role.

Notwithstanding the above considerations, the derived slow flows at all three stream monitoring sites showed no seasonal discrepancies with the recessions and so these problems appeared to be unimportant in the current study.

Seasonality of the stream flow separation

The seasonality of the catchment's flow characteristics is highlighted by seasonal differences in the runoff:rainfall ratio and the relative fractions of slow flow and quick flow (Tables 4.4 and 4.5). In general, slow flow assumes greater importance at all sites during the summer, at the expense of quick flow. This is explicable in terms of seasonal changes in soil moisture, but the exact mode of influence of such changes depends upon the physical nature of the flow components in question. The seasonality of the flow separation is considered below, with respect to the two physical flow paths thought possibly to be responsible for quick flow in the study catchment.

At all three stream gauging sites, a great proportion of the winter flow volume is quick flow. Even at hourly resolution, as used in the graphs of the three-month periods, the quick flow does not fully recede between winter rainfall events. This illustrates the wetness of the catchment at this time of the year. Two major types of flow path are possibly responsible for quick flow in the Goss Moor catchment - overland flow and flow through the region of

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soil just above the steady state water table. The latter may involve throughflow or groundwater ridging (Burt, 1992), or may be a result of a fragmentation zone with joints and fissures in the upper layers of the bedrock. There is good evidence in the literature (see Section 2.4.2) that such fragmentation occurs around Goss Moor, and assuming that the fissures in this zone remain, to a degree, unfilled, then it may provide a highly transmissive flow path down the hill slopes.

For those areas in which the quick flow equates with overland flow, the long term presence of quick flow corresponds to the maintenance of large areas of saturated soil in which the soil water tension is insufficient to cause infiltration. This is caused by a high frequency of rainfall, as suggested in Section 4.2. The proportion of rainfall which forms groundwater recharge is lower in such circumstances than in times of less frequent rainfall and so quick flow constitutes a greater fraction of the subcatchment outflow, as shown in Table 4.4. This factor and the seasonal variation in rates of evapotranspiration both contribute to a runoff:rainfall ratio which is higher in winter than in summer, as shown in Table 4.5.

For areas in which percolation rates are high enough to maintain strong soil water tension and hence high infiltration rates despite a high rainfall frequency, the quick flow does not correspond to overland flow but comes from throughflow, via conduits in the soil or the fragmentation zone of the granite, or groundwater ridging. In such cases, the quick flow fraction of the subcatchment outflow and the runoff:rainfall ratio are higher in winter than in summer, just as with the areas of overland flow, but for different reasons: water percolates down more slowly and suffers greater longitudinal dispersion in the summer than in the winter due to the lower soil moisture content and consequent loss of conductivity. This reduces the abruptness of recharge to the underlying saturated zone and therefore reduces the probability of groundwater ridging events. More gradual recharge to the fragmentation zone of the granite will produce less retardation of the flow response than in a Darcian porous aquifer layer, but some retardation will result nevertheless, due to the irregular and partially sedimented nature of the conduits between the weathered fragments. In addition,
infiltrating water is used to maintain the soil moisture profile against evapotranspiration, rather than to recharge the saturated zone, during the summer.

Relevant catchment characteristics

The review of slow flow separation methods in Section 4.5.3 demonstrated the high degree of uncertainty in determining the exact mathematical form of the slow flow recession and therefore in establishing which process is producing the slow flow. This leaves the identity of the slow flow open to interpretation according to independently determined physical constraints. Such constraints include water balance restrictions, piezometric variations, modelling results and chemical balance calculations. In addition, the geomorphology of the catchment may be used as an indicator of what flow paths to expect.

1) Hill Slopes

As described in Section 2.6.1, the Goss Moor catchment forms a depression with peripheral slopes of between 5% and 15% and gradients of less than 1% in the central wetland. Section 2.4.2 details the evidence concerning the occurrence of weathering in the upper layers of the surrounding granite hills. From this evidence, a zone of fragmented and jointed rock with a significant basal dip towards the centre of the catchment is fairly certain to surround the wetland at higher elevations and to act as the main aquifer in the outer catchment. While its storage capacity may be in some doubt, a tendency to provide rapid conduction of water down the hill slopes can be expected.

The high slopes found near the catchment boundary generally encourage the swift descent of water into the central region of the catchment. However, the speed of water transmission down such slopes may vary according to the nature of the aquifer and the resident soils. Among those areas in which most rainfall infiltrates, highly permeable soils together with a transmissive aquifer enhance the speed of response: water percolates rapidly through the soil into the aquifer, in which it swiftly responds to the large potential gradient towards the basin. Reduction of the permeability of the soil or the weathered layer would, in general, result in a slower response to rainfall, although highly conductive soils might maintain ease of transmission despite a poorly permeable regolith. Areas of soil of very low
permeability would encourage a substantial amount of the incident rainfall to run off as overland flow. The same effect would arise from areas of perched saturation caused by blockages or topographic hollows in the hillslope regolith.

The soils in the Goss Moor catchment outside the wetland, described in Section 2.5, include the Manod and Moorgate soil associations at the catchment periphery, composed mainly of freely draining podzols. These soils may allow swift percolation into the weathered granite layer which, due to the steep slope of its impermeable base, may rapidly transmit water downslope towards the wetland and may remain unsaturated for most of the year. The steep slopes on the boundary may also encourage overland runoff from areas of perched saturation, iron pan (Hexworthy soil association) and poorly permeable soils (Hafren soil association). Such runoff is unlikely to remain above ground if it encounters areas of more permeable soil, and so may join the transient subsurface flow body.

2) Central System of River Pools

As shown in Figure 4.11, the rate of flow of water past gauging station C1 often exceeded that past the catchment outlet C6 during storms. This illustrates the effect of the retardation of flow in the intervening stretch of river, which was therefore a primary cause of the observed differences in flow characteristics between the two stations. The principal area in which retardation may occur must be the central river pool system, shown in Figure 2.1, which provides reservoir storage for the river water. Therefore, the observed level of slow flow in the catchment outflow may not entirely be attributable to storage units outside the river channel. It is noted, nonetheless, that the hydrograph recorded at gauging station C5, receiving water from the northern section of the catchment, exhibited as much smoothing as did that at the catchment outflow C6, even though involving no instream pool storage.

3) Drainage Ditches and Wetland Water Store

The interception of hill slope runoff by drainage ditches and streams provides the fastest flow path from hill slope to river and therefore is responsible for the sharpest variations of river quick flow. The difference between the flow characteristics at C5 and C1 may be partly explained by the fact that the subcatchment of C1 contains more hillside drainage
ditches than that of C5. However, if not intercepted by a drainage ditch, hillslope runoff may eventually reach the zone of permanent aquifer saturation to become part of the gradually varying subsurface store. This storage body extends through the lower regions of the catchment, including the lower parts of the subcatchments draining to both upstream gauging stations. The storage level therefore exceeds the water level at each station and must provide at least some of the slow flow.

Overland flow is the dominant rapid flow mechanism in the permanently saturated wetland. As soon as subsurface runoff from the hill slopes contributes to a rise of head in the surrounding parts of the alluvial aquifer, it may encourage an increase in flow from springs or seepage zones at the edge of the wetland which contribute to wetland surface flow (Chappell, 1990). In this way, it may contribute to the quick flow in the river. However, the gentle slopes at this level in the catchment and the possibly large hydraulic roughness of the wetland vegetation and terrain do not suggest particularly high rates of overland flow, and so the very rapid response of the hill slope flow is lost upon entering the wetland. The degree of retardation may be enough to produce slow flow. Apart from the retardation of flow in the central system of river pools, this is the main reason why the slow flow from the study catchment cannot definitely be attributed to groundwater flow beneath the wetland.

4.6 OBSERVED GROUNDWATER HEADS

4.6.1 Introduction

The present section analyses the response of the wetland groundwater heads to seasonal and shorter-term changes in rainfall and evapotranspiration. The behaviour of the water table in different parts of the wetland is related to the proximity of drainage channels and to variations in the local sediment type. In addition, the proximity of the water table to the ground surface is considered as an important factor in limiting water table fluctuations.
4.6.2 The Stratigraphy of the Alluvial Aquifer

The geology of the Goss Moor catchment has been described in Section 2.4. Kaolinised pelite forms a base of impermeable laminated clays for the alluvial deposits in the valley bottom. A frost-heave mechanism has disturbed the upper few metres of this altered pelite, possibly changing its permeability. Mixing downwards into these disturbed pelitic sediments, the alluvial deposits present a highly heterogeneous field of sand, gravel, silt and clay deposits more than two metres thick in most places, with an average thickness of about four metres and up to a maximum of nine metres thick, originating from a braided stream environment (Camm, 1981).

As described in Section 3.4.2, groundwater piezometers were installed at twenty sites in the wetland. Most of these sites contained a single piezometer penetrating to 1 m, but other sites housed nested pairs at 1 m and 3.5 m, or single piezometers to 3.5 m. The positions of these sites on the wetland are shown in Figure 4.32. Also shown in the figure are the positions of the 71 boreholes drilled by Billiton Exploration (U.K.) Ltd. as described in Section 2.4.1. In the present study, the downhole stratigraphy could not be recorded for every piezometer. However, those logs which were recorded are shown in Figures 4.34 - 4.37 along with the logs for the closest of the boreholes drilled by Billiton Exploration. The key to the stratigraphic symbols is shown in Figure 4.33. In comparing the two records, it can be seen that although similar sedimentary lithofacies are frequently shared between the neighbours, these do not often occur at the same level in the stratigraphic sequence. The stratigraphic units are therefore unlikely to be continuous over the distances of around 300 m between these drilling sites.

4.6.3 Spatial and Temporal Piezometric Variations

The response of the water table to rainfall in an unconfined aquifer depends upon several factors including stratigraphy and drainage demands from rivers and vegetation. The water table fluctuations observed at Goss Moor provide an example of the modes of influence of these factors in a highly heterogeneous aquifer.
The observed groundwater heads are shown in Figures 4.38 - 4.57. Shallow piezometers are suffixed “s”, while the deeper piezometers bear the suffix “d”. Many of the records show a sudden fall of between 0.1 m and 0.45 m in the water level at 20/3/93. This was due to the extraction of a water sample for chemical analyses. Recovery of the water column from this drawdown generally took two or three weeks, but in piezometers P4d and P6d a full recovery did not occur, presumably through the destruction of a well skin. The sampling drawdown at these two sites was greater than elsewhere, indicating the possibility that some other piezometers might have similarly failed to recover if they had been subjected to greater drawdowns. The presence of a well skin complicates the interpretation of piezometric fluctuations. However, the above recoveries show that over durations of several weeks the piezometer water column can equilibrate with the water table, thus reflecting seasonal variations in water table with sufficient accuracy for hydrological interpretation.

Figure 4.58 shows the contours of the average water table elevation during summer 1994, virtually indistinguishable from the ground surface elevations shown underneath in Figure 4.59.

The seasonal averages of the water table depth, calculated using weekly data from the 1 metre piezometers, are contoured in Figures 4.60 - 4.63, showing further evidence of drainage by the river near the centre of the wetland and near the eastern piezometer site, P16s. Figures 4.64 - 4.67 show the changes in water table elevation between seasons. As expected, the water table rises from autumn to winter. This is followed by a slight decline through spring and a noticeable drop in summer. These seasonal changes are greatest near the river, demonstrating the control of the river level over nearby groundwater flow. As shown in Figure 4.50, the stage variations of the central river pool are similar in magnitude to those of the water table height at P13s. The pool surface elevation also occasionally exceeds the water table height. This indicates the control of the water table in this area by the stage of the pool rather than by precipitation. Here, flow from the river into the surrounding alluvial deposits may be as frequent as groundwater discharge to the river and
so not only do the central pools provide storage as a surface water feature, they may also induce river water storage in the surrounding alluvium.

The effect of river drainage is reduced with increasing distance from the river, as shown in the seasonal contour plots. In the areas of the wetland where the groundwater head is near or above ground level, seasonal control of water table height passes to precipitation and the limiting elevation of the ground surface.

The time series of water table elevations shown in Figures 4.38 - 4.57 illustrate again the effects of both river and ground surface constraint on the water table in the wetland. For example, the water table at piezometer P16s fluctuates in a manner similar to that at P13s, due to the control exerted on the groundwater by the nearby river.

The effects of ground surface constraint on the water table in the wetland can be seen throughout the piezometer time series records. Examples are at P2s, P5s, P9s, P10s and P11s, where the water table rose to the ground surface during the winter. Where the piezometers were nested (P5, P7, P8, P9 and P10, all near the southern edge of the wetland), the head in the deeper piezometer exceeded that at the shallower level during the winter and spring, indicating an upward discharge of groundwater, perhaps as a result of higher groundwater heads around the outside of the wetland. The head difference at the piezometers was generally reversed during late summer and early autumn 1994, indicating recharge to the aquifer after the reduction of groundwater heads in the wetland and its periphery. In general, the intersection of the piezomeric surface with ground level in autumn/winter indicated that the alluvium was saturated and could not accept any further recharge. Thus, excess rain water would either run off or be stored on the ground surface, as observed over most of the wetland.

The piezometer nest at site P4 shows the only exception to this behaviour. It is situated near a stream (see Figure 3.1) which is thought to have been draining the locale, reducing the groundwater head at the level of the deeper piezometer and pulling the water table down once surface saturation had been lost. Due to the downward groundwater head
gradient, surface water found at this site in the wetland could have come only from rain or overland flow, and was recharging the stream by percolation through the shallow saturated zone.

During autumn, winter and spring, the water table at most of the piezometers fluctuated visibly in response to rainfall on a weekly time scale, implying that some drainage of the water table was taking place between rainfall events. Apart from lateral flow through the substrata into channels or lower wetland areas, the most significant cause of this drainage in the wetland during the study year was evapotranspiration. ET is also a major factor in the sharp decline of the water table during the less rainy summer period.

The sediment found at piezometers P11s and P12s was very poorly permeable clay, probably with a high specific retention. This combination of properties in the locality of each site undoubtedly retards drainage, leaving the local water table hydraulically isolated from changes in river elevation. However, the addition or removal of water from the water column within such areas by rainfall or ET would produce large changes in water table height, due to the low specific yield in the local sediment. Hence the summer decline and recovery of the piezometric surface at these two sites was enhanced, in contrast with other streamside sites such as P13s, P16s and P4s where the fluctuations occurring during individual rain events are greater than the seasonal variations. These are two extremes in the local behaviour of the alluvium, highlighting the variability in its transmissive and retentive properties. The behaviour of the aquifer as a whole probably falls somewhere between the two extremes.

Figures 4.68 and 4.69 show the variations in stage in the surface water pools S1, S2 and S3. S1 and S2 form part of a system of cascading pools near the catchment outlet (see Section 3.4.2). This system is not part of the river but is very close by and drains through a culvert into the river a short distance downstream of the catchment outlet. S2 is the highest pool in the system and maintains a steadier level than S1, partly due to relative piezometric independence from the river water level. Since the regional rainfall is around 800 mm/annum greater than potential evaporation (Section 2.2), such pools must get rid of a
surplus of water by draining to either groundwater or the channel system. Furthermore, these pools receive a certain amount of overland flow, with catchment areas several times greater than their own surface areas, and must drain away most of this additional contribution. However, groundwater is unlikely to constitute the greatest drainage flow from these pools due to the predominance of clay in the surrounding alluvium. The greater oscillations of the winter water level in S1 may be primarily due to a surface outlet with less conveyance than that of the surface outlet of pool S2. Drainage and replenishment of both pools is evident in late summer and early autumn.

On the southern side of the wetland, pool S3 has no surface outflows and therefore must drain away excess contributions by seepage to groundwater alone. However, the pool has no surface inflows and so is required to drain only effective precipitation. The pool's levels are stable, presumably as a result of control by groundwater levels which are themselves being limited by the ground surface.

The similarity observed in the magnitudes and timing of fluctuation of river pool stage and nearby water table elevations suggests that river water may at times flow into neighbouring parts of the alluvium and be stored there. This would happen during times of high spate, thus providing a mechanism for the reduction of downstream peak flows. The observations may also suggest significant groundwater flow to the river from the wetland aquifer at times of lower river stage. The possibility of such strong interaction between the river and the aquifer depends upon reasonably high permeabilities in the alluvium, which possibly exist in the region of piezometer P13s but have been found absent at P11s and P12s. Numerical modelling of the wetland aquifer in Chapter 6 provides an assessment of the possible groundwater flows to the river given the observed water table elevations and permeabilities determined from piezometer slug tests in the wetland.
4.7 SUMMARY

This chapter has considered the characteristics of rainfall, evapotranspiration, stream flow and the wetland water table in the study catchment. Evapotranspiration has been shown to be stronger from the wetland than from the surrounding catchment due to a difference in vegetation type and waterlogging. Seasonal variations in ET from the wetland are commensurately greater. Seasonal variations in rainfall regime and evapotranspiration have been shown to be responsible for changes in stream flow as characterised by the relative contributions of slow flow and quick flow. The possible mechanisms by which this happens, discussed in Section 4.5.4, involve the influence of the soil moisture status on the generation of two different possible types of quick flow.

A comparison of the Goss Moor outflow duration curve with those of other headwater wetland catchments showed that Goss Moor maintains significantly steadier stream flows than many other headwater wetlands. Within the Goss Moor catchment, the variability and peakiness of stream flow was reduced between entry and exit of the wetland. This was shown by flow duration curves, by the spectral characteristics of the flows and by flow recession analysis at different locations within the site.

During the study year, variations in water table height near the river were found to be controlled by river stage. This raised the possibility of bankside storage during high river spate and subsequent stabilisation of downstream river flows, as well as the possibility of significant slow flow from the alluvial aquifer. Further from the river, the water table remained close to the ground surface during the winter, thus reducing its fluctuations. The water table fell below ground surface for about four months during the summer, as a result of drainage by evapotranspiration and possibly by the river.

River slow flow has been found to be uncertain in origin since three possible sources, namely unconfined groundwater flow to the river, wetland surface water flow and water storage in river pools, have been identified. It remains to be shown in Chapter 7, using the results of numerical groundwater modelling performed in Chapter 6, whether the first of the
above three processes is a viable source of slow flow given the hydraulic properties of the alluvium.
5.1 INTRODUCTION

5.1.1 Aims of the Water Budget Evaluation

As discussed in Chapter 1, the determination of some form of water budget for the study area is a major goal in many programmes of scientific catchment research. Water budgets involve accounting for all water flowing into, out of or residing in the study area over a specified period of time. Independent measurements and calculations are made of the flows via different pathways such as subsurface diffuse flow, surface diffuse flow, subsurface channel flow, surface channel flow, atmospheric precipitation and evapotranspiration, together with the volumes of water stored in aquifers and channels, on the ground surface and in vegetation (Woo and Rowsell, 1993; Zimmermann et al., 1999).

The water budget may then be used to test hypotheses about the hydrological processes occurring in the study area, or simply to establish the current water balance before predicting the effects of perturbations. Thus, the aim of purely hydrological research may be to explain the balance between the different flows and stored volumes in terms of processes whose occurrence in the area is otherwise unestablished. For example, Haldorsen et al. (1996) investigated the potential interaction of subpermafrost groundwater in relation to glacier hydrological regimes in Svalbard by the measurement of mine drainage. Also by way of example, Moreno et al. (1996) calculated the monthly water balance of 4 oak forests, using measurements of quantities such as rainfall, interception, runoff and soil moisture. They used this water balance to show that the forests’ water consumption was
limited, or alternatively was maintained by deeper groundwater, during spells of low rainfall.

Alternatively, the significance of processes known to be occurring in the catchment is evaluated using water budget calculations. For example, Meyer and Gee (1999) used flux-based measurements of drainage and field capacity to improve future drainage estimates for waste repository caps.

The present study intends to determine the role of the alluvial aquifer in the wetland's hydrology. Towards this goal, the present chapter formulates and examines the hypothesis that groundwater flow through the alluvium constitutes a significant part of the total flow through the catchment. The fluxes with which this shallow groundwater flow is to be compared are the rainfall, the evapotranspiration and the total stream flow, whose characteristics aside from sheer quantity were discussed in the preceding chapter. By inspecting the cumulative volumes of water input by rainfall and output by evapotranspiration (ET) and stream flow over the year, implications are found for the groundwater inputs and outputs of the catchment. Additionally, the seasonal variations in the various water fluxes are brought to light in this chapter.

5.1.2 Water Budget Methodology

The mode of investigation described in this thesis involves the examination of the water budget at two different spatial scales, along with the modelling of the wetland's groundwater regime. The water budget evaluated in this chapter provides a catchment-scale context for the groundwater modelling in Chapter 6, which in turn provides input to the wetland-scale water budget evaluated in Chapter 7. The basis for this sequence of analysis is discussed below.

Two simplifications were made for the purposes of the catchment water budget calculation:

(1) the study area receives no surface water through its perimeter, and
(2) the study area receives no groundwater through its perimeter.
Condition (1) is validated by the surface topography of the catchment (see Figure 2.8).

Condition (2) depends on the assumption that the groundwater catchment does not extend beyond the surface catchment. The consideration of uncertainties in catchment area is one way of compensating possible inflows and outflows to a catchment, the alternative method being to maintain a rigid catchment boundary which is the same for surface and groundwater and to estimate inflows or outflows through its perimeter by measurement or other means. In the present chapter, the problem of peripheral groundwater inflow is addressed by considering the possibility of modifying the boundaries.

Should significant discrepancies exist between the surface and the groundwater catchment areas, the assigned precipitation (and evapotranspiration) volume would stand in need of correction. This would be achieved by partitioning the meteoric fluxes of areas which the Goss Moor groundwater catchment shared with other surface catchments, and vice versa for similar areas of Goss Moor’s surface catchment, using recharge or runoff models such as that of Thornthwaite and Mather (1955), Penman and Grindley (Lerner et al., 1990) and Kachroo (1992). Section 3.3.3 of the current thesis considers the use of the Penman-Grindley model in estimating actual evapotranspiration from the areas of pasture within Goss Moor’s surface catchment.

As mentioned in Section 2.6, china clay mines, settling ponds and waste mounds lie on the southern boundary of the Goss Moor catchment. A lack of topographic information for these features means that the exact whereabouts of the boundary is more uncertain on the southern than on the northern side of the catchment. The northern boundary is well defined and considered to be identical for surface water and groundwater. In the south, the uplands of the St. Austell Granite extend a few kilometres southwards from the catchment’s periphery, maintaining an average elevation which is similar to that of the surface water boundary. This implies that, if the groundwater boundary were to be different from the surface water boundary in this region, it would be further out.
However, if the aquifer thickness on the catchment boundary was small in comparison with the changes in elevation on either side of the catchment perimeter, then the impermeable base of the aquifer on the catchment boundary would closely follow the surface relief and flow across the surface catchment divides from neighbouring upland basins would be prevented. For this situation to apply at the southern boundary of the Goss Moor catchment, the weathered layer of granite or slate constituting the aquifer would therefore have to be less than a few tens of metres in depth. The strength of this assumption cannot be assessed with certainty due to a lack of geological section data on the catchment boundary. However, the prevalence of large granite boulders in the uplands of the St. Austell Granite (introduced in Section 2.4.2) suggests that a deep fractured zone exists on the granite surface. This zone may extend to depths of several tens of metres below ground level. The assumption that the peripheral aquifer thickness is small in comparison with the changes in ground surface elevation therefore may not be strictly true, but an effective conductivity which declined with depth due to a reduction in jointing would be envisaged for the aquifer and therefore would improve the applicability of the assumption.

As described in Section 2.6.2, the catchment produces no diffuse surface water output because of obstruction by a disused railway embankment. This simplification applies also to the wetland-scale water budget calculated in Chapter 7. Simplifications of this nature reduce the number of unknown quantities in the water budget equation, allowing more specific although not necessary more accurate estimation of other quantities such as groundwater outflow.

Conditions (2) relates to the input of groundwater to the catchment through its perimeter, asserting that this is likely to be insignificant in the water budget of the whole catchment. In contrast, the possibility of groundwater output is acknowledged and is addressed in the water budget analysis below. Although the main groundwater output of interest is that through the alluvium in the lower elevations of the catchment, the presence of the Castle-Dinas Wolfram mine (see Section 2.4.2) with its outwardly draining drainage adit some 90 m below the top of Castle Downs involves a strong possibility of groundwater losses from a source area around the crown of this hill. Because the flows from this adit remain
unknown, uncertainty must increase in the amount of groundwater output through the wetland alluvium. Furthermore, the results of the catchment water budget must now be interpreted in the near-certain knowledge that an unknown but non-zero groundwater output is occurring.

The catchment defined in this study contains the wetland in such a position that the downslope boundary of the wetland forms a large subsection of the catchment's downslope boundary, as seen in Figure 2.8. The groundwater losses of the wetland are therefore a major part of the catchment's groundwater output. Hence the value of the catchment groundwater losses determined in the initial water budget evaluation may be, subject to the above observations on mine drainage, an initial estimate of groundwater losses in the wetland water budget. In addition, the two land areas share the same river outflow and also the same river slow flow, so that the channels traversing the wetland are the last stage in channel flow from the entire catchment. This is relevant to the wetland water budget in Chapter 7.

The discussion so far has shown how certain results from the catchment water budget will be relevant to the groundwater modelling and the wetland water budget. However, the inapplicability of conditions (1) and (2) to the wetland area leaves undetermined the values of groundwater and surface water inflow. This and other problems in the more detailed wetland water budget are resolved with data output from the wetland groundwater model.

5.2 THE CATCHMENT WATER BUDGET EQUATION

The quantities evaluated for the catchment water budget are:

- rainfall
- evapotranspiration
- stream outflow
- net groundwater output + overall storage gain.
All of these quantities were evaluated in daily intervals, for a total duration of one year.
The period of evaluation was 1st September 1993 - 31st August 1994.

The water budget of the catchment obeys the equation

\[ P - E - R = G_{out} - G_{in} + \Delta Stor, \]  

where

\[ P \] is the rainfall volume (m³ per m² of catchment),
\[ E \] is the evapotranspired volume (m³ per m² of catchment),
\[ R \] is the river outflow volume (m³ per m² of catchment),
\[ G_{out} \] is the outgoing groundwater volume (m³ per m² of catchment),
\[ G_{in} \] is the incoming groundwater volume (m³ per m² of catchment), and
\[ \Delta Stor \] is the overall storage volume gained (m³ per m² of catchment).

Although diffuse surface flow exists near the catchment outlet, it does not cross the catchment boundary due to obstruction by a disused railway embankment (discussed in Section 2.6.2). Hence it does not take part in the catchment water budget.

The groundwater inflows, outflows and storage gain in the study site were not measured or independently determined, so requiring derivation as the unknown right hand side of Equation 5.1. Since the combined term contains three unknown quantities, various interpretations may be placed on the overall value. These are discussed in Section 5.3.1.
5.3 INTERPRETATION OF WATER BUDGET

5.3.1 Groundwater Flows and Stored Water

As stated in the introduction to this chapter, a primary aim of the catchment water budget is to evaluate and compare the major water fluxes averaged over the catchment as a whole. One of these fluxes in particular, the groundwater flow out of the catchment, is initially assumed representative of the direct groundwater output from the wetland, although as explained in Section 5.1.2 it may be somewhat greater than the wetland groundwater output due to the presence of mine drainage from Castle Downs. Additionally, the river slow flow is hypothetically taken to come from wetland groundwater flow which, together with the aforementioned direct groundwater output, forms the total groundwater output from the wetland, as shown in Equation 5.2:

\[ \text{Loss}_{gw} = B + G_{out} = R - Q + G_{out} \] (5.2)

where
- \( \text{Loss}_{gw} \) is the total groundwater loss from the catchment (m\(^3\) per m\(^2\) of catchment),
- \( B \) is the river slow flow from the catchment (m\(^3\) per m\(^2\) of catchment), and
- \( Q \) is the river quick flow from the catchment (m\(^3\) per m\(^2\) of catchment).

This is compared with the other major water fluxes, such as rainfall and evapotranspiration, to assess the importance of groundwater flow in the wetland. The assumptions involved are examined here, and further investigation of their validity through numerical modelling is anticipated for Chapters 6 and 7.

The groundwater flow out of the catchment, \( G_{out} \), forms part of the term \( G_{out} - G_{in} + \Delta \text{Stor} \) on the right hand side of Equation 5.1. Since the combined term contains three unknown quantities, \( G_{out} \) can be obtained only by using certain assumptions to account for the other two unknowns.

Figure 5.1 shows the cumulative volumes of flux developing over the budget period.
Through September, October and November 1993, the term $G_{out} - G_{in} + \Delta Stor$ (termed “net groundwater output + storage gain” in Figure 5.1 and henceforth called the water surplus) remained roughly constant as the onset of more frequent rain and reduction in ET began to reverse its summer decline. December, January and February saw the greatest increase in the water surplus because of low ET and high rainfall. From mid-April onwards, the surplus was depleted by increasing ET with less replenishment from rain. During the budget year, the water surplus values varied from -0.057 m to +0.189 m relative to the starting value, covering a range of 0.246 m.

Simplified interpretations of the water surplus may be applied to these data in order to show the significance of the variations. This is also a test of the validity of each simplification. Firstly, assuming that $G_{out} = G_{in} = 0$ at all times, then $\Delta Stor$ may become as large as 0.246 m. In a soil with a realistic storativity of 0.1 this would raise the water table by 2.46 m, which far exceeds the range of water table variations observed in the wetland. Due to relatively steep slopes and a potentially deep transmission zone, the surrounding hill sides would be a more likely location for such storage fluctuations. A little surface water storage on the wetland would also be involved in this interpretation.

Secondly, it is noted that neither $G_{out}$ nor $G_{in}$ alone can explain the variations in water surplus since neither can go negative in order to reverse the direction of change. $\Delta Stor$ is the only quantity which can possibly account on its own for the changes in water surplus.

Before considering combinations of more than one such quantity, a hydrological constraint is noted, namely that $\frac{d}{dt}(G_{in})$, $\frac{d}{dt}(G_{out})$ and $\Delta Stor$ all change in the same direction. For example when $\Delta Stor$ is increasing, so are $\frac{d}{dt}(G_{in})$ and $\frac{d}{dt}(G_{out})$, and when $\frac{d}{dt}(G_{out})$ is decreasing, so are $\frac{d}{dt}(G_{in})$ and $\Delta Stor$. This constraint is a consequence of the fact that every hydrological output is an increasing function of the source storage and that any water stores supplying the catchment would be subject to the same seasonality as the study catchment. The constraint is implicit in the following.

Combinations formed by two of the three quantities are as follows:
\[
\text{water surplus} = G_{\text{out}} - G_{\text{in}}
\]
\[
\text{water surplus} = \Delta \text{Stor} - G_{\text{in}}
\]
\[
\text{water surplus} = \Delta \text{Stor} + G_{\text{out}}.
\]

The first case is unphysical since increases in \(\Delta \text{Stor}\) (representing rising groundwater heads) are required in order to produce increases in \(G_{\text{out}}\).

In addressing the second combination, use is made of condition (2) from Section 5.1.2, bearing in mind the qualifications involved in its use and the consequent uncertainty in the whereabouts of the groundwater catchment boundary in relation to the surface catchment boundary. Notwithstanding such qualifications, the assumption that \(G_{\text{in}} = 0\) is made in the present study, given no accurate information on bedrock elevations or on rainfall-recharge characteristics in the upland waste areas of the St. Austell granite. Vertical groundwater flow into or out of the Goss Moor catchment is negligible because clay solifluction products and partially kaolinised granite or slate at deeper levels form an impermeable base to the aquifer.

The third combination considers the water surplus divided into two components: accumulated groundwater outflows and accumulated storage. This is the most likely combination applicable to the studied water budget year 1/9/93 - 31/8/94, given the reasoning outlined above that \(G_{\text{in}} = 0\). Examination of Figure 5.1 reveals that the water surplus is near zero at the end of the 12 month water budget period after rising through autumn and winter and falling through spring and summer. Since the accumulated outflow must be positive and can never decrease, it may have been either be near-zero itself, in combination with a storage accumulation which alone accounted for the rising and falling of the water surplus, or it may have been significantly greater than zero, in combination with a storage accumulation which became significantly negative as the year progressed. Situations midway between these two extremes are probable. For example, the drainage of the Castle-an-Dinas Wolfram Mine, introduced in Sections 2.4.2 and 5.1.2, is considered to be part of \(G_{\text{out}}\) in the above calculation. Most importantly, groundwater drainage out through the wetland alluvium on the western edge of the catchment is part of \(G_{\text{out}}\) and
remains undetermined by the catchment water budget calculations due to uncertainty in storage gain or deficit over the water budget year.

In Chapter 6, numerical modelling of the groundwater flow through the wetland alluvium will be conducted, incorporating measured piezometric boundary conditions, alluvium permeability and other physical characteristics of the wetland hydrological system. In addition to estimating the groundwater flow to the stream network for comparison with the spectrally filtered slow component of the stream flow, as stated in Section 4.5.3, this modelling will also ascertain the amount of groundwater lost through the wetland aquifer's western boundary during the water budget year. The above catchment water surplus fluctuations will therefore be placed in context with the groundwater hydrology of the wetland.

5.3.2 Intercomparison of Catchment Water Losses

The various seasonal input and output volumes are shown in Figure 5.2 and Table 5.1. Rainfall and stream flow show the expected seasonal behaviour, increasing in winter and decreasing in summer. As discussed in Section 4.5.4, slow flow dominates the stream flow during the summer since the generation of quick flow is subdued by depleted soil moisture. Evapotranspiration shows the expected increase during summer. Figure 5.3 shows each seasonal input or output volume as a percentage of its yearly total, so that the seasonal variability of each flow can be compared with the others. Quick flow is seen to be more variable than all other fluxes, including rain. Slow flow and ET show the least variability. However, for every flux there is one three-month period in which more than forty percent of the year's total volume is passed. While stream flow is the dominant output from the catchment for nine months of the year, as shown in Figure 5.2, ET takes precedence during summer since its state of greatest vigour is then coinciding with severely attenuated stream outflow. The gain in water surplus is also shown in the figure, showing that a gain in the amount of stored water occurs during winter while during the summer months the reduced rainfall and increased ET combine to reduce storage.
5.4 FURTHER LIMITATIONS OF THE CATCHMENT WATER BUDGET

The catchment water budget is limited in three main respects: firstly, in the accuracy of the measurements upon which it is based; secondly, in the accuracy of the stream flow separation procedure described in Chapter 4; and thirdly, in the validity of the assumption that the catchment is a closed groundwater basin. The third shortcoming has been discussed in Section 5.3.1. The remaining two are mentioned here for the sake of caution.

5.4.1 Errors in Measured Water Budget Components

The possible sources of error in the measured water budget components are as follows:

1) Differences in albedo and therefore in net radiation between the various land cover types were neglected. Other evaporation-related differences not taken into account were in humidity, temperature and wind speed. In Section 3.3.3 some consideration is given to these variations.

2) The heat flux into the ground, or absorbed into biochemical storage, has been assumed negligible. Consequently, the sensible and latent heat fluxes may have been overestimated, leading to further overestimation of evapotranspiration.

3) A small area of mica waste tips was parameterised as pasture, thus slightly increasing the estimate of overall catchment ET.

4) As described in Section 3.2.3 and Appendix A, large errors were expected in the stage-discharge relation derived at the catchment outlet. Expected deviations in flow due to digitising errors ranged from 0.016 cumecs at low stages to around 0.039 cumecs at high stages. Substantial bias in the rating curve at the catchment outlet was expected, probably exceeding the error associated with digitisation. The sign of this bias in the river flow error was unknown, and may have varied with flow levels, but should be consistent over time.

5) Due to the presence of unmapped waste tips on the southeastern periphery of the catchment, the watershed boundary is undefined in this area. This results in a small
error of unknown sign in the catchment area, potentially producing a bias between the stream flow and the areally calculated quantities, ET and rainfall.

5.4.2 Errors in the Stream Flow Separation

In Section 4.5.4, it was mentioned that the stream flow separation procedure neglected the possible seasonal variability in the slow flow parameters due to contraction of the draining reservoir and the consequent migration of its centre of mass through areas of varying hydraulic characteristics. However, errors in the annual slow flow volume were minimised by adjustment of the digital filter for an overall optimum fit with the observed recessions, while resultant seasonal errors of the slow flow may have been mitigated by the random, small-scale nature of the heterogeneity in the alluvium.

Seasonal variations in evapotranspiration were also found capable of imparting seasonal errors to the estimated slow flow. However, discrepancies between the estimated and the observed recessions showed no discernible seasonal variations and so errors from both sources are assumed negligible for the purposes of this study.

5.4.3 Dependence of the Water Balance on the Area of Analysis

Variability within the area or length of evaluation of a spatial average may give rise to conditions which contrast with those apparent at the scale of averaging. Hence conclusions drawn from the averaged quantities may be invalid for certain zones within the averaging region. Conversely, overall apparent conditions may be caused by the domination of one type of zone over all other zones. This is relevant to the study of Goss Moor, since the wetland, comprising almost one quarter of the total catchment area, clearly contrasts with the surrounding catchment in terms of vegetation, soils, stratigraphy, topology and surface water coverage.

The estimated slow flows suffer from the problem introduced above, that the heterogeneity within the catchment area may affect the slow flow separation, making it unrepresentative.
of certain parts of the catchment. This is an expression of the fact that the location of slow flow source area in a non-uniform catchment cannot be determined from the river hydrograph alone. Similarly, the proportion of slow flow in Goss Moor which is attributable to the wetland area cannot be determined in the catchment analysis. It is most probable that the partitioning of rainfall into groundwater recharge and surface or near-surface flow, related to the production of slow flow and quick flow, differs significantly between the wetland and other parts of the catchment since, as explained in Section 2.4 and Section 4.5.4, the gradients, soils and geology of the central wetland are divergent from that found on the hill slopes of the outer catchment. Likewise, Section 4.3 has shown the contrast in strength of ET between the wetland and the catchment. Finally, the wetland may differ in the significance of storage fluctuations as part of its water balance. Consequently, it is necessary to perform water budget calculations for the wetland area itself. The hydrological character of the wetland will then be better understood. However, the wetland is subject to a greater variety of water influxes than the catchment, so complicating the water budget calculation. This problem and that of identifying the nature of the slow flow are resolved through the modelling of the wetland groundwater regime in Chapter 6, together with the estimation of groundwater flows and storage fluctuations from the model output in Chapter 7.

5.5 SUMMARY

This chapter evaluated the seasonal and cumulative volumes of water passed into or out of the catchment by rainfall, stream flow and evapotranspiration, contributing to the assessment of the significance of each flux within the catchment water balance. Additionally, the examination of imbalances between these surface inputs and outputs allowed an assessment of the possibility of contributions by groundwater.

The “water surplus”, defined in Section 5.3.1 as the sum of the net groundwater output and storage gain of the catchment, was found to fluctuate but returned to its original level over
the study year. This could be explained by variations in both groundwater losses and catchment water storage. The groundwater losses were noted to include drainage from the Castle-an-Dinas Wolfram Mine and possible outflows through the western edge of the wetland alluvium. Potential errors in estimated river flow (Section 3.2.3) and other errors in rainfall and ET were acknowledged.

While stream flow was the dominant output from the catchment between September and May, its severe curtailment during the summer allowed evapotranspiration to take precedence between June and August. Due to the lack of replenishment by rainfall, these three months showed the greatest storage depletion which was therefore attributable mainly to ET.

Finally, it was noted that the catchment water balance, although benefiting from the assumption of zero external contributions to the catchment, could not elucidate conditions in the wetland itself. A water budget evaluation for the wetland itself was therefore proposed, using results from numerical modelling of the wetland aquifer to further quantify the interaction between the wetland and the river. This would be carried out in Chapters 6 and 7.
CHAPTER 6
NUMERICAL MODELLING OF WETLAND GROUNDWATER FLOW

6.1 AIMS OF THE NUMERICAL GROUNDWATER MODELLING

The goal of the work described in this chapter was to develop a numerical model of the groundwater flow in the Goss Moor wetland aquifer. This model was calibrated with respect to the observed water table behaviour by adjusting the specified parameters and stimuli within reasonable limits. Once calibrated, the model was used to further examine the behaviour of the wetland water table during the study year. However, this use of the model was secondary to the aim of estimating the flows between the wetland groundwater body and its surroundings. The estimation of such flows was to be carried out in Chapter 7 and so the main purpose of the present chapter was simply to establish the calibrated model ready to be used in this analysis. In Chapter 7, such flow estimation will complement earlier analysis performed in Chapters 4 and 5, resulting in the integration of all findings into a coherent assessment of the hydrology of the wetland.

6.2 SELECTION OF MODELLING SYSTEM

The present study was concerned with the simulation of saturated groundwater flow over an area of several square kilometres and a time scale of several months. Given that the main aim of the modelling was effectively to test the significance of horizontal saturated groundwater flow in the water budget of the wetland, the chosen modelling system would be specialised for the simulation of groundwater flow. However, it was desirable to use a modelling system with substantial flexibility in the treatment of external influences on the
groundwater domain, in order to account for any unusual conditions arising out of the
wetland-groundwater interaction.

The balance of forces in a fluid-saturated porous medium gives rise to a simple relation
between the velocity and the head gradient of the fluid, known as Darcy’s Law (see, for
example, Bear and Verwey, 1987). The following equation of motion expresses Darcy’s
Law for a water-saturated medium:

\[
V = K \cdot \frac{d\phi}{dx}
\]  

(6.1)

where

- \( V \) is the discharge per unit cross-sectional area of medium (m/day),
- \( K \) is the hydraulic conductivity or aqueous permeability of the medium (m/day),
- \( \phi \) is the hydraulic head in the water (m), and
- \( x \) is the spatial coordinate in the direction of \( V \) (m).

Invoking the principle of conservation of mass, the fluid velocity may be eliminated from
Darcy’s Law to give an equation of groundwater motion in terms of the groundwater head
and its derivatives. For a compressible fluid such as water in an incompressible porous
matrix, Darcy’s Law thus produces the following relation (see, for example, Freeze and
Cherry, 1979):

\[
\frac{\partial}{\partial x}\left(K_x \cdot \frac{\partial \phi}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_y \cdot \frac{\partial \phi}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_z \cdot \frac{\partial \phi}{\partial z}\right) = S \cdot \frac{\partial \phi}{\partial t}
\]  

(6.2)

where

- \( x, y \) and \( z \) are orthogonal, Cartesian spatial coordinates, parallel to the porous
  medium’s principal conducting directions (m),
- \( t \) is the elapsed time (days),
- \( K_x, K_y \) and \( K_z \) are the components of conductivity in each orthogonal direction
  (m/day), and
- \( S \) is the specific storage of the porous medium (m\(^{-1}\)).
This equation, or its equivalents in other coordinate systems, may be used as the basis of computer simulation for many saturated groundwater flow situations not involving processes such as matrix compression, preferential flow, multiphase flow, heat transport and others. Boundary conditions, initial conditions and source/sink terms, as described in Section 6.4, are also necessary for the solution of the flow problem, allowing the position of a water table to be incorporated into the solution and to consider the effects of rivers, rainfall, evapotranspiration and other hydrological features.

Since adequate solution of the saturated groundwater flow problem is provided by most numerical solution techniques, including finite difference and finite element methods, these features were not important in the selection of an appropriate modelling program. The features required in the groundwater modelling program were as follows:

- 2-D (horizontal) or 3-D saturated flow
- Steady state and transient simulations
- Unconfined flow (non-linear equation)
- Effects of water table drainage at the ground surface
- Effects of drainage by a river
- Effects of areally distributed evapotranspiration
- Heterogeneous aquifer properties
- Unspecified flow allowed through boundary (prescribed head boundary)
- Flow (including zero flow) prescribable at boundary
- Heterogeneous ground surface elevation.

Burden (1998) provided an assessment of several currently available groundwater flow modelling programs, such as 3DFEMFAT (Yeh et al., 1994), AQUA3D (Vatnaskil Consulting Engineers, 1998), AQUIFEM-N (Anderson and Woessner, 1992) and MODFLOW (McDonald and Harbaugh, 1988). These and other programs were all found to provide most of the above facilities.
For the present study, several factors suggested that MODFLOW would be particularly suitable. Firstly, it allowed some accounting for the effects of water table drainage at the ground surface. This facility would be most important in the modelling of a wetland such as Goss Moor. Secondly, MODFLOW source code and comprehensive documentation were available which would provide greater control over the model runs for calibration purposes and allow full modification of input and output to/from other programs. In this respect, MODFLOW provided greater versatility. It may also be important to have access to the source code on occasions when the model user is unsure of the exact interpretation of the simulation results. Furthermore, MODFLOW has been in use over several years and so has benefited from attention to any conceptual or coding problems which may have arisen during its initial development (Ashley, 1994).

An important requirement of the current study is that the numerical model should have a proven record in simulating the exchange of water between an aquifer and a stream. MODFLOW has been successfully used in the past to simulate stream-aquifer exchanges in study areas ranging from a few hectares to several hundred square kilometres in extent. For example, Squillace (1996) employed the software to model a vertical section of aquifer extending 440 m horizontally and 15 m vertically from a river channel in Iowa, USA. The cell dimensions were 10 m (horizontal) \( \times \) 1 m (vertical). The model was intensively calibrated against transient data from a river stage recorder and a transect of piezometer nests and showed, via a particle tracking postprocessor, the daily movement of bank storage water during and after river peak flow events. On a wide scale, Modica et al. (1998) used MODFLOW and its particle tracking postprocessor in steady state to determine residence times of groundwater in the 264 km\(^2\) Cohasey River basin, USA. They calibrated the model with hydraulic heads from 43 observation well locations and with river flow data. Chlorofluorocarbon distributions obtained by field sampling provided some validation of the model. The study illustrated the potential for this type of stream-aquifer model to identify the source areas of different stretches of a river. Others who have used MODFLOW successfully to model stream-aquifer interaction include Sophocleous and Perkins (1993) and Christensen et al. (1998).
Various approaches have been adopted in the past towards the modelling of groundwater systems which exchange water with wetlands. These approaches are objective-dependent. For instance, Siegel (1988a) used a steady state vertical slice model of groundwater flow to produce theoretical support for a conceptualisation of the recharge-discharge function of wetlands on a glacial recession moraine in Alaska. A river was located downslope of the wetlands, and the model transect extended from this river past the wetlands and further on up the valley slope into an upslope recharge zone. The water table elevations were held invariable at elevations suggested by piezometer nest readings. The simulated heads indicated that the wetlands were mainly involved in local recharge-discharge systems and were bypassed by deeper groundwater flow on its way to the river. This study was conducted using an unspecified numerical modelling program.

MODFLOW has been used for similar theoretical investigations regarding wetland groundwater flow systems. An example is the study by Gilvear et al. (1993) who used it to assess the instability in the flow system beneath a small groundwater-fed fen in East Anglia, UK (see also, Section 1.3.2). Significantly, they improved the representativeness of the model for the wetland area through the use of MODFLOW’s Drain Package to allow for the emergence of groundwater at the land surface.

Turning to groundwater model applications for prediction, Hensel and Miller (1991) used MODFLOW to simulate the effects of newly constructed wetland ponds on the groundwater flows in the vicinity of a lowland river. Some of the ponds, replenished by water pumped from an upstream reach of the river, drained water through their beds and doubled the local groundwater discharge to the stream. These ponds would have dried up if their artificial replenishment were stopped. Others, however, were underlain by low permeability sediments and so had negligible drainage. A significant factor in the applicability of the model was the use of MODFLOW’s River Package to represent the wetland ponds.

Other studies employing MODFLOW in wetland groundwater planning include that of Mohanty et al. (1994), who used it to evaluate the effects of agricultural drainage well
closure in a 4.7 km$^2$ catchment in Iowa, USA and that of Stewart et al., assessing schemes for the augmentation of seepage to a developing wetland in a nature reserve in northern Texas, USA. These case studies together with the specifications listed above, and in particular the research advantages of source code availability, suggested that MODFLOW was a suitable program for the task in hand.

6.3 METHODOLOGY OF THE GROUNDWATER MODELLING

The development of a model of the wetland groundwater flow regime was divided into three stages:

1) Specification of flow domain and stimuli
2) Steady state calibration
3) Transient calibration.

The first stage specified the shape of the aquifer, its unconfined nature, the course and depth of the river and the elevation of the ground surface for drainage of a high water table. The values of rainfall and evapotranspiration (ET) data were also collected and averaged over the appropriate time intervals for both equilibrium and transient simulations. In the steady state calibration (stage 2), the stability of the specified groundwater system was established, in preparation for the transient calibration, along with the range of uniform aquifer properties which would permit agreement with the observed equilibrium heads. The effects of alterations in aquifer properties and in system stimuli upon the model's representation of observed variations in water table elevations were then examined in the transient calibration (stage 3). This analysis would provide a more detailed assessment of the nature of the groundwater system in Goss Moor, based upon the success or failure of various measures taken to fit the observed data. Eventually, the calibrated transient model would provide estimates of various groundwater flows for incorporation into a wetland water budget for the study year. Examination of this budget would quantify the interactions between the aquifer, the wetland and the outer catchment. This work is described in Chapter 7.
6.4 CONSTRUCTION OF THE MODEL DOMAIN AND GRID

6.4.1 The Shape and Boundary Conditions of the Model Domain

This section focuses on the general considerations involved in the allocation of the area to be modelled and the specification of conditions at points within that area which control the flow field. Where appropriate, related parameters and statistics such as river bed depth and the distribution of aquifer thickness will be discussed in Section 6.4.3, following on from the clarification of the domain discretisation. Further related parameters come into consideration during the calibration of the transient model, described in Section 6.7.

Figure 6.1 displays the model domain and associated surface physiographic features within the study site. Attention is brought to the disused railway embankment on the western side of the moor, which served as the western boundary of both the surface catchment and the area of geological sampling by Billiton Exploration (see Section 2.4.1). Although both the wetland and the alluvium extend south-westward beyond this feature, it provided a convenient western limit for the model domain on two counts: firstly, that it would thus keep the area of evaluation of the wetland water budget entirely within the study catchment, and secondly, that it would also keep the model domain to an area of known geology. In other parts of the study site, the model domain was constructed to represent the wetland aquifer as closely as possible. Since the wetland and its aquifer were almost coincident in areal extent, the model therefore covered most of the wetland area within the catchment, although its areal coverage was limited to that of the Billiton geological data, again limiting the model domain to an area of known geology.

The alluvium beneath Goss Moor was not sufficiently thick to justify the construction of a three-dimensional model, with the result that the aquifer was considered in only two dimensions, using Dupuit's approximation of horizontal groundwater flow to facilitate the calculation of the water table elevation. Nevertheless, the elevation of the base of the
alluvium, and hence the transmissivity of the model domain, was allowed to vary with location according to the borehole bedrock soundings.

Mathematically, the calculation of spatial field values can be performed only over bounded regions of space in which the behaviour of the field is constrained at the boundary. Such constraints may also be specified within the region to further control the behaviour of the field. Bear and Verruijt (1987) discuss such constraints, collectively termed “boundary conditions”, for groundwater flow, including the condition of prescribed boundary heads (Dirichlet condition), prescribed normal flux (Neumann condition) and an equivalence between the normal flux and the superiority of the boundary head over an external, prescribed head (Cauchy condition). A combination of these three boundary conditions was used in defining the boundary of the present model.

The boundary conditions of the model domain can be seen in Figure 6.2. As discussed in Section 3.4.2, the siting of many of the piezometers installed on the wetland was influenced by the need to prescribe Dirichlet conditions over large sections of the wetland boundary. The groundwater head measured at each piezometer on the periphery of the wetland was extended out from the piezometer in a line parallel to the land surface contours. Along the northern and southern edges of the domain, each section of constant head boundary was extended until coinciding with the line of slope running through the end of the next constant head section. The two sections were then joined along the line of slope with a zero flow Neumann boundary section. The Neumann boundary condition was used also along two other sections of the model perimeter, both on the eastern side of the domain. These sections corresponded to the contact between the alluvium and the remnant outcroppings of unkaolinised peelite at Toad Hole and Tregoss Moor, both assumed to be impermeable. All sections of Neumann-condition boundary used in the present study could thus be termed zero flux or “no-flow” boundaries. This included the base of the aquifer since it was assumed to be impermeable and therefore allowed no normal gains or losses.

For steady state simulations, the head prescribed at each Dirichlet boundary section remained invariable at the annual average water table elevation measured at the relevant
piezometer. Conversely, during transient simulations, the Time-Variant Specified-Head Package was used to extend the original functionality of MODFLOW, allowing weekly variation of the assigned heads in accordance with observed water table variations.

Most of the model perimeter was defined in the fashion outlined above, using either the constant head or the no-flow condition, the only exception being a short section of the south-western boundary, coinciding with the river Fal on its exit from the catchment. As with other stretches of the river, this section was represented with the Cauchy condition between the normal groundwater flux and the head of the river water surface.

The piezometric record discussed in Chapter 4 (Section 4.6.3 and Figures 4.38 - 4.57) showed the tendency of the wetland water table to rise to the ground surface during the wet season. On Goss Moor, this behaviour was integral to the manifestation of the wetland itself, being accompanied by surface water ponding and probably by surface drainage. Modelling this situation poses the problem that the elevation of the water table, which was hitherto free to increase with the gains of water imposed on it, must now be limited to that of the ground surface and some account be made for the supplied water which would otherwise raise it above ground level. The ground surface does not always feature as a boundary condition in the groundwater flow equation to be solved, and therefore must be dealt with in a contingent fashion.

To represent the ground surface in the present study, a MODFLOW facility known as the Drain Package (McDonald and Harbaugh, 1988) was used. With this facility, the ground surface is interpreted as a groundwater sink whose rate of uptake of water is proportional to the head difference exerted on the surface by the groundwater column, as expected by Darcy's Law. When the water table falls back below the sink level, the uptake ceases since no head difference is exerted. Since this sink covers the whole area of the domain, it offers a large cross-sectional area of flow and is thus capable of withdrawing enough water from the domain to suppress any further rise in the water table. During such activity, the land surface drain function may be likened to a Cauchy boundary condition.
6.4.2 Discretisation of the Domain

Finite difference modelling schemes subdivide the region of interest into rectangular cells for piecewise approximation of the groundwater potential field. There are two widely used techniques, known as the block-centred and the vertex-centred methods, for relating the positions of the nodes (the points at which the field values are determined) to the cells (the rectangles in which the properties of the medium are each given one local value). MODFLOW uses the block-centred method, in which the nodes fall at the centroids of the cells. A consideration of the nodal values necessary for the specification of boundary conditions shows that this scheme then effectively places the constant head boundaries in the middle of the boundary cells and the constant flux boundaries at the borders of the boundary cells. This issue, along with that of accurately representing the position of the river and tributary channels, affected the placement of the model grid over the study area and the choice of cell size. For the present study, it was finally decided to use a cell size of 50x50 m, subject to the considerations of discretisation error described below. The model grid, shown in Figure 6.3, had 49 rows and 70 columns aligned with the National Grid coordinate axes.

Problems of numerical instability are prevented in MODFLOW by the use of "unconditionally" stable backward differencing to represent temporal variations. The Slice-Successive Over-Relaxation package was chosen in order to solve the resulting system of linear equations. A time step of 0.5 days was found to provide insurance against problems in converging to a numerical solution.

While the adoption of backward differencing suppresses the growth of errors in the simulation, the faithfulness of the numerical system to the physical system still depends upon the coarseness of the discretisation (Smith, 1985; Sewell, 1988). It is therefore important to choose a sufficiently small cell size and time step length to ensure that the solution of the numerical system will be sufficiently close to reality. Authors such as Smith (1985) and Sewell (1988) have provided general symbolic expressions for the dependence of the discretisation-incurred error on the step sizes, showing that spatial curvature and higher-
order temporal rates of change in the water table increase the potential for larger step sizes to cause greater error. The discretisation error is thus dependent on the space- and time-derivatives which are a-priori unknown, rendering unreliable the estimation of error bounds for any particular simulation and consequently making it difficult to base the choice of step size on such bounds. In the present study, the 50 m grid cell dimension was chosen primarily on the basis of the ease of positioning of surface channels and boundary conditions in the model, as mentioned above. Since the true head distribution on Goss Moor remains always unknown, error bounds were not calculated for the study domain, but a trial simulation of a steady state unconfined groundwater system in one dimension, for which the analytical solution was known, was carried out with the chosen cell size (50 m) to determine whether the discretisation errors for the wetland aquifer domain were likely to be acceptable.

This one-dimensional system is shown in Figure 6.4. Steady net recharge is applied to an unconfined aquifer of permeability 0.1 m/day with an impermeable horizontal base. The aquifer is bounded at one end by an impermeable unit and at the opposite end by a constant head representing a river or reservoir. The water table elevations given by the finite difference solution scheme differ from the analytical solution by only a small percentage (less than 2%) and a small absolute value (0.50 m at a distance of 1000 m from the river), thereby vindicating the chosen grid cell dimension. In the two-dimensional flow problem for Goss Moor, additional constraints such as land surface drainage and the constant head boundary sections around the edge of the domain would further reduce the finite difference errors. Thus, the 50 m cell size allowed sufficient accuracy for the determination of water table elevations in the wetland aquifer.

6.4.3 Discretised Domain Characteristics

This section describes the allocation of values to the dimensions of the aquifer and the stream channels in the model domain. The method of calculation of the channel bed and land surface drainage conductances is also presented.
Land Surface Elevation
Spot heights supplied in a GIS data set by English Nature (© English Nature 1993) were interpolated to produce a grid of the wetland surface elevation every 50 m. In order to ease the solution of the numerical equations, the grid was smoothed by averaging each cell value with those of its neighbours. The resulting distribution of elevations at the “upper surface” of the model domain is shown in Figure 6.5.

Aquifer Thickness
The 73 bedrock soundings obtained from the mineral survey by Billiton Exploration were interpolated and smoothed similarly to the land surface elevations, to produce a grid with values every 50 m. The original and the smoothed grids of the aquifer thickness are shown in Figures 6.6 and 6.7. The smoothed grid of thickness was subtracted from the smoothed grid of land surface elevations to give aquifer base elevations for input to MODFLOW.

Aquifer Permeability and Specific Yield
As elaborated in Section 2.4.3 and Section 4.6.2, braided channel alluvium, consisting of a disorderly pattern of channel-sized deposits, was found to underlie much of the wetland. The comparison of stratigraphic sequences between nearby boreholes indicated that the sedimentary units were discontinuous even over short distances. Although some spatial trends in the stratigraphy of the aquifer might be expected, due to a transition from alluvial sediments around the river to frost heave and solifluction products on the wetland periphery, for simplicity no broad structure was assumed and the hydraulic properties of the aquifer were assumed to be effectively uniform over the whole model domain. For the calibration of the steady state model, various values of permeability, within the range determined by in situ slug tests (Section 3.5), were tried. The calibration of the transient model (Section 6.7) considered permeabilities within the same range and took into account the geometric mean of the slug test sample. The heterogeneity and highly varied nature of the surficial sediments in the Goss Moor wetland also gave rise to large uncertainties in the appropriate value of specific yield. However, Brassington (1988) showed that the specific yield of most sediments is below 0.3, and the present study adopted the approach of keeping the specified value as low as possible (less than 0.4), as far as was allowed by
numerical stability constraints and the need to allow for the possible action of evapotranspiration (see Section 6.7.3).

Stream Channels

Three different MODFLOW packages were considered for use in representing the stream channels in the model: the River Package, the Drain Package and the General Head Boundary Package. All three modules implement a Cauchy boundary condition, but differ in the degree to which they modify the basic behaviour of the numerical boundary (McDonald and Harbaugh, 1988). In the present study, the choice of module for simulating the effects of the streams was governed by the need to use the Drain Package to represent the wetland’s ground surface, which is considered in a later subsection. Although both the streams and the ground surface required Cauchy boundary conditions, they could not be represented with the same package since their effects on the groundwater heads in the coarsely spaced model cells were distinct in the vicinity of the streams. Therefore, having chosen the Drain Package to represent the ground surface drainage, the River Package was chosen for the streams. While the Drain Package discontinues drainage upon the falling of the water table below the level of the drain, the River Package may allow some leakage from the river to the aquifer when the groundwater head drops below the river stage. However, this leakage was inappropriate in the present context of a headwater stream, and so such behaviour was disabled by setting RBOT = STAGE for all river segments (see McDonald and Harbaugh, 1988).

In MODFLOW’s River Package, stream channels are represented by Cauchy boundary conditions placed at user-chosen finite difference model nodes. The stream water level serves as the external reference head in the Cauchy boundary condition for each cell. Unlike other modelling codes in which the river is simply a constant head condition imposed upon the chosen points in the groundwater domain, this allows the head at the underlying nodes to vary with conditions elsewhere in the domain. Thus, even in a model with only one layer, continuous groundwater flow from one side of the river to the other is possible. In addition, the conductance of the river bed material is specified separately from the conductivity of the aquifer in the local cell. Initially, this can be of benefit in simulations
where the model cells are much wider than the river, since it obviates the need to modify the aquifer conductivity over an inappropriately wide area of the domain. However, the introduction of areal (e.g., meteoric) recharge to the aquifer poses a problem in flowpath definition for the near-stream zone when using a coarse model grid. This problem is addressed below, following explanatory notes on the calculation of bed conductance.

In the present study, the formulation of the hydraulic conductance of the stream bed follows the methods of Miles (1985, 1987) who modified an equation developed by Herbert (1970) for the description of head losses in radial groundwater flow to a partially penetrating stream. Under the assumption of a flat water table and a semi-circular stream cross-section as shown in Figure 6.8, the conductance, in m²/day, of the semi-cylindrical stream bed layer for radial flow is found to be:

\[
C_A = \frac{\pi \cdot L \cdot K_A}{\ln\left(\frac{r+t}{r}\right)} \quad (6.3)
\]

Assuming that the region of radial flow extends out to the underlying model node, as in Figure 6.8, then a similar formula can be used for the conductance, in m²/day, between the model node and the outer surface of the stream bed sediments:

\[
C_B = \frac{\pi \cdot L \cdot K_B}{\ln\left(\frac{T-b+d}{\frac{T}{2} \cdot r+t}\right)} \quad (6.4)
\]

The equivalent inner stream radius was related to the wetted perimeter of the rectangular-section stream as follows:

\[
r = \frac{w+2d}{\pi} \quad (6.5)
\]

The nomenclature of the above equations is given below:
L \quad \text{is the length of the stream reach (m)},

K_A \quad \text{is the hydraulic conductivity of the stream bed sediment (m/day)},

K_B \quad \text{is the directionally averaged hydraulic conductivity of the aquifer in the region of radial flow (m/day)},

r \quad \text{is the equivalent inner radius of the stream (m)},

t \quad \text{is the thickness of the stream bed sediment (m)},

T \quad \text{is the distance of the aquifer base below ground level (m)},

b \quad \text{is the distance of the stream bottom below ground level (m)},

d \quad \text{is the depth of water in the stream (m), and}

w \quad \text{is the stream channel width (m)}.

Referring back to Section 3.5.4, in which significant anisotropy was anticipated in the hydraulic conductivity of the wetland aquifer, and to Figures 4.34 - 4.37 showing noticeable small-scale layering in the alluvium, it was found necessary to provide for the effect of anisotropy upon the radial groundwater flow into the stream, following the method of Miles (1987). This involved the arithmetic averaging, in accordance with an arrangement of conductances in parallel, of the anisotropic hydraulic conductivity over the range of angles of approach from the aquifer into the stream bed, achieved using angle increments of 1 degree of arc. This procedure was used in the calculation of $K_2$ only, since anisotropy would not be involved in the flow through the relatively thin stream bed layer. As suggested by Miles (1987), the conventional ellipse of direction in which the principal axes are $\sqrt{k_h}$ and $\sqrt{k_v}$, giving $k(\theta) = k_h \cdot \cos^2(\theta) + k_v \cdot \sin^2(\theta)$, was replaced with an ellipse in which the principal axes are $k_h$ and $k_v$, giving $k(\theta) = \sqrt{(k_h \cdot \cos(\theta))^2 + (k_v \cdot \sin(\theta))^2}$. This allowed for the fact that the groundwater flow into the stream would not be truly radial, having originated in a shallow aquifer. Although Miles (1987) does not give theoretical justification of the exact nature of the modification, it can be seen that the modified formula biases $k(\theta)$ towards $k_h$, thus accounting for a flow system in which most flow happens to be coming from a near-horizontal direction. In the above formulae,

$\theta$ \quad \text{is the angle of flow with respect to the horizontal (degrees)},

$k(\theta)$ \quad \text{is the hydraulic conductivity in the direction $\theta$ (m/day)},
\( k_h \) is the horizontal hydraulic conductivity of the aquifer (m/day), and
\( k_v \) is the vertical hydraulic conductivity of the aquifer (m/day).

\( C_A \) and \( C_B \) were combined in series to obtain the total conductance between the model node and the stream water body.

Having formulated the effective hydraulic conductance between the model node and the river water as above, it was necessary to address a problem in flowpath definition for the near-stream zone when using a coarse model grid. This problem arises on the introduction of areal (e.g., meteoric) recharge to the aquifer: the finite difference modelling system introduces all such flow inputs/outputs directly to the model nodes. Direct recharge to each node beneath the stream would then encounter only the below-stream and stream bed resistances en route to the stream water, whereas in reality there was also aquifer resistance encountered en route from the various parts of the 50x50 m model cell. To correct for this, the conductance, \( C_X \) (m²/day), corresponding to the average streamward flowpath length through the 50x50 m square of aquifer was calculated and combined in series with \( C_A \) and \( C_B \). The resulting value was input to the model as the river bed conductance. \( C_X \) was calculated as shown below, referring to Figure 6.9.

Representative recharge is assumed to begin flowing towards the channel at a distance of \( \Delta x/4 \) metres from the stream centre.

\[
\bar{l} = \frac{1}{h} \left( h \cdot \frac{\Delta x}{4} - \frac{\pi \cdot h^2}{4} \right) = \frac{1}{4} \left( \Delta x - \pi \cdot h \right)
\]

(6.6)

where

\[
h = \frac{T}{2} \quad \text{if} \quad r + t < \frac{T}{2} - b + d \]

and

\[
h = r + t \quad \text{if} \quad r + t \geq \frac{T}{2} - b + d
\]
If the latter condition was true, $C_B$ was set very high at $10^6 \text{ m}^2/\text{day}$, thus effectively removing the corresponding cylindrical zone of the aquifer. This was equivalent to having repositioned the model node on the lower edge of the stream bed layer, below its original location.

$I$, having been evaluated, was inserted into the following equation for $C_X$:

$$C_X = \frac{k_h \cdot 2 \cdot L \cdot (T - h)}{\Delta x} + \frac{k_h \cdot 2 \cdot L \cdot h}{I}$$

(6.7)

Further symbols used in the above formula are defined as follows:

- $\Delta x$ is the model cell width (m), and
- $l$ is the length of a horizontal flowpath for recently recharged groundwater towards the outer surface of the stream bed layer (m).

These formulae can be expected to overestimate the conductance since, firstly, they assume that the water table is at ground surface, and secondly, they do not account for the lowering of the water table on the approach to the river. However, such errors are likely to be minimal in the case of a wetland such as Goss Moor in which high water tables are maintained due to the low permeability of the substrate. The Hooghoudt-Ernst formula (quoted by, for example, Crebas et al., 1984) would allow a more accurate calculation in cases with a free water table, but its application would be problematic in cases where the water table meets the ground surface as in the present study.

The introduction of supplementary flow resistance to the stream bed also increased the resistance encountered by water flowing towards the stream from farther parts of the aquifer. Compensation was made for this by recalculating the value of hydraulic conductivity assigned to each river-bearing model cell, after the new stream bed conductances had been allocated. This revision effectively set the model cell's aquifer block conductance to a value which, when placed in series with $C_X$, would reset the streamward
flowpath resistance (from neighbouring cells to the outer surface of the stream bed layer) to its original value. The revised conductivity, $K_{rev}$ (m/day), was thus:

$$K_{rev} = \frac{\Delta x}{L \cdot T \cdot \left(\frac{\Delta x}{k_h \cdot L \cdot T} - \frac{4}{C_X}\right)}$$ (6.8)

The above completes the description of calculations involved in setting the river bed conductances. These calculations were implemented in two ways: for the steady state model calibration, they were included in modifications to MODFLOW allowing iterative resimulation, while for the transient model calibration, they were implemented in a spreadsheet which contributed to the MODFLOW input files for single simulations. The modifications made to the MODFLOW source code for the steady state calibration are shown in Appendix C. Input files required for the extended functionality of the modified code are listed in Appendix D. Surface water pools within the river Fal and at the western end of the wetland were also simulated with the Cauchy boundary condition, using the River Package and the General Head Boundary Package, respectively. The conductances assigned to the bed sediments of these water bodies were calculated similarly to the above, but with geometrical differences. This procedure was also used for the particularly wide sections of the river on the approach to/exit from the central in-stream pools, and is described in the next subsection of the present thesis. Below, the current subsection now moves on to consider the model's required information on the characteristics of the streams in Goss Moor.

In the present study, detailed topographic survey data for the stream channel were available for approximately half of the 3.7 km main river course within the modelled area. This data resulted from a survey of the channelised reach extending 1.6 km downstream from Tregoss Bridge (see Figure 2.1), undertaken for a different research project. Other parts of the stream network were sparsely covered by spot measurements of stream channel dimensions. From such information, along with a GIS data set supplied by English Nature (© English Nature 1993), a representation was constructed of the stream network and its various
channel dimensions. The locations of the river segments in the model grid are shown in Figure 6.3, while Figures 6.10 - 6.16 show profiles of the ground surface elevation, assigned bed elevation and assigned water depth over the 7 branches of the modelled stream network. Table 6.1 gives channel width and bed thickness data at upstream and downstream ends of the reaches.

The bed thicknesses were generally set to between 0.5 m and 0.2 m, based on values found in the literature. The assigned bed thickness was greatest in the higher-order channels, and least at locations close to source. Table 6.2 lists the thicknesses of the stream bed and hyporheic zone found at various sites in Europe and North America together with relevant drainage characteristics: catchment area, mean discharge, channel width and bed gradient. Where possible, the nature of the geological substrate is also indicated. The data shown fail to indicate any significant correlation between the listed catchment characteristics and the stream bed thickness. However, they suggest the validity of the adopted range of bed thicknesses in the present study.

Many authors in the groundwater flow modelling literature have addressed flows between groundwater and rivers with poorly permeable bed sediments (e.g. Chin, 1991; Christensen et al., 1998). Such studies are often concerned with large rivers in which there is less flow turbulence, allowing increased deposition of the finer fractions of the sediment load. The rivers concerned may also be recharging the groundwater (an unlikely situation in the headwater environment of Goss Moor), and thus become a focus for a large section of the literature concerned with water resources and water quality investigations. In this context, it is usual to refer to the stream beds as “clogged stream beds” or “channel linings”. An example of the retardation attributed to such layers is evidenced by Sophocleous et al. (1995) who assume 0.1 and 0.01 to be typical ratios between stream bed conductivity and aquifer conductivity. However, other sections of the literature are more relevant to the present study, since the streams in the Goss Moor catchment are predominantly gravel bedded, as reported below, and fall in a headwater floodplain environment in which the grain sizes and permeabilities characterising contemporary in-channel sediments are greater
than those of the ambient floodplain (including overbank) deposits (He and Walling, 1997; Marriott, 1998).

A gravel river bed may potentially have a hydraulic conductivity of the order of $10^3$ m/day, or even greater. However, finer sediments are commonly deposited in the gravel pores, such finer sediments coming in size classes from sand right down to the clay fraction. Recent research into suspended sediment deposition (Stone and Walling, 1997; Phillips and Walling, 1999) has shown that, although individual silt/clay particles are unlikely to settle out of suspension in stream water, such fine sediments may be deposited within composite particles (flocs) which have settling velocities higher than those of their component grains. This raises the possibility of drastic reductions in stream bed permeability by deposition, but only in cases where substantial amounts of fines are retained by colmation (retention processes leading to the clogging of bed sediments just below the armour layer - see Brunke, 1999). A study of bed sediment storage within subcatchments of the river Ouse, Yorkshire (Walling et al., 1998) has shown that in certain small catchments, one could expect to find resuspendable sediments comprising only a small percentage (~1% by dry weight) of the top five centimetres of the gravel bed. This indicated that the bed sediment would not reach the level of impermeability exhibited by fully formed clays or silts, although some reduction of permeability could not be ruled out: the presence of sand in the suspended sediment load of the stream would more strongly bias the bed conductivity towards that of suspended load deposits (Schärlchli, 1992; Brunke, 1999). Since many streams include sandy fractions which may be deposited contemporaneously with the gravel, reductions in permeability may be highly significant in some cases. In laboratory flume experiments conducted by Schärlchli (1992), the hydraulic conductivity near the upper surface of a sandy gravel channel bed was noted to fall from around 4 m/day down to around 0.4 m/day during a deposition event. Brunke (1999) reports intra-bed permeabilities of between 1 and $10^4$ m/day, calculated from particle size distributions measured at various levels within the gravel bed of a prealpine river.

The grain size distribution and layering of the bed sediments in most of the Goss Moor streams remained largely unknown in the present study, although some information was
available for the main river channel. Bate (1997) sampled the bed deposits in the
channelised stretch of river downstream from Tregoss Bridge, finding arithmetic mean grain
sizes characteristic of fine-medium pebbles. This statistic undoubtedly reflects the largest
sizes of particle found in this stretch of the river, belying a wider range of smaller grain sizes
also found in the bed. In a restricted sampling programme of short duration, Ashford
(1996) found the modal grain size to be approximately 0.7 millimetres in sediments
transported within 0.05 m of the stream bottom. D10 was found to be about 0.13
millimetres, and so the silt/clay fraction (comprising particle sizes less than 0.062
millimetres) constituted less than ten percent of this bed load. D90 was found to be around
2 millimetres, somewhat smaller than the prevalent pebbles of the bed surface and
confirming the relative immovability of these stones under normal flows. The bank material,
as sampled by Stokes (1996), featured the poorly sorted clay, silt and sandy gravel deposits
of the wetland alluvium. Although the clayey deposits exhibited strong cohesion, bank
collapses due to stream erosion of less competent strata would frequently introduce such
fines into the sediment supply (Stokes, 1996).

Given the wide range of finer sediments found in storage and supply for the monitored
reach of the main channel, and the geological evidence (see Section 2.4.3) for similar
conditions of sediment supply in other areas of the wetland, deposition and colmation may
play an important role in the behaviour of the bed deposits of the wetland's streams. Such
processes would be counterbalanced by decolmation and resuspension during spates
(Brunke, 1999; Phillips and Walling, 1999; Walling et al., 1998), thus limiting the overall
reduction in bed permeability.

Concluding the above deliberations on stream bed permeability, it is evident that the
literature on the subject provides an extremely wide range of possible values. The sediment
types found in the Goss Moor channels may be sufficiently similar to those used by Schälichli
(1992), to allow adoption of a hydraulic conductivity within the suggested range of 0.4 - 4
m/day. However, in the absence of direct measurement of bed permeability for the present
study, alternative criteria would also be useful in assigning a value to this parameter in the
groundwater model. It will be seen in Section 6.6.3 that, for the Goss Moor wetland,
further guidance can be found through an analysis of the sensitivity of the steady state groundwater model accuracy to variations in this parameter and in the hydraulic conductivity of the wetland aquifer.

**Surface Water Pools**

The model of the alluvial groundwater flow system included representations of stable surface water levels in excavated pools (see Section 2.6), both in the central reaches of the river and at the downslope, western boundary of the wetland. The Cauchy boundary condition was used, and so the River Package, the Drain Package and the General Head Boundary Package were feasible for this purpose. The choice between these MODFLOW modules was governed by the need to keep groundwater exchanges with the in-stream pools separate from exchanges with the western pools, in the wetland water budget of Chapter 7. In order to facilitate this, the River Package was used for the in-stream pools, while the General Head Boundary Package, rather than the Drain Package, was used for the western boundary pools in order not to interfere with the land surface drainage budget. The representation of the western boundary pools affected only four model cells, as seen in Figure 6.3. In the wetland water budget of Chapter 7, the exchanges between these pools and the groundwater were included with the total flow through the outer boundary of the aquifer.

The conductances assigned to the bed sediments of these water bodies were calculated in a similar way to those of the stream channels, with some geometrical differences. Rather than a semi-cylindrical model for the bed layer, a cuboid model was adopted, as shown in Figure 6.17, such that the conductance of the bed layer, $C_A$ (m²/day), was given by:

$$C_A = \frac{K_s \cdot w \cdot L}{t} \quad (6.9)$$

and the conductance of the layer of aquifer material between the underlying model node and the bottom of the bed layer, $C_B$ (m²/day), was given by:
In the above equations,

\[ C_B = \frac{k_v \cdot w \cdot L}{\left( \frac{T}{2} - b + d \right) - (d + t)} \]  

(6.10)

In the above equations,

- \( L \) is the length of the pool (m),
- \( K_i \) is the hydraulic conductivity of the pool bed sediment (m/day),
- \( k_v \) is the vertical hydraulic conductivity of the aquifer (m/day),
- \( t \) is the thickness of the pool bed sediment (m),
- \( T \) is the distance of the aquifer base below ground level (m),
- \( b \) is the distance of the pool bottom below ground level (m),
- \( d \) is the depth of water in the pool (m), and
- \( w \) is the pool width (m).

As with the stream nodes described in the last subsection, the resistance encountered by local areal recharge en route to the surface water body was in need of modification. However, the problem was complicated in the case of the pools by the large amount of recharge captured in reality by the pool itself. Consequently, areal rainfall and its interaction with the aquifer were represented in two ways according to the width of the pool in question: firstly, if the pool was as wide as the model cell, then both the groundwater recharge and the evapotranspiration in that cell were set to zero and supplementary flowpath resistance was made negligible. This accounted for the fact that rainfall and evaporation within the boundary of the cell would affect only the pool water, and not the groundwater. Such treatment was consistent with the representation of the pool as a Cauchy boundary. However, if the pool was less wide than the cell, recharge and ET were maintained. This situation was used to represent the wider sections of the river where the semi-cylindrical stream geometry would be inappropriate. In this case, considerable error could be expected from neglecting to reduce recharge according to river width where the river was particularly wide. Some compensation was attempted for this error in cells where the river was over half as wide as the cell, by reducing the supplementary flowpath resistance to a negligible amount in order to hasten the removal of the excessive recharge.
Flowpath resistances from neighbouring cells were unaffected by this rather crude modification, since they were maintained in the same way as those to the narrower parts of the stream (see previous subsection).

The average poolward flowpath length for recharge arriving on the cell was taken from the midpoint between the edge of the pool and the edge of the cell, as shown in Figure 6.17, and the flowpath conductance was divided into two blocks in series. The end of the flowpath was taken to be \( \frac{1}{4} \) pool widths inwards from the edge of the pool. The conductance of this flowpath, \( C_X \) (m\(^2\)/day), was given by:

\[
\frac{1}{C_X} = \frac{1}{C_{block\alpha}} + \frac{1}{C_{block\beta}} = \frac{w}{4k_hL'T} + \frac{\Delta x - w}{8k_hL'T}
\]

\[\Rightarrow C_X = \frac{8k_hL'T}{\Delta x + w}\]

(6.11)

As stated above, the maximum pool width for which supplementary flowpath resistance was introduced was \( \Delta x/2 \). These formulae can be expected to overestimate the conductance since they assume that the water table is at ground surface and do not account for the lowering of the water table on the approach to the pool. However, such errors are likely to be minimal in the case of a wetland such as Goss Moor in which high water tables are maintained due to the low permeability of the substrate. Further errors are likely to result from the crude characterisation of the flowpath.

\( C_X \) was combined in series with \( C_A \) and \( C_B \), the resulting conductance being specified as the pool bed conductance. It is noted that the pool beds were likely to comprise much finer sediments than those of the stream beds. This was addressed, not by specifying a separate value of hydraulic conductivity for the pool beds, but by specifying a small value for the pool bed thickness, thus allowing the large thickness of the pool bed to be taken up by aquifer material. The assumption used in this approach was that the fine pool sediments had
a permeability which was closest to that of the aquifer material. Finally, the value of hydraulic conductivity assigned to each pool-bearing model cell was revised after the calculation of the pool bed conductances, in order to preserve the original resistance encountered by poolward flow from neighbouring cells. These revisions followed the procedure outlined in the previous subsection.

The calculations of pool bed conductance were implemented in two ways: for the steady state model calibration, they were included in modifications to MODFLOW allowing iterative resimulation, while for the transient model calibration, they were implemented in a spreadsheet which contributed to the MODFLOW input files for single simulations. The modifications made to the MODFLOW source code for the steady state calibration are shown in Appendix C. Input files required for the extended functionality of the modified code are listed in Appendix D.

**Land Surface Drainage**

MODFLOW's Drain Package, which was used to simulate the drainage of high water tables by groundwater emergence at the ground surface, functions similarly to the River Package. The external reference head for each cell was set to the ground surface elevation and the drain conductance set as high as possible in order to prevent the water table from rising too far above this level. However, too high a conductance was occasionally found to cause instability in the simulations, presumably due to sudden abstraction from high water tables during numerical iteration. Consequently, a value of 900 m²/day was found to be the highest possible conductance in the steady state calibration. The drain conductance is related to the model cell dimensions similarly to the river bed conductance, with the difference that the cell width replaces the channel bed width.

Every cell in the model domain was allocated a ground surface drain, including cells already with river segments, but with the exception of cells covered by pools. The Drain Package and the River Package operated independently of each other, were given distinct water table criteria for activation in the model, and had different effects on the modelled heads. The River Package is addressed in a previous subsection, while the operation of the Drain Package as a boundary condition is described in Section 6.4.1.
6.5 SYSTEM STIMULI (RECHARGE AND EVAPOTRANSPIRATION)

MODFLOW simulates both recharge and evapotranspiration by direct local addition/withdrawal of water from the model domain at a user-specified rate. Since MODFLOW deals with only the saturated groundwater zone, processes such as interception of rainfall, infiltration through the unsaturated soil zone and stomatal reaction to climatic/soil moisture conditions must be simulated by the user before inputting the results to the groundwater model. The present section summarises the approach taken to account for processes of this kind in the Goss Moor study, referring to discussion in other parts of this thesis. However, it is noted that the evapotranspiration regime was varied during the calibration of the transient model (Section 6.7.3) and so is further discussed in that section.

Rainfall was assumed to be uniform over the model domain, as discussed in Section 3.3.2, and was input directly to the model domain as recharge. For steady state simulations, the annual average rainfall determined over the period 1/6/93 - 31/5/94 was used while for transient simulations with a 1-week stress period, weekly averages were used. Some provision for the evaporation of intercepted rainwater was made in the ET calculations. However, no other interception-related process was represented, thus omitting any delay in recharge due to storage of rainwater during throughfall. However, this delay would be apparent neither in the steady state nor at the weekly time scale of the groundwater simulations undertaken here. The delay due to the infiltration process was also omitted by assuming it to be shorter than the averaging periods used for the model inputs, since the proximity of the wetland's water table to the ground surface would allow recharge very soon after the incidence of rainfall.

Evapotranspiration was allowed to vary from place to place in the model domain according to the type of land coverage, as shown in Figure 6.18. Four different classes of land cover were considered: wet willow carr, wet heath, open water and pasture, the ET rate for each
class being calculated from field measurements, taking into account seasonal changes in climate and plant activity, as described in Section 3.3.3. Steady state simulations used the annual averages of these ET rates, while transient simulations used weekly averages. As discussed in Section 3.3.3, the estimates of evapotranspiration were subject to errors arising from the assumption of a uniform boundary layer meteorology over the heterogeneous vegetation cover of the wetland. Estimates were most accurate for the wet heath over which local measurements were taken, deviating from the true rate above the aerodynamically and energetically different willow canopy. The direction of this deviation was unknown. With respect to the spatial zonation of ET, the true spatial distribution of ET rates was likely to have fuzzier boundaries than those in the model, due to the disturbance of the surface layer equilibrium and the wind field at the edges of vegetation stands. This type of error may have been comparable to that involved in the 50x50 m discretisation of the ET zones.

As shown in Section 3.3.3, the calculated ET rates included some representation of evaporation from intercepted rainwater clinging to plant leaves. Also included in the ET flux assigned to the model was evaporation from water on the ground surface.

It is useful to distinguish perched from non-perched surface water on the wetland, and from this standpoint to make a further distinction, this time between water which is flowing over the ground surface and water which is trapped in topographic depressions. With the exception of non-perched depression storage, all such surface water can be considered to be separate from the groundwater domain, whether by virtue of its non-Darcian flow or because it is perched. The evaporation from such surface water should not be abstracted directly from the groundwater body since the groundwater plays little or no part in directly replenishing the region of abstraction. For non-perched depression storage, evaporation might be abstracted from the groundwater domain as long as the unit specific yield of the surface pools could also be represented in the model. Initially, such considerations were neglected in the present study, indiscriminately abstracting a quantity of water which was representative of evaporation from the whole wetland surface. However, the transient calibration involved measures which might account for the presence of an unsaturated soil
zone both as a decoupling zone between groundwater and evaporating surface water and as an alternative water source for plant uptake.

When the water table was at ground level, land surface drainage cells were used to constrain the water table below ground level, removing such water as might be considered never to have infiltrated, or to be upwelling and flowing away over land. The drains thus modified the local recharge during times of high water tables. However, the amount of water withdrawn by the land surface drainage function was dependent upon the groundwater head and therefore upon all forced source/sink terms such as the ET. Some compensation would therefore be possible for a poorly specified ET, but only when the drain cells were active. Upon the decline of the water table and consequent deactivation of the drain cells during the summer, the local accretion to the groundwater body would be fully determined by the specified recharge and evapotranspiration.

6.6 STEADY STATE MODEL CALIBRATION

6.6.1 Introduction and Aims

The purpose of the steady state calibration in the present study was to establish the approximate range of uniform aquifer hydraulic conductivities \((k_h, \text{ m/day})\) and stream bed hydraulic conductivities \((K_A, \text{ m}^2/\text{day})\) which would permit a stable groundwater flow system with the appropriate water table shape, given the observed climatic and boundary conditions. This is the first stage of the procedure in which the conceptual model, once coded for numerical expression, produces data to be compared in various ways with the observed data. At such an early stage, the probability that some basic aspects of the model may need to be revised is relatively high. Once a recognisable head distribution is obtained, therefore, the shape of the model domain, the boundary conditions and the system stimuli may need to be reassessed for their effect upon the head distribution. This is effectively an examination of the degree of consistency between the various model inputs, including the calibration reference data. Having resolved any serious inconsistencies in this way, the
steady state calibration may be completed and the resulting information on the aquifer and river bed properties then serves as the basis from which to investigate the system's response to fluctuating conditions in the transient calibration.

6.6.2 The Steady State Calibration Procedure

For recharge to the groundwater domain, the steady state calibration used the annual average of the daily rainfall measured over the period 1/9/93 - 31/8/94. This period matches the water budget period, allowing correspondence between the results of the steady state simulations and the conditions in the water budget period.

Iterative solution of the difference equations for the equilibrium flow system requires a first guess of the groundwater head distribution. The present study used the ground surface elevation as this initial estimate. Initially, convergence to a numerical solution was attempted by removing all time dependence from the equations by setting the specific yield \( S \) in Equation 6.2 to zero. This approach was unsuccessful and a method of convergence based upon a transient flow system was adopted in its place. Unlike the steady state approach, the transient method takes the initial guess not as an estimate of the solution, but rather as initial conditions from which the flow system may evolve. \( S_0 \) is set to a non-zero value (reducing the head fluctuations which occur during iteration) and all boundary conditions and system stresses are left at their equilibrium values. The water table then equilibrates over a long period of 1000 days towards the resulting steady state configuration.

In the steady state calibration, each hydraulic parameter was assumed to be uniform over the modelled area. The values of \( k_h \) and \( K_A \) were optimised by repeated modification and re-simulation, using the reduction of the sum of squared errors between the simulated and the observed equilibrium water table elevations as the criterion for success. For residual variances equal to or smaller than the error variance in the observed data, further optimisation would be unnecessary, as described below.
The steady state water table elevations against which to compare the computed groundwater heads were calculated as weighted averages of the observed piezometer water levels. At each piezometer site, each successive measurement was given a weight proportional to the length of the interval over which it was the closest measurement in time. (Since the water table on Goss Moor was measured over a period of less than one year, each observed time series was wrapped around to provide coverage of one full year and to allow the calculation of such weights for the first and last observations.) Such a simple scheme may result in over-representation of isolated measurements, but conversely, also prevents their under-representation if they have legitimate values which are significantly different from the rest.

Since a large fraction of the piezometer sites were used to specify constant head boundaries (for which the averaging procedure was the same as the above), only 10 sites remained with which to calibrate the model. These sites have been identified in Figure 6.3.

The calibration values obtained as above from the observed groundwater heads contained inaccuracy due to (i) errors in measurement and (ii) uncertainty in the time-average due to the finite number of measurements.

(i) Although, with only 10 sampling sites, the amount of information available is limited, the purpose of these measurements is to find the true water table shape at the particular time of measurement. However, there are uncertainties in the elevation of each piezometer above the common datum. In the present study, the height of each piezometer lip above ground level was measured. Knowing the ground level elevation above Ordnance Datum, all piezometer tubes might then be related to a common datum and hence the true shape of the water table determined. However, ground level elevations above OD were determined photogrammetrically and so incurred errors which were large relative to those of other surveying methods. The variance of the errors in observed water table data with respect to the common datum might therefore be as large as 0.5 m.
(ii) The weighted mean of the water table heights at each site is distributed normally according to the central limit theorem (Wackerly et al., 1996), with a variance equal to \(1/(n-1)\) of the weighted sample variance, where \(n\) is the sample size. The variance in the weighted water table mean at each site is given in Table 6.3. On average, this variance was found to be approximately 0.035 \(m^2\), corresponding to a standard deviation of 0.187 \(m\). The present study constructed a \(\chi^2\) statistic (dimensionless) for each steady state model simulation, to be minimised for calibration, according to the equation

\[
\chi^2 = \frac{(n-1) \cdot S^2}{\sigma_0^2},
\]

(6.12)

where

- \(\sigma_0^2\) is the variance in the error between the time-averaged observed head and the true equilibrium head \((m^2)\), and
- \(S^2\) is the variance in the error between the time-averaged observed head and the simulated head \((m^2)\).

Although the \(\chi^2\) statistic is often used for the testing of the hypothesis that \(S^2 = \sigma_0^2\), the lack of a satisfactory estimate of \(\sigma_0^2\), due to unknown errors in piezometer tube elevations, prevented any such objective analysis in the present study. Nevertheless, a provisional \(\chi^2\) value, in which \(\sigma_0^2\) incorporated only error (ii), was calculated for each simulation and used as the least squares objective function in the calibration.

The \(\chi^2\) statistic compares experimental with expected variance. The experimental variance \(S^2\) incorporates a set of correlated errors between the simulated and the true water table height which, unless they are zero, invalidate the assumption of independence required to apply the \(\chi^2\) distribution. However, if these errors were to become zero, then \(S^2 = \sigma_0^2\), marking the point below which it is meaningless (although not impossible) to reduce the discrepancy between the modelled and the observed data. This point corresponds to \(\chi^2 = 9\) for the present data.
The modifications made to the MODFLOW source code in order to iterate simulations for a range of values of \( k_h \) and \( K_A \) in the steady state calibration are shown in Appendix C. Input files required for the extended functionality of the modified code are listed in Appendix D together with example output.

6.6.3 Results of the Steady State Calibration

As mentioned in Section 6.6.1, some adjustment of the domain shape, boundary conditions or system stimuli is often necessary during the early stages of model calibration. The present study was no exception to this tendency and a small extension of the model domain’s south-west boundary to meet the river Fal, with conversion from a Dirichlet to a Cauchy boundary condition, was found necessary to improve the fit to observed data in this area. The original model domain is not displayed or discussed in this thesis, but is mentioned here to authenticate this account of the calibration process. All results mentioned henceforth come from the improved model domain.

Figure 6.19 shows the dependence of the above \( \chi^2 \) ratio on the base 2 logarithms of a range of hydraulic conductivity values for the aquifer, \( k_h \), and for the stream bed sediments, \( K_A \). Each grid point signifies a simulation with the corresponding values of \( k_h \) and \( K_A \). The range of \( k_h \) values on this grid, from 0.0313 to 32.00 m/day, compares with the range 0.0004 to 13.00 m/day estimated from the \textit{in situ} slug tests in Section 3.5.4. The river bed \( K_A \) values covered a wider range from 0.00024 to 128.00 m/day since preliminary model runs suggested that the model was relatively insensitive to this parameter.

The minimum \( \chi^2 \) value obtained in this optimisation was 37.7, well above the limiting value of 9.0 determined above. This value was determined assuming no errors in the ground surface elevations (Section 6.6.2). It is possible to examine the amount of worsening in such expected errors which would allow one to conclude that the model fit could be improved no further. Using Equation 6.12, the standard deviation in ground elevation errors which would reduce the minimised \( \chi^2 \) value down to 9.0 was found to be about 0.33 m. Bearing in mind that the moor was extensively covered with closely spaced hummocks
of vegetation which were often greater than \( \frac{1}{2} \) m in height, the expected error in the photogrammetrically determined ground surface elevations would probably equal or exceed this value. Thus, the estimates of the equilibrium heads determined from the observed data had reached the limit of their usefulness and could no longer serve to distinguish between rival parameterisations of the model domain.

The optimised aquifer conductivity was \( k_h = 4.9 \) m/day, equivalent to that of a silty sand (Freeze and Cherry, 1979; Brassington, 1988). This result does not agree with the predominance of clay in the Goss Moor sediments. From the response surface shown in Figure 6.19, it is evident that the modelled water table was highly insensitive to values of channel bed sediment conductivity, \( K_a \), greater than approximately 4 m/day. This implied a threshold of conductivity above which the channel bed sediment layers, due to their limited thicknesses, were no longer as important as the surrounding aquifer in restricting groundwater drainage to the streams.

The equilibrium water table contours output by the optimum model are displayed in Figure 6.20, along with the water table depth below ground in Figure 6.21. The latter reveals that the water table was at or near ground surface over most of the wetland, which explains the features of the \( \chi^2 \) surface in Figure 6.19. Reduction of \( k_h \) lessened lateral groundwater drainage and so raised the water table up to the simulated ground surface. Once at model ground level, the water table was unable to rise much further and so ceased to respond to further reductions in the parameter value. Hence the flattening out of the \( \chi^2 \) surface on that side. On the other side of the parameter plane, the extent to which the water table might descend in response to an increased permeability was not limited, and so \( \chi^2 \) was able to increase beyond the value reached on the left-hand side. The minimum value of \( \chi^2 \), corresponding to an optimised water table, lay between the two directions of error. Figures 6.22 - 6.25, showing the modelled equilibrium head superimposed upon the piezometric time series at P2s, P3s, P11s and P14s show that, for the most part, the optimised equilibrium water table fell within the annual range of variation of the measured water table.
The shallow gradient of the $\chi^2$ surface under reductions in aquifer $k_h$ from the optimum value was consistent with the possibility that lower $k_h$ values were also appropriate. This situation was not mirrored for higher values of $k_h$, as described above. Thus, the optimisation provided an upper limit for the aquifer's hydraulic conductivity, but did not constrain possible lower estimates of this parameter. In view of the predominance of clay in most of the sedimentary deposits at Goss Moor, such lower values were considered highly probable rather than merely possible. Comparison with the results of the slug tests in Section 3.5 corroborated this perspective. The measured conductivities were mainly below 0.2 m/day, with an overall range of 0.0004 to 13 m/day. Assuming a lognormal distribution of conductivity values, a rough estimate of the overall effective $k_h$ of the wetland deposits could be obtained by taking a geometric mean of the slug test sample values (de Marsily, 1986). This mean turned out to be approximately 0.05 m/day. The calibration optimum of 4.9 m/day was therefore thought to be unrepresentative of the true value.

In general, the optimum parameter values of any mathematical model are correlated (Brooks et al., 1994; Spear et al., 1994), suggesting that it may be possible to represent the modelled system adequately using more than one combination of parameter values. In assigning model parameters such as the balance of rainfall and ET or such as the head of the river water, there remains the possibility of inaccuracy. Correction of these parameters would displace the water table from its original position for any given value of aquifer $k_h$ or channel bed $K_A$, thus necessitating reoptimisation to agree with the measured equilibrium groundwater heads. The extent to which such changes might affect the optimum conductivities was investigated by repeating the optimisation twice with the value of $(P-E)$ altered successively by -20 % and +25 %. River water head was also altered, in two separate re-optimisations, by ±0.20 m. In this way, a more realistic value might be obtained for the optimum aquifer conductivity.

However, the above alterations in net rainfall and river stage caused only relatively minor changes in the best fit values of aquifer $k_h$ and river bed $K_A$. The increase in net rainfall from 80% of the observed value to 125% of the observed value caused $k_h$ to increase from 4 m/day to 6.06 m/day, reflecting the need to maintain equilibrium without raising and
steepening the water table. The optimisation of stream bed conductivities was unaffected by the increases in net rainfall: water table height remained insensitive to $K_A$ values greater than about 4 m/day. If the optimum aquifer permeability had increased to a much greater extent, then the optimum stream bed permeabilities would also have risen as the bed resistance to increased flows regained importance. Changes in the stream water heads mainly affected $K_A$. By lowering the stream water heads by 0.20 m, the minimum value of $K_A$ for which the water table was optimum increased from 4 m/day to 32 m/day, whereas when stream water heads were raised by 0.20 m, this value of $K_A$ decreased to 1 m/day. The optimum aquifer conductivity $k_h$ remained approximately constant, at 4.8 m/day, when stream levels were lowered, but decreased slightly to 4 m/day upon the raising of the stream levels. In response both to alterations of net rainfall and to alterations of stream levels, the variability of the aquifer permeability was insufficient to improve the level of agreement with the sedimentary characteristics and slug test results.

The response of the optimum stream bed conductivity to alterations in stream levels was counter to initial expectations. By lowering the stream level, one would normally expect that stream bed permeability should decrease in order to prevent an accompanying fall in the water table. However, later problems in the transient calibration (Section 6.7.3) suggested that the observed levels at a few sites near the river (Pl3s, P14s and P16s), where permeabilities may have been locally higher and unaccounted-for processes may have been in operation, were unduly influencing the overall $\chi^2$ surface in the steady state calibration. These particular calibration sites favoured a water table which was lower overall than that favoured by other sites. The above imposed changes in stream level had the effect of modifying the balance between the influence of these near-river sites and that of the other sites in the domain. A lowering of stream levels encouraged the influence of sites P13s, P14s and P16s on the overall $\chi^2$ surface by reducing the model error at these locations. The optimisation then augmented this reduction in error by increasing $K_A$. This view is supported by the reduction of $\chi^2$ from 37.7 to 34.3 under the stream level lowering. The reduction of $K_A$ for raised stream levels was due to a lessening of the influence of the aforementioned calibration sites on the $\chi^2$ surface, causing the optimum $\chi^2$ to rise to 41.5.
For comparison, the reduction of net rainfall by 20% reduced $\chi^2$ from 37.7 to 37.5, while the increase in net rainfall by 25% reduced $\chi^2$ to 37.2.

Furthermore, the steady state calibration optimum of aquifer conductivity, $k_h$, was probably the maximum value of aquifer conductivity which would not cause the water table to drop below ground surface at sites away from the river. Lower values of $k_h$ were not favoured because of the influence of the near-river sites. Thus, the calibration optimum of $k_h$ probably bore little relation to the true overall effective conductivity of the wetland aquifer.

6.6.4 Conclusion

The calibration of the groundwater model to steady state conditions in the wetland aquifer demonstrated the feasibility of representing the observed water table behaviour with a Darcian flow model. Control of the water table at many calibration points by the ground surface produced an insensitivity to the decrease of specified permeability at these sites. Thus, the objective function responded only to changes in error at a few unrepresentative sites, leading to an overestimated aquifer permeability. Thus, contrary to initial expectations, the steady state calibration was unable to provide a realistic estimate of the aquifer permeability. However, the calibration proceeded without problems in stability, demonstrating that the modelled water table was stable for a range of values of permeability which included realistic as well as unrealistic values. Finally, the steady state calibration has provided information relevant to the discussion of bed sediment characteristics in Section 6.4.3. It has shown that the assignment of a high permeability to the stream bed layer in the model, in accordance with the observed gravelly nature of the stream beds in Goss Moor, results in the best fit to the observed water table elevations near the river. Moreover, as long as the conductivity of the bed sediment is above a certain threshold, its exact value is unimportant, reflecting the negligible resistance which it then offers to groundwater drainage to the river in both model and reality.
6.7 TRANSIENT MODEL CALIBRATION

6.7.1 Introduction and Aims

Having established the stability of a two-dimensional, saturated, rain-recharged Darcian flow domain with the geometry and equilibrium conditions of the Goss Moor aquifer, the question remained whether the observed temporal variations followed the expected behaviour for the same system. The aim of the following sections was to use the established system model to reproduce the observed temporal behaviour. In attempting this calibration, the success or failure of the various measures applied would indicate the underlying character of the observed groundwater system and its stimuli. After calibration, an examination of the transient model's behaviour would reveal the ways in which the groundwater body was influenced by rainfall, ET and drainage to stream channels.

6.7.2 Methods Employed in the Transient Calibration

In transient simulations, time-varying stimuli are imposed on the model. Numerical models achieve this by dividing the simulated period into intervals (called "stress periods" in MODFLOW), each interval having a different local value of the imposed stress. Among the first considerations in producing a transient groundwater simulation is therefore the adoption of an appropriate length of stress period, involving the assessment of input variability and output requirements.

Although the proximity of the water table to the ground surface removes the influence of unsaturated zone dispersion and delay from infiltrating water, the prevalence of waterlogged ground during the winter suggests a gradual and continuous, rather than event-specific supply of rain water to the water table. At times during the summer when the water table falls below ground level, infiltrating rain water is assumed to disperse and delay en route to the saturated zone, thus maintaining the smooth character of rain contribution to the groundwater body. Considering also the weekly spacing of the water table measurements, it was decided to average the applied recharge and ET into weekly stress periods. As in the
steady state calibration, the transient simulations used a time step of 0.5 days, therefore with 14 time steps per stress period.

The transient meteorological stimuli to the system were therefore the weekly means of rainfall and ET, ET being at the potential rate at all times due to high wetland soil moisture (see Section 3.3.3). The Dirichlet, or “constant head” boundary conditions for the transient calibration were varied with the observed peripheral water table, as explained in Section 6.4.1. A review of Figure 6.2 serves as a reminder that the outer boundary consisted of mainly no-flow and “constant” head sections, with a small section of head-dependent flow conditions at the south-western corner. River stages were held steady at the same level as in the steady state calibration.

The hydraulic conductivity of the stream bed layers, $K_A$, was set to 2 m/day, in keeping with the range 0.4 - 4 m/day determined by Schülchli (1992) (see Section 6.4.3). While the sensitivity tests of the steady state model in Section 6.6.3 showed that a lowering of the stream levels in the model would result in an increase of the optimum $K_A$ to 32 m/day, such a high value would not be necessary in the transient calibration since stream levels were to be held at their surveyed values. Furthermore, the higher optimum values of $K_A$ for the steady state model were found in the context of higher values of aquifer conductivity, $k_h$ which would not be used in the transient calibration. As explained below, the maximum value of $k_h$ to be considered in the transient calibration was 0.5 m/day. Thus, the adopted value of $K_A$ was higher than the maximum $k_h$ and so would ensure a negligible role for the gravelly stream beds in restricting groundwater drainage.

Water table measurements against which to compare the model output were available for the period 18/10/93 - 12/10/94, as seen in Figures 4.38 - 4.57. Initial conditions for this period were determined by allowing the water table to equilibrate under the steady state stimuli for 1000 days and then performing a 4½ month lead-in simulation, with transient stresses, beginning on 1/6/93 and continuing through to the start of the calibration period.
The transient calibration was approached through manually run numerical experimentation and trial and error, in order to ascertain the influence on the water table behaviour of various factors, such as conductivity, storativity and evapotranspiration. The larger number of parameters involved in this investigation made it unrealistic to attempt an automated global optimisation such as was performed for the steady state calibration.

6.7.3 Transient Calibration Procedure and Results

From the discussion in Section 4.6.3, it can be deduced that among the limited set of groundwater monitoring sites established in the wetland, three main categories of water table behaviour were observed. In the first such category, the water table remained close to ground level all year round, fluctuating little from week to week and falling during the summer by 0.35 m at most. Sites P1s, P2s, P9s, P10s, P18s, P19s and P20s were included in this class. As has been mentioned before, the ground surface at such locations was evidently limiting the height to which the water table might rise.

The second grouping of water table responses exhibited high, ground-limited water tables all through the winter and spring, but with an abrupt decline of at least 0.50 m during early summer followed by a rapid recovery during early autumn. This category included sites P11s, P12s and, to a lesser degree, P3s.

In the third category of groundwater behaviour, shown at piezometers P13s, P14s and P16s, the water table did not reach ground level during the monitoring period, possibly because of drainage by the river. During winter, the water table consequently was able to fluctuate more intensely in response to rainfall. During summer, all three sites showed a deep decline of the water table, although at P13s and P16s the lowest reach of this descent could not be observed since it continued below the bottom of the piezometer.

The objective of the transient calibration was to confirm what factors would influence the model’s ability to respond in each of the above three ways and hence to adjust these factors so that such responses were manifest in the model.
As expected given the results of the steady state calibration, the behaviour of the modelled
water table was insensitive to parameter alterations in winter, when a surplus of mass
accretion maintained the water table within the influence of ground surface drainage.
Therefore, the calibration concentrated on the representation of observed summer variations
at all sites except those in the third of the above behavioural categories.

The steady state calibration (Section 6.6) failed to provide a feasible value for the effective
hydraulic conductivity of the aquifer, since the proximity of the annual average water table
to the ground surface made the objective function insensitive to changes in the value of the
parameter. Therefore, the value determined in the steady state calibration was not used in
the transient calibration. Instead, the range of values considered for the transient calibration
was dictated solely by the values of hydraulic conductivity determined in the in situ slug
tests (Section 3.5). Since very few slug test measurements were available on the wetland,
this posed the problem of validating the adopted value of permeability.

Recently, several authors (Konikow and Bredehoeft, 1992; Beven, 1993; Aschenbrenner
and Ostin, 1995) have focused attention on similar problems of defining, with limited
available data, the geological structure and the distributions of the flow parameters in the
region of interest. These uncertainties in site characterisation are seen to weaken the
validity of any conclusions drawn in distributed modelling studies. The ultimate purpose of
the present modelling effort was to estimate the magnitude of groundwater flows to the
river. Given eventual success in the calibration, determined by reasonable agreement of the
modelled potentials with the observed water table heights, the permeability of the aquifer
would be the most influential of the sediment parameters in the determination of this flux.
Since it remained largely undetermined, it was viewed as the most critical aspect of the
aquifer characterisation. The present study adopted a simple approach, after researchers
such as Peck et al. (1988) and Brooks et al. (1994), to show the range of flow estimates
obtainable given uncertainties indicated by the variability in measured values of this
parameter.
As explained in Section 6.4.3, the $k_h$ field was considered to be uniform over the model domain. Therefore, the calibration was defined by bracketing a plausible range of permeabilities between two fixed values for which the adjustment of other parameters might proceed independently. On completion of the calibration, the fluxes through the groundwater domain could be bracketed using the calibrated head distribution with the two alternative characterisations. The two chosen values of permeability were $k_h = 0.005 \text{ m/day}$ and $k_h = 0.500 \text{ m/day}$, obtained by scaling the geometric mean of the slug test results, $k_h = 0.050 \text{ m/day}$ (Table 3.3).

Recharge (which was set equal to rainfall) and ET were initially left unaltered from their measured values. During winter, the action of the model's drainage cells would remove surplus recharge from the water table as soon as it reached the ground surface. In the summer, the inclusion of surface water evaporation as part of the overall weekly ET value would help to account for rainwater which did not reach the water table. As discussed in Section 6.5, the calculated values of ET also made some account of the evaporation of rainwater intercepted by vegetation.

$S_y$ was initially assigned a low value of 0.05 on account of the high specific retention of clays such as those found in the Goss Moor aquifer. Even lower values were thought to be equally appropriate, but the numerical model became unstable and failed to converge with such values, irrespective of the assigned value of $k_h$. Figures 6.26 - 6.31 show the water table variations modelled, with $S_y = 0.05$, 0.1 and 0.4, and $k_h = 0.050 \text{ m/day}$, compared with the observed water table at a representative selection of the monitoring sites. For $S_y = 0.1$, the model behaviour is instead shown using $k_h = 0.005$ and $k_h = 0.5 \text{ m/day}$. At most sites away from the river, the modelled water table came close to the ground surface, and therefore to the observed water levels, during the winter. However, it sank too low in the summer when modelled with $S_y = 0.05$. This occurred with both bracketing values of conductivity. The observed streamside depression of the water table was not fully reproduced by the model, leading to discrepancies at calibration sites P13s, P14s and P16s.
Increasing $S_y$ to 0.1 reduced the summer decline in modelled potentials, although not so much as to match the observed behaviour. A high $S_y$ value of 0.4 curtailed the water table decline but also subdued the water table recovery in early autumn.

There were site-to-site differences in the degree to which each value of specific yield suited the observed water table behaviour. A high $S_y$ allowed the model a closer fit at P2s and P10s, whereas other sites (P11s and P12s) favoured a lower $S_y$. The high specific yield of 0.4 is close to the total porosity for clay and exceeds it for sediments containing significant amounts of larger grains, such as sand or gravel. The gravity-induced drainage of a water table is unlikely to remove such a high proportion of the pore water. In fact, clayey sediments are left at near saturation by gravitational drainage. As mentioned earlier, this was the reason for the use of a low specific yield ($S_y = 0.05$) for the initial transient simulations. However, drainage by evapotranspiration can remove greater amounts of pore water. In this way, the effective $S_y$ is greater when ET is applied than when gravity is the sole draining force.

In the absence of gravitational drainage, the unconfined storativity represents the amount of water (as a proportion of bulk soil volume) drawn by plant roots from beneath the water table. As the water table falls, the water left in the unsaturated zone continues to be depleted in order to satisfy evapotranspirative demand. Thus, not all transpired water is taken from the saturated zone and the assignment of all evapotranspirative losses to the groundwater model is erroneous. However, some upward capillary flow (flow under tension) of water from the water table might occur as a result of plant uptake from the unsaturated zone. Thus, the true amount of evapotranspirative demand on the water table cannot be known without considering unsaturated flow. This reflects recent research on unsaturated soil water fluxes in wetlands (Gerla and Matheney, 1996). A modelling scheme for flow with varying soil moisture content is necessary for the proper treatment of such situations, together with knowledge of the distribution and osmotic potentials of the plant roots, but these problems are not to be addressed in the present study.
In the present calibration, an intermediate value of storativity, 0.1, was therefore thought to be reasonable. Given that clay porosity reaches 0.5 in some cases, this intermediate value constitutes a non-negligible fraction of the total porosity, and yet does not approach such high values as to suggest shrinkage of the clay. Referring again to Figures 6.26 - 6.31, this resulted in rough agreement between the modelled and the observed water tables, although the model displayed an excessive decline summer water table decline at sites such as P2s and P10s. This over-decline was less significant at sites such as P11s and P12s where the observed water table was, in any case, low during the summer.

As discussed above, this over-decline could be explained partly by the assignment of unsaturated zone ET to the groundwater domain. Additionally, the specified ET included surface water evaporation during dry periods, when no rainwater was available for evaporation. In the next stage of calibration, compensation for this was provided by attenuating the specified ET progressively with water table descent, using MODFLOW’s “ET extinction” facility (McDonald and Harbaugh, 1988).

Observed water levels at locations such as P2s and P10s declined minimally during the summer. Thus, according to the view of plant-soil water interaction outlined above, evapotranspiration would be redirected from the groundwater domain at a very early stage in the water table descent, equivalent to the operation of a very high storativity (approximately 0.4) in the top few tens of centimetres of soil. Alternatively, the presence of surface water at such sites during the summer might be taken as evidence of overland flow from the periphery of the wetland or the entrapment of rainwater from previous weeks. Such ponded water would be expected to infiltrate and recharge the groundwater body, to account for which the net accretion of the modelled groundwater domain must be made more positive by reducing the specified evapotranspiration.

At other locations, less adjustment of the net accretion of the saturated domain would be required since the observed water table descended to greater depths. Thus, a non-uniform pattern of ET attenuation was applied in order to match the observed water table behaviour. Figures 6.32 - 6.40 show the modelled responses in comparison with observed water tables.
after such adjustment of the specified net accretion, with $S_y = 0.1$ and for $k_h = 0.005$ and 0.5 m/day.

Following such adjustments, further improvement of the model fit was deemed unnecessary, although certain shortcomings in model behaviour were still apparent. The most obvious of these problems was the continued overestimation of groundwater head near the river. As seen in Figure 6.38, the augmentation of aquifer permeability from 0.005 to 0.5 m/day could not significantly improve the model’s response at riverside sites such as P13s. The problem remained unsolved in the present study, although the presence of extensive deep pools nearby (Figure 6.2), which had been dug during gravel extraction operations earlier in the 20th century, suggested the possibility of associated drainage channels which remained unaccounted on the map and in the groundwater model. Such drainage channels may have been filled in after the gravel extraction, the nature of the infill determining whether any water-transmitting function was retained and thus whether drainage of the nearby water table might be enhanced beyond its undrained state.

The second remaining problem with model behaviour was due to the boundary conditions. At two measurement sites, P10s and P16s, the changes in modelled groundwater head induced by the exploratory adjustments of the hydraulic conductivity were contrary to initial expectations. The reduction of transmissivities between a stream and its groundwater drainage area would raise groundwater heads in that area, this being apparent in many areas of the present model domain. However, the behaviour of the model at P10 and P16 was also influenced by nearby sections of constant head boundary, shown in Figure 6.31. At sites P10s and P16s (see Figures 6.27 and 6.31) showed that during summer, when there was little recharge, a high permeability permitted inflows from the nearby constant head boundary, thereby maintaining water levels, whereas a low permeability prevented such inflows. This effect occurred despite the temporal variation of the boundary heads to match the seasonal fluctuations of the observed water table, and was likely to occur also in other peripheral parts of the model domain. It lends further significance to the constant head boundary sections in the calibration of the transient model and shows that, if the alluvium was assumed to be more conductive, a more boundary-fed flow system would result in the
model. Because the model boundary heads were varied in time to match observed heads, and thus were not acting as artificially replenished sources, this aspect of the flow system could also be assumed to reflect reality. Such findings demonstrate the significance of the estimated value of hydraulic conductivity in characterising the groundwater hydrology of the wetland.

6.7.4 Interpretation of Transient Model Response

The pattern of the modelled water table depth using $k_h = 0.050$ m/day (the geometric mean of the slug test results in Table 3.3) is shown every 4 weeks from 18/10/93 to 19/9/94 in Figures 6.41 - 6.53. From a water table which was uniformly high on 18/10/93, moderate rainfall brought the water table right up to the ground surface over most of the wetland within the next 4 weeks, remaining at ground level throughout the winter and spring until after April. In the beginning of May, the summer recession began and continued until the end of July. Incipient depressions in the water table on 2/5/94 (Figure 6.48) are seen to originate beneath the areas of strongest evapotranspiration (see Figure 6.18 for comparison), corresponding to the areas colonised by willow. During the simulated period, the stream channels did not appear to induce any depression of the water table. Groundwater flows to the stream network would naturally be greatest in response to the highest water tables, when recharge of the aquifer by rainfall may have been compensating for lateral discharges, and so the high water tables alone were insufficient evidence of negligible groundwater drainage to the stream network. However, the pattern of evapotranspirative drainage emerging in the May water table (compare Figure 6.48 with Figure 6.18) showed that replenishment by rainfall had by this time become too little to compensate for the more powerful drainage processes. Such considerations suggest that the stream channels probably were not draining the alluvium. The evaluation of groundwater fluxes as part of the wetland water budget in Chapter 7 would be necessary to confirm this.

According to the results of the slug tests described in Section 3.5.4, $k_h = 0.050$ m/day was the best estimate of the overall hydraulic conductivity of the alluvial aquifer. However,
conductivities estimated at individual sites ranged from 0.0004 to 13.0 m/day, over a total sample size of 8. This was by no means a fully representative sampling of the aquifer material, leaving a broad uncertainty in the geometric average of 0.050 m/day. Consequently, the drainage characteristics of the aquifer noted in the simulation above were subject to similar uncertainties. To investigate this, the calibrated model was run again with hydraulic conductivity scaled up or down by a factor of 10 (giving $k_h = 0.500$ or $0.005$ m/day), and the behaviour of the water table re-examined. The accuracy of the model was shown to be highly sensitive to changes in storativity in the transient calibration (Section 6.7.3), and so the optimum value $S_y = 0.1$ was retained during these resimulations. All other model parameters were also left unchanged such that any alterations in water table behaviour were due the change in conductivity alone.

For $k_h = 0.005$ m/day, water table behaviour was very similar to that for $k_h = 0.050$ m/day. Therefore, the water table “snapshots” obtained using this value are not shown. As expected, this reduced value of conductivity simply reinforced the aquifer’s inability to drain to the stream channels.

The water table “snapshots” for $k_h = 0.500$ m/day are shown in Figures 6.54 - 6.66. These plots show a marked increase in the influence of the streams on groundwater flow in the aquifer, as seen by substantial drawdown of the water table alongside the channels through most of the year. The influence of incipient summer evapotranspiration is visible in the water table depressions in May, but less so than with $k_h = 0.050$ m/day, being obscured by a broadening zone of drawdown reflecting groundwater depletion by drainage to the streams. Examination of the aquifer water budget in Chapter 7 will confirm the substantially greater groundwater drainage when considering this value of permeability. Since this simulation was shown to agree with the observed behaviour of the water table at the 9 calibration sites (Section 6.7.3), the simulated behaviour was equally as valid in terms of calibration results, as that simulated with $k_h = 0.050$ m/day, except in that the chosen value of hydraulic conductivity was less realistic given the predominance of clay in the aquifer.
However, it may be that the transient model calibration was overly insensitive to changes in hydraulic conductivity, in the same way that the steady state calibration overestimated this parameter (Section 6.6.3). Unlike in the steady state model, the ground surface did not constantly drain the water table in the transient model and so would not have neutralised the utility of the piezometric calibration criterion for the whole study period. However, at the time in summer when this criterion became useful, the water table was also beginning to respond to evapotranspiration at least as readily as to lateral drainage. This would weaken the piezometric criterion, leading to the observed insensitivity and making it advisable to favour other criteria in choosing the appropriate permeability.

Summarising the above discussion, the groundwater simulations have indicated the possible ways in which the water table may have evolved over the study period. Taking the more realistic values of aquifer permeability, $k_h = 0.050$ or $0.005$ m/day, the stream channel network would seem to have little influence on the water table, suggesting that there was not much groundwater drainage to the streams. Evapotranspiration and reduced rainfall were seen to depress the water table in the summer. However, due to uncertainty in the overall permeability of the aquifer, a permeability value of 0.500 m/day was also plausible. Simulation with this value of conductivity revealed the possibility of substantial drawdown of the water table by groundwater drainage to the streams. In this case, the influence of the streams upon the water table was comparable to that of the transpiring vegetation and surface evaporation during summer.

Finally, the modelling did not account for spatial variations in the hydraulic properties of the alluvium. While a good fit was obtained in the transient calibration for many of the observation sites, the model could not reproduce the water table depression observed closer to the river at sites P13, P14 and P16. One possible explanation for this was that the sediments were much more permeable between these sites and the river. One other explanation in the case of P13 and P14 may have been the presence of unsurveyed drainage channels connecting the various excavated pools in the central area of the wetland. In view of the undoubted heterogeneity of the aquifer as described in Section 2.4.3 and the borehole logs (Figures 4.34 - 4.37), the above simulations should be regarded as broadly
representative but lacking in detail. In reality, drainage to the stream network may occur in some parts of the wetland's groundwater flow system, but not in others.

6.8 SUMMARY

A numerical model of the groundwater flow in the alluvial aquifer beneath Goss Moor was developed and calibrated. In Chapter 7, output from this model would provide estimates of the fluxes between the alluvial groundwater body and its surroundings, thus contributing towards characterisation of the hydrology of the wetland.

The USGS finite difference computer program MODFLOW was used to simulate the groundwater flow regime in the wetland aquifer. MODFLOW solves the partial differential equation for groundwater flow given a specified aquifer shape, stream courses, other boundary conditions, sediment hydraulic properties and the regime of recharge and evapotranspiration. Importantly for wetland groundwater modelling, MODFLOW was able to drain the model water table at the ground surface.

The modelled area of 4.5 km² was divided into 50 m x 50 m grid cells for the numerical calculations. Data such as aquifer thickness, boundary potentials and spatial distribution of ET were obtained from 73 pre-existing borehole logs, from current piezometric records and from a GIS conservation data set. The hydraulic conductivity of the aquifer was assumed uniform over the modelled domain and was kept within the range of values determined by \textit{in situ} slug tests in Chapter 3, while the permeability of the stream bed deposits reflected the range of values possible for gravelly stream beds. The model was calibrated, initially in steady state, with respect to measured water table elevations at 9-10 piezometer sites in the wetland.

The automated steady state calibration used a least squares criterion to optimise the values of aquifer and stream bed permeability. However, the objective function was found to be insensitive to decreasing permeability due to the proximity of the water table to the ground
surface at many of the calibration points. This allowed the dominance of errors at certain locations near the river where measured water tables were unusually low, and so caused some overestimation of the aquifer conductivity. The permeability determined in the steady state calibration therefore was not representative of the true effective aquifer permeability and was not used in the subsequent transient calibration. Similar problems might be expected in modelling groundwater flow for other wetlands. Furthermore, instability was found in the numerical system which necessitated solving the steady state equations as a transient problem. This instability may have been due to the amplification of iterative water table fluctuations by wetland ground surface drainage nodes. This would also be expected in other, similar wetland applications. Nevertheless, the steady state calibration finally established the stability of the groundwater system under long term average conditions given the assumed aquifer/stream geometry and reasonable values of aquifer/stream bed permeability.

The transient calibration was conducted by manually run trial and error. The determination of confidence limits for groundwater flow magnitudes was enabled from the beginning of the transient calibration. This involved bracketing a feasible range of aquifer permeabilities with two fixed values for which the adjustment of other parameters could proceed independently. From the range of in situ slug test results, $k_h = 0.005 \text{ m/day}$ and $k_h = 0.500 \text{ m/day}$ were chosen.

Time-varying recharge (which was set equal to rainfall) and ET were initially unaltered from their measured values. The uniform storativity of the aquifer was adjusted to a value of 0.1, giving the best agreement between the modelled and the observed summer decline in water table heights. Although a relatively high value for the predominantly clayey sediments under gravitational drainage, this allowed for drainage by evapotranspiration. Furthermore, some account was taken of the characteristics of plant root systems. It was assumed that water extraction by roots in the upper soil was at least as strong as that occurring at greater depths. Hence, as the water table descended from the ground surface, the vegetation would continue to extract water from the deepening unsaturated zone. Saturated zone ET was
therefore reduced with increasing water table depth. Similar reductions in ET might alternatively result from recharge from ponded surface water.

After the transient calibration, one shortcoming remained in the behaviour of the model: overestimation of groundwater levels near the river. This problem remained unresolved in the present study. The model exhibited one further anomaly in its behaviour, which was not detrimental to the calibration but illustrated the flow system's mutable character in both the model and in reality. This involved the regulation by the permeability of the aquifer, of the influence of peripheral heads upon the flow system.

Finally, head distributions were output from the calibrated transient model using alternative conductivities of 0.005, 0.050 and 0.500 m/day. Given the predominance of clay over the aquifer in general, the two lower values were more realistic and appeared to show that the influence of the stream channels on the wetland water table was negligible in comparison with that of high evapotranspiration. During the summer, depressions developed in the water table in areas of willow, whereas little streamside drawdown could be seen. However, the model showed that significant streamside drawdown was possible using the higher value of conductivity. Although perhaps not a likely parameterisation for the aquifer as a whole, the higher value would probably be representative of restricted regions within the alluvium. Thus, this increased drawdown caused by drainage to streams might be expected to occur in some parts of the real wetland.
CHAPTER 7

THE WATER BUDGET OF THE GOSS MOOR WETLAND

7.1 INTRODUCTION AND AIMS

The primary aim of the present chapter was to estimate the exchanges of water between the wetland and its surroundings. In Chapter 4, the river outflow from the Goss Moor catchment was divided into fast and slow response components. For the slow component, three possible sources were identified, namely unconfined groundwater flow to the stream channels, wetland surface water flow and water storage in river pools. If wetland groundwater flow should prove to be a major source of river slow flow, it would then also significantly affect the wetland’s surface water balance through either recharge or discharge. Estimation of the amount of wetland groundwater flowing to the river, using output from the numerical model calibrated in Chapter 6, therefore forms an important part of the present chapter. Additionally, the model output will be used to quantify the outflow from the wetland aquifer through its lateral boundary, in order to help interpret the zero annual catchment “water surplus” found in Chapter 5. Further comparison and combination of measured and modelled flows from earlier chapters will also help to build a coherent assessment of the hydrology of the wetland.

7.2 EVALUATION OF WETLAND WATER BUDGET

This section presents a daily water budget equation for the wetland, indicating the methods used to calculate the value of each quantity contained therein. Output from the groundwater model provides several of these values. The evaluation of the water budget
equation illustrates the concepts used in considering the wetland and involves the assessment of certain shortcomings in the available data.

Figure 7.1 shows the positions of channel flow measurements and their approximate surface catchments within the extent of the groundwater model domain or wetland. The only flows represented on the map are the stream flows $C_1, \ldots, C_6$ and $C_{2X}, \ldots, C_{5X}$. Other flows within the model domain (wetland) area are identified in Table 7.1 along with certain wetland and subcatchment dimensions, whose values are then listed in Table 7.2.

It is noted that only 11 or so channel flow meterings were available at each of the sites C2, C3 and C4. Moreover, these measurements were taken in the period from January 1995 to April 1997, long after monitoring finished in the present study, for a different research programme, and so gave no direct indication of the flows during the water budget period. Nevertheless, this scanty collection of data could be used to estimate, very roughly, the daily flows in the water budget period at each of these 3 sites.

Table 7.3 gives the metered flows, their means, standard deviations and coefficients of variation. Every such measurement fell within the calendar months December to April. Also shown for comparison are similar statistics for gauging site C6, derived from the hourly flow record over the same 5-month calendar period using a random number generator to simulate the scarce sampling regime of the metered data.

The ratio of each mean spot-metered flow with the mean randomly sampled flow at site C6 gave the factor by which to scale the daily flows at C6, to obtain an estimate of the daily flows at each other site. Bearing in mind the differing site positions within the drainage network, it would not have been surprising to find that these estimated hydrographs exhibited smoother peaks and higher slow flows than the true hydrographs at each site. For sites C3 and C4, this proposition was supported by their higher coefficients of variability within the observed data. However, flows at site C2 appeared to have a variability similar to that of the continuously recorded flows at C6, such that the errors in its derived hydrograph might be considered relatively low.
As mentioned in Section 4.4.1, gauging station C5 was situated on a distributary of the major channel connection between the northern wetland area and the rest of the wetland. The recorded flows from this subcatchment were therefore accompanied by a similar unrecorded volume of flows. In the estimation of the wetland water budget, this was taken into account by doubling the values of flow recorded at C5.

As seen in the map, the wetland was divided into 5 surface subcatchments, numbered from 2 to 6, for the purpose of this water budget analysis. Each subcatchment was delineated using contours available from the Goss Moor GIS database (© English Nature 1993), and constitutes the topological intersection between the surface catchment of each stream flow measurement site and the model domain area. Leaving aside the consideration of subcatchments 2 - 5 and their contributions to the measured stream flows, it would be possible to calculate a daily water budget for subcatchment 6 alone, which constitutes the major part of the wetland area. However, in order to evaluate the water budget of the whole wetland, it would be necessary to back-calculate the stream inflows of subcatchments 2 - 5 (which were not measured) from their stream outflows and other fluxes. This task was undertaken as follows.

Equation 7.1 expresses the budget of daily inflows and outflows to the stretches of channel in subcatchment 6. This part of the wetland channel network receives stream inputs at points C1, C2, C3, C4 and C5. The only output from this sub-network is via the stream outflow at point C6, while stream storage is neglected. Additional inputs are from bed and bank seepage of groundwater (RIV_{0}), from overbank entry of rainwater which has not infiltrated the wetland aquifer (DRN_{0}) and from overbank entry of water which has flowed across the wetland surface from the surrounding hill slopes (S_{IN-L}). The uninfiltreted rainwater and the water traversing from the hill slopes contribute to an increase in the surface water column of the wetland, to surface evaporation and to evapotranspiration from the unsaturated soil zone as well as to the stream flow:
\[ C_6 = \sum_{i=1}^{5} C_i + RIV_6 + DRN_6 - E_{RES6} + S_{IN} \cdot L_6 - \Delta D \cdot A_6 \]  

(7.1)

Nomenclature is given in Table 7.1.

Rearrangement of Equation 7.1 would provide an estimate of the peripheral surface water inflow term \( S_{IN} \). However, the rate of increase of the wetland surface water depth remained unknown and so the undetermined term became \( (S_{IN} \cdot L_6 - \Delta D \cdot A_6) \). For the time being, \( \Delta D \) was assumed to be zero. Additionally, the resulting value of \( S_{IN} \) was taken to apply uniformly over the whole length of the wetland perimeter. These assumptions permitted the use of this value in the calculation of the channel inflows to the remaining four subcatchments, as shown in Equation 7.2.

\[ C_{iX} = C_i - RIV_i - DRN_i + E_{RES_i} - S_{IN} \cdot L_i + \Delta D \cdot A_i \]  

(7.2)

where \( i = 2, \ldots, 5 \) and \( \Delta D \cdot A_i = 0 \).

Having estimated the channel inflow to each subcatchment, it was then possible to combine all calculated and measured fluxes in a single daily water budget equation for the whole wetland:

\[ C_6 + G_{OUT} + E \cdot A + \Delta D \cdot A + \Delta H \cdot S_y \cdot A = P \cdot A + C_1 + \sum_{i=2}^{5} C_{iX} + G_{IN} + S_{IN} \cdot (L - L_0) \]  

(7.3)

Rearrangement would give the average increase in water table height over the whole wetland, \( \Delta H \), for each day.

It was remembered that errors in the values of \( S_{IN} \) and \( C_{iX} \) were incurred due to the neglect of the wetland surface water storage during their calculation. This resulted in three
erroneous terms in Equation 7.3, as listed in Table 7.4. The first error term in this table equals the sum of the second and third, remembering that \[ \sum_{i=2}^{6} L_i + L_0 = L \]. Consequently, no surface water storage error would be incurred as long as the surface storage, channel inflows and peripheral inflows were combined to make one inseparable term, as shown in Equation 7.4:

\[
C_6 + G_{OUT} + E \cdot A + \Delta H \cdot S_y \cdot A = P \cdot A + C_1 + G_{IN} + \left[ \sum_{i=2}^{5} C_{iX} + S_{IN} \cdot (L - L_0) - \Delta D \cdot A \right]
\]

(7.4)

The results of these calculations are discussed in Section 7.3.

7.3 INTERPRETATION OF WETLAND WATER BUDGET

7.3.1 The Water Budget of the Groundwater Model Domain

As explained in Section 6.7.3, a feasible range of aquifer permeabilities was spanned with two values, \( k_h = 0.005 \) and 0.5 m/day, which were applied in the groundwater model to provide the equivalent of confidence limits for the magnitudes of the groundwater fluxes through the aquifer.

The adopted value of permeability affected the ability of the model to incorporate the specified rainwater into the flow of groundwater. A lower permeability (and therefore a lower hydraulic diffusivity) encourages the storage rather than the flow of incoming water, assuming that the water table is currently below ground level and therefore free to rise. The water table would thus rise further for a given input of rainwater and less groundwater would flow laterally, under lower permeabilities. In the present formulation of MODFLOW, drainage cells remove any increases in groundwater storage above ground level. Thus, surplus rainwater does not remain in the groundwater domain. A lower
permeability domain thus retains less of the available rainwater, since it involves more frequent ground level emergence of the water table. As described below, these processes produced measurable differences between the water budgets for the two bracketing values of hydraulic conductivity. While the ranking of the budgeted flows such as drainage to river and evapotranspiration remained identical, large differences in flows were obtained between the two alternative characterisations.

Tables 7.5a and 7.5b show seasonal and annual statistics for the flows through the groundwater model domain. Assuming an aquifer permeability of 0.005 m/day, it is seen that 33% of the rain annually joined the groundwater body. The remaining 67% of the rain was removed from the model domain in the DRN term. That part of the rainwater which joined the groundwater body formed 99.9% of the total input to the aquifer's saturated zone, with inflows of groundwater (GR) through the Dirichlet or Cauchy boundaries accounting for the remaining 0.1%. The overall inflow to the aquifer was then used by evapotranspiration and drainage to the river in the approximate proportions 100% and 0%, respectively. By comparison, a permeability of 0.5 m/day allowed 40% of the rain to join the groundwater body, with 60% removed in the DRN term. The infiltrating rain accounted for 96% of the total input to the saturated zone, the remaining 4% being provided by groundwater inflows (GR) through the boundaries. Evapotranspiration and drainage to the streams then accounted for approximately 76% and 23% of the input, respectively, with about 1% taken up by outflow of groundwater (GOUT) through the perimeter of the domain. In both cases, outflows (GOUT) of groundwater through the Dirichlet or Cauchy boundary sections comprised a minimal part of the overall system of flows. This was because groundwater drained preferentially to the river since, as seen in Figures 6.41 - 6.53, the low level of the river exerted greater influence on the modelled water table shape than did any of the external boundaries. This suggested (due to topography discussed in Chapter 5) that few catchment groundwater losses besides the drainage of the Castle-an-Dinas Wolfram Mine could be expected and so the overall groundwater output from the Goss Moor catchment was small in comparison to river flows, evapotranspiration and seasonal storage fluctuations. Consequently, the “water surplus” calculated in Chapter 5 was attributable mainly to storage in the catchment.
Figures 7.2a and 7.2b compare the magnitudes and variations of $RIV$, $G_{IN}$ and $G_{OUT}$, for the two limiting values of permeability. $G_{IN}$ and $G_{OUT}$ are seen to be minimal in comparison with $RIV$, as explained above. For $k_h = 0.005$ m/day, $RIV$ discharge exhibited clipping at peak flows due to the intersection of the streamside water table with the ground surface and resultant curtailment of the hydraulic gradient. As discussed in Sections 6.4.1 and 6.4.3, one side effect of this was the removal of surplus recharge from the model. Such surplus could be regarded as never having entered the aquifer in reality, since the low conductivity minimised lateral flows within the saturated zone. In contrast, no curtailment of $RIV$ was apparent for $k_h = 0.500$ m/day, since the streamside water table did not meet the ground surface for this value of $k_h$ (see Figures 6.54 - 6.66). Most importantly, $RIV$ was several orders of magnitude lower in the case of $k_h = 0.005$ m/day, in comparison with its value for $k_h = 0.500$ m/day, and accounted for virtually none of the water leaving the aquifer. As shown in Table 7.5a, the removal of surplus recharge left sufficient water in the aquifer for substantial evapotranspiration to occur ($0.524$ m over one year), and so would have permitted significant groundwater drainage to the river if the permeability of the aquifer had been higher.

For $k_h = 0.500$ m/day, the drawdown of the water table near the stream did not preclude intersection with the ground surface at greater distances from the stream during autumn/winter (Figures 6.54 - 6.66). With a higher permeability, the presence of lateral flows allowed the possibility that the model's ground surface drainage partly corresponded to some re-emergence of water which had entered the aquifer elsewhere. However, this would be unlikely to occur without the presence of a region with no ground surface intersection upslope of the emergence zone. This is due to the fact that the ground surface drain normally takes almost all of the water leaving a model cell since it offers a high conductance in comparison with the routes through the other cell faces, and so allows very little lateral flow to neighbouring cells. Furthermore, the depression of the water table by evapotranspiration does not produce a valid source area for an emergence zone. In the Goss Moor model, a configuration of the water table in which water tables upslope of an emergence zone were below the ground surface and yet were not subject to strong
evapotranspiration, was not found even for a high value of $k_h$ (see Figures 6.54 - 6.66). Thus, problems of interpretation for the DRN term did not arise and it could be assumed to represent water which did not in reality enter the flow domain.

Since $G_{IN}$ and $G_{OUT}$ were both minimal, the only significant input to the aquifer was from vertical recharge, given in the tables by $P-DRN$, while the only significant outputs occurred through evapotranspiration ($E$) and drainage to the stream channels ($RIV$). The time-cumulative flow budget of the groundwater model could therefore be plotted using only these three quantities, as seen in Figures 7.3a and 7.3b. However, $RIV$ was negligible in the case of $k_h = 0.005$ m/day (Figure 7.3a), whereas it was noticeable for $k_h = 0.005$ m/day (Figure 7.3a). This highlights again the uncertainties in the groundwater flow budget due to uncertainty in the aquifer permeability.

The discrepancy between the cumulative rainfall value and the combined output terms is the level of wetland aquifer storage above its level at the beginning of the accumulation period. The fluctuation of this storage is seen to account for very little of the water flowing into or out of the model domain.

As mentioned at the beginning of the present section, the uncertainty in the aquifer permeability was not great enough to affect the ranking of the groundwater budget components. At both ends of the assumed range of conductivities, the model water budget shows that evapotranspiration rather than drainage to the river was the main cause of water table decline over the study year. $RIV$ may have exceeded $E$ on occasion during winter, when $RIV$ was at its greatest due to high water tables and the ET was reduced by lower available energy and leaf cover and higher ambient relative humidity. However, this was not the case at the seasonal time scale (Tables 7.5a and 7.5b).

As seen in Figure 7.3a or 7.3b, the wetland aquifer could not absorb all rain-water during winter, and so the removal of water through drainage to the river, although greater than in other seasons of the year, was more than counterbalanced by a surplus of supply. The
effects of groundwater drainage to the river and of evapotranspiration were thus more noticeable during summer when less rain was available.

The observed and modelled water table variations showed that autumn was the time of greatest increases in wetland groundwater storage. As shown in Tables 7.5a and 7.5b, between 0.031 and 0.037 m (m$^3$ per m$^2$ of modelled wetland area) of water (22% of the recharge) went into storage during autumn. Storage in winter was steady at its maximum level, reflecting the oversupply of water mentioned above. The storage then returned to lower levels with changes of between -0.012 and -0.018 m during spring and with a change of -0.027 m in summer. All such values were effectively averages over the whole wetland area. Storage depths might be expected to differ between sites, as illustrated by the observed water table variations in Figures 4.38 - 4.57.

While P-DRN accurately represented the recharge to the numerical model, it may not have been faithful to the true recharge, particularly in the summer period. There were two reasons for this. Firstly, the calculation of recharge relied upon the operation of the drainage cells. These cells were mainly inactive during the summer period, removing no surplus water from the groundwater flow system and thus allowing overestimation of recharge. Secondly, as explained in Section 6.5, interception of rain-water by vegetation was not explicitly simulated. All rain-water was allowed to enter the aquifer whose water budget was then corrected by a slight increase in evapotranspiration due to a reduced resistance against water vapour diffusion from wet leaves. This may have produced some overestimation of both recharge and evapotranspiration from the aquifer.

Regarding the first of these two recharge problems, the influence of further factors such as the presence of ponded water, for which no data were available, precluded any attempt to partition the rainfall into recharge and runoff during the summer. MODFLOW did not provide for the modelling of surface water detention/depression storage. The agreement of the modelled potentials with the observed water table was therefore taken to indicate that recharge was correct in the model. The second recharge problem was taken to be
unimportant since the simulated evapotranspiration did not change greatly during rainfall, indicating that losses due to evaporation from intercepted water were small.

The temporal variation of groundwater storage in the model domain, $\Delta H \cdot S \cdot A$ ($m^3$) was obtained by rearranging Equation 7.4. It is shown in Figures 7.4a and 7.4b, in comparison with the overall catchment storage fluctuations. The values are normalised with respect to area. The diagrams show that the modelled wetland groundwater store has a threshold beyond which it cannot increase. This is due partly to the limited storativity of the aquifer. An additional factor determining the possible storage is the limited elevation of the top of the modelled aquifer above the drainage level. The existence of the wetland may be due to a combination of these two factors with the low permeability and low hydraulic gradients in the alluvium, together with the low slope of the ground surface in the floodplain.

Figures 7.5a and 7.5b, in which the storage fluctuations are expressed as equivalent catchment depth, show that the wetland groundwater storage accounted for only a very small part of the total volume of catchment water stored/released during the study year. The main components of storage in the catchment remained unidentified in the present study. Wetland surface water storage might be of greater significance than wetland groundwater storage in this respect, while storage in the outer catchment also could not be ignored.

7.3.2 The Water Budget of the Wetland Stream Channel Network

The purpose of the present section was to compare stream flows entering and leaving the wetland with the flow contributions from rain falling on the wetland itself, in order to clarify the role of the wetland in the generation of stream flow.

Tables 7.6a and 7.6b give seasonal and annual statistics for various components of the total catchment outflow, including the contribution made to such flow by rain-water which had fallen on the wetland. Figures 7.6a and 7.6b compare such flows over the water budget year. In Figure 7.6a, groundwater drainage is too small to be visible. Its values lie very
close to the x-axis. The tables involve stream flow components which were separated in Chapter 4 using a digital filter. Since the parameters of the filter were varied from site to site (reflecting the differences in the recession constant between the sites), a component so determined at one site may not have fully corresponded to the same component at another site. For example, quick flow at the outlet (site C6) could have involved a certain proportion of the slow flow at site C5 due to the higher recession constant ascribed to slow flow at the latter location. Therefore, the values could be used to give only a general indication of the balance of flows.

Incoming quick flows were assumed to include the total flows at sites C2, C3 and C4 as well as the separated quick flows at sites C1 and C5. Incoming slow flows included only the separated slow flows at sites C1 and C5. As in previous chapters, the terms “quick flow” and “slow flow” are used only for in-channel flows in the present chapter.

There were certain differences between the flows presented in Tables 7.6a and 7.6b, which arose from the alternative values of aquifer permeability. With $K = 0.005$ m/day, $RIV$ decreased by 98% of its value using $K = 0.500$ m/day. This decrease was equivalent to 0.03 m of water per year over the catchment area. $DRN$ increased by a similar absolute volume, amounting to an increase of around 16%. However, the total volume of wetland-sourced river flow, $= RIV + DRN - E_{RES}$, changed very little. The synopsis of the wetland water budget would therefore be essentially the same for $K = 0.500$ m/day and $K = 0.005$ m/day: the total river outflow included 51% slow flow, as determined in Section 4.5.4, the remainder being attributed to quick flow. Wetland-sourced water accounted for between 17% and 18% (between 15% and 17% from $DRN - E_{RES}$ with between 0% and 3% from $RIV$) of the total outflow.

Wetland surface water is likely to exhibit greater reservoir storage effects than the in-channel water body, even accounting for such features as the in-stream pools near the centre of Goss Moor. That is, outflow from the densely vegetated wetland surface would probably exhibit greater time-dispersal of a given input impulse than would outflow from the channel system. In Section 4.4.1 it was observed that rapid peaks in flow were
smoothed away in transit from the upstream gauging station, C1, to the catchment outlet, C6, and that therefore the channel provided substantial dispersion of flow-momentum. The wetland ground surface would be expected to provide even greater pulse-integration. Although the areas of wetland within a short distance of the stream channels would deliver water from rain events earlier than the remainder of the wetland, such areas comprised only a small part of the overall wetland area due to the low channel density of the drainage network in the wetland, and therefore would source relatively little of the wetland runoff.

An examination of the difference between incoming and outgoing quick flows (= outgoing quick flow - quick flows entering subcatchment 6) shed further light on whether such streamside wetland runoff would be necessary to help account for all of the outgoing quick flows. Figure 7.7 shows the daily variations in this difference over the water budget year. The main source of momentary deviation from zero appeared to be due to the flashier nature of quick flow at C1, since most of the peaks were initially negative and then became positive as upstream quick flow receded to a level below the downstream response. The seasonal total deviations reached negative values as low as -0.03 m and positive values as high as 0.02 m, as shown by the values tabulated in Table 7.6a. These seasonal deviations were thus slightly greater in magnitude than the seasonal totals of wetland groundwater flow to the river. Negative seasonal deviations possibly indicated that some of the incoming quick flow may have been retarded in-stream so as to fall within the slow flow response at the catchment outlet. Positive seasonal deviations would indicate that some quick flow contribution from wetland subcatchment 6 had occurred.

Since in-stream retardation of flows would be greater during summer when the river stage was lower, while wetland runoff would naturally be encouraged in winter, the above seasonal deviations would be expected to be negative in summer and positive in winter, with less pronounced deviations in spring and autumn. However, this was not the case, the deviation being most negative during winter and only slightly negative during summer. This suggested that the deviations were due to errors in the measurement and separation processes rather than due to any true hydrological process. Additionally, the annual total deviation was the smallest of all annual total flows considered. Therefore, a close event-
averaged correspondence between upstream and downstream quick flows was not inconsistent with the data, given the number of possible errors in flow measurement and calculation. For the purposes of calculating a wetland budget, it was therefore assumed that the wetland-sourced water would contribute only to slow flow. Furthermore, given the conservation of quick flows within the channel, a major contribution from the wetland would be required to maintain slow flow. This is evident from Table 7.6a/b, where the surplus of outgoing over incoming slow flow is greater than the wetland-sourced flow in subcatchment 6. The occasional exceedance of catchment slow flow by this contribution, seen in Figures 7.6a/b, was due to poor parameterisation of the effects of wetland surface storage as explained below.

The limitations in accounting for the storage of surface water on the wetland, introduced in Section 7.2, particularly affected the accuracy of the daily values of wetland-sourced flows to the river. However, the character of the resulting errors could be taken into account when inspecting Figures 7.6a/b and the tabulated values.

The estimate of wetland-sourced river flow was given by the expression

\[ C_6 - C_1 - \left[ \sum_{i=2}^{5} C_{ix} + S_{IN} \cdot (L - L_0) - \Delta D \cdot A \right] - G_{IN} + G_{OUT} \]

and so included the total gain in wetland surface water storage. (This expression is equivalent to \( RIV + DRN - E_{RES} \)). The true contribution to the river would therefore be somewhat less than that calculated during times when the wetland surface was gaining water, and somewhat more at times of surface water depletion. This fact may have partially explained the wetland-sourced flow's occasional exceedance of the catchment slow flow and its negative values from June to August 1994, seen in Figures 7.6a/b.

However, in immediate terms, the negative summer values of wetland-sourced flow were caused by demand from wetland surface/vadose zone evapotranspiration exceeding supply. The question remained whether sufficient wetland surface storage of directly incident rainwater could be accumulated during the winter months whose depletion would then redress this balance of supply and demand during the summer. If not, then the observed data would
imply that overland inflows from the surrounding hill slopes were required to augment the wetland water store so that it could satisfy the evapotranspirative demand. This question could not be answered definitely in the scope of the present study. However, as seen in Figures 7.6a/b, the peaks in the calculated wetland-sourced flow often preceded the corresponding peaks in river slow flow by up to 2 weeks. This further suggested that the calculated wetland-sourced flow to the river was too rapid and therefore that much scope remained for the incorporation of a slowly draining storage component into the calculation. Thus, overland flow contributions from the neighbouring slopes might not have been necessary to explain the measured balance (local to the wetland surface) between rainfall, evapotranspiration and flow to the river, as far as could be seen on this narrow time scale. However, it is shown later in the present section that the annual wetland water balance did require such inputs.

Due to the involvement of wetland surface water storage described above, the seasonal distribution of overland inflows from the surrounding hill slopes could not be estimated. However, assuming that no great overall surface water storage change took place over the complete study year, it was possible to estimate the annual peripheral overland contribution to the wetland.

Of the slow flow leaving wetland subcatchment 6, 36% came from tributary slow flows. These were denoted source (1). Thus, 64% of the total outgoing slow flow (0.367 m³ per m² of entire catchment) remained to be explained. Three further sources of water would combine to account for this fraction of the outgoing slow flow. Overall, the sources were:

- source (1): slow flows contributed by tributaries,
- source (2): that proportion of the quick flows incoming to subcatchment 6 which had dispersed in-stream to become slow flow,
- source (3): the flow (overland and subsurface) which was sourced in subcatchment 6, and
- source (4): water transmitted across subcatchment 6 (via the surface storage body) from the surrounding hill slopes.
Source (2) included dispersed quick flows from the other four wetland subcatchments. Based on the assumed close event-averaged correspondence between upstream and downstream quick flows and the possible measurement errors for such flows, it was reasonable to adopt the small annual difference between these flows as the annual total of source (2). This value was \(-0.019 \text{ m}^3\) per \text{m}^2 of entire catchment \(= 0.019\) m, or 3\% of the total outgoing slow flow (see Table 7.6a). This was equivalent to 1\% of the annual river outflow.

For source (3), all seasonal totals, including the negative summer total, could be summed to give the annual total of between 0.129 and 0.132 m\(^3\) per m\(^2\) of entire catchment, or 22-23\% of the total outgoing slow flow. This was equivalent to 11-12\% of all river outflow. The negative summer total was due to the evapotranspiration term, \(-E_{\text{res}}\), in the defining expression given earlier. Its inclusion under source (3) implied that all surface/soil evapotranspiration detracted from directly incident rain-water rather than from water introduced to the subcatchment across its outer boundary. Compensation for this misrepresentation would be introduced later.

An initial estimate of the annual flow volume of source (4) was found by subtracting sources (2) and (3) from the remaining outgoing slow flow. The resulting annual river flow volume attributable to peripheral flow into the wetland from the surrounding hill slopes was approximately 0.22 m, or 39\% of the total outgoing slow flow. This was equivalent to 20\% of the annual river outflow. Because no evapotranspiration was subtracted from this figure, it represented the input to the wetland rather than the corresponding output from the wetland to the stream channels.

Although it was impossible to determine accurately the time-dependence of this flow from the surrounding uplands, it would be expected to follow a seasonal pattern similar to that of all flows so far measured or simulated in the present study. For example, the stream flows generally reached their minimum during July and August of the study year. The flow at C1 almost wholly accounted for the minimal stream outflow from the wetland at C6 during late
July 1994 (see Figure 4.8). Therefore, the contribution of the other tributaries, of the wetland and of any outer source areas contributing via the wetland was also minimal during this time. Earlier in the present section, a contribution from wetland-sourced flows was found to be necessary in order to account for a large proportion of the slow flow leaving subcatchment 6. Consequently, the wetland exhibited sufficient flow retardation to smoothen the input from rainfall so as to provide a gradual flow recession. (The true wetland-sourced flow was undoubtedly more slowly varying than the estimates shown in Figure 7.6a/b, as explained above.) This implied that the flows from source (4) above, which also contributed to the slowly varying river flow, need not have been subject to dispersion/retardation before entering the wetland. However, the possibility of prior dispersion remained and would correspond with the potential for substantial storage in the weathered layer of the surrounding hills.

For \( k_h = 0.005 \text{ m/day} \), the relative annual magnitudes of the main inputs and outputs of subcatchment 6 (which covered the main part of the wetland) could be summarised as shown in Figure 7.8a and Table 7.7a. Tributaries of the river (including the outflows of the other four wetland subcatchments) accounted for 63% of all inputs to subcatchment 6, while rain accounted for 19%. The remaining 18% of the input was by peripheral overland/near surface flow. Evapotranspiration accounted for only 8% of the outputs from this subcatchment, while the remaining 92% of the output was via river flow. Channel flow grew from an input of 0.75 to an output of 1.12 (m\(^3\) per m\(^2\) of entire catchment), gaining by one half of its incoming value. This gain was provided by drainage from the wetland surface in the case of \( k_h = 0.005 \text{ m/day} \), with negligible subsurface lateral drainage.

For \( k_h = 0.500 \text{ m/day} \), all relative annual magnitudes are shown in Figure 7.8b and Table 7.7b. They remained nearly the same as with \( k_h = 0.005 \text{ m/day} \), except that \( \text{RIV}_6 \) now composed 1.3% of the total output from the subcatchment.

In Figure 7.9a and Table 7.8a, the fluxes into and out of the surface and substrata of wetland subcatchment 6 using \( k_h = 0.005 \text{ m/day} \) are considered in isolation from the stream flows. Rain and surface/near surface flows each account for 50% of the input to this area.
20% of the output was via ET, 0% via groundwater flow to the streams and 80% via wetland surface runoff to the stream channels.

Figure 7.9b and Table 7.8b show the same fluxes for $k_h = 0.500$ m/day. Again, rain and surface/near surface flows each account for 50% of the input to the subcatchment. 20% of the output was via ET, 3% via groundwater flow to the streams and 77% via wetland surface runoff to the stream channels. The increased aquifer permeability therefore raised the groundwater drainage losses from zero to 3% of the annual losses from the wetland surface/substrata. Based upon this result, the uncertainty in aquifer conductivity would appear not to be reflected in the hydrological behaviour of the wetland. However, the increase in $k_h$ is known to cause a significant alteration in the modelled water table behaviour (Section 6.7.4), that is the development of a drawdown zone around the stream channels and the enhancement of drawdowns across the wetland during the summer. The fact that the annual water budget does not reflect the change in water table behaviour is explained by the budgetary importance of the winter surplus of rainfall and of evapotranspiration during the summer, neither of which have been affected by the change in permeability.

The approximate equality of rain and diffuse peripheral inflows as annual inputs to the subcatchment implied that the annual contributions of these two sources to the river outflow should also be roughly equal. This was because, once on the wetland, the water from one such source could not be distinguished from water from the other source, evapotranspiration thus acting equally on all such water. The contributions of sources (3) and (4) above to the river flow could thus be revised and set equal. The need for such a revision could also be seen from the inconsistency in accounting for ET in the initial calculations, as explained previously in the current section. As a result, the contributions to the river outflow were as follows:
36% of outgoing slow flow (18% of total river outflow).

03% of outgoing slow flow (11½% of total river outflow).

31% of outgoing slow flow (15% of total river outflow).

31% of outgoing slow flow (15% of total river outflow).

Total output from the subcatchment was approximately equal to total input, primarily because the assumption of invariant storage level was used in the calculation of the peripheral overland inflow. The water stored on the wetland ground surface was assumed to return to its previous level after 12 months and therefore the inequality between (tributary slow flows + rain) and (outgoing slow flow + ET) could be resolved using peripheral overland inflows alone. Quick flows were not involved in this calculation. The equality of the outgoing quick flow with the tributary quick flows, which contributed to the equality of total outflow and total inflow, was a feature of the measured and spectrally filtered data.

Finally, one observation remains to be made on the analysis used in the present chapter. Among the important results of the analysis was the finding that the wetland contribution to the river channel was composed primarily of flow which varied slowly enough to account for the stream flow recession observed at the catchment outflow. In Section 7.3.2 an appraisal was made of the probable physical origins of the slow flow, finding that it was most probably sourced from wetland runoff rather than from the smoothening of upstream quick flows. This appraisal was undertaken as a result of the large proportion of outgoing slow flow which was unaccounted for by upstream slow flows. If this proportion had turned out to be small, a different appraisal would have had to be made - that of the probable physical origins of a large amount of quick flow. Therefore some of the analysis carried out in the present chapter may be viewed as contingent and conditional on the results of the stream flow separation conducted in Chapter 4. Furthermore, the results of the numerical groundwater flow modelling in Chapter 6 were also important by eliminating groundwater flow from the set of physical processes which were considered in relation to the augmented river slow flow.
7.4 SUMMARY

In the present chapter, components of the flow of water through the wetland were compared which had been individually determined in previous chapters from measurements and numerical simulations. The aim was to establish the relative magnitudes of such flows. These flows included:

- channel flow introduced to the wetland river channel via tributaries,
- wetland surface water flow originating from rain falling on the wetland,
- wetland surface water flow introduced from the hill slopes surrounding the wetland,
- wetland groundwater flow to the stream channels,
- flow through the horizontal boundaries of the wetland aquifer,
- river flow leaving the wetland, and
- evapotranspiration from the wetland.

Fluxes through the modelled groundwater domain were listed for two bracketing values of aquifer permeability: $k_h = 0.005$ m/day and $k_h = 0.500$ m/day. Significant differences in the water budget of the aquifer were found between the two cases. Recharge increased from 33% of rainfall for $k_h = 0.005$ m/day to 40% of rainfall for $k_h = 0.500$ m/day. The ratio of annual evapotranspiration to annual river uptake was 100:0 for $k_h = 0.005$ m/day and 76:23 for $k_h = 0.500$ m/day. Boundary flows into/out of the aquifer were negligible in both cases. The fact that the outflow of alluvial groundwater was negligible suggested, due to topography discussed in Chapter 5, that the catchment as a whole lost little water via groundwater flow. It was also shown that the level of water storage in the wetland aquifer was limited by low specific yield and the proximity of the aquifer drainage level to the top of the aquifer. The overall catchment storage was not subject to such a limit. Wetland groundwater storage variations thus made up a very small part of the overall catchment storage fluctuations.
Inflows to and outflows from the wetland via stream channels were compared with contributions from rain falling on the wetland itself, in order to clarify the role of the wetland in the generation of stream flow. A close correspondence between event-averaged upstream and downstream quick flows was compatible with the assumption that the wetland contributed little to quick flows. Furthermore, a large proportion of the slow flow leaving the catchment was unaccounted for by incoming slow flows. It was therefore appropriate to regard all wetland runoff as constituting this part of the river slow flow. Thus, the separated stream flows of Chapter 4 were combined in the present chapter, providing evidence that the wetland runoff was slowly varying.

However, the absence of surface water storage measurements on the wetland itself presented difficulties in estimating wetland-sourced river flow. A simple assumption of zero wetland surface storage was examined for its effects upon the estimation of this component of flow. The resulting estimates of wetland-sourced river flow exceeded the outgoing slow flow at times of high rainfall and became negative at times of low flow. The estimated flow peaks also preceded those of the in-channel slow flow by up to two weeks.

Such inaccuracy also affected the seasonal totals of flow. Thus, confidence could be placed in only the annual total of flow from the wetland to the stream. Flow sourced from wetland subcatchment 6, which covered the main part of the wetland, was found to constitute 31% of the annual outgoing slow flow (15% of all river outflow). Tributary flows (including the outflows of the other four wetland subcatchments) accounted for 39% of the annual outgoing slow flow (19% of all river outflow). The remaining 31% of outgoing slow flow (15% of river outflow) was attributed to water which had flowed across wetland subcatchment 6 from the adjoining uplands.

Finally, the permeability-induced changes in the water budget of the aquifer did not have a great effect on the water budget of the wetland as a whole. The annual inputs and outputs of the wetland surface and substrata in subcatchment 6 (which covered the main part of the wetland) were divided up as follows.
Inputs:
⇒ rain 50%
⇒ peripheral overland flow 50%

Outputs:
⇒ evapotranspiration 20%
⇒ groundwater flow to river 00-03%
⇒ wetland surface runoff to channels 77-80%.

In contrast with the water table behaviour and the aquifer water budget, the annual water budget of the wetland's surface and substrata was thus relatively insensitive to a 100-fold increase in the permeability of the aquifer. This insensitivity was explained by the budgetary importance of the winter surplus of rainfall and of evapotranspiration during the summer, neither of which had been affected by the change in permeability.
CHAPTER 8

SUMMARY AND CONCLUSION

8.1 REVIEW OF RESEARCH AIMS

The aim of this study was to furnish an understanding of the water fluxes and storages occurring at the subcatchment scale in Goss Moor, a large lowland wetland in Cornwall. A major aspect of the hydrological regime in need of investigation was the rapidity of depletion of wetland water during the receding period of the annual cycle, in relation to the same recession in other parts of the catchment. Such knowledge would underpin subsequent approaches to the control and mitigation of ecohydrological changes known to be occurring in the wetland. Thus, the study aimed to answer the following three complementary questions.

A) How much and what type of flow is contributed to the wetland surface and to its substrata?
B) What are the relative water demands from the various drainage processes on the wetland surface and on its substrata?
C) Does the wetland suffer from more rapid depletion during the summer than other parts of the catchment? (The answer would highlight whether remediative work should be performed on the wetland as well as on other parts of the catchment.)

The hydrological characterisation was achieved using variables measured directly on site, using spectrally derived stream flow components and using flows output from a calibrated numerical model of the transient groundwater potentials beneath the wetland. The device of groundwater modelling and the distributed spectral analysis of stream flows were included
partly as further exploration of the applicability of such techniques for investigating wetland hydrology.

8.2 SYNOPSIS OF RESULTS

The final section of each of the first seven chapters in this thesis provides a detailed summary of the results of that chapter in isolation. The present section complements and draws together these summaries by focusing on the most important aspects of the results.

The geographical context of the study features Goss Moor as a wetland of 5 km² with a sparse stream network and a single channel outlet, situated in a broad alluvial basin surrounded by low, weathered granite uplands. While the kaolinised metamorphic rocks underlying the alluvial deposits isolated the wetland’s local groundwater from deeper groundwater systems, a fragmentation zone beneath the surface of the surrounding granite hills was suspected to be involved in the transmission of water to the wetland from the outer catchment. The wetland vegetation was characterised as wet heathland with extensive patches of willow carr, while the neighbouring hill slopes supported improved pasture. Surface water pools occupied occasional excavated pits scattered across the wetland, providing detention storage of some of the water flowing across the wetland surface. Such surface water was prevented from directly leaving the catchment by a disused railway embankment defining the downslope boundary. The climate was oceanic with most precipitation (which included virtually no snow) occurring during winter. This precipitation amounted to around 1.3 - 1.4 m per annum, in comparison with annual losses of around 0.5 m by potential evaporation.

The study concentrated on the hydrology of the wetland itself, excluding surrounding parts of the catchment from the quantitative analysis due to limited time and resources. This left some uncertainty in the interpretation of the quantitative results and limited the detail of possible answers to the questions above. Where appropriate, the extent of such limitations is considered below.
The hydrometric measurement programme involved the recording of stream flows and water table depths at various locations across the wetland. In addition, rainfall and evapotranspiration were monitored at one site on the wetland and the hydraulic conductivity of the alluvial sediment was determined from water level recovery tests performed in several of the piezometers. The stratigraphy of the wetland aquifer was obtained from the work of previous researchers on the moor.

The results of many of these measurement operations highlighted the difficulties involved in producing unbiased hydrological data. Estimates of river flow were possibly subject to errors ranging from a few percent to greater than 15 percent, while evapotranspiration estimates suffered from the effects of sampling a spatially variable field of net radiation, wind speed, temperature and humidity at only one location. Water table elevations were referenced to datum within a margin of as much as 0.5 m, while the small number of locations at which the water table was observed did not give a complete picture of the water table behaviour. Furthermore, a small sample size and the possibility of well skin effects during the slug tests made it impossible to hydraulically characterise the wetland aquifer in detail, despite the availability of previous stratigraphic data, and produced great uncertainty in the aquifer's overall effective permeability. All such problems increased uncertainty in the catchment and wetland water budgets and subsequently restricted the analysis to the very simplest of achievements which, nonetheless, would fulfil the overall aims of the study. These achievements are outlined below.

1) The Goss Moor stream outflow was characterised by its daily flow duration curve and this curve was compared with those of other wetland catchments in Britain and North America in order to provide a wider perspective for the results of the present study. For example, it was thus possible to see how different the wetland's hydrological regime might have been given modified topography and geology.

River flows from Goss Moor which were exceeded 90% of the time had higher normalised values than those from other British and North American wetland headwater catchments.
Similarly, at 10% exceedance the normalised values of the Goss Moor outflow were slightly lower than those from the other wetland headwater catchments. In this way, the Goss Moor catchment produced steadier outflows than the other wetland catchments, suggesting that it had a greater than usual degree of storage.

2) Evapotranspiration (ET) rates from the wetland were compared with rates from the pasture of the outlying catchment. Thus, the significance of any differences between the ET rates from the wetland and those from the pasture was assessed. It was found that the wetland ET rate was higher, mainly due to the presence of extensive surface water which produced evaporation additional to the evapotranspiration of the vegetation.

3) The slowly and quickly varying components of stream flow were separated by spectral filtering at sites upstream and downstream of the wetland. The criterion used in defining the slow flow was that it should reproduce the later stages of each stream flow recession, whereas the quick flow was defined as the residual of the total flow after removing the slow flow. These definitions were equivalent to those used in other methods of slow flow or base flow separation. Although the spectrally separated components could not be theoretically attributed to particular source processes or catchment flow paths, use of the above criterion implies that the processes responsible for the quick flow have virtually ceased to operate during the later stages of the flow recession. The identification of the processes which produce quick or slow flow must be carried out through additional techniques such as hydrochemical analysis or hypothesis testing by numerical modelling. The latter of these two techniques was chosen in the present study due to its potential for providing insight into other aspects of the wetland’s hydrology, as explained under item (5) below.

An initial comparison was made of the relative rates of variation and proportions of the total flow ascribable to these two components at the upstream and downstream ends of the wetland, assessing the possibility of wetland-derived effects on the stream flow characteristics. It was found that the recession constant downstream of the wetland was greater than that upstream, indicating that slow flow had become somewhat faster in its
variations. The proportion of total flow ascribable to slow flow was also slightly greater at the downstream site. These findings provided the first, tentative evidence that the wetland contributed slow flow to the stream, although the possibility remained that much of this slow flow had arisen through retardation of quick flows in the in-stream pools in the central area of the wetland. Such in-channel dispersion of flow was also evident in the fact that the downstream quick flows varied far more smoothly than those upstream.

4) Using the observed balance between rainfall, river outflow and ET for the catchment as a whole, the significance of groundwater outflows from the catchment in comparison with the seasonal changes in catchment storage was assessed. It was necessary to assume that the catchment received no groundwater inputs through its lateral perimeter, thus equating the groundwater catchment with the surface catchment. There was some uncertainty associated with this assumption due to the relatively constant high elevations just beyond the southern boundary of the catchment. No definite estimate could be obtained for the amount of groundwater outflow since the overall change in catchment storage was unknown for the study period. However, the groundwater outflow volume could not be any greater than the storage deficit accrued by the end of the study year.

The catchment storage fluctuations could be discerned despite a lack of information on the presence of any overall trend in storage. These fluctuations showed that storage was maintained at high levels over the winter and spring, in accord with the comparison of flow duration curves referred to in item (1) above.

5) A numerical model of the transient groundwater flows through the wetland alluvium was developed, using uniform hydraulic conductivity since the number of locations at which permeability tests were conducted was not sufficient to permit the construction of a heterogeneous aquifer model. During calibration of the model, insight was gained into the effects of drainage to the stream, of limitation of the water table by the ground surface and of evapotranspiration on the behaviour of the wetland water table.
The streamside drawdown of the water table varied with the assigned hydraulic conductivity $k_h$ of the aquifer. Whereas virtually no drawdown occurred around the stream for the best estimate value $k_h = 0.050$ m/day, a distinct streamside drawdown zone was observed for $k_h = 0.500$ m/day. Due to the predominance of clay in the wetland alluvium, the tenfold increase of $k_h$ over the best estimate was unlikely for the aquifer as a whole. Nevertheless, it was not inconceivable that restricted areas of the aquifer might exhibit similarly augmented permeabilities. In places where such areas were in hydraulic connection with a stream channel, the water table would be expected to show drawdown similar to that noted in the high permeability model. However, it is mentioned in item (6) below that the storage of water on the wetland surface may have persisted over summer, thus providing continued recharge to the underlying aquifer. MODFLOW could not simulate this process, but if such facilities had been available then the riverside drawdown zone in the water table of the high permeability model might not have developed during the summer to the same extent.

The modelled water table stayed at the ground surface for long periods during the winter and spring, irrespective of the assigned value of aquifer permeability. This was due to a surfeit of recharge over drainage, which overrode other influences on the groundwater body. However, higher evapotranspiration rates and lower rainfall produced water table decline during the summer.

It was found necessary to compensate, during the calibration of the groundwater flow model, for the ability of roots to extract water at high tension and for the development of an unsaturated soil zone during the summer. The storativity of the clayey sediments was increased from their very low weight-induced specific yield up to 0.1, in order to account for high-tension extraction of water by roots. (Aquifer storage fluctuations nevertheless remained very small in comparison with the fluctuations in catchment storage.) Furthermore, it was assumed that water extraction by roots in the upper soil was at least as strong as that occurring at greater depths. As the water table descended from the ground surface, the vegetation would continue to extract water from the deepening unsaturated zone. Therefore, saturated zone ET was made to decrease with increasing water table depth and so the groundwater body was less affected by ET during the summer, despite the
seasonal increase in total ET. ET from the groundwater domain was slightly less during June-August than during March-May. Gerla and Matheney (1996) found similar effects in a prairie wetland in North Dakota, USA. Bradley (1996) similarly used MODFLOW's ET extinction facility as a means of reducing modelled wetland groundwater losses. Due to the availability of surface water as replenishment, extraction by roots from the unsaturated zone in a wetland would also induce less upward capillary flow from the water table than it would in other environments.

The numerical model also quantified the groundwater flows ascribable to evapotranspiration, stream uptake and aquifer boundary exchanges. These data were incorporated into the wetland water budget described under item (6) below, in which were clarified the result of the catchment water budget considered under item (4) above and the likely sources of the slowly varying component of stream flow.

6) Flows and fluxes from the ET calculations, the stream flow separation and the numerical groundwater flow model were combined with minor stream flows measured at other locations in the wetland to produce a water budget for the wetland. There were two unknown quantities in the wetland water budget: the peripheral diffuse surface inflow and the change in storage volume of wetland surface water. It was impossible to estimate the surface water storage changes due to a lack of measurements. However, on assuming the annual storage change to be zero, the annual diffuse surface inflow could be estimated. The result is considered below after examining the other terms in the wetland water budget.

A large throughput of water traversed the wetland site within the stream channel network. On an annual basis, this accounted for 62% of all water entering the main wetland area (subcatchment 6). Upon leaving the wetland site, the river had gained flow by one half of its incoming amount. This gain came to the river from the wetland surface/substrata.

The spectrally filtered slowly and quickly varying components of the stream flows upstream and downstream of the wetland were compared quantitatively within the framework of the wetland water budget. Although altered in character, quick flows were found to be
conserved in traversing the wetland, and so the 50% gain in river flow from the wetland was largely slow flow. The budgetary output from the numerical model was used to evaluate the potential for alluvial groundwater flow to produce this river slow flow and was also used to provide some appraisal of the water budget significance of uncertainty in the aquifer permeability.

The effect of uncertainty in the value of $k_h$ was significant only in the aquifer water budget. For $k_h = 0.005 \text{ m/day}$, roughly 33% of the rain falling annually on the wetland entered the aquifer. However, this generated virtually no groundwater flows to the stream channels and was all lost to evapotranspiration. In contrast, for $k_h = 0.500 \text{ m/day}$, around 40% of rainfall became recharge of which 23% drained to the channels.

In the context of the main stream flows, such uncertainties in the groundwater drainage to the stream were insignificant. As a result of the 100-fold increase in aquifer permeability, the proportion of catchment slow flow attributable to groundwater seepage rose from zero to 5%. Therefore, groundwater drainage could not account for the observed slow flow leaving the wetland. Thus the gain in slow flow was routed primarily by diffuse surface flow from the wetland.

The wetland water budget was used to calculate the amount of water which had fallen as rain on the wetland and which was now available for such surface drainage (and also for seepage) to the river. The estimates of this wetland-sourced river flow did not account for detention on the wetland surface and so exhibited rates of variation similar to those of the rainfall. The peaks of this flow preceded the corresponding river slow flow peaks by up to 2 weeks, implying a 2 week interval of concentration for the wetland surface runoff. Water traversing the wetland surface from the peripheral hillsides would have a similar delay in reaching the stream. Most importantly, the comparison between the estimated wetland runoff, which varied abruptly from week to week, and the smoothly varying slow river flow to which it contributed illustrated that a considerable degree of flow dispersion and storage occurred on the wetland surface.
Groundwater flows through the lateral boundaries of the wetland aquifer were negligible, meaning that the only significant inputs to the wetland surface and substrata were rain and diffuse surface/near surface inflow. Using the assumption that the annual total input and output of the wetland were equal, implying no overall change in storage, the annual peripheral diffuse surface inflows to the wetland were estimated as the residual when all other inputs were subtracted from the sum of the outputs. This gave a total inflowing volume of around 1.5 m, similar to the input of 1.6 m from rainfall. The two annual influxes combined to give around 3.2 m of water over the area of the wetland.

Evapotranspiration (ET) constituted 20% of all losses from the wetland surface/substrata. 0-3% of the losses were attributable to groundwater drainage by the streams. Diffuse surface runoff to the stream channels accounted for 77-80% of the wetland losses.

Losses from the entire catchment were also dominated by river flow during all seasons except summer, when ET took precedence. On the other hand, the main output from the wetland surface/substrata as a whole was surface runoff. However, ET held sway in the wetland alluvium, as shown by the numerical model in which it accounted for between roughly 76% and 100% of the recharge.

The amount of water stored in the catchment was found to fluctuate, rising by 0.2 m in autumn/winter and falling in spring/summer. By comparison, the storage fluctuation in the modelled wetland substrata was small (~0.02 m) due to low aquifer storativity and limited thickness above the drainage level. These two factors, along with low permeability, probably were significant in the formation of the wetland. Even with a relatively high permeability of 0.500 m/day, the winter water table remained at ground surface and was unaffected by any outflows since recharge was readily available from a winter surplus of surface water.

In summary and in answer to the main questions of the study, the above findings reveal that rain and peripheral surface inflows were roughly equal inputs to the wetland. These inputs were stored mainly on the wetland surface, being discouraged from infiltration by low-
permeability sediments. Despite having higher evapotranspiration than the surrounding areas, the wetland's dominant loss was via surface runoff to the river channel. This runoff provided most of the slow component of river flow from the catchment. Since half of this wetland runoff originated from rain which had fallen directly on the wetland, the wetland could be seen to provide significant retardation of flows by virtue of its densely vegetated surface. Accounting for part of the stream recession flow, it was one of the more storative flowpaths in the catchment. However, the degree to which runoff from the hill slopes onto the wetland periphery exhibited any flashiness remained undetermined in the present study due to masking by the wetland intervening between the base of slope and the locations of the stream flow recorders.

Finally, the relative magnitudes of the wetland inputs by rainfall and peripheral inflow would be partially a function of the climatic conditions during the period of monitoring. During the present study, this monitoring period had higher than average rainfall. Whereas the long term average annual rainfall was 1.4 m, the study site received 1.8 m of rain during the water budget year, an increase of 30%. Upon reversion to more normal annual weather conditions, the concentration of flow from the outer catchment into the wetland area would assume greater significance in comparison with the input from rainfall. This should be kept in mind when considering the water budget results of the present study.

8.3 RESEARCH METHODS AND FURTHER DEVELOPMENT

This section considers the value and limitations of each separate method of analysis in the present study and suggests enhancements which might lead to improved characterisation of the site's hydrology.

8.3.1 Analysis of Stream Flow Characteristics

Separation of stream flow into two different components, one of which varied slowly enough to reproduce the observed stream flow recession, was achieved using a spectral
filter. The intention was to infer the storage attributes of the contributing flow paths directly from the stream flow characteristics. The method provided not only the components separated with respect to the recession, but also produced a recession constant as a property of the slowly varying component which could be used to slow flows at different locations and times.

The flow separation was applied at sites both downstream and upstream of the wetland, thus quantifying the effect of the wetland upon the stream flow characteristics. This distributed use of the spectral flow separation also proved useful in estimating volumes of flow attributable to the wetland. Despite considerable alteration of the behaviour of each flow component in the transferral downstream, the use of the flow recessions as indicators of the discontinuation of quick flow-generating processes (Linsley et al., 1982) seemed to prevent any crossing over between the quick flow and the slow flow. This may be further investigated by modelling the in-channel dispersal of the upstream flows with a physically based numerical model of the river channel and in-stream pools, with no lateral inputs, and spectrally separating the downstream model output to see if the flow components are conserved.

8.3.2 Numerical Modelling

The numerical modelling undertaken in the present study illustrated the effects of domain geometry on the exchanges between the wetland groundwater and its surroundings. The calibration procedure for the transient groundwater flow regime addressed several characteristics of the flow system such as hydraulic conductivity, specific yield and evapotranspiration, testing and modifying the assumptions initially used with regard to such quantities. Thus, the effects of drainage by the river and of evapotranspiration upon the water table were demonstrated.

Steady state optimisation:
The calibrated steady state model established the stability of the groundwater system under long term average conditions given the assumed geometric and parametric properties of the aquifer. Two major problems were encountered during the steady state calibration:

1) Steady state calibration with water table drainage at the ground surface posed a problem in the convergence of the solution vector, due to the removal of excessive quantities of water in response to high heads during iteration. This situation was resolved by redefining the steady state solution as the final equilibrium of a transient problem in which the iterative water table fluctuations were reduced by a high storage coefficient. Other methods of representing contact of the water table with the ground surface, for example by switching the water table from a variable to a constant head, might not cause this problem.

2) Limiting of the water table height also caused insensitivity of the steady state objective function to decreasing permeability. This resulted in consistent overestimation of the aquifer permeability by the optimisation routine. The permeability estimated in this way was unrealistically high given the clayey nature of the wetland sediments.

Both of the above problems arose from the application of the model to a wetland and so similar problems might be expected in other steady state wetland groundwater modelling exercises.

Geological parameterisation:

Because the locations of groundwater head measurement were few and far between, little consequence could be attributed to any spatial parameter value distributions determined by calibration. However, this was not important for the purpose of investigating the main factors in water table behaviour. Nevertheless, with regard to improving the spatial validity of the hydraulic conductivity and storativity fields, geological information from the Billiton UK commercial mineral survey of Goss Moor might be used to greater advantage. The 73 borehole logs available from this survey give great stratigraphic detail, recording
sedimentary units as thin as 0.1 m. However, the hydraulic interpretation of such lithologies can be problematic when the data was originally gathered for mineral assay rather than for hydrogeological assessment.

At an early stage in the present study, some progress was made towards the incorporation of the stratigraphic data into the numerical model, through 2-dimensional Dirichlet tessellation around the drilling locations and quantification of lithotype continuity between the resulting tiles. This approach was most suited for supplying spatial information to a 2-dimensional numerical model, and yet was also capable of providing approximate structural information in 3-dimensions. A vertically averaged spatial pattern for the hydraulic conductivity was thus incorporated into the 2-dimensional Goss Moor model. However, on steady state calibration this heterogeneous field performed slightly worse than the uniform conductivity field and so was abandoned in favour of the latter. Such problems illustrate the aforementioned difficulty in interpreting lithological information. Additionally, greater sophistication in spatial interpolation might produce better results. In particular, a geostatistical treatment of the spatial distribution of lithotypes and their parameters might prove more successful than a search for continuity between sampling points as widely spaced as the Billiton boreholes. Such improvements would be implemented by incorporating kriging into the automated calibration procedure for the steady state system.

The above possible enhancements in geological parameterisation are a fully wrought modelling refinement which might not yield great improvements in model fit given the available data. As an alternative means of improvement, spatial variations in model fit indicate where supplementary data might prove useful. In the present study, the area to the north of the river and near the central pool system contained low water tables which could not be reproduced with conductivities within the range determined by slug tests in other locations. This suggests that supplementary drilling and conductivity measurements would be appropriate in that area, along with sampling of the river and pool bed sediments to obtain a better idea of the conductances between these water bodies and the aquifer. Investigation of the area around the ponds for evidence of remnant drainage ditches might also be fruitful.
8.3.3 Water Budget

Both the catchment-scale and the wetland-scale budgets were a means of combining and comparing measured and modelled fluxes so as to allow inferences on the nature of the hydrological system. For instance, the catchment water budget provided a tentative indication that no groundwater entered or left the catchment, while a comparison between the slow component of river flow and the wetland water available for runoff gave an estimate of peripheral contributions to the wetland surface.

Such inferences were conditional on the validity of certain assumptions about the water budgets which may be placed in two categories. Firstly, having assessed the geology, topography and climate of the catchment, constraints were imposed on the type of system under consideration. In the present study this involved denying any input of deep groundwater into the shallow aquifers of the catchment. The denial of any snowpack storage was a similar, although indisputable, assumption. The second major assumption commonly used in water budgeting is that the errors in measured and modelled fluxes are not so great as to change the relative magnitudes of the terms in the equation. Errors ranging from a few percent to more than 15 percent in measured stream flow are possible, the Goss Moor measurements being no exception (Section 3.2.3). The discrepancy in annual rainfall between Goss Moor and the supplementary gauging site at Roche was unknown, as was the error in potential ET due to differences in wind speed and net radiation with St. Mawgan airfield.

8.3.4 Further Process Investigations

Modelling of root uptake and infiltration:

In this study, the observed water table stayed higher during the summer than that simulated using potential ET and observed rainfall as recharge. Compensatory reductions in the specified model ET were attributed to preferential extraction by roots from the unsaturated soil left above the descending water table. Furthermore, it was possible that surface water
infiltrated to replenish the depleted soil water. These processes might be investigated by monitoring surface water depth, soil moisture/tension and water table depth in the wetland, at a site removed from the vicinity of the river and its associated horizontal groundwater drainage. The observed data would be used in the calibration of a numerical model of variably saturated vertical seepage with vertically distributed ET and a coupled surface water budget. Such work would improve the applicability of numerical models to wetlands.

**Monitoring of hillslope and wetland surface flows:**

The speed of flow over the wetland surface itself could profitably be compared with that of the runoff from the surrounding hillslopes, thus appraising the degree to which the wetland offers flow retardation of such runoff. As demonstrated by Chappell (1990), hillslope flow velocities over distances of 10 - 100 m could be evaluated with a combination of tracer tracking and piezometric techniques on a very short time scale. Distributed measurements with an electromagnetic flow meter would serve to estimate wetland surface flow velocities.

### 8.4 MANAGEMENT IMPLICATIONS

The present study was not directly concerned with assessing hydrological management scenarios at Goss Moor. However, the results of the investigation shed light on certain possible approaches for preventing the desiccation of the wetland. All these approaches involve the augmentation of the wetland’s or the catchment’s propensity to store rather than transmit water. They are outlined below:

1. **Removal of Willow Scrub**

   Evapotranspiration rates were estimated to be greatest from the areas of willow scrub in Goss Moor. However, the difference between the ET rate from willow and that from wet heath was minimal and so the removal of the willow carr and reintroduction of other communities of wetland vegetation might not cause any significant reduction in the ET losses from the wetland. However, the understorey in the willow carr was sparse in comparison with other ground-level vegetation (Jan
Davies - English Nature, *pers. comm.*). This suggests that removal of the willow would eventually cause retardation of surface water flows over large areas of the moor, through the reintroduction of a denser mat of vegetation on the ground surface. The resulting situation with slower overland flow would then resemble circumstances before the willows gained a hold on the site.

(2) **Alteration of hill slope drainage network**

From the point of view of wetland conservation, the ideal change to the drainage of the outer catchment would be the blocking up of the ditch network so as to encourage the maximum possible amount of diffuse flow down onto the wetland surface. However, this approach would cause wetter hill slope soils which would be unacceptable to residents and landowners. Alternatively, the existing ditches could be redirected away from their delivery points on the main channel network and onto the periphery of the wetland. Although resulting in a flashier supply of water than would the complete blockage of the drains, this would at least ensure that the available water went to the wetland surface rather than the channel network.

(3) **Raising of river levels**

By installing weirs in the wetland reaches of the river, the river stage would be increased and the water table drawdown caused by groundwater drainage would be reduced. This scheme would be most useful for summer conditions when, as shown by the observed piezometer responses and the transient groundwater simulations in the present study, the riverside water table was at its lowest.

(4) **Investigation and reduction of wetland substrate drainage near central ponds**

The low water tables near the central river ponds could not be reproduced using aquifer permeabilities determined from measurements elsewhere in the wetland. This may have been caused by drainage to ditches which were not shown on the map of the area, and which may have been incompletely filled in. If gravels of high permeability were used for the infill, then drainage of the local groundwater would
continue. Location and refilling of such ditches would raise the water table and improve summer soil moisture conditions in this part of the wetland.

When implementing such improvement schemes, it is recommended that hydrological monitoring be continued in the wetland in order that the effects of the above management techniques can be properly assessed. Such analysis must also take into account the variability of the climatic stresses on the hydrological system of Goss Moor. Because the hydrological monitoring for the present study was conducted over a wetter-than-average period, the water balance of the wetland may have been somewhat different from normal. The lower rainfall of a normal year would probably increase the significance of lateral surface flow from the hill slopes in the wetland water budget. This would affect the efficacy of each of the different hydrological management treatments outlined here. Furthermore, the performance of these management measures would vary from year to year along with the fluctuating climatic balance. In order to assess the interactions between climatic and management influences on the hydrology of Goss Moor, a record of the wetland’s water balance over a representative range of annual weather conditions would be essential. Therefore, daily monitoring of rainfall and evapotranspiration should continue for several years along with weekly water table readings and continuous recording of the wetland’s stream inflows and outflows.

8.5 FINAL REMARKS

It has been shown that distributed measurements of hydrometric variables such as water table depth, stream flows and evapotranspirative flux can form the basis for a successful investigation of wetland hydrology. In the analysis of such data, the combination of stream flow separation with numerical modelling of groundwater flow in a wetland-scale water budget, allows inference towards the hydrological processes occurring in the wetland. This approach is generally suitable for investigating the hydrological balance of other headwater valley wetlands such as those found in the south-west of England.
APPENDIX A

DATA RETRIEVAL AND CONVERSION OF STAGE TO DISCHARGE FOR GAUGING STATIONS ON GOSS MOOR

The flow past the gauging station is determined according to an equation of the form

\[ Q = \alpha \cdot h^\beta \]  

(A1)

where

- \( Q \) is the stream discharge \((m^3s^{-1})\),
- \( h \) is the recorded stage \((m)\),
- \( \beta \) is a constant (dimensionless), and
- \( \alpha \) is a constant \((m^{3\beta} s^{-\beta})\).

The stage, \( h \), is here defined as the elevation of the stream water surface above the downstream flow control. Water surface level is originally measured with respect to the zero point on a fixed measurement staff, and so stage is obtained by offsetting the original measurements by the height of staff zero above the control.

In general, there is an error in the discharge obtained from measurement \( i \). This error is denoted \( dQ_i \). There are two main sources of this error: random inaccuracies in digitising the stage, \( dH_i \), and uncertainty in prediction of \( Q \) using the stage-discharge relation, \( dq_i \).

The total variance in the determined flow values, \( V_Q \), is given by

\[ V_Q(h) = \frac{1}{N - 1} \sum_{i=1}^{N} (dQ_i)^2 \]  

(A2)

while the flow variance resulting from the \( dH_i \)'s is denoted...
\[ V_h \{h\} = \frac{1}{N-1} \sum_{i=1}^{N} \left( \frac{\partial Q}{\partial H} \right)_i^2 \{h, \alpha, \beta\} \cdot (dH)_i^2 \]  

(A3)

and that resulting from the \( dq_i \)'s is denoted

\[ V_q \{h\} = \frac{1}{N-1} \sum_{i=1}^{N} (dq_i)^2 \]  

(A4)

The derivation of the stage-discharge data did not involve the process of digitisation, and so the errors in prediction of \( Q_i \) from the stage-discharge relation do not contain any component due to digitisation inaccuracies. That is, \( dq_i \) is independent of \( dH_i \).

Due to its dependence upon \( \alpha \) and \( \beta \), \( \left( \frac{\partial Q}{\partial H} \right)_i \{h, \alpha, \beta\} \cdot dH_i \) is also dependent upon \( dq_i \).

However, each \( \left( \frac{\partial Q}{\partial H} \right)_i \{h, \alpha, \beta\} \cdot dH_i \) can be approximated by \( \left( \frac{\partial Q}{\partial H} \right)_i \{h\} \cdot dH_i \), in which \( \alpha \) and \( \beta \) are assumed invariant. The deviation \( \left( \frac{\partial Q}{\partial H} \right)_i \{h\} \cdot dH_i \) is independent of \( dq_i \), and can be substituted into the expression for \( V_h \). In this case, it can then be assumed that

\[ V_q \{h\} = V_h \{h\} + V_e \{h\} \]  

(A5)

which is valid for all distributions of the values concerned.

From Equation A3,

\[ V_h \{h\} = \frac{1}{N-1} \sum_{i=1}^{N} \left( \frac{\partial Q}{\partial H} \right)_i \{h, \alpha, \beta\} \cdot (dH)_i^2 \]

\[ = \frac{1}{N-1} \sum_{i=1}^{N} \left( \alpha^2 \cdot \beta^2 \cdot h^{2-2} \right)_i \cdot (dH)_i^2 \]

Since the digitising errors in stage, \( dH_i \) are independent of \( \alpha \), \( \beta \) and \( h_i \), then

\[ V_h = (\alpha \cdot \beta \cdot h^{\beta-1}) \cdot \frac{1}{N-1} \sum_{i=1}^{N} (dH)_i^2 \equiv (\alpha \cdot \beta \cdot h^{\beta-1})^2 \cdot W_h \]  

(A6)

where

\[ W_h \] is the variance of the errors in digitising stage \( (m^2) \).

This equation was used to determine the variance in estimated flow due to digitisation errors, \( V_h \), from \( W_h \).
The relation between \( Q \) and \( h \) was determined by regression of \( \ln(Q) \) on \( \ln(h) \), for the gauging stations at sites C6, C5 and C1. As recommended by Herschy (1971), the offset between the downstream control level and the zero level of the measuring staff was determined by choosing the value which gave the straightest line in a graph of \( \ln(Q) \) against \( \ln(h) \). The offset and the values of \( \alpha \) and \( \beta \) were found to be as follows.

At C6: control level = staff zero + 0.00 m.

For all \( h \), \( \alpha = 3.766, \beta = 1.433 \).

At C5: control level = staff zero - 0.16 m.

The flow at this location was controlled by gabions installed on the channel bed after installation of the gauging station. Since the gabions were not impervious, some flow remained possible when \( h \) fell below the dominant control level.

For \(-0.10 < h < 0.02\), a linear relation was assumed: \( Q = 0.0061\cdot(h + 0.10) \).

For \( h < 0.02\), \( \alpha = 5.668, \beta = 1.700 \).

At C1: control level = staff zero + 0.16 m.

The channel at this location had a compound cross section with an undercut hollow along the base of one bank. This resulted in an irregularly shaped stage-discharge curve.

For \( 0.00 < h < 0.36 \), \( \alpha = 21.32, \beta = 4.950 \).

For \( 0.36 < h < 0.43 \), a linear relation was assumed: \( Q = -0.206 + 0.951\cdot h \).

For \( h > 0.43 \), \( \alpha = 5.534, \beta = 3.921 \).

Figures A1 - A3 show the resulting graphs of discharge against stage for the three gauging stations.

The stage charts, reading time in hours along the x-axis and staff level in metres along the y-axis, were digitised to obtain a record of hourly flows for all stations. The factor by which to scale y-axis distance in metres to obtain staff level in metres was 10.00 at C6, 5.00 at C5 and 5.00 at C1.

A two-tailed 90% confidence interval of ±1 mm for the errors in the digitised stage chart.
ordinates was assumed. Assuming normally distributed digitising errors, this was converted to an expected stage variance, \( W_H \), and from this into \( V_H \) for each station, as follows:

At C6: \( W_H = (10 \times 0.001/Z_{0.05})^2 = 3.70 \times 10^{-5} \text{ m}^2 \).

For typical values of \( h \) (0.2 < \( h \) < 0.6 m, giving 0.4 < \( Q \) < 2.0 m\(^3\)s\(^{-1}\)), the resulting standard deviation, \( V_H^{h'} \), is between 0.016 and 0.039 m\(^3\)s\(^{-1}\).

At C5: \( W_H = (5 \times 0.001/Z_{0.05})^2 = 9.24 \times 10^{-6} \text{ m}^2 \).

For typical values of \( h \) (0.04 < \( h \) < 0.2 m, giving 0.02 < \( Q \) < 0.37 m\(^3\)s\(^{-1}\)), the resulting standard deviation, \( V_H^{h'} \), is between 0.0031 and 0.0095 m\(^3\)s\(^{-1}\).

At C1: \( W_H = (5 \times 0.001/Z_{0.05})^2 = 9.24 \times 10^{-6} \text{ m}^2 \).

For typical values of \( h \) (0.3 < \( h \) < 0.6 m, giving 0.05 < \( Q \) < 0.75 m\(^3\)s\(^{-1}\)), the resulting standard deviation, \( V_H^{h'} \), is between 0.0028 and 0.015 m\(^3\)s\(^{-1}\).

Having evaluated \( V_H \) as above, it is appropriate to evaluate \( V_q \) by analysis of the uncertainty in the regression line for ln(\( Q_i \)) versus ln(\( h \)). Firstly, a confidence interval for the prediction of ln(\( Q_i \)) is calculated assuming that the stream flows and their errors were distributed lognormally.

Since the exceedance probabilities in a statistical distribution are preserved upon taking logarithms, it is then simple to translate this into the corresponding confidence interval for prediction of \( Q_i \). The variance of a statistical distribution is not preserved upon taking logarithms, and so \( V_q \) is not normally calculable. Thus, \( V_q \) and \( V_H \) cannot be added to obtain \( V_Q \).

However, the calculation of a similar confidence interval from \( V_H \) allows comparison of the effects of digitising errors and rating curve errors on the uncertainty in predicting \( Q_i \). Statistical principles are given by Wackerly et al. (1996).

In the present study, such analysis was impossible as a result of the minimal number of measurements available for statistical inference.
APPENDIX B

FORTRAN 77 PROGRAM TO CALCULATE EVAPOTRANSPIRATION USING THE PENMAN-MONTEITH EQUATION

Program to calculate daily evapotranspiration from single-level measurements of net radiation, temperature, vapour pressure and wind speed, using the Penman-Monteith equation. For periods when no humidity measurements are available, the actual vapour pressure is estimated using a regressed linear relation with the saturated vapour pressure. When net radiation is unavailable, it is calculated from daily hours of sunshine, using empirical equations determined by Kasten et al. (1982), and Monteith and Unsworth (1990).

sun_hrs : number of hours of sunshine in the day, in hours
tanchi : tan of the angle of inclination of Earth's axis away from the orbital normal (dimensionless)
phi : pi/2 - latitude, in radians
psi : orbital angle from summer solstice, in radians
beta : angle of inclination of Earth's axis towards the Sun (dimensionless)
theta : angle of rotation from solar noon, in radians
thetaup : theta at sunrise
thetadn : theta at sunset
sumcosxi : sum of cosines of angles xi for certain values of theta, where xi is the angle of the Sun from zenith (dimensionless)
acoscosxi : average calculated from sumcosxi (dimensionless)
global : global radiation flux at ground level, in W/m2
reflect : radiation flux reflected back from ground, in W/m2
netrad : net radiation flux in W/m2
tdry : temperature of dry thermometer in Kelvin
twet : temperature of wet thermometer in Kelvin
winspeed : wind speed in m/s
rain : rainfall in m
svp : saturation vapour pressure at temp dry, in Pa
delta : slope of sat. vap. press. w.r.t. temp, in Pa/K
avp : actual, ambient vap. press., in Pa
d : zero level displacement, in m
z : roughness length, in m
z : level of the measurements, in m
lai : leaf area index (dimensionless)
bai : bark area index (dimensionless)
alpha : exponential index in radiation extinction (dimensionless)
pi : geometrical constant (1/1) (dimensionless)
rlw : stomatal resistance of a single willow leaf, in s/m
rlhp : leaf stomatal resistance for heath or pasture, in s/m
rst : bulk stomatal resistance ( = rlw/lai ), in s/m
rav : bulk aerodynamic resistance, in s/m
denom : 1 + rst / rav
rho : density of air, in kg/m3
rhowat : density of water, in kg/m3
specht : specific heat capacity of air, in J/kg K
latht : latent heat of vaporisation of water, in J/kg
energy : delta * netrad / latht
energy1 : energy * exp( -alpha * ( lai + bai ) / pi )
gamma : psychrometric constant, = specht * pressure / 0.622*latht
vapdef : ( rho * specht / latht )*( satvp - actvp )
numerat : vapdef / rav
evap : evaporative flux from waterlogged ground beneath veg., in m/day
evap2 : evapotranspirative flux from veg., or evaporative flux from open water, in m/day
etot : evapl + evap2 ( note that evapl = 0 if open water ), in m/day

ARRAYS ARE DIMENSIONED FOR UP TO 32 MONTHS OF DAILY DATA
program pennen5

implicit none
double precision netrad(1000), tdry(1000), twet(1000),
& sun_hrs(1000),
& winspeed(1000), rain(1000), svp, delta, avp,
& energy, d, z0, z, rst, rav, denom, rho, specht,
& latht, gamma, vapdef, numerat, evapi, evap2,
& etot, open, en, del, num, den, rlx, rhp,
& la(4,100), bai(4,100), alpha, pi, energy1,
& tanchi, phi, psi, beta, theta(37), chetaup,
& thetaad, suncoxi, avcoxi, global,
& estrnrad(1000), reflect, rhomat, rootc4(100),
& wilpt4(100), psmd, smd, rech, aevt, f
integer dfin(500), dinit(500), i, j, nper, type,
& cday(100), fday(100), sect(500), nsect, k,
& sect(1000), day(1000)
data netrad, tdry, twet, sun_hrs, winspeed, rain,
& energy, d, z0, z, rst, rav, denom, vapdef,
& numerat, evapi, evap2, etot, en, del, num, den,
& lai, bai, energy1, psi, beta, theta, thetaup,
& thetaad, suncoxi, avcoxi, global, estrnrad,
& reflect, rootc4, wilpt4, psmd, smd, rech, aevt
& /8066*0.00d+00/
data dfin, dinit, i, j, nper, type, sday, fday, sect,
& sect, k, act, day /3706*0/
parameter ( rho=1.2923d+00, specht=1.005d+03,
& latht=2.466d+06, gamma=6.639d+01, rlw=3.64d+01,
& rlhp=1.30d+02, alpha=5.00d-01, f=1.00d-01,
& pi=3.141592654d+00, rhowat=9.990d+02,
& tanchi=4.37757116d-01, phi=6.91354005d-01 )
common /blockl/ svp, delta, avp
epen(en, del, num, den) = ( en + num )/( del + gamma*den )
ope n(10, file='format.dat', status='old')
ope n(11, file='no_synth.dat', status='old')
ope n(12, file='temp_req.dat', status='old')
ope n(13, file='allsynth.dat', status='old')
ope n(14, file='indices.dat', status='old')
ope n(15, file='opnwater.dat', status='old')
ope n(16, file='willow.dat', status='old')
ope n(17, file='wetheath.dat', status='old')
ope n(18, file='pasture.dat', status='old')
do 30 i=15,18
do 10 j=1,1000
read (i,*,end=20)
10 continue
20 backspace i
write (i,*)
30 continue
nsect = 0
read (10,*)
read (10,*)
do 40 k=1,500
read (10,999,end=111) dinit(k), dfin(k), sect(k)
nsect = nsect + 1
40 continue
111 close (10)
write (*,*) • nsect = ' , nsect
write (*,*)
read (11,*)
read (12,*)
read (13,*)
do 300 k=1,nsect
do 290 j=dinit(k), dfin(k)
act(j) = sect(k)
if ( sect(k) .eq. 1 ) then
read (11,998) day(i), netrad(i), tdry(i), twet(i),
& winspeed(i), rain(i)
elseif ( sect(k) .eq. 2 ) then
read (12,997) day(i), netrad(i), tdry(i), winspeed(i),
& rain(i)
else
read (13,996) day(i), sun_hrs(i), tdry(i), winspeed(i),
& rain(i)
endif
if ( day(i) .ne. i ) then
write (*,*) • ' Section ', k, ' does not tally.'
write (*,*) • i = ', i, ' ; day(i) = ', day(i)
stopt238
```plaintext
theta(j) = -2.35619449d+00 + 1.30899693d-01*(j-1)

psi = (i-172)*2.00d+00*pi/3.65d+02
beta = atan(tanchi*cos(psi))
thetaup = -(pi - acos(tan(beta)/tan(phi)))

theta(j) = theta(j) - thetaup + sumcosxi

Reflect = (1.00d+00 - 7.50d-01)*global
estntrad(i) = global - Reflect

netrad(i) = 9.27248d-01*estntrad(i) - 7.5533d-01
```

endif
write (type+14,*)
240 continue

write (*,*) • Calculating.....'
write (*,*)
do 888 type=1,4
write (*,*) ' land type =', type
write (*,*)
do 777 j=1,nper
do 666 i=sday(j),fday(j)

if ( sct(i) .eq. 1 ) then
call vpress1(tdry(i),twet(i))
else.
call vpress2(tdry(i))
endif
energy = delta*netrad(i)/latht
energyl = 0.00d+00
if ( type .eq. 1 ) then
C
NO WATERLOGGED GROUND BENEATH OPEN WATER! HENCE EVAP = 0.
evapl = 0.00d+00
C PENMAN PARAMETERS FOR OPEN WATER:
rst = 0.00d+00
d = 0.00d+00
z0 = 1.37d-03
z = 2.00d+00
else
if ( type .eq. 4 ) then
C
NO WATERLOGGED GROUND IN PASTURE. HENCE EVAP = 0.
evapl = 0.00d+00
C PARAMETERS FOR PASTURE:
rst = rlhp/lai(type,j)
d = 5.00d-02
z0 = 5.00d-03
z = 2.00d+00
else
C PRIESTLEY - TAYLOR (1972) FORMULA FOR EVAPORATION WITH
C MINIMAL ADVECTION
denom = 1.00d+00
numerat = 0.00d+00
energyl = energy * exp( (-1.00d+00)*alpha*
& ( lai(type,j) + bai(type,j) )/pi )
& / rhowat )*0.64d+04
C NOW THE WIND-AFFECTED CANOPY EVAPORATION IS CALCULATED:
if ( type .eq. 3 ) then
C PARAMETERS FOR WET HEATH:
rst = rlhp/lai(type,j)
d = 2.50d-01
z0 = 2.50d-02
z = 2.00d+00
else
C PARAMETERS FOR WILLOW SCRUB:
rst = rlp/lai(type,j)
d = 2.00d+00
z0 = 2.00d-01
z = 5.50d+00
endif
if ( rain(i) .gt. 0.00d+00 ) then
C EVAPORATION IS FROM INTERCEPTED WATER ON LEAF SURFACES
rst = 0.00d+00
endif
endif
C AERODYNAMIC RESISTANCE INCORPORATING THOM + OLIVER (1977)
C CORRECTION
if ( z-d .le. z0 ) then
write (*,*) ' Measurement level (z) too low.'
stop
endif
rav = \( \frac{4.72d+00 \times \log((z-d)/z0)}{1.00d+00 + \text{rast/rav}} \)

denora = 1.00d+00 + \text{rast/rav}

vapdef = \rho \times \text{specht} \times (\text{svp-avp})/\text{latht}

numerat = vapdef/rav

gn = energy - energy1

evap2 = (open(\( \text{energy, delta, numerat, denora } \)/rhowat)*8.64d+04

etot = evapl + evap2

**ACTUAL ET CALCULATION USING PENMAN-GRINDLEY**

**SOIL MOISTURE ACCOUNTING MODEL (LERNER et al., 1990)**

if ( type .eq. 4 ) then
  if ( smd.lt.rootc4(j) .or. rain(i).ge.etot ) then
    aevt = etot
  elseif ( smd .lt. wilpt4(j) ) then
    aevt = rain(i) + f*(etot-rain(i))
  else
    aevt = rain(i)
  endif
  psmd = smd + aevt - rain(i)
  rech = -1.00d+00*min(psmd,0.00d+00)
  write (type+14,992) i, rav, svp, avp, evapl, evap2, aevt, rain(i), smd, rech
  & psmd, xmd, rech
else
  aevt = etot
  write (type+14,991) i, rav, svp, avp, evapl, evap2, aevt
endif

666 continue
777 continue
888 continue
do 900 type=1,4
  close (type+14)
900 continue
write (*,*)
write (*,*) ' Output files written'
write (*,*) ' End of run.'

999 format ( t6,i6, t16,i6, t29,i6 )
998 format ( t6,i6, t14,f7.2, t24,f7.2, t34,f7.2, t45,f6.2, t53,f8.5 )
997 format ( t6,i6, t14,f7.2, t24,f7.2, t34,f7.2, t45,f6.2, t53,f8.5 )
996 format ( t6,i6, t14,f7.2, t24,f7.2, t34,f7.2, t45,f6.2, t53,f8.5 )
995 format ( 215, 8f5.2, 2f6.3 )
994 format ( t2,a9, t12,a9, t22,a9, t32,a9, t42,a9, t52,a9, 
      t62,a9, t72,a9, t82,a9, t92,a9 )
993 format ( t2,a9, t12,a9, t22,a9, t32,a9, t42,a9, t52,a9, 
      t62,a9, t72,a9, t82,a9, t92,a9 )
992 format ( t6,i6, t14,f7.2, t22,f9.3, t32,f9.3, t43,f8.5, t53,f8.5, 
      t63,f8.5, t75,f6.3, t85,f6.3, t95,f6.3 )
991 format ( t6,i6, t14,f7.2, t22,f9.3, t32,f9.3, t43,f8.5, t53,f8.5, 
      t63,f8.5 )

stop
end

**SUBROUTINE VPRESS1 CALCULATES SVP, DELTA AND AVP USING**

**RICHARDS (1971) FORMULA**

**SUBROUTINE VPRESS1**

double precision gamma, dtemp, wtemp

double precision svp, delta, avp

gamma = 6.639d+01

if ( dtemp .le. 0.00d-00 ) then
  write (*,*) ' dtemp =', dtemp
endif
if ( dtemp .eq. 0.00d+00 ) then
  stop
endif

treduc = 1.00d+00 - ( 3.7315d+02 / dtemp )

svp = 1.01325d+00 * 
& \exp(1.33185d+01*treduc - 1.9760d+00*(treduc**2) 
& - 6.445d-01*(treduc**3) 
& - 1.299d-01*(treduc**4) 
& delta = 3.7315d+02 *
SUBROUTINE VPRESS2 CALCULATES SVP, DELTA AND AVP USING RICHARDS (1971) FORMULA

subroutine vpress2( dtemp )
double precision gamma, dtemp, svp, delta, avp
parameter ( gamma=6.639d+01 )
common /blockl/ svp, delta, avp
if ( dtemp .le. 0.00d+00 ) then
  write (*,*) • dtemp = ', dtemp
if ( dtemp .eq. 0.00d+00 ) then
  stop
  endif
endif
treduc = 1.00d+00 - ( 3.7315d+02 / dtemp )
svp = 1.01325d+05 * 
  exp( 1.33185d+01*treduc - 1.9760d+00*(treduc**2) 
  & - 6.445d-01*(treduc**3) 
  & - 1.299d-01*(treduc**4) )
delta = 3.7315d+02 * 
  svp * ( 1.33185d+01 - 1.952d+00*treduc 
  & - 1.9335d+00*(treduc**2) 
  & - 5.196d-01*(treduc**3) ) 
  / (dtemp**2)
avp = -5.36028d+01 + 1.017814d+00*svp
return
end
APPENDIX C

MODIFICATIONS TO THE U.S.G.S. GROUNDWATER FLOW SIMULATION PROGRAM, MODFLOW, FOR PURPOSES OF STEADY STATE CALIBRATION AND NEAR-STREAM FLOW PATH PARAMETERISATION

C.1 THE CENTRAL PROGRAM

**********
 MAIN CODE FOR MODULAR MODEL -- 9/1/87
 BY MICHAEL G. MCDONALD AND ARLEN W. HARBAGH

VERSION 1033 15JUNE1994 -- added TLK1 as documented in USGS report
VERSION 1247 05JUNE1995 -- added DEAS as documented in USGS report
VERSION 0917 20SEP1996 -- added REAS as documented in USGS report
VERSION 1026 18NOV1994 -- added file opening routine for use on
personal computers

Calibration of steady state, uniform-K model for Goss Moor, C.Ishemo 1999
All code inserted by C.Ishemo is in lower case.

aquc ; conductance of aquifer zone between model node and bottom surface of
stream/pool bed layer, in m2/day
aquidirk : conductivity of aquifer zone between model node and bottom surface of
stream bed layer, averaged over all directions between horizontal and
vertical, in m/day
aquihrzk : conductivity of aquifer along the horizontal principal axis, in m/day
aquikrev : revised conductivity of aquifer in stream-bearing model cell, in
m/day
aquivrzk : conductivity of aquifer along the vertical principal axis, in m/day
bankfall : distance of stream bottom below ground level, in m
bedc : conductance of stream/pool bed layer, in m2/day
bedthick : thickness of stream/pool bed layer, in m
cellwid : width of model cell, in m
interim : an intermediate variable in the calculations
log2ak : logarithm (base 2) of the aquifer’s horizontal conductivity
(log2rk : logarithm (base 2) of the conductivity of the stream bed sediments
m_col(i) : MODFLOW column index of stream reach i
m_row(i) : MODFLOW row index of stream reach i
nreaches : total number of stream reaches and pool sections in the model
outc : conductance of average flowpath from outer parts of the model cell to
the lower surface of the stream/pool bed layer, in m2/day
radeff : effective inner radius of stream, in m
radielc : effective conductance between model node and stream/pool water, in
m2/day
reachlen : length of stream reach or pool section within the model cell, in m
riverk : conductivity of stream bed sediments, in m/day
thick_mx : distance of aquifer base below ground level, in m
totc : final value of conductance to be used in RIV1 or GHBL module, in
m2/day
wdepth : depth of water in stream/pool, in m
width : width of water surface in stream/pool, in m

Also added allocatable X array, and DE4 solver.

SPECIFICATIONS:

COMMON X(1500000)

block data
real aquihrzk, aquikrev(200), totc(200), log2ak, log2rk
integer m_col(200), m_row(200), nreaches
common /goss0/ m_col, m_row, nreaches
common /goss1/ aquihrzk, aquikrev
common /goss2/ totc
common /goss3/ log2ak, log2rk, riverk
data aquihrzk, aquikrev, riverk, totc, log2ak, log2rk
& /404*0.00/
data m_col, m_row, nreaches /401*0/

DIMENSION X(:)
ALLOCATABLE :: X
COMMON /FLWCOM/LAYCON(80)
COMMON /FLWAVG/LAYAVG(80)
common /goss0/ m_col, m_row, nreaches
common /gossl/ aquihrzk, aquikrev
common /goss2/ tote
common /goss3/ log2ak, log2rk, riverk
CHARACTER*4 HEADNG, VBNM
CHARACTER*4 HEADNG(32), VBNM(4,20), VBVL(4,20), lUNIT(24)
CHARACTER*80 FNAME
CHARACTER*4 CUNIT(24)
character*2 labels(2), aquifer, river

DATA CUNIT/'BCF ', 'WEL ', 'DRN ', 'RIV ', 'EVY ', 'SLX ', 'GHB ',
'HFB ', 'RES ', 'FRN ', 'SIP ', 'CHD ', '0C ', 'PCG ', 'GFD ',
'HFB ', 'RES ', 'STR ', 'ESS ', 'CD ', 'RO ', 'RCH ', 'SIP ', 'DRN ',
'DRG ', 'BCF ', 'RIV ', 'EVY ' /

set string for use if RCS ident command

versn = '$Id: modflow.f,v 2.6 1996/09/20 15:37:18 rsregan Exp $'
versn = '(K^)MODFLOW - Modular 3-D Finite-Difference GW Flow Model'
versn = '@(#)MODFLOW - USGS TWRI, Book 6, Chapter Al, McDonald and Harbaugh'
versn = '@(#)MODFLOW - Version: 2.5 1995/06/23 new DE45 module'
versn = '@(#)MODFLOW - Version: 2.6x 1996/09/20 new RES module'

LENX=15000000

INBASE=5
INUNIT=99

open (67,file='rivparam.esv',status='old')
read (67,*) nreaches
read (67,*)
read (67,*) ( m_col(i), m_row(i), reachlen(i), bankfall(i),
& wdepth(i), width(i), bedthick(i), thick_mx(i),
& i=1,nreaches )
close (67)

for do 1050 riverk=1,14

log2ak = (real(aquifer) - 6.00)/1.00
log2rk = (real(river) - 7.00)*2.00
aquihrzk = 2.00**log2ak
riverk = 2.00**log2rk
aquivrtk = aquihrzk/10.00

aquidirk = 0.00

do 1001 i=0,90
aquidirk = aquidirk +
& sqrt( (aquihrzk*cos(3.141592654*real(i)/180.00))**2.00 +
& (aquivrtk*sin(3.141592654*real(i)/180.00))**2.00 )
1001 continue

aquidirk = aquidirk/91.00

do 1002 i=l,nreaches
rad eff(i) = ( wwidth(i) + 2.00*wdepth(i) )/3.141592654
if ( wwidth(i) .lt. cellwid/4.00 ) then
    bedc(i) = 3.141592654*reachlen(i)*riverrk/
& log ( (r adef f(i)+bedthick(i))/r adef f(i) )
else
    aquc(i) = aquivr tk*w width(i)*reach len(i)/
& max(real(int(min(0.5*cellwid/wwidth(i),1.90))),0.000001) )
endif
1002 continue

OPEN(DNIT=INUNIT,FILE='2d03.nam',STATUS='OLD')
C
C3 DEFINE PROBLEM.Rows, Columns, Layers, Stress Periods, Packages
CALL BAS2DF(ISUM,HEADING,NPER,ITMUNI,TOTIM,NCOL,NROW,NLAY,1
& NODES,INBAS,ICUN,TUNIT,TUNIT,CUNIT)
C
C4— ALLOCATE SPACE IN "X" ARRAY.
CALL BASIAL(ISUM,LENX,LCHNEW,LCHOLD,LCIBOU,LCCR,LCCV,LCTRT,LCRHS,LCOFL,
& INBAS,ISTRT,NCOL,NROW,NLAY,LUNIT,LUNIT)
IF (IUNIT(1) .GT. 0) CALL BCF3AL(ISUM,LENX,LCSC1,LCY,1
& LCBOT.LCCVR,LCVTR,UNIT1(1),LES,2
& NCOL,NROW,LAY,UNIT,ICBCB,LCVTRD,INDFLL,LCVTRD,3
& WETFCT,INWET,INWET,HDRY,LCCR,LCHIER)
IF (IUNIT(4) .GT. 0) CALL BCF3AL(ISUM,LENX,LCRHS,LCHIER,1
& LCRAF,LCCVR,LCVTR,LCCR,LCCR,HDRY,LCRHS,LCRHS,2
& LCCVR,LCLR,LCLL,LCLR,LCLL,LCSL,NORM1,NM1,NM2,NM3,3
& NWM1,NWM2,NWM3,IFISAC,ISAC,UNIT4(4),IUNIT)
IF (IUNIT(2) .GT. 0) CALL BCF3AL(ISUM,LENX,LCVTR,LCRHS,1
& LCRAF,LCCVR,LCVTR,LCCR,LCCR,HDRY,LCRHS,LCRHS,2
& LCCVR,LCLR,LCLL,LCLR,LCLL,LCSL,NORM1,NM1,NM2,3
& NWM1,NWM2,NWM3,IFISAC,ISAC,UNIT4(4),IUNIT)
IF (IUNIT(3) .GT. 0) CALL DRAL(ISUM,LENX,LCDRAI,1
& HDMRAIN,HDRAIN,UNIT3(3),IUNIT,TDRN)
IF (IUNIT(5) .GT. 0) CALL RCHIAL(ISUM,LENX,LCRMS,1
& LCRMS,LCRMS,UNIT5(5),ICUNIT,ICUNIT)
C
1002 continue

OPEN(UNIT=INUNIT,FILE='2d03.nam',STATUS='OLD')
C
C3------DEFINE PROBLEM.Rows, Columns, Layers, Stress Periods, Packages
CALL BAS2DF(ISUM,HEADING,NPER,ITMUNI,TOTIM,NCOL,NROW,NLAY,1
& NODES,INBAS,ICUN,TUNIT,TUNIT,CUNIT)
C
C4—— ALLOCATE SPACE IN "X" ARRAY.
CALL BASIAL(ISUM,LENX,LCHNEW,LCHOLD,LCIBOU,LCCR,LCCV,LCTRT,LCRHS,LCOFL,
& INBAS,ISTRT,NCOL,NROW,NLAY,LUNIT,LUNIT)
IF (IUNIT(1) .GT. 0) CALL BCF3AL(ISUM,LENX,LCSC1,LCY,1
& LCBOT.LCCVR,LCVTR,UNIT1(1),LES,2
& NCOL,NROW,LAY,UNIT,ICBCB,LCVTRD,INDFLL,LCVTRD,3
& WETFCT,INWET,INWET,HDRY,LCCR,LCHIER)
IF (IUNIT(4) .GT. 0) CALL BCF3AL(ISUM,LENX,LCRHS,LCHIER,1
& LCRAF,LCCVR,LCVTR,LCCR,LCCR,HDRY,LCRHS,LCRHS,2
& LCCVR,LCLR,LCLL,LCLR,LCLL,LCSL,NORM1,NM1,NM2,NM3,3
& NWM1,NWM2,NWM3,IFISAC,ISAC,UNIT4(4),IUNIT)
IF (IUNIT(2) .GT. 0) CALL BCF3AL(ISUM,LENX,LCVTR,LCRHS,1
& LCRAF,LCCVR,LCVTR,LCCR,LCCR,HDRY,LCRHS,LCRHS,2
& LCCVR,LCLR,LCLL,LCLR,LCLL,LCSL,NORM1,NM1,NM2,3
& NWM1,NWM2,NWM3,IFISAC,ISAC,UNIT4(4),IUNIT)
IF (IUNIT(3) .GT. 0) CALL DRAL(ISUM,LENX,LCDRAI,1
& HDMRAIN,HDRAIN,UNIT3(3),IUNIT,TDRN)
IF (IUNIT(5) .GT. 0) CALL RCHIAL(ISUM,LENX,LCRMS,1
& LCRMS,LCRMS,UNIT5(5),ICUNIT,ICUNIT)
C5-------IF THE "X" ARRAY IS NOT BIG ENOUGH THEN STOP. 
C6-------READ AND PREPARE INFORMATION FOR ENTIRE SIMULATION. 
C7-------SIMULATE EACH STRESS PERIOD.
IF (IUNIT(17).GT.0) CALL RESIRP(X(LCIRES),X(LCIRSL),X(LCBRES),
1 X(LCCRES),X(LCBBRE),X(LCHRSE),X(LCIBOU),X(LCDELR),X(LCDELC),
2 NRES, NRESOP, NPTS, NCOL, NROW, NLAY, PERLEN, DELT, NSTP, TSMULT,
3 IUNIT(17), IOUT)
IF (IUNIT(18).GT.0) CALL STRIRP(X(LCSTRM),X(ICSTRM),NSTREM,
1 MXSTRM,IUNIT(18),IOUT,X(LCTBAR),NDIV,NSS,STR1
2 NTRIB,X(LCIVAR),ICALC,IPFLGS)
IF (IUNIT(20).GT.0) CALL CHDIRP(X(LCCHDS),NCHDS,MXCHD,X(LCIBOU),CHD
1 X(LCHNEW),X(LCHOLD),PERLEN,PERTIM,DELT,NSTP,TSMULT,IUNIT(20),IOUT)
C
C7C SIMULATE EACH TIME STEP.
DO 200 KSTP=1,NSTP

C7C1 CALCULATE TIME STEP LENGTH. SET HOLD=HNEW.
CALL BAS1AD(DELT,TSMULT,TOTIM,PERTIM,X(LCHNEW),X(LCHOLD),KSTP,
1 NCOL, NROW, NLAY)
IF(IUNIT(6).GT.0) CALL TLK1AD(X(LCRAT),X(LCZCB),X(LCALPH),X(LCBET),
1 X(LCRM1),X(LCRM2),X(LCRM3),X(LCRM4),X(LCTL),X(LCTLK),X(LCSLU),
2 X(LCSLD),NM1,NM2,NUMC,NTM1,DELTM1,X(LCHNEW),X(LCTOP),
3 NROW,NCOL,NLAY)
IF(IUNIT(17).GT.0) CALL RES1AD(X(LCHRES),X(LCHRSE),X(LCIRES),
1 X(LCBRES),X(LCDELR),X(LCDELC),NRES,IRESPT,NCOL,NROW,
2 PERLEN,PERLEN,PERLEN,DELT,TLKTIM,UNIT(6),IOUT)
C
C7C2 ITERATIVELY FORMULATE AND SOLVE THE EQUATIONS.
DO 100 KITER=1,MXITER

C7C2A FORMULATE THE FINITE DIFFERENCE EQUATIONS.
CALL BAS1FM(X(LCHCOF),X(LCRHS),NODES)
IF(IUNIT(1).GT.0) CALL BCF3FM(X(LCHCOF),X(LCRHS),X(LCHNEW),
1 X(LCHRES),X(LCHRSE),X(LCIBOU),X(LCCR),X(LCCC),X(LCCV),
2 X(LCHY),X(LCTRPY),X(LCBOT),X(LCTOP),X(LCSC1),X(LCSC2),
3 X(LCCDTR),X(LCCDTC),X(LCDELR),X(LCDELC),DELT,ISS,KKITER,KKSTP,KSTP,
4 NCOL, NROW, NLAY, IOUT, X(LCCVWD), WETFCT, IWETIT, IHDFWT, HDRY,
5 X(LCHNEW),X(LCHCOF),X(LCRHS),KSTP)
IF(IUNIT(2).GT.0) CALL WELIFM(NWELLS,MXWELL,NCOL,NROW,NLAY)
IF(IUNIT(4).GT.0) CALL RIV1FM(NRIVER,MRIVER,X(LCRIVR),
1 X(LCHCOF),X(LCRHS),X(LCHNEW),X(LCHOLD),NCOL,NROW,NLAY)
IF(IUNIT(5).GT.0) CALL EVT1FM(NEVT0P,X(LCEVT),X(LCEVTR),
1 X(LCEXDP),X(LCSURF),X(LCRHS),X(LCHCOF),X(LCHNEW),
2 X(LCHOLD),NCOL,NROW,NLAY)
IF(IUNIT(6).GT.0) CALL HFB1FM(X(LCHNEW),X(LCCR),X(LCCC),
1 X(LCCD),X(LCHCHG),X(LCLHCH),X(LCRCHG),X(LCLRCH),KSTP,
2 NITER,RCLOSE,ICNVG,KKSTP,KKPER,IPCALC,IPROSIP,MXITER,NSTP,
3 NCOL,NROW,NLAY)
C MODIFICATIONS FROM HILL(1990):
C 01JUL90 OMITTED TWO OCCURRENCES OF ICD=0
C 01SEPT1990 OMITTED IPCGCD, STEPL, IUNIT(5), AND IP
C 01SEPT1991 ADDED NNP,IP,SN,SP,SR
IF(IUNIT(13).GT.0) CALL PCG2AP(X(LCHNEW),X(LCHOLD),X(LCCR),
1 X(LCCC),X(LCHCHG),X(LCLHCH),X(LCRCHG),X(LCLRCH),KSTP,
2 NITER,RCLOSE,ICNVG,KKSTP,KKPER,IPCALC,IPROSIP,MXITER,NSTP,
3 NCOL,NROW,NLAY)
C7C2B MAKE ONE CUT AT AN APPROXIMATE SOLUTION.
IF (IUNIT(9).GT.0) CALL SLP1AP(X(LCHNEW),X(LCHOLD),
1 X(LCCR),X(LLLL),X(LCCD),X(LCHCHG),X(LCLHCH),X(LCRCHG),X(LCLRCH),KSTP,
2 NITER,RCLOSE,ICNVG,KKSTP,KKPER,IPCALC,IPROSIP,MXITER,NSTP,
3 NCOL,NROW,NLAY)
C
247
C

C7CA---IF CONVERGENCE CRITERION HAS BEEN MET STOP ITERATING.
IF (ICNVG.EQ.1) GO TO 110
100 CONTINUE
KSTP=KXITER
110 CONTINUE
C

C7C3---DETERMINE WHICH OUTPUT IS NEEDED.
CALL BASIOC (NSTP, KKSTP, ICNVG, X(LCIOFL), NLAY, 1, IBUDEL, ICBCFL, IHDELFL, IUNIT(12), IOUT)
C

C7C4---CALCULATE BUDGET TERMS. SAVE CELL-BY-CELL FLOW TERMS.
MSUM=1
IF (IUNIT(6).GT.0) CALL TLKIBD(X(LCRM1), X(LCRM2), X(LCRM3), X(LCRM4), NM1, NM2, ITLKSV, DELTM1, TLKTIM, IOUT)
248
C

C7C6---IF ITERATION FAILED TO CONVERGE THEN STOP.
IF (ICNVG.EQ.0) STOP
200 CONTINUE
300 CONTINUE
C

C7CA---PRINT AND OR SAVE SUBSIDENCE, COMPAC TION. AND CRITICAL HEAD.
IF (UNIT(19).GT.0) CALL IBSIOT(NCOL, NROW, NLAY, PERTIM, TOTIM, KSTP, KPER, NRES, NRESOP, 3, NCOL, NROW, NLAY, DELT, IRSUB, ICBCFL, JOUT)
C

C7C7---WRITE RESTART RECORDS
C7C7A---WRITE RESTART RECORDS FOR TRANSIENT-LEAKAGE PACKAGE
IF (IUNIT(6).GT.0) CALL TLKIOT(X(LCRM1), X(LCRM2), X(LCRM3), X(LCRM4), NM1, NM2, ITLKSV, DELTM1, TLKTIM, IOUT)
1
write (*, '(al3,i2.2,al,i2.2,a6)')
& ' Simulation ('.aquifer', ',', ' ,river', ',') done'
dallocate (x,statsierr)
if(ierr.ne.0) then
   write(iout,*) ' Unable to DEALlocate the x array'
stop
end if

close (inbas)
close (iout)
do 1040 igoss=l,24
   if ( iunit(igoss) .ne. 0 ) then
      close (iunit(igoss))
   endif
1040 continue
1050 continue
1060 continue
C
C END PROGRAM
STOP
C
END

C.2 THE "READ AND PREPARE" SUBROUTINE OF THE BLOCK-CENTRED FLOW MODULE

SUBROUTINE BCF3RP(IBOUND, HNEW, SCI, HY, CR, CC, CV, DELR, DELC, BOT, TOP,
   1  SC2, TRFY, IN, ISS, NCOL, NROW, NLAY, NODES, IOUT, WETDRY, IWDFLG, CVWD)
C
C VERSION 1276 9JULY1992 BCF3RP
Q ******************************************************************
C READ AND INITIALIZE DATA FOR BLOCK-CENTERED FLOW PACKAGE,
C VERSION 3
Q ******************************************************************
C
C !!!!!!I!!!!!!!!!!!!!!! 1 !!!!!!!!!!!!!!!!!!!!! I !!!!!!!!!!!!! !
C Calibration of uniform-K model for Goss Moor, C.Ishemo 1999
C All code inserted by C.Ishemo is in lower case.
C !! 1 !! 1 !!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!! !
C
C aquihrzk : conductivity of aquifer along the horizontal principal axis, in m/day
C aquikrev : revised conductivity of aquifer in stream-bearing model cell, in
C m/day
C m_col(i) : MODFLOW column index of stream reach i
C m_row(i) : MODFLOW row index of stream reach i
C nreaches : total number of stream reaches and pool sections in the model
C
C SPECIFICATIONS:

 CHARACTER*4 ANAME
 REAL aquihrzk, aquikrev(200)
 INTEGER igoss, jgoss, kgoss, m_col(200), m_row(200),
 & nreaches
 DIMENSION HNEW(NODES), SCI(NODES), HY(NODES), CR(NODES), CC(NODES),
 & CV(NODES), ANAME(1,3), DELR(NCOL), DELC(NROW), TOP(NODES),
 & IBOUND(NODES), WETDRY(NODES), CVWD(NODES)
 COMMON /FLWCOM/LAYCON(80)
 COMMON /gossO/ m_col, m_row, nreaches
 COMMON /gossl/ aquihrzk, aquikrev
 DATA igoss, jgoss, kgoss /3*0/
 DATA ANAME(1,1), ANAME(2,1), ANAME(3,1), ANAME(4,1), ANAME(5,1),
 & ANAME(6,1) / 'PRIM', 'ARY', 'STOR', 'AGE', 'COEF' /
 DATA ANAME(1,2), ANAME(2,2), ANAME(3,2), ANAME(4,2), ANAME(5,2),
 & ANAME(6,2) / 'TRAN', 'SHIS', 'AL', 'ONG', 'ROWS' /
 DATA ANAME(1,3), ANAME(2,3), ANAME(3,3), ANAME(4,3), ANAME(5,3),
 & ANAME(6,3) / 'H', 'YD', 'COND', 'AL', 'ONG', 'ROWS' /

249
DATA ANAME(1,4), ANAME(2,4), ANAME(3,4), ANAME(4,4), ANAME(5,4),
1  ANAME(6,4) /'VERT', 'HYD', 'CON', 'D/TH', 'HICKNESS' /
DATA ANAME(1,5), ANAME(2,5), ANAME(3,5), ANAME(4,5), ANAME(5,5),
1  ANAME(6,5) /'BO', 'TOP' /
DATA ANAME(1,6), ANAME(2,6), ANAME(3,6), ANAME(4,6), ANAME(5,6),
1  ANAME(6,6) /'BO', 'TOP' /
DATA ANAME(1,7), ANAME(2,7), ANAME(3,7), ANAME(4,7), ANAME(5,7),
1  ANAME(6,7) /'SE', 'COND', 'ARY', 'STOR', 'AGE', 'COEF' /
DATA ANAME(1,8), ANAME(2,8), ANAME(3,8), ANAME(4,8), ANAME(5,8),
1  ANAME(6,8) /'COL', 'MN T', 'G RO', 'W AN', '1SOT', 'ROPY' /
DATA ANAME(1,9), ANAME(2,9), ANAME(3,9), ANAME(4,9), ANAME(5,9),
1  ANAME(6,9) /' ' /
DATA ANAME(1,10), ANAME(2,10), ANAME(3,10), ANAME(4,10), ANAME(5,10),
1  ANAME(6,10) /'DEL' /
DATA ANAME(1,11), ANAME(2,11), ANAME(3,11), ANAME(4,11), ANAME(5,11),
1  ANAME(6,11) /' ' /

C1------CALCULATE NUMBER OF NODES IN A LAYER AND READ TRPY, DELR, DELC
NID=NCOL*NROW
C
CALL U1DREL(TRPY,ANAME(1-,8) ,NLAY, IN, IOUT)
CALL U1DREL(DELR,ANAME(1,9),NCOL,IN,IOUT)
CALL U1DREL (DELC, ANAME (1,10), NROW, IN, IOUT)

C2------READ ALL PARAMETERS FOR EACH LAYER.
K1=0
K2=0
DO 200 K=1,NLAY
  KK=K
C
C2A------FIND ADDRESS OF EACH LAYER IN THREE DIMENSION ARRAYS.
  IF (LAYCON(K) .EQ.1 .OR. LAYCON(K) .EQ.3) KB=KB+1
  IF (LAYCON(K) .EQ.2 .OR. LAYCON(K) .EQ.3) KT=KT+1
  LOC=1+(K-1)*NIJ
  L0CB=1+(KB-1)*NIJ
  L0CT=1+(KT-1)*NIJ
C
C2B------READ PRIMARY STORAGE COEFFICIENT INTO ARRAY SCI IF TRANSIENT
  IF (ISS.EQ.O) CALL U2DREL(SC1(LOC),ANAME(1,1),NROW,NCOL,KK,IN,IOUT)
C
C2C------READ TRANSMISSIVITY INTO ARRAY CC IF LAYER TYPE IS 0 OR 2
  IF (LAYCON(K) .EQ.3 .OR. LAYCON(K) .EQ.1) GO TO 100
  CALL U2DREL(CC(LOC) ,ANAME(1,2) ,NROW,NCOL,KK,IN,IOUT)
  GO TO 110
C
C2D------READ HYDRAULIC CONDUCTIVITY(HY) AND BOTTOM ELEVATION(BOT)
C2D IF LAYER TYPE IS 1 OR 3
  100 CALL U2DREL(HY(LOCB) ,ANAME(1,3) ,NROW,NCOL,KK,IN,IOUT)
  CALL U2DREL(BOT(LOCB) ,ANAME(1,5) ,NROW,NCOL,KK,IN,IOUT)

  DO 1017 IGROSS=1,NROW
    DO 1016 JGROSS=1,NCOL
      DO 1015 KGROSS=1,NCOL
        IF (M_ROW(KGROSS) .EQ.IGROSS.AND.IN_COL(KGROSS) .EQ.JGROSS) THEN
          HY(LOCB-1+(IGROSS-1)*NCOL+JGROSS) = AQUIKREV(KGROSS)
        ELSE
          HY(LOCB-1+(IGROSS-1)*NCOL+JGROSS) = AQUIHRZK
        ENDIF
      1015 CONTINUE
    1016 CONTINUE
  1017 CONTINUE
C
C2E------READ VERTICAL HYCOND/THICK INTO ARRAY CV IF NOT BOTTOM LAYER
C2E READ AS HYCOND/THICKNESS -- CONVERTED TO CONDUCTANCE LATER
  110 IF(K.EQ.NLAY) GO TO 120
  CALL U2DREL(CV(LOC) ,ANAME(1,4) ,NROW,NCOL,KK,IN,IOUT)
C
C2F------READ SECONDARY STORAGE COEFFICIENT INTO ARRAY SC2 IF TRANSIENT
C2F AND LAYER TYPE IS 2 OR 3
  120 IF (ISS.EQ.O) CALL U2DREL(SC2(LOC) ,ANAME(1,7) ,NROW,NCOL,KK,IN,IOUT)
C
C2G------READ TOP ELEVATION(TOP) IF LAYER TYPE IS 2 OR 3
C2G READ IN MILLITS
  CALL U2DREL(TOP(LOCB) ,ANAME(1,6) ,NROW,NCOL,KK,IN,IOUT)

  DO 1017 IGROSS=1,NROW
    DO 1016 JGROSS=1,NCOL
      DO 1015 KGROSS=1,NCOL
        IF (M_ROW(KGROSS) .EQ.IGROSS.AND.IN_COL(KGROSS) .EQ.JGROSS) THEN
          TOP(LOCB-1+(IGROSS-1)*NCOL+JGROSS) = AQUIHRZK
        ELSE
          TOP(LOCB-1+(IGROSS-1)*NCOL+JGROSS) = AQUIHRZK
        ENDIF
      1015 CONTINUE
    1016 CONTINUE
  1017 CONTINUE
C
C2H------READ WETDRY CODES IF LAYER TYPE IS 1 OR 3 AND WETTING
C2H CAPABILITY HAS BEEN INVOKED (IWDFLG NOT 0)
  130 IF (LAYCON(K) .NE.3 .AND. LAYCON(K) .NE.1) GO TO 200
  CALL U2DREL(WETDRY(LOCB) ,ANAME(1,11) ,NROW,NCOL,KK,IN,IOUT)
  CONTINUE
C
C3------PREPARE AND CHECK BCF DATA
  CALL SBCF3N(HNEW, IBOUND, SCI, SC2, CR, CC, CV, HY, TRPY, DELR, DELC, ISS, 1
  NCOL, NROW, NLAY, IOUT, WETDRY, IWDFLG, CWD)
C
C4------RETURN
RETURN
END
C.3 THE "READ AND PREPARE" SUBROUTINE OF THE RIVER MODULE

SUBROUTINE RIV1RP(RIVR, NRIVER, MXRIVR, IN, IOUT)
C
C-----VERSION 1319 25AUG1982 RIV1RP
C ***********************************************
C READ RIVER HEAD, CONDUCTANCE AND BOTTOM ELEVATION
C ***********************************************
C
C calibration of uniform-K model for Goss Moor, C.Ishemo 1999
C All code inserted by C.Ishemo is in lower case.
C
m_col(i) : MODFLOW column index of stream reach i
m_row(i)  : MODFLOW row index of stream reach i
nreaches : total number of stream reaches and pool sections in the model
totc : conductance assigned to stream/pool bed layer, accounting for
       flowpath of recharge within river-bearing model cell, in m2/day
C
C SPECIFICATIONS:
C
DIMENSION RIVR(6,MXRIVR)
C
C real tote(200)
integer igoss, m_col(200), m_row(200), nreaches
common /gossO/ ra_col, m_row, nreaehes
common /goss2/ tote
data igoss /1*0/
C
C C1-------READ ITMP(NUMBER OF RIVER REACHES OR FLAG TO REUSE DATA)
READ(IN,8)ITMP
8 FORMAT(I10)
C
C C2-------TEST ITMP.
IF(ITMP.GE.0)GO TO 50
C
C C2A IF ITMP <0 THEN REUSE DATA FROM LAST STRESS PERIOD.
WRITE(IOUT,7)
7 FORMAT(1Ho,'REUSING RIVER REACHES FROM LAST STRESS PERIOD')
GO TO 260
C
C C3-------IF ITMP=> ZERO THEN IT IS THE NUMBER OF RIVER REACHES
50 NRIVER=ITMP
C
C C4-------WHAT NRIVER=MXRIVR THEN STOP.
IF(NRIVER.LE.MXRIVR)GO TO 100
WRITE(IOUT,99)NRIVER, MXRIVR
99 FORMAT(1Ho,'NRIVER(', 14, ' ) IS GREATER THAN MXRIVR( ' ,14, ' )')
C
C C4A-------ABNORMAL STOP.
STOP
C
C C5-------PRINT NUMBER OF RIVER REACHES IN THIS STRESS PERIOD.
100 WRITE(IOUT,1)NRIVER
1 FORMAT(1Ho,'/IX,15, ' RIVER REACHES')
C
C C6-------IF THERE ARE NO RIVER REACHES THEN RETURN.
IF(NRIVER.EQ.0) GO TO 260
C
C C7-------READ AND PRINT DATA FOR EACH RIVER REACH.
WRITE(IOUT,3)
3 FORMAT(1Ho,15X,'LAYER',5X,'ROW',5X,'COL'
1,' STAGE CONDUCTANCE BOTTOM ELEVATION RIVER REACH'
2/IX,15X,80('-')) DO 250 II=1,NRIVER
READ(IN,4)K,I,J,RIVR(4,II),RIVR(5,II),RIVR(6,II)
DO 249 igoss=1,nreaches
if ( m_row(igoss).eq.I .and. m_col(igoss).eq.J ) then
rivr(5,ii) = tote(igoss)
endif
249 continue
WRITE(IOUT,5)K,I,J,RIVR(4,II),RIVR(5,II),RIVR(6,II)
5 FORMAT(1Ho,15X,T4,9H18,G13.4,G14.4,G19.4,110)
RIVR(1,II)=K
RIVR(2,II)=I
RIVR(3,II)=J
250 CONTINUE
C
C
C.4 THE "READ AND PREPARE" SUBROUTINE OF THE GENERAL HEAD BOUNDARY MODULE

SUBROUTINE GHBIRP (BNDS, NBOUND, MXBDN, IN, IOUT)
C
C-------VERSION 1651 02FEB1983 GHBIRP
C READ DATA FOR GHB
C******************************************************************
C READ ITMP (# OP GENERAL HEAD BOUNDS OR FLAG TO REUSE DATA.)
READ(IN,8) ITMP
8 FORMAT(1I0)
C
C2-------TEST ITMP
IF (ITMP.GE.0) GO TO 50
C
C2A------IF ITMP<0 THEN REUSE DATA FROM LAST STRESS PERIOD
WRITE(IOUT,7)
7 FORMAT(1O8,'REUSING HEAD-DEPENDENT BOUNDS FROM LAST STRESS',10,' PERIOD')
GO TO 260
C
C3------IF ITMP=0 THEN IT IS THE # OF GENERAL HEAD BOUNDS.
50 NBOUND=ITMP
C
C4------IF MAX NUMBER OF BOUNDS IS EXCEEDED THEN STOP
IF (NBOUND.LE.MXBND) GO TO 100
WRITE(IOUT,99) NBOUND, MXBDN
99 FORMAT(1O8,'NBOUND(',14,') IS GREATER THAN MXBND(',14,')')
C
C4A------ABNORMAL STOP
STOP
C
C5------PRINT # OF GENERAL HEAD BOUNDS THIS STRESS PERIOD
100 WRITE(IOUT,1) NBOUND
1 FORMAT(10H,1X,15X,'HEAD-DEPENDENT BOUNDARY NODES')
C
C6------IF THERE ARE NO GENERAL HEAD BOUNDS THEN RETURN.
IF (NBOUND.EQ.0) GO TO 260
C
C7------READ & PRINT DATA FOR EACH GENERAL HEAD BOUNDARY.
WRITE(IOUT,3)
3 FORMAT(1I0,15X,'LAYER',5X,'ROW',5X,1,'COL ELEVATION CONDUCTANCE BOUND NO.',1X,15X,60(1H,'-'))
DO 250 II=1,NBOUND
READ (IN,4) K,I,J,BNDS(4,II),BNDS(5,II)
4 FORMAT(3I10,2F10.0)
DO 249 igoss=1,nreaches
   IF ( m_row(igoss).eq.i .and. m_col(igoss).eq.j ) then
      bnds(5,ii) = tote(igoss)
   endif
249 continue
WRITE (IOUT,5) K,I,J,BNDS(4,II),BNDS(5,II)
5 FORMAT(1H,15X,'14,18,213,4,214,4,28)
END
C.5 SUBROUTINE FOR OUTPUT OF CALIBRATION RESULTS

subroutine chi2unik(hnew, ncol, nrow, aquifer, river)

Calculates chi-squared statistic of errors in water table elevation obtained in
each steady state run of the Goss Moor groundwater model. Also informs the user
of the locations of the water table observation points and outputs the model's
deviation at each individual observation point. C. Ishemo 1999

aquihrzk : conductivity of aquifer along the horizontal principal axis, in m/day
aquikrev : revised conductivity of aquifer in stream-bearing model cell, in m/day
cell : pointer array giving row and column indices of each model cell
containing an observation point
chi2 : chi-squared statistic of errors in water table elevation
(dev : deviation of predicted from observed water table elevation, in m
hobs : observed steady state water table elevation, in m
hpred : predicted steady state water table elevation, in m
log2ak : logarithm (base 2) of the aquifer's horizontal conductivity
(log2rk : logarithm (base 2) of the conductivity of the stream bed sediments
riverk : conductivity of stream bed sediments, in m/day
rms : FINAL VALUE = root mean squared deviation of predicted from observed
steady state water table, in m
xgoss : Easting, relative to South-West corner of model grid, of water table
observation point, in m
ygoss : Northing, relative to South-West corner of model grid, of water table
observation point, in m

imPLICIT none
REAL aquihrzk, aquikrev(200), xgoss(10), ygoss(10), hobs(10),
& rms, chi2, hpred(10), dev(10), log2ak, log2rk, riverk
INTEGER cell(10,2), i, ncol, nrow
DOUBLE PRECISION hnew(ncol, nrow)
CHARACTER*6 site(10)
CHARACTER*2 aquifer, river
CHARACTER*1 text
COMMON /gossl/ aquihrzk, aquikrev
COMMON /goss3/ log2ak, log2rk, riverk
DATA xgoss, ygoss, hobs, hpred, rms, dev, chi2 /52*0.00/
DATA cell, i /21*0/
DATA site /10*' '/
DATA text /1*' '/

xgoss(1) = 810.00
xgoss(2) = 755.00
xgoss(3) = 958.00
xgoss(4) = 1017.00
xgoss(5) = 1989.00
xgoss(6) = 2478.00
xgoss(7) = 1590.00
xgoss(8) = 1587.00
xgoss(9) = 1923.00
xgoss(10) = 3262.00

ygoss(1) = 779.00
ygoss(2) = 338.00
ygoss(3) = 205.00
ygoss(4) = 208.00
ygoss(5) = 632.00
ygoss(6) = 731.00
ygoss(7) = 632.00
ygoss(8) = 1243.00
ygoss(9) = 1285.00
ygoss(10) = 998.00

site(1) = 'P2s'
site(2) = 'P3s'
site(3) = 'P5s, d'
site(4) = 'P6d'
site(5) = 'P10s, d'
site(6) = 'P12s'
site(7) = 'P13s'
site(8) = 'P13s'

253
site(9) = 'P14s'
site(10) = 'P16s'

hobs(1) = 120.4119
hobs(2) = 121.5834
hobs(3) = 123.2691
hobs(4) = 123.2532
hobs(5) = 127.3908
hobs(6) = 130.3360
hobs(7) = 124.1981
hobs(8) = 122.8925
hobs(9) = 125.7456
hobs(10) = 136.2047

cell(1,1) = 34
cell(1,2) = 17
cell(2,1) = 43
cell(2,2) = 45
cell(3,1) = 20
cell(3,2) = 45
cell(4,1) = 21
cell(4,2) = 37
cell(5,1) = 40
cell(5,2) = 35
cell(6,1) = 50
cell(6,2) = 30
cell(7,1) = 34
cell(7,2) = 25
cell(8,1) = 32
cell(8,2) = 24
cell(9,1) = 39
cell(9,2) = 20

do 1020 i=1,10
  hpred(i) = real(hnew(cell(i,1),cell(i,2)))
  dev(i) = hpred(i) - hobs(i)
  rms = rms + dev(i)**2
continue

chi2 = rms/0.034788
rms = rms/9.00
rms = sqrt(rms)

open (65, file='a'//aquifer//'r'//river//'r'.txt', status='new')
write (65, '(al8, f10.3, a6)') 'Aquifer Xhoriz = ', aquihrz, ' m/day'
write (65, '(al8, f10.3, a6)') 'River Bed K = ', riverk, ' m/day'
write (65, ' (al8, f10.3, a6) ') 'SITE', ' X', ' Y', ' OBS', ' PRED', ' DEV'
write (65, '*' )
write (65, 9999) ( sited(i), xgoss(i), ygoss(i), hobs(i), hpred(i),
  dev(i), i=1,10)
write (65, '*')
write (65, '*')
write (65, '*')
write (65, '*')
write (65, '*')
write (65, '*')
write (65, '*')
write (65, ' R.M.S. Deviation = ', rms)
write (65, ' Chi Squared (sigma2=0.348) = ', chi2)
close (65)

open (66, file='aqu_riv.dat', status='old')
do 1030 i=1,1000
  read (66, ' (al') end=7777) text
1030 continue
7777 backspace 66
write (66, ' (t5, f6.2, t15, f6.2, t35, f16.6) ') log2ak, log2rk, chi2
close (66)

9999 format ( 3(al, a4), 3(l6x, a4) )
9998 format ( t5, a6, t14, f7.1, t24, f7.1, t37, 1pe14.4, &
  t51, 0pe20.13, t71, e20.13 )
return
end
APPENDIX D

SAMPLES OF THE SUPPLEMENTARY INPUT AND OUTPUT FILES OF THE STEADY STATE CALIBRATION PROGRAM FOR U.S.G.S. MODFLOW

D.1 STREAM GEOMETRY DATA (INPUT FILE RIVPARAM.CSV, EXPORTED FROM MICROSOFT EXCEL)

Variable names are given in program listing in Section C.1.

File starts on next line:

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<th>m_row</th>
<th>reachlen</th>
<th>bankfall</th>
<th>wdepth</th>
<th>wwidth</th>
<th>bedthick</th>
<th>thick_mx</th>
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**D.2 MODEL FIT OBTAINED IN INDIVIDUAL SIMULATION (DATA OUTPUT IN FILE A??R???.TXT WHERE "?" SIGNIFIES WILDCARD)**

- X is the easting of the observation point relative to the western edge of the model grid (m).
- Y is the northing of the observation point relative to the southern edge of the model grid (m).
- OBS is the annual (weighted) average of the observed groundwater heads above O.D. (m).
- PRED is the steady state groundwater head above O.D. in the closest model grid cell (m), and
- DEV is the deviation of PRED above OBS (m).

**File starts on next line:**

- Aquifer Khoriz = 4.000 m/day
- River Bed K = 16.000 m/day

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R.M.S. Deviation = 0.379956

Chi squared (\(\sigma^2=0.348\)) = 37.3491

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D.3 VARIATION OF MODEL FIT CAUSED BY ADJUSTMENTS OF PERMEABILITY (DATA OUTPUT IN FILE AQU_RIV.DAT)

- \(12a\) is the base 2 logarithm of the assigned aquifer permeability in \(m/day\) (dimensionless).
- \(12r\) is the base 2 logarithm of the assigned river bed permeability in \(m/day\) (dimensionless), and
- \(\text{chi}^2\) is the chi-squared statistic of simulation errors with respect to measurement errors (dimensionless).
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THE HYDROLOGY OF A MAJOR VALLEY WETLAND
AT GOSS MOOR, CORNWALL

Carl A.L. Ishemo, Ph.D. Thesis

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Figure 7.8a Relative Annual Magnitudes of Inputs and Outputs of Wetland Subcatchment 6 when Aquifer Permeability = 0.005 m/day.

Figure 7.8b Relative Annual Magnitudes of Inputs and Outputs of Wetland Subcatchment 6 when Aquifer Permeability = 0.500 m/day.

Figure 7.9a Relative Annual Magnitudes of Inputs and Outputs of Wetland Surface and Substrate in Subcatchment 6 when Aquifer Permeability = 0.005 m/day.

Figure 7.9b Relative Annual Magnitudes of Inputs and Outputs of Wetland Surface and Substrate in Subcatchment 6 when Aquifer Permeability = 0.500 m/day.

APPENDIX A

Figure A1 Stage-Discharge Relation at Gauging Station C6.
| Figure A2 | Stage-Discharge Relation at Gauging Station C5 | 150 |
| Figure A3 | Stage-Discharge Relation at Gauging Station C1 (Goss Moor Upstream Inflow) | 151 |
| PLATE 1 | Looking South-Eastwards from the Centre of Goss Moor towards the Hensbarrow Downs China Clay Area. | 152 |
| PLATE 2 | Looking Northwards from St. Dennis Crown across Goss Moor. | 153 |
### Table 2.1 Seasonal Rainfall and Potential Evaporation at Goss Moor

<table>
<thead>
<tr>
<th>Data from</th>
<th>OCT</th>
<th>NOV</th>
<th>DEC</th>
<th>JAN</th>
<th>FEB</th>
<th>MAR</th>
<th>APR</th>
<th>MAY</th>
<th>JUN</th>
<th>JUL</th>
<th>AUG</th>
<th>SEP</th>
<th>ANNUAL TOTAL</th>
</tr>
</thead>
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<tr>
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<td>125.2</td>
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<td>104.9</td>
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<td>Highest value (mm)</td>
<td>260.0</td>
<td>168.7</td>
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<td>Lowest value (mm)</td>
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<td>[Camm, 1981]</td>
<td>Mean Seasonal Rainfall (mm)</td>
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<td></td>
</tr>
</tbody>
</table>

**Data:** June '93 to May '94. (University of Plymouth and Met. Office)

<p>| Rainfall (mm) | 165.5 | 106.0 | 213.5 | 247.3 | 244.0 | 96.0 | 140.5 | 101.0 | 92.1 | 159.7 | 30.8 | 135.5 | 1732.9 |
| P.E. (mm) | 42.4 | 19.0 | 23.6 | 25.3 | 20.0 | 30.6 | 71.9 | 63.7 | 70.2 | 89.2 | 65.8 | 53.4 | 575.1 |
| Seasonal Rainfall Totals (mm) | 1072.3 | | | | | | | | | | | | 1732.9 |
| Seasonal P.E. Totals (mm) | 660.6 | | | | | | | | | | | | 575.1 |</p>
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<th>Instrument</th>
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<th>Site I.D.'s</th>
<th>Period of Measurement</th>
<th>Sampling Schedule</th>
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<td>River Flow</td>
<td>stage recorder</td>
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<td>C1, C5, C6</td>
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<tr>
<td>Stream Flow</td>
<td>current meter</td>
<td>3</td>
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<td>January 1995 - April 1997</td>
<td>occasional sampling</td>
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<td>stage board</td>
<td>4</td>
<td>S1 - S4</td>
<td>March 1993 - October 1994</td>
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<td>piezometer</td>
<td>26</td>
<td>P01s - P05s, P07s - P14s,</td>
<td>October 1993 - October 1994</td>
<td>weekly</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>P16s - P20s, P05d - P10d,</td>
<td></td>
<td></td>
</tr>
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<td></td>
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<tr>
<td>Air Temperature 1 m Above</td>
<td>temperature-dependent resistor</td>
<td>1</td>
<td>M</td>
<td>January 1994 - May 1994</td>
<td>continuous recording</td>
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<td>Ground</td>
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<tr>
<td>Actual and Saturated Vapour</td>
<td>wet/dry thermistor</td>
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<td>M</td>
<td>January 1994 - May 1994</td>
<td>continuous recording</td>
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<td></td>
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</tr>
<tr>
<td>Wind Speed 2 m Above Ground</td>
<td>anemometer</td>
<td>1</td>
<td>M</td>
<td>January 1994 - May 1994</td>
<td>continuous recording</td>
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<td>Wind Direction</td>
<td>potentiometric wind vane</td>
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<td>M</td>
<td>January 1994 - May 1994</td>
<td>continuous recording</td>
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<td>continuous recording</td>
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<td>piezometer-based slug test</td>
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<td>1994</td>
<td>single measurement</td>
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<td>Alluvium</td>
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<td>none</td>
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<td>single measurement</td>
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<td>Exploration (U.K.) Ltd.)</td>
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Table 3.1  Hydrometric Measurements Taken; Instruments Used and Periods of Sampling. Site I.D.'s correspond to those shown in Figure 3.1.
Table 3.2 Meteorological Parameters Used to Distinguish Types of Vegetation in the Calculation of Evapotranspiration on Goss Moor.

Stand height was measured in metres.

<table>
<thead>
<tr>
<th></th>
<th>Willow Scrub</th>
<th>Wet Heath</th>
<th>Pasture</th>
<th>Open Water</th>
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<td>d</td>
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<td>0.05</td>
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<td>$z$</td>
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<td>2.00</td>
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<tr>
<td>$r_1$</td>
<td>36.4</td>
<td>130.0</td>
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</table>

<table>
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<td>February</td>
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<td>August</td>
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<td>1.0</td>
<td>3.5</td>
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<tr>
<td>September</td>
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<tr>
<td>October</td>
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<tr>
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<td>0.0</td>
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<td>-</td>
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</tr>
<tr>
<td>December</td>
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<td>0.0</td>
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</table>

These data were obtained or derived from the following sources:
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<thead>
<tr>
<th>Piezometer</th>
<th>Material</th>
<th>K (B-R)</th>
<th>K₀(CBP)</th>
<th>Kᵣ adjusted for well skin</th>
<th>Kᵣ</th>
<th>Sᵣ (CBP)</th>
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<tbody>
<tr>
<td>P04s</td>
<td>silty clay</td>
<td>0.006</td>
<td>0.040</td>
<td>-</td>
<td>0.006</td>
<td>5x10⁻⁴</td>
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<td>P05d</td>
<td>peaty clay</td>
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<td>0.156</td>
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<td>0.0009</td>
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<td>P09d</td>
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<td>5x10⁻⁴</td>
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<td>-</td>
<td>4x10⁻⁵</td>
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<td>-</td>
<td>0.0003</td>
<td>0.002</td>
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<td>P16s</td>
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<td>0.606</td>
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<td>0.61</td>
<td>5x10⁻⁴</td>
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</table>

Table 3.3 Hydraulic Conductivities and Specific Storages Derived from Slug Tests on Goss Moor. The values of horizontal conductivity thought to be most representative, for reasons given in the text, are emboldened. The geometric mean of these values is 0.049 m/day.

- \( K \) is isotropic hydraulic conductivity (m/day)
- \( K_h \) is horizontal hydraulic conductivity (m/day)
- \( K_v \) is vertical hydraulic conductivity (m/day)
- \( S \) is specific storage (m⁴)

<table>
<thead>
<tr>
<th>Material</th>
<th>Permeability</th>
<th>Vertical Compressibility (mN⁻¹)</th>
<th>Expected Specific Storage (m⁴)</th>
<th>Specific Storage (m⁴)</th>
<th>Piezometers</th>
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<tr>
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<td>2x10⁻⁶</td>
<td>0.02</td>
<td>0.002, 0.001, P11s, P12s, P05d</td>
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<tr>
<td></td>
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<tr>
<td>Stiff Clay</td>
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<td>2x10⁻⁵</td>
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<tr>
<td></td>
<td>lower limit</td>
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<td>1.3x10⁻⁵</td>
<td>0.0013</td>
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<td>Medium-Hard Clay</td>
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<td>0.4</td>
<td>6.9x10⁻⁸</td>
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<tr>
<td>Loose Sand</td>
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<td>10⁻⁷</td>
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<td></td>
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<td>5.2x10⁻⁸</td>
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<tr>
<td>Dense Sand</td>
<td>upper limit</td>
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<td>2x10⁻⁸</td>
<td>0.0002</td>
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<td>1.3x10⁻⁸</td>
<td>0.00013</td>
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</tr>
<tr>
<td>Dense, Sandy Gravel</td>
<td>upper limit</td>
<td>0.4</td>
<td>10⁻⁸</td>
<td>0.0001</td>
<td>5x10⁻⁴, 5x10⁻⁴</td>
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<tr>
<td></td>
<td>lower limit</td>
<td>0.25</td>
<td>5.2x10⁻⁸</td>
<td>5.2x10⁻⁵</td>
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<tr>
<td>Fissured Rock</td>
<td>upper limit</td>
<td>0.5</td>
<td>6.9x10⁻¹⁸</td>
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<tr>
<td></td>
<td>lower limit</td>
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<td>3.3x10⁻¹⁶</td>
<td>3.5x10⁻⁸</td>
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<td>Sound Rock</td>
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<td>3.3x10⁻¹⁶</td>
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<td></td>
<td>lower limit</td>
<td>0.05</td>
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Table 3.4 Comparison of Goss Moor Specific Storages Calculated by the CBP Method, with Values derived from Data given by Domenico and Schwartz (1990).
<table>
<thead>
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<th>Month</th>
<th>Total Rainfall in Study Month (m)</th>
<th>Total Long Term Average Rainfall (m)</th>
<th>Percentage of L.T.A. in Study Month</th>
</tr>
</thead>
<tbody>
<tr>
<td>September</td>
<td>0.217</td>
<td>0.106</td>
<td>204</td>
</tr>
<tr>
<td>October</td>
<td>0.118</td>
<td>0.134</td>
<td>88</td>
</tr>
<tr>
<td>November</td>
<td>0.155</td>
<td>0.150</td>
<td>103</td>
</tr>
<tr>
<td>December</td>
<td>0.245</td>
<td>0.162</td>
<td>151</td>
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<tr>
<td>January</td>
<td>0.208</td>
<td>0.165</td>
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</tr>
<tr>
<td>February</td>
<td>0.274</td>
<td>0.119</td>
<td>230</td>
</tr>
<tr>
<td>March</td>
<td>0.139</td>
<td>0.118</td>
<td>117</td>
</tr>
<tr>
<td>April</td>
<td>0.100</td>
<td>0.079</td>
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<tr>
<td>May</td>
<td>0.129</td>
<td>0.080</td>
<td>161</td>
</tr>
<tr>
<td>June</td>
<td>0.040</td>
<td>0.080</td>
<td>50</td>
</tr>
<tr>
<td>July</td>
<td>0.061</td>
<td>0.085</td>
<td>72</td>
</tr>
<tr>
<td>August</td>
<td>0.123</td>
<td>0.093</td>
<td>132</td>
</tr>
<tr>
<td>12 Month Total</td>
<td>1.809</td>
<td>1.371</td>
<td>132</td>
</tr>
</tbody>
</table>

Table 4.1 Monthly Rainfall Totals at Roche Weather Station (Environment Agency) during 1993/94
### Volumes of Water Lost by Evapotranspiration

<table>
<thead>
<tr>
<th>Range of Dates</th>
<th>scrub (m³)</th>
<th>heath (m³)</th>
<th>pool (m³)</th>
<th>wetland (m³)</th>
<th>pasture (m³)</th>
<th>TOTAL (m³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1/9/93 - 30/11/93</td>
<td>249,771</td>
<td>188,687</td>
<td>17,592</td>
<td>456,050</td>
<td>1,575,484</td>
<td>2,031,534</td>
</tr>
<tr>
<td>1/12/93 - 28/2/94</td>
<td>163,706</td>
<td>103,289</td>
<td>6,928</td>
<td>273,934</td>
<td>787,742</td>
<td>1,061,676</td>
</tr>
<tr>
<td>1/2/94 - 31/5/94</td>
<td>404,513</td>
<td>302,871</td>
<td>28,205</td>
<td>735,590</td>
<td>2,562,958</td>
<td>3,298,549</td>
</tr>
<tr>
<td>1/6/94 - 31/8/94</td>
<td>539,921</td>
<td>443,705</td>
<td>44,270</td>
<td>1,926,856</td>
<td>3,471,920</td>
<td>4,498,816</td>
</tr>
<tr>
<td>1/9/93 - 31/8/94 (1 year)</td>
<td>1,359,913</td>
<td>1,038,551</td>
<td>97,006</td>
<td>2,492,470</td>
<td>8,398,104</td>
<td>10,890,574</td>
</tr>
</tbody>
</table>

### Percentages of Total Evapotranspiration Losses

<table>
<thead>
<tr>
<th>Range of Dates</th>
<th>scrub %</th>
<th>heath %</th>
<th>pool %</th>
<th>wetland %</th>
<th>pasture %</th>
<th>TOTAL %</th>
</tr>
</thead>
<tbody>
<tr>
<td>1/9/93 - 30/11/93</td>
<td>12.29</td>
<td>9.29</td>
<td>0.87</td>
<td>22.45</td>
<td>77.55</td>
<td>100</td>
</tr>
<tr>
<td>1/12/93 - 28/2/94</td>
<td>15.42</td>
<td>9.73</td>
<td>0.63</td>
<td>25.80</td>
<td>74.20</td>
<td>100</td>
</tr>
<tr>
<td>1/3/94 - 31/5/94</td>
<td>12.26</td>
<td>9.18</td>
<td>0.86</td>
<td>22.30</td>
<td>77.70</td>
<td>100</td>
</tr>
<tr>
<td>1/6/94 - 31/8/94</td>
<td>11.98</td>
<td>9.86</td>
<td>0.98</td>
<td>22.83</td>
<td>77.17</td>
<td>100</td>
</tr>
<tr>
<td>1/9/93 - 31/8/94 (1 year)</td>
<td>12.46</td>
<td>9.54</td>
<td>0.89</td>
<td>23.34</td>
<td>76.66</td>
<td>100</td>
</tr>
</tbody>
</table>

Table 4.2 Seasonal Volumes and Percentages of ET Water Loss from the Goss Moor Catchment

### Contributions of Gauged Tributaries to Annual Volume of Flow at the Wetland Outflow during the Study Period (Sept. 1993 - Aug. 1994)

<table>
<thead>
<tr>
<th>Volume of Flow (m³ per m² of entire catchment)</th>
<th>C5</th>
<th>C1</th>
<th>Other Contributions</th>
<th>C6</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.126</td>
<td>0.371</td>
<td>0.524</td>
<td>1.122</td>
<td></td>
</tr>
<tr>
<td>% Contribution</td>
<td>11.3%</td>
<td>33.1%</td>
<td>55.6%</td>
<td>100%</td>
</tr>
</tbody>
</table>

Table 4.3 Contributions of Gauged Tributaries to Annual Volume of Flow at the Wetland Outflow during the Study Period (Sept. 1993 - Aug. 1994)
### Table 4.4 Stream Flow Separation Parameters and Results

<table>
<thead>
<tr>
<th></th>
<th>C5</th>
<th>C6</th>
<th>C1</th>
</tr>
</thead>
<tbody>
<tr>
<td>recession constant, c (hours⁻¹)</td>
<td>0.0022</td>
<td>0.0018</td>
<td>0.0025</td>
</tr>
<tr>
<td>cutoff frequency, s (hours⁻¹)</td>
<td>0.0090</td>
<td>0.0035</td>
<td>0.0001</td>
</tr>
<tr>
<td>damping factor, β</td>
<td>8.0</td>
<td>2.5</td>
<td>7.0</td>
</tr>
<tr>
<td>scaling factor</td>
<td>0.00012000</td>
<td>0.00003900</td>
<td>0.000000016</td>
</tr>
<tr>
<td><strong>first available data at:</strong></td>
<td>12:00, 16/7/93</td>
<td>17:00, 12/5/93</td>
<td>11:00, 18/8/93</td>
</tr>
<tr>
<td><strong>first input to filter at:</strong></td>
<td>16:00, 16/7/93</td>
<td>20:00, 19/5/93</td>
<td>11:00, 18/8/93</td>
</tr>
<tr>
<td><strong>filter length (hours):</strong></td>
<td>1112</td>
<td>2500</td>
<td>1636</td>
</tr>
<tr>
<td><strong>first output from filter at:</strong></td>
<td>01:00, 1/9/93</td>
<td>01:00, 1/9/93</td>
<td>15:00, 25/10/93</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th>Slow flow (m³)</th>
<th>Quick flow (m³)</th>
<th>Slow flow (m³)</th>
<th>Quick flow (m³)</th>
<th>Slow flow (m³)</th>
<th>Quick flow (m³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1/9/93 - 30/11/93</td>
<td>108481</td>
<td>394783</td>
<td>2810640</td>
<td>3204351</td>
<td>——</td>
<td>——</td>
</tr>
<tr>
<td>1/12/93 - 28/2/94</td>
<td>398034</td>
<td>1414276</td>
<td>5420004</td>
<td>6294685</td>
<td>1040803</td>
<td>2635861</td>
</tr>
<tr>
<td>1/3/94 - 31/5/94</td>
<td>140478</td>
<td>349502</td>
<td>3683223</td>
<td>2446292</td>
<td>1127042</td>
<td>863677</td>
</tr>
<tr>
<td>1/6/94 - 31/8/94</td>
<td>8681</td>
<td>18076</td>
<td>956234</td>
<td>323630</td>
<td>439917</td>
<td>313347</td>
</tr>
<tr>
<td><strong>12 month total</strong></td>
<td>655675</td>
<td>2176638</td>
<td>1287010</td>
<td>12268978</td>
<td>2853273</td>
<td>4155282</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th></th>
<th>Slow flow %</th>
<th>Quick flow %</th>
<th>Slow flow %</th>
<th>Quick flow %</th>
<th>Slow flow %</th>
<th>Quick flow %</th>
</tr>
</thead>
<tbody>
<tr>
<td>1/9/93 - 30/11/93</td>
<td>21.55</td>
<td>78.45</td>
<td>46.72</td>
<td>53.27</td>
<td>——</td>
<td>——</td>
</tr>
<tr>
<td>1/12/93 - 28/2/94</td>
<td>23.14</td>
<td>76.85</td>
<td>46.26</td>
<td>53.73</td>
<td>28.34</td>
<td>71.65</td>
</tr>
<tr>
<td>1/3/94 - 31/5/94</td>
<td>28.67</td>
<td>71.32</td>
<td>60.09</td>
<td>39.91</td>
<td>56.61</td>
<td>43.38</td>
</tr>
<tr>
<td>1/6/94 - 31/8/94</td>
<td>31.95</td>
<td>68.04</td>
<td>74.71</td>
<td>25.28</td>
<td>57.86</td>
<td>42.13</td>
</tr>
<tr>
<td><strong>12 month mean</strong></td>
<td>23.14</td>
<td>76.85</td>
<td>51.19</td>
<td>48.80</td>
<td>40.69</td>
<td>59.30</td>
</tr>
</tbody>
</table>

### Table 4.5 Seasonal Ratios of Runoff: Rainfall at Gauging Stations in Goss Moor

<table>
<thead>
<tr>
<th></th>
<th>C5</th>
<th>C6</th>
<th>C1</th>
</tr>
</thead>
<tbody>
<tr>
<td>1/9/93 - 30/11/93</td>
<td>0.15394</td>
<td>0.35028</td>
<td>---</td>
</tr>
<tr>
<td>1/12/93 - 28/2/94</td>
<td>0.31913</td>
<td>0.40066</td>
<td>0.72343</td>
</tr>
<tr>
<td>1/3/94 - 31/5/94</td>
<td>0.16107</td>
<td>0.32063</td>
<td>0.48449</td>
</tr>
<tr>
<td>1/6/94 - 31/8/94</td>
<td>0.0209</td>
<td>0.09592</td>
<td>0.40644</td>
</tr>
<tr>
<td></td>
<td>Rain</td>
<td>Slow flow</td>
<td>Quick flow</td>
</tr>
<tr>
<td>------------------</td>
<td>------</td>
<td>-----------</td>
<td>------------</td>
</tr>
<tr>
<td><strong>Flux Volumes Into/Out of Catchment (m)</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1/9/93 - 30/11/93</td>
<td>0.408</td>
<td>0.125</td>
<td>0.143</td>
</tr>
<tr>
<td>1/12/93 - 28/2/94</td>
<td>0.701</td>
<td>0.242</td>
<td>0.281</td>
</tr>
<tr>
<td>1/2/94 - 31/5/94</td>
<td>0.341</td>
<td>0.164</td>
<td>0.109</td>
</tr>
<tr>
<td>1/6/94 - 31/8/94</td>
<td>0.151</td>
<td>0.043</td>
<td>0.014</td>
</tr>
<tr>
<td><strong>Flux Volumes as Percentages of Yearly Totals</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1/9/93 - 30/11/93</td>
<td>25.5</td>
<td>21.8</td>
<td>26.1</td>
</tr>
<tr>
<td>1/12/93 - 28/2/94</td>
<td>43.8</td>
<td>42.1</td>
<td>51.3</td>
</tr>
<tr>
<td>1/3/94 - 31/5/94</td>
<td>21.3</td>
<td>25.6</td>
<td>19.9</td>
</tr>
<tr>
<td>1/6/94 - 31/8/94</td>
<td>9.4</td>
<td>7.4</td>
<td>2.6</td>
</tr>
<tr>
<td><strong>Flux Volumes Into/Out of Catchment as Percentage of Rainfall Volume</strong></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1/9/93 - 30/11/93</td>
<td>100</td>
<td>30.8</td>
<td>35.1</td>
</tr>
<tr>
<td>1/12/93 - 28/2/94</td>
<td>100</td>
<td>34.5</td>
<td>40.1</td>
</tr>
<tr>
<td>1/3/94 - 31/5/94</td>
<td>100</td>
<td>48.2</td>
<td>32.0</td>
</tr>
<tr>
<td>1/6/94 - 31/8/94</td>
<td>100</td>
<td>25.3</td>
<td>9.6</td>
</tr>
<tr>
<td><strong>Table 5.1</strong> Seasonal Variations of Inputs and Outputs of the Goss Moor Catchment</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
### Table 6.1 Channel Width and Bed Thickness Values Assigned to Streams in Groundwater Model

<table>
<thead>
<tr>
<th>Reference in Literature</th>
<th>Stream Type</th>
<th>Bed Sediment Thickness (m)</th>
<th>Hyporheic Zone Thickness (m)</th>
<th>Bed Gradient (%)</th>
<th>Mean Discharge (m³/s)</th>
<th>Channel Width (m)</th>
<th>Catchment Area (km²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Angadi and Hood, 1998</td>
<td>Appalachian streams</td>
<td>0.27, 0.37, 0.38</td>
<td>-----</td>
<td>6.7</td>
<td>0.002</td>
<td>-----</td>
<td>0.1</td>
</tr>
<tr>
<td></td>
<td>Appalachain streams</td>
<td>0.15, &gt;0.5, &gt;0.5</td>
<td>-----</td>
<td>4.6</td>
<td>0.010</td>
<td>-----</td>
<td>0.36</td>
</tr>
<tr>
<td></td>
<td>Appalachain streams</td>
<td>0.82, 1.00, 1.60</td>
<td>-----</td>
<td>2.7</td>
<td>0.043</td>
<td>-----</td>
<td>1.78</td>
</tr>
<tr>
<td></td>
<td>Appalachain streams</td>
<td>0.76, 1.00, &gt;1.60</td>
<td>-----</td>
<td>1.2</td>
<td>0.139</td>
<td>-----</td>
<td>----</td>
</tr>
<tr>
<td>Hendricks and White, 1991</td>
<td>Sand-bottomed river, N. Michigan</td>
<td>0.50</td>
<td>-----</td>
<td>-----</td>
<td>-----</td>
<td>7. - 8.</td>
<td>----</td>
</tr>
<tr>
<td>Sack et al., 1992</td>
<td>1st order Coastal Plain streams</td>
<td>0.35 - 0.40</td>
<td>-----</td>
<td>0.08</td>
<td>0.05 - 0.08</td>
<td>2.5</td>
<td>----</td>
</tr>
<tr>
<td></td>
<td>1st order Coastal Plain streams</td>
<td>&lt;0.6 (floodplain clay)</td>
<td>-----</td>
<td>0.03</td>
<td>0.15 - 0.16</td>
<td>3.0</td>
<td>----</td>
</tr>
<tr>
<td>Schmid-Araya, 1994</td>
<td>Alpine gravel stream</td>
<td>0.70</td>
<td>0.60</td>
<td>1.0 - 6.7</td>
<td>0.80</td>
<td>-15.</td>
<td>20.</td>
</tr>
<tr>
<td>Hill and Lymburner, 1998</td>
<td>Upland, headwater streams</td>
<td>0.5 - 1.0</td>
<td>0.10 - 0.20</td>
<td>-----</td>
<td>baseflow = 0.012 - 0.015</td>
<td>-4.8</td>
<td>----</td>
</tr>
<tr>
<td></td>
<td>Upland, headwater streams</td>
<td>&lt;0.0 (floodplain gravel)</td>
<td>&lt;0.20</td>
<td>-----</td>
<td>baseflow = 0.11 - 0.12</td>
<td>----</td>
<td>----</td>
</tr>
</tbody>
</table>

### Table 6.2 Bed Sediment and Hyporheic Zone Thicknesses of Various Streams

<table>
<thead>
<tr>
<th>P2s</th>
<th>P3s</th>
<th>P5s</th>
<th>P6s</th>
<th>P10s</th>
<th>P10d</th>
<th>P11s</th>
<th>P12s</th>
<th>P13s</th>
<th>P14s</th>
<th>P16s</th>
<th>Average variance</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mean Elevation (m above OD)</td>
<td>120.412</td>
<td>121.583</td>
<td>123.224</td>
<td>123.314</td>
<td>123.253</td>
<td>127.415</td>
<td>127.366</td>
<td>130.356</td>
<td>124.198</td>
<td>122.892</td>
<td>125.745</td>
</tr>
<tr>
<td>Variance of Mean (m²)</td>
<td>0.006</td>
<td>0.014</td>
<td>0.012</td>
<td>0.002</td>
<td>0.040</td>
<td>0.003</td>
<td>0.092</td>
<td>0.039</td>
<td>0.037</td>
<td>0.057</td>
<td>0.080</td>
</tr>
</tbody>
</table>
Quantities with no numerical indices are defined over the total modelled wetland area. Those with numerical indices are defined over the appropriate wetland subcatchment or boundary section shown in Figure 7.1.

### Table 7.1 Quantities in the Wetland Water Budget and their Symbols

<table>
<thead>
<tr>
<th>Surface Catchment</th>
<th>Area</th>
<th>Subcatchment Perimeter ( \cap ) Model Perimeter</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Symbol</td>
<td>Value (m²)</td>
</tr>
<tr>
<td>Subcatchment 2 (Toad Hole Drain)</td>
<td>( A_2 )</td>
<td>442 000</td>
</tr>
<tr>
<td>Subcatchment 3, receiving from Tregoss Ridge</td>
<td>( A_3 )</td>
<td>164 500</td>
</tr>
<tr>
<td>Subcatchment 4, receiving from Tregoss Ridge</td>
<td>( A_4 )</td>
<td>93 000</td>
</tr>
<tr>
<td>Subcatchment 5 (Culvert beneath A30 road)</td>
<td>( A_5 )</td>
<td>642 500</td>
</tr>
<tr>
<td>Subcatchment 6 (outflow = total outflow)</td>
<td>( A_6 )</td>
<td>3 180 500</td>
</tr>
<tr>
<td>Total Wetland Area (outflow = subcatchment 6 outflow)</td>
<td>( A )</td>
<td>4 522 500</td>
</tr>
<tr>
<td>none (no-surface-inflow section of wetland boundary)</td>
<td>---</td>
<td>---</td>
</tr>
<tr>
<td>Monitoring Site</td>
<td>C2</td>
<td>C3</td>
</tr>
<tr>
<td>----------------</td>
<td>-----</td>
<td>-----</td>
</tr>
<tr>
<td>Date of Metering</td>
<td></td>
<td></td>
</tr>
<tr>
<td>26/1/95</td>
<td>0.178</td>
<td>0.307</td>
</tr>
<tr>
<td>15/12/96</td>
<td>—</td>
<td>0.095</td>
</tr>
<tr>
<td>6/1/97</td>
<td>—</td>
<td>0.048</td>
</tr>
<tr>
<td>13/1/97</td>
<td>0.106</td>
<td>0.046</td>
</tr>
<tr>
<td>20/1/97</td>
<td>0.038</td>
<td>0.046</td>
</tr>
<tr>
<td>29/1/97</td>
<td>0.030</td>
<td>0.041</td>
</tr>
<tr>
<td>24/2/97</td>
<td>0.067</td>
<td>0.114</td>
</tr>
<tr>
<td>3/3/97</td>
<td>0.113</td>
<td>—</td>
</tr>
<tr>
<td>10/3/97</td>
<td>0.060</td>
<td>0.079</td>
</tr>
<tr>
<td>18/3/97</td>
<td>0.080</td>
<td>0.068</td>
</tr>
<tr>
<td>24/3/97</td>
<td>—</td>
<td>—</td>
</tr>
<tr>
<td>14/4/97</td>
<td>0.030</td>
<td>0.029</td>
</tr>
<tr>
<td>28/4/97</td>
<td>0.041</td>
<td>0.039</td>
</tr>
<tr>
<td>Mean</td>
<td>0.077</td>
<td>0.083</td>
</tr>
<tr>
<td>Standard Deviation</td>
<td>0.045</td>
<td>0.079</td>
</tr>
<tr>
<td>Coefficient of Variation</td>
<td>0.58</td>
<td>0.95</td>
</tr>
<tr>
<td>Ratio of Site Mean to C6 Mean</td>
<td>0.062</td>
<td>0.066</td>
</tr>
</tbody>
</table>

Table 7.3 Relating Spot-Metered Flows at Sites C2, C3 and C4 to Continuously Recorded Flows at Site C6

<table>
<thead>
<tr>
<th>Term in Wetland Water Budget</th>
<th>Error in the Value Assigned to the Term, due to Neglect of Surface Water Storage</th>
</tr>
</thead>
<tbody>
<tr>
<td>$+ \Delta D \cdot A$</td>
<td>$-\Delta D \cdot A$</td>
</tr>
<tr>
<td>$+ S_{IN} \cdot (L - L_0)$</td>
<td>$\left[ \frac{\Delta D \cdot A_6}{L_6} \cdot (L - L_0) \right]$</td>
</tr>
<tr>
<td>$+ \sum_{i=2}^{S} C_{2i}$</td>
<td>$-\Delta D \cdot \sum_{i=2}^{S} A_i + \Delta D \cdot A_6 \cdot \left( \frac{\sum_{i=2}^{S} L_i}{L_6} \right)$</td>
</tr>
</tbody>
</table>

Table 7.4 Surface Water Storage Errors, Term by Term, in Equation 7.3.
Table 7.5a Fluxes and Storage Gain in Transient Model of Wetland Aquifer with Aquifer Permeability = 0.005 m/day.

<table>
<thead>
<tr>
<th>Date</th>
<th>Rain Removed</th>
<th>Calibrated ET</th>
<th>Recharge</th>
<th>DRN</th>
<th>RIV</th>
<th>GIN</th>
<th>GOUT</th>
<th>Storage Gain</th>
</tr>
</thead>
<tbody>
<tr>
<td>1/93 - 30/11/93</td>
<td>0.407</td>
<td>0.137</td>
<td>0.106</td>
<td>0.270</td>
<td>0.0006</td>
<td>0.0001</td>
<td>0.0000</td>
<td>0.031</td>
</tr>
<tr>
<td>1/12/93 - 28/2/94</td>
<td>0.702</td>
<td>0.071</td>
<td>0.066</td>
<td>0.631</td>
<td>0.0007</td>
<td>0.0001</td>
<td>0.0000</td>
<td>0.002</td>
</tr>
<tr>
<td>1/3/94 - 31/5/94</td>
<td>0.343</td>
<td>0.169</td>
<td>0.180</td>
<td>0.175</td>
<td>0.0007</td>
<td>0.0001</td>
<td>0.0000</td>
<td>-0.012</td>
</tr>
<tr>
<td>1/6/94 - 31/8/94</td>
<td>0.144</td>
<td>0.143</td>
<td>0.170</td>
<td>0.001</td>
<td>0.0003</td>
<td>0.0001</td>
<td>0.0000</td>
<td>-0.027</td>
</tr>
<tr>
<td>1/9/93 - 31/8/94</td>
<td>1.597</td>
<td>0.524</td>
<td>0.524</td>
<td>1.976</td>
<td>0.0623</td>
<td>0.0003</td>
<td>0.0001</td>
<td>-0.096</td>
</tr>
</tbody>
</table>

Table 7.5b Fluxes and Storage Gain in Transient Model of Wetland Aquifer with Aquifer Permeability = 0.500 m/day.
<table>
<thead>
<tr>
<th>Subcat.</th>
<th>Total Outflow, $C_t$</th>
<th>Outgoing Quick flow - Quick flows Incoming to Subcat. 6</th>
<th>Total Slow flow</th>
<th>Outgoing Slow flow - Slow flows Incoming to Subcat. 6</th>
<th>Wetland-sourced River Flow</th>
<th>DRN - $E_{RES}$</th>
<th>RIV</th>
<th>Wetland-sourced River Flow in Subcat. 6</th>
<th>DRN - $E_{RES}$</th>
<th>RIV - $E_{RES}$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1/9/93 - 30/11/93</td>
<td>0.263</td>
<td>0.017</td>
<td>0.125</td>
<td>0.082</td>
<td>0.053</td>
<td>0.055</td>
<td>0.0001</td>
<td>0.037</td>
<td>0.036</td>
<td>0.0001</td>
</tr>
<tr>
<td>1/12/93 - 28/2/94</td>
<td>0.524</td>
<td>-0.031</td>
<td>0.242</td>
<td>0.158</td>
<td>0.127</td>
<td>0.127</td>
<td>0.0001</td>
<td>0.091</td>
<td>0.091</td>
<td>0.0001</td>
</tr>
<tr>
<td>1/3/94 - 31/5/94</td>
<td>0.237</td>
<td>0.004</td>
<td>0.164</td>
<td>0.104</td>
<td>0.034</td>
<td>0.034</td>
<td>0.0001</td>
<td>0.024</td>
<td>0.024</td>
<td>0.0001</td>
</tr>
<tr>
<td>1/6/94 - 31/8/94</td>
<td>0.038</td>
<td>-0.005</td>
<td>0.043</td>
<td>0.024</td>
<td>-0.019</td>
<td>-0.019</td>
<td>0.0001</td>
<td>-0.019</td>
<td>-0.019</td>
<td>0.0000</td>
</tr>
<tr>
<td>1/9/93 - 31/8/94</td>
<td>1.142</td>
<td>-0.019</td>
<td>0.576</td>
<td>0.367</td>
<td>0.195</td>
<td>0.195</td>
<td>0.0005</td>
<td>0.122</td>
<td>0.122</td>
<td>0.0005</td>
</tr>
</tbody>
</table>

Flow Volumes as Percentage of Catchment Outflow

| Subcat. | Total Outflow, $C_t$ | Outgoing Quick flow - Quick flows Incoming to Subcat. 6 | Total Slow flow | Outgoing Slow flow - Slow flows Incoming to Subcat. 6 | Wetland-sourced River Flow | DRN - $E_{RES}$ | RIV | Wetland-sourced River Flow in Subcat. 6 | DRN - $E_{RES}$ | RIV - $E_{RES}$ |
|---------|---------------------|--------------------------------------------------------|----------------|------------------------------------------------------|                             |                |-----|----------------------------------------|----------------|----------------|
|         |                     |                                                        |                |                                                     |                             |                |     |                                        |                |                 |
| 1/9/93 - 30/11/93 | 100       | 6.52                                                 | 47.71          | 31.15                                                | 19.97                        | 19.92           | 0.05 | 13.89                                  | 13.87           | 0.03           |
| 1/12/93 - 28/2/94 | 100       | -5.93                                                | 46.15          | 30.09                                                | 24.31                        | 24.28           | 0.03 | 17.29                                  | 17.27           | 0.02           |
| 1/3/94 - 31/5/94  | 100       | 1.40                                                 | 59.38          | 37.39                                                | 12.35                        | 12.30           | 0.05 | 8.56                                   | 8.53            | 0.03           |
| 1/6/94 - 31/8/94  | 100       | -15.92                                               | 73.22          | 40.69                                                | -31.97                       | -32.09          | 0.12 | -32.19                                 | -32.26          | 0.07           |
| 1/9/93 - 31/8/94  | 100       | -2.75                                                | 51.19          | 32.62                                                | 17.42                        | 17.37           | 0.03 | 14.72                                  | 14.74           | 0.02           |

Flow Volumes as Percentage of Catchment Slow flow

| Subcat. | Total Outflow, $C_t$ | Outgoing Quick flow - Quick flows Incoming to Subcat. 6 | Total Slow flow | Outgoing Slow flow - Slow flows Incoming to Subcat. 6 | Wetland-sourced River Flow | DRN - $E_{RES}$ | RIV | Wetland-sourced River Flow in Subcat. 6 | DRN - $E_{RES}$ | RIV - $E_{RES}$ |
|---------|---------------------|--------------------------------------------------------|----------------|------------------------------------------------------|                             |                |-----|----------------------------------------|                |                 |
|         |                     |                                                        |                |                                                     |                             |                |     |                                        |                |                 |
| 1/9/93 - 30/11/93 | ___       | 13.66                                                 | 65.29          | 41.85                                                | 41.75                        | 0.10            | 29.12 | 29.07                                  | 0.06            |
| 1/12/93 - 28/2/94 | ___       | -12.82                                                | 65.19          | 52.56                                                | 52.61                        | 0.06            | 37.46 | 37.42                                  | 0.03            |
| 1/3/94 - 31/5/94  | ___       | 2.35                                                  | 62.97          | 20.79                                                | 20.71                        | 0.08            | 14.11 | 14.16                                  | 0.05            |
| 1/6/94 - 31/8/94  | ___       | -21.74                                                | 55.57          | -43.67                                               | -43.83                       | 0.16            | -43.97 | -44.06                                 | 0.09            |
| 1/9/93 - 31/8/94  | ___       | -23.66                                                | 63.85          | 34.02                                                | 33.94                        | 0.08            | -33.09 | -32.94                                 | 0.04            |

Flow Volumes as Percentage of Wetland-Sourced Flow

| Subcat. | Total Outflow, $C_t$ | Outgoing Quick flow - Quick flows Incoming to Subcat. 6 | Total Slow flow | Outgoing Slow flow - Slow flows Incoming to Subcat. 6 | Wetland-sourced River Flow | DRN - $E_{RES}$ | RIV | Wetland-sourced River Flow in Subcat. 6 | DRN - $E_{RES}$ | RIV - $E_{RES}$ |
|---------|---------------------|--------------------------------------------------------|----------------|------------------------------------------------------|                             |                |-----|----------------------------------------|                |                 |
|         |                     |                                                        |                |                                                     |                             |                |     |                                        |                |                 |
| 1/9/93 - 30/11/93 | ___       | ___                                                   | 100            | 99.76                                                | 0.24                         | ___            | ___ | ___                                    | ___            | ___            |
| 1/12/93 - 28/2/94 | ___       | ___                                                   | 100            | 99.89                                                | 0.11                         | ___            | ___ | ___                                    | ___            | ___            |
| 1/3/94 - 31/5/94  | ___       | ___                                                   | 100            | 99.60                                                | 0.40                         | ___            | ___ | ___                                    | ___            | ___            |
| 1/6/94 - 31/8/94  | ___       | ___                                                   | 100            | 100.37                                               | -0.37                        | ___            | ___ | ___                                    | ___            | ___            |
| 1/9/93 - 31/8/94  | ___       | ___                                                   | 100            | 99.76                                                | 0.24                         | ___            | ___ | ___                                    | ___            | ___            |

Table 7.6a Components of Flow in the River on its Exit from the Catchment, with Aquifer Permeability = 0.005 m/day
<table>
<thead>
<tr>
<th></th>
<th>Total Outflow, ( C_2 )</th>
<th>Outgoing Quick flow - Quick flows Incoming to Subcat. 6</th>
<th>Total Slow Flow</th>
<th>Outgoing Slow flow - Slow flows Incoming to Subcat. 6</th>
<th>Wetland-sourced River Flow</th>
<th>DRN( - ) E(_{RES})</th>
<th>RIV</th>
<th>Wetland-sourced River Flow in Subcat. 6</th>
<th>DRN( <em>6 ) (- ) E(</em>{RES})</th>
<th>RIV(_6)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Total Outflow, ( C_2 )</td>
<td>Outgoing Quick flow - Quick flows Incoming to Subcat. 6</td>
<td>Total Slow Flow</td>
<td>Outgoing Slow flow - Slow flows Incoming to Subcat. 6</td>
<td>Wetland-sourced River Flow</td>
<td>DRN( - ) E(_{RES})</td>
<td>RIV</td>
<td>Wetland-sourced River Flow in Subcat. 6</td>
<td>DRN( <em>6 ) (- ) E(</em>{RES})</td>
</tr>
<tr>
<td>1/9/93 - 30/11/93</td>
<td>0.263</td>
<td>0.017</td>
<td>0.125</td>
<td>0.082</td>
<td>0.0517</td>
<td>0.0440</td>
<td>0.0077</td>
<td>0.035</td>
<td>0.031</td>
<td>0.004</td>
</tr>
<tr>
<td>1/12/93 - 28/2/94</td>
<td>0.524</td>
<td>-0.031</td>
<td>0.242</td>
<td>0.158</td>
<td>0.1279</td>
<td>0.1164</td>
<td>0.0115</td>
<td>0.090</td>
<td>0.083</td>
<td>0.006</td>
</tr>
<tr>
<td>1/6/94 - 31/5/94</td>
<td>0.277</td>
<td>0.004</td>
<td>0.164</td>
<td>0.104</td>
<td>0.0360</td>
<td>0.0276</td>
<td>0.0084</td>
<td>0.023</td>
<td>0.019</td>
<td>0.004</td>
</tr>
<tr>
<td>1/6/94 - 31/8/94</td>
<td>0.058</td>
<td>-0.009</td>
<td>0.043</td>
<td>0.024</td>
<td>-0.0177</td>
<td>-0.0204</td>
<td>0.0027</td>
<td>-0.019</td>
<td>-0.021</td>
<td>0.001</td>
</tr>
<tr>
<td>1/3/94 - 31/8/94</td>
<td>-0.019</td>
<td>0.075</td>
<td>0.267</td>
<td>0.156</td>
<td>0.1670</td>
<td>0.1676</td>
<td>0.0162</td>
<td>0.012</td>
<td>0.012</td>
<td>0.016</td>
</tr>
</tbody>
</table>

Table 7.6b Components of Flow in the River on its Exit from the Catchment, with Aquifer Permeability = 0.500 m/day
## Table 7.7a Annual Inflows and Outflows for Wetland Subcatchment 6, with Aquifer Permeability = 0.005 m/day

<table>
<thead>
<tr>
<th></th>
<th>Inflows to Subcatchment 6</th>
<th>Outflows from Subcatchment 6</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Rain</td>
<td>Overland Inflows</td>
</tr>
<tr>
<td>m&lt;sup&gt;3&lt;/sup&gt; per m&lt;sup&gt;2&lt;/sup&gt; of entire catchment</td>
<td>0.227</td>
<td>0.216</td>
</tr>
<tr>
<td>m&lt;sup&gt;3&lt;/sup&gt; per m&lt;sup&gt;2&lt;/sup&gt; of subcat. 6</td>
<td>1.597</td>
<td>1.517</td>
</tr>
<tr>
<td>As % of Total</td>
<td>19.06</td>
<td>18.11</td>
</tr>
</tbody>
</table>

## Table 7.7b Annual Inflows and Outflows for Wetland Subcatchment 6, with Aquifer Permeability = 0.500 m/day

<table>
<thead>
<tr>
<th></th>
<th>Inflows to Subcatchment 6</th>
<th>Outflows from Subcatchment 6</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Rain</td>
<td>Overland Inflows</td>
</tr>
<tr>
<td>m&lt;sup&gt;3&lt;/sup&gt; per m&lt;sup&gt;2&lt;/sup&gt; of entire catchment</td>
<td>0.227</td>
<td>0.219</td>
</tr>
<tr>
<td>m&lt;sup&gt;3&lt;/sup&gt; per m&lt;sup&gt;2&lt;/sup&gt; of subcat. 6</td>
<td>1.597</td>
<td>1.541</td>
</tr>
<tr>
<td>As % of Total</td>
<td>19.01</td>
<td>18.34</td>
</tr>
</tbody>
</table>

Table 7.7a

Table 7.7b
### Table 7.8a Annual Inflows and Outflows for Wetland Subcatchment 6 excluding Stream Flows, with Aquifer Permeability = 0.005 m/day

<table>
<thead>
<tr>
<th></th>
<th>Inflows to Subcatchment 6 excluding Stream Flows</th>
<th>Outflows from Subcatchment 6 excluding Stream Flows</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Rain</td>
<td>Overland Inflows to subcat. 6</td>
</tr>
<tr>
<td>m³ per m² of entire catchment</td>
<td>0.227</td>
<td>0.216</td>
</tr>
<tr>
<td>m³ per m² of subcat. 6</td>
<td>1.597</td>
<td>1.517</td>
</tr>
<tr>
<td>As % of Total</td>
<td>51.28</td>
<td>48.72</td>
</tr>
</tbody>
</table>

### Table 7.8b Annual Inflows and Outflows for Wetland Subcatchment 6 excluding Stream Flows, with Aquifer Permeability = 0.500 m/day

<table>
<thead>
<tr>
<th></th>
<th>Inflows to Subcatchment 6 excluding Stream Flows</th>
<th>Outflows from Subcatchment 6 excluding Stream Flows</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Rain</td>
<td>Overland Inflows to subcat. 6</td>
</tr>
<tr>
<td>m³ per m² of entire catchment</td>
<td>0.227</td>
<td>0.219</td>
</tr>
<tr>
<td>m³ per m² of subcat. 6</td>
<td>1.597</td>
<td>1.541</td>
</tr>
<tr>
<td>As % of Total</td>
<td>50.89</td>
<td>49.11</td>
</tr>
</tbody>
</table>
Figure 1.1  Location of the Goss Moor Catchment in South-West England
Figure 2.1 Surface Features of the Goss Moor Wetland
Figure 2.3 The Grid of Boreholes Drilled by Billiton Exploration (UK) Ltd. on Goss Moor and the Directions of the Vertical Sections shown in Figures 2.4 - 2.6.
Figure 2.4  Vertical Section A-A' through the Goss Moor Wetland Aquifer, from Camm (1981). See Figure 2.3 for the Direction of this Section.
Figure 2.5  Vertical Section B-B' through the Goss Moor Wetland Aquifer, from Canran (1981). See Figure 2.3 for the Direction of this Section.
Figure 2.6  Vertical Section C-C' through the Goss Moor Wetland Aquifer, from Camin (1981). See Figure 2.5 for the Direction of this Section.
Figure 2.7 Soils of the Goss Moor Catchment
Figure 2.8  Surface Features of the Goss Moor Catchment. Reproduced from: Ordnance Survey, 1974. 1:50 000 First Series, Sheet 200: Newquay and Bodmin. Ordnance Survey, Southampton.
Figure 3.2 Subcatchments Draining to each Stream Gauging Site
Figure 3.3  Configuration of Stage Recorders used at Sites C1, C5 and C6
Figure 3.4  Daily Rainfall Totals at Goss Moor versus Daily Rainfall Totals at St. Mawgan

Figure 3.5  Daily Rainfall Totals at Goss Moor versus Daily Rainfall Totals at Roche
Figure 3.6  Net Radiation Measured at Goss Moor versus Net Radiation calculated from St. Mawgan Data

\[ \text{Goss} = -0.755 + 0.927 \times \text{StMawgan} \]

\[ R^2 = 0.90 \]

Figure 3.7  Daily Mean Wind Speed at Goss Moor versus Daily Mean Wind Speed at St. Mawgan

\[ \text{Goss} = 0.048 + 0.416 \times \text{StMawgan} \]

\[ R^2 = 0.92 \]
Goss = 13.15 + 0.95*StMawgan

\[ R^2 = 0.87 \]

1 : 1 line

regression line

Upper 95% confidence limit for gradient = 1.02
Lower 95% confidence limit for gradient = 0.88

Figure 3.8  Daily Mean Air Temperature at Goss Moor versus Daily Mean Air Temperature at St. Mawgan

Actual = -53.60 + 1.02*Saturation

\[ R^2 = 0.96 \]

Figure 3.9  Daily Mean Actual Vapour Pressure versus Daily Mean Saturation Vapour Pressure, at Goss Moor
Figure 3.10  Piezometer Slug Test in an Unconfined Unit

Figure 3.11  Piezometer Slug Test in a Confined Stratum
Figure 3.12  Cooper et al. (1967) Slug Test Recovery Function Fitted to Response of Piezometer P04s

Figure 3.13  Cooper et al. (1967) Slug Test Recovery Function Fitted to Response of Piezometer P05d
Figure 3.14  Cooper *et al.* (1967) Slug Test Recovery Function Fitted to Response of Piezometer P07d

Figure 3.15  Cooper *et al.* (1967) Slug Test Recovery Function Fitted to Response of Piezometer P09d
Figure 3.16  Cooper et al. (1967) Slug Test Recovery Function Fitted to Response of Piezometer P10d

Figure 3.17  Cooper et al. (1967) Slug Test Recovery Function Fitted to Response of Piezometer P11s
Figure 3.18  Cooper et al. (1967) Slug Test Recovery Function Fitted to Response of Piezometer P12s

Figure 3.19  Cooper et al. (1967) Slug Test Recovery Function Fitted to Response of Piezometer P16s
Conductivity = 6.59x10^{-6} \text{ m/s} = 0.00569 \text{ m/day}

Figure 3.20 Bouwer and Rice (1976) Slug Test Recovery Function Fitted to Response of Piezometer P04s

Conductivity = 5.76x10^{-7} \text{ m/s} = 0.0497 \text{ m/day}

Figure 3.21 Bouwer and Rice (1976) Slug Test Recovery Function Fitted to Response of Piezometer P05d
Figure 3.22 Bouwer and Rice (1976) Slug Test Recovery Function Fitted to Response of Piezometer P09d

Figure 3.23 Bouwer and Rice (1976) Slug Test Recovery Function Fitted to Response of Piezometer P10d
Figure 3.24  Bouwer and Rice (1976) Slug Test Recovery Function Fitted to Response of Piezometer P16s

Conductivity = $7.96 \times 10^{-7}$ m/s = 0.0687 m/day
Figure 4.1  Daily Rainfall and Antecedent Precipitation Index over the Study Year (1993/94)

Figure 4.2  Number of Rain Days Per Month During Study Year (1993/94)
Figure 4.5  Daily Mean Rates of Evapotranspiration from the Willow Canopy and from the Surfaces Underlying the Willow Canopy, Normalised with respect to the Open Water Evaporation

Figure 4.6  Seasonal Variations of Daily Evapotranspired Volumes from Pasture and from Wetland Areas in the Goss Moor Catchment
Figure 4.7  Daily Evapotranspiration Rates from beneath the Canopies of Willow Scrub and Wet Heath Vegetation, compared with those found above Pasture and Water
Figure 4.8  Daily Average Stream Flows at Sites C1, C5 and C6, and Local Rainfall
Figure 4.9  Comparison of Flows at C6 with Flows Summed from C1 and C5

Figure 4.10  Contributions to Cumulative Stream Outflows from the Wetland
Figure 4.11  Hourly Rainfall and Stream Flows in the Wetland

Figure 4.12  Duration Curves for Daily Average Flows over One Year at Sites C1, C5 and C6
Figure 4.13  Duration Curves for Daily Average Flows at the Outflow of the Goss Moor Catchment (Site C6) and other Headwater Wetland Catchments. (After Burt, 1995 and Devito et al., 1996.)

Figure 4.14  Spectral Power Densities (Normalised with respect to Total Variance) of Hydrological Variables on Goss Moor (Daily Data)
Figure 4.15  Phase Spectra of Stream Flows and Evapotranspiration with respect to Rainfall (Daily Data)

Figure 4.16  Coherencies of Stream Flows and Evapotranspiration with respect to Rainfall (Daily Data)
Figure 4.17 Phase Spectra of Stream Flows with respect to Evapotranspiration (Daily Data)

Figure 4.18 Coherencies of Stream Flows with Evapotranspiration (Daily Data)
Figure 4.19  Impulse Response Function used for Filtering Stream Flow at Site C6

Figure 4.20  Hourly Flows at Site C6 in Log-Linear form, illustrating Exponential Recession
Figure 4.21 Structure of Dawdy-O'Donnell Rainfall-Runoff Model (After Fleming, 1975)
Figure 4.22  Amplitude Spectra (= √[Normalised Spectral Power Density]) of the Exponential Recession and the Complete Flow Record at Site C6. Also shown is the Gain Spectrum of the Filter used for Flow Separation at this Site.
Figure 4.23  Daily Means of Stream Flow separated into Quick Flow and Slow Flow at Site C6
Figure 4.24  Hourly Means of Stream Flow during Winter 1993/94 separated into Quick Flow and Slow Flow at Site C6

Figure 4.25  Hourly Means of Stream Flow during Summer 1994 separated into Quick Flow and Slow Flow at Site C6
Figure 4.26
Daily Means of Stream Flow separated into Quick Flow and Slow Flow at Site C5
Figure 4.27  Hourly Means of Stream Flow during Winter 1993/94 separated into Quick Flow and Slow Flow at Site C5

Figure 4.28  Hourly Means of Stream Flow during Summer 1994 separated into Quick Flow and Slow Flow at Site C5
Figure 4.29 Daily Means of Stream Flow Separated into Quick Flow and Slow Flow at Site CI
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Distance Upstream of Confluence (m)

Elevation above O.D. (m)

- Ground Surface
- Stream Water Surface
- Stream Bed Surface
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Figure 6.18: Annual Average Evapotranspiration Rates (m/day) in the Model Domain

Legend:
- 0.0025 - 0.0029
- 0.0029 - 0.0033
- 0.0033 - 0.0037
- 0.0037 - 0.0041
- Surface Water Features
- Model Boundary

Scale:
0 0.5 1 1.5 2 2.5 3 Kilometers
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Figure 6.29 Agreement between Modelled Water Table Fluctuations and Observed Behaviour at Site PI2s: Dependence upon Storativity of Model Domain. For $S_y = 0.05$ or 0.4, $k_h = 0.050 \text{ m/day}$. For $S_y = 0.1$, a: $k_h = 0.005 \text{ m/day}$; b: $k_h = 0.500 \text{ m/day}$
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Figure 6.57  Wetland Water Table Elevations above Ground Surface on 10/1/1994, Simulated with the Calibrated Transient Model, Aquifer Permeability = 0.500 m/day
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Figure A2  Stage-Discharge Relation at Gauging Station C5
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Plate 2  Looking Northwards from St. Dennis Crown across Goss Moor. On the Western Side of the Wetland can be seen the Disused Railway Embankment which forms the Downstream Boundary of Goss Moor’s Surface Catchment. Castle Downs and Belowda Beacon can be seen on the Northern Side of the Wetland.