THE UNIVERSITY OF HULL

HYDROLOGICAL RESPONSES TO MOORLAND LAND-USE CHANGE

being a Thesis submitted for the Degree of

DOCTOR OF PHILOSOPHY

in the University of Hull

by

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September 1986

SUMMARY

Summary of Thesis submitted for Ph.D. degree

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on

HYDROLOGICAL RESPONSES TO MOORLAND LAND-USE CHANGE

Most documented investigations of the effects of land-use change on hydrological systems have considered the modification of forest areas. In this thesis, a headwater area in the North York Moors is used to examine the consequences of maintaining a land management regime which has received comparatively little observation in this context: controlled heather burning (muirburn). The effects of coniferous afforestation are also evaluated for selected variables. Particular attention is given to the responses of soil moisture and evapotranspiration and the relationship between these two components.

Simulated soil moisture deficits derived from empirical models are tested against measured values. Predictions based on Penman-Monteith evapotranspiration and 'layer' moisture deficits, along with an optimised soil-drying parameter, were found to simulate observed conditions most closely. A land-use change from open heather moorland to burnt ground promoted reductions both in evapotranspiration levels, especially at potential demand, and in moisture deficits. In contrast, following afforestation, deficits were maintained or enhanced throughout the year, with higher moisture losses to interception than found under heather, due to the higher aerodynamic resistance of the latter. Predictions of actual evapotranspiration, determined from soil moisture models, were generally found to be reliable estimates of those 'observed' from the moorland water balance.

Antecedent catchment conditions and storm characteristics were used in analysis of runoff distribution over time, quantified in terms of 'unit hydrographs' and linear regression models. Land-use effects were manifested most significantly in a doubling of hydrograph peak discharge following muirburn, the lower measured soil moisture deficits under a burnt catchment rendering more water available for storm runoff. A secondary, underlying control, that of a slower response from a wet catchment, lent support to evidence for the existence of variable source areas.

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ACKNOWLEDGEMENTS

I would like to thank Dr R.R. Arnett for supervising the project and maintaining encouragement throughout. Thanks also go to Drs. J.S. Pethick and K.S. Richards of Hull University Geography Department, and to Dr. R. Brown of the North York Moors National Park, for providing discussion and guidance.
R.J. Aspinall, M. A. Fullen and the staff of the North York Moors National Park assisted with field work, while the Egton Estates Company arranged and managed the muirburn.

My thanks are also extended to the staff of the Computer Centre and Main Library of the University of Birmingham, for allowing my use of their facilities during the later stages of completion of the thesis; to Mr J.B. Fisher who reproduced the illustrative material and to Ms. J. Finn and Ms. J. Jones who, respectively, commenced and completed the typing.

I am grateful to the Director of the Institute of Hydrology for supplying automatic weather station data. The project was funded by a University of Hull Research Scholarship (Faculty of Science).

Finally, I thank David, for his unfailing moral support from beginning to end.

CHAPTER 1

INTRODUCTION

Complete understanding of man's impact on hydrogeomorphological systems and, ultimately, of the controlling processes, remains a prerequisite to acceptable land management practices. The importance of land-use planning and control is examined in the present report in terms of its role in moorland hydrology. Man's activities through animal grazing, and controlled and accidental vegetation burning in moorland environments have been significant influences on the general decline of organic surfaces through erosion, states from which some areas may never fully recover. Accelerated channel formation, runoff and soil erosion are now characteristic of several localities in the region used for analysis here, the North York Moors National Park, particularly as a result of the drought of 1976. Losses to agriculture and forestry are also prevalent in this area, where about one-fifth of the moorland zone of 1951 has been reclaimed (Parry *et al.*, 1981). The British moorland environment, important for ecological and recreational reasons is examined here with respect to existing land management systems and their implications for the hydrology of a 'dry' open *Calluna* moorland.

Investigation of the hydrological repercussions of land-use change has formed the basis of a profusion of research projects, the 'experimental catchment' approach being of paramount importance in this context. Justification for some of these studies is questionable, however (Hewlett *et al.*, 1969; Ward, 1971) and, inevitably, this extensive range of enquiries has promoted a variety of conclusions, and discrepancy and ambiguity remain. Such a multitude of experiments has arisen from the concern that results are not transferable between environments. This criticism of catchments as experiments is largely unwarranted, however, since they are not designed to depict a complete environmental system, but rather to concentrate on the behaviour of specific factors which can be varied under experimental control (Church, 1984). With increasing pressure for utilisation of upland environments in Great Britain this type of study is as relevant now as in the late nineteenth century when many of these experiments began, and despite an increasing emphasis on techniques such as model simulations and mathematical syntheses of input-output relationships for prediction, catchment studies remain valuable both in their own right and for the provision of data for modelling applications.

In terms of hydrology, the 'complete system' view has gained recognition over recent years, although for practical purposes many of the methods of hydrological analysis relying on black-box, time-invariant and linear assumptions are still applied. The present study attempts to integrate the geographical approach to hydrology, typically that of analysing interactions within and between systems, and that of the engineer, concerned with the application of immediate results to practical problems (Ward, 1975), by applying some of the engineer's tools in a geographical context.

Most work on the hydrological consequences of modified rural land-use patterns has concentrated, particularly in Great Britain, on the effects of afforestation or deforestation and even during the most recent years comparative studies have focussed upon forest/grassland combinations. By contrast, natural medium-height species have been largely neglected in this respect, so that comprehension of water use under moorland vegetation is relatively poor and the full hydrological implications of moorland reclamation schemes have yet to be realised. These crops may, in specific ways, prove more difficult to monitor than taller vegetation, for example, in the physical measurement of interception, but the results of their removal or substitution remain similarly important.

Further, typical expositions on the relationships between vegetation and hydrology have centred around the effects on runoff. Simplified and restricted interpretations of controls on runoff proposed by Horton (1933) have been replaced by more realistic and more widely applicable theories of dynamic source area runoff generation, saturated overland flow, and subsurface features and processes, incorporating analysis of potential flow gradients and hydraulic conductivity. Although the consequences of land-use change for runoff response are incorporated in the present study, the impact on other hydrological components is also emphasised. In particular, the potential significance of the soil moisture and evapotranspiration terms of the catchment water balance is acknowledged and the implication of the magnitudes of these variables is examined for moorland hydrology.

By means of excesses and deficits, the soil moisture regime acts as a significant control in determining land-use type and distribution (Thornthwaite and Mather, 1955) whilst, through man's intervention, involving irrigation or land drainage programmes it is also one of the most easily varied terms of the water balance. Successful evaluation of a valid evapotranspiration term reaffirms the probability of constructing a reliable water balance, since this is perhaps the most difficult variable to estimate accurately. Evaporative loss may be controlled by man in order to alter water yields, either by modification of prevailing land-use, or more directly through specific application of chemicals. The importance of these two water balance components, soil moisture and evapotranspiration, and the relationships between them form a major focus of the dissertation.

Maintenance of experimental control in comparative studies is a fundamental requirement for definition of the essential effects of physical change. This stipulation is more difficult to fulfil under field conditions than in a laboratory environment and calls for careful experimental design. Paired catchment experiments facilitate calibration although problems of comparability may not be easily resolved. Complications due to heterogeneity are less prominent for a single catchment approach, especially in a small basin, although the need for control is equally important for results to be conclusively related to vegetation cover type. Variations in any further attribute of the experiment must be eliminated or accounted for in interpretations, whilst, 'carry-over' effects, treatments applied to one experimental unit affecting observations made on another unit, should be either prevented by physical control or explained in experimental analysis (Cox, 1958). Thus, for example, the effects of vegetation replacement may vary from region to region in accordance with age of crop or previous land-use.

In an attempt to quantify the little-documented hydrological catchment responses to modification of medium-height upland vegetation, this dissertation examines the normal management practice of controlled heather burning, 'muirburn'. A single watershed approach is adopted, wherein open heather moorland conditions are compared with those of the same area following burning. The hydrology of a headwater region of a moorland river valley is interpreted in terms of surface and subsurface runoff controls, soil moisture characteristics, actual evapotranspiration and the water balance of the area. The subsurface moisture regime and evapotranspiration/soil moisture relationships of the forested zone of the study area are analysed for comparative purposes. The area is physically homogeneous in all respects except land-use, and is chosen such that the results have implications for wider areas of the North York Moors.

As well as being a study in its own right, the research project makes a contribution to a larger moorland research programme co-ordinated by North York Moors National Park personnel. The longer-term study, initiated in 1976 from concern with a declining peat cover, has attempted to determine the environmental consequences of current moorland management practices, and to suggest alternative management schemes wherever necessary (North York Moors National Park, 1979).

The current project is intended to provide definition and explanation of the hydrological regimes characteristic of managed moorland, through the establishment of an intensive monitoring programme on an instrumented micro-catchment, using environmental modelling where possible. The effects of controlled heather burning and coniferous afforestation on soil moisture regimes are considered in relation to the potential consequences for water yield and water use. The significance of spatial and temporal patterns of moisture characteristics for the rainfall-runoff conversion is examined in accordance with current thinking on runoff generation, and differences in subsurface flow concentrations under these covers are investigated. Although the study is concerned with 'plot-scale' comparisons, no attempt is made to analyse detailed moisture fluxes either in the soil profile or through the plant, since these smaller scale investigations require detailed monitoring systems and are less relevant to a broad hydrological catchment study.

Sensitivity of catchment response to medium-height vegetation removal is also quantified in terms of its surface runoff expression. Analytical tools are used which both compare runoff regimes prior to and following catchment alteration, and account for changes in controlling factors other than that of land-use. Special attention is paid in this respect to changes in the distribution of runoff over time: the flood hydrograph. In exposing land-use/runoff relationships and interpreting hydrograph changes, it is necessary to recognise variations in physical controls by storm and antecedent catchment conditions, some of which are vegetation induced. Similarly, catchment state is an important control over relationships between gross rainfall and runoff volumes, which are analysed here through ratio values.

In aiming to construct an accurate annual water budget for the moorland area, the study represents an attempt to account for all water entering the moorland headwater, routes taken and amounts leaving. Particular aspects worthy of study include the resolution of interrelationships between separate components of the water balance and interpretation of the water balance equation through physical controls on plant water use. 'Observed' values of evapotranspiration, derived by difference from the water budget, provide effective indices with which to assess the reliability of model-predicted estimates.

The two subsequent chapters of this dissertation describe, respectively, the physical features of the area selected for study in relation to the remainder of the North York Moors region, and the methods and instrumentation appropriate for acquisition of data needed to implement the proposals outlined above. Included in Chapter 2 is further description of the experimental approach adopted and reasons for catchment selection. Experimental design, monitoring programmes and sampling schemes employed (both continuous and discrete) are discussed in Chapter 3, along with the merits and constraints of available measurement techniques in relation to experimental requirements and the physical characteristics of the area. Instrument design is described and methods and problems of equipment installation, maintenance and operation are reviewed, although discussion of data analysis procedures is left to individual succeeding chapters.

Following consideration of the traits and applicability of the two main categories of soil moisture model, theoretical and empirical, in Chapter 4, those selected for use in the present study are discussed separately in terms of their structure and operation. The 'drying curve' and 'root constant' concepts are introduced and the controversy surrounding the soil moisture constant 'field capacity' is outlined. Model predictions of soil moisture deficit, computed using standard and optimised drying specification parameters, as well as different types of evaporation and soil moisture data, are assessed both graphically and quantitatively through a calculated error term. Timing of predicted summer runoff is evaluated with respect to observed flood periods.

In Chapter 5, changing runoff responses are analysed. A development history of runoff concepts precedes a review of existing evidence for modified runoff characteristics consequent upon land-use change. Subsurface flow responses are evaluated and spatial variations in soil moisture distribution within and between each land-use plot are investigated with particular reference to theories of variable source areas. Potential methods of approach to flood evaluation and forecasting are considered, prior to analysis of land-use induced changes in temporal storm runoff distribution, using a deterministic response model, the 'unit hydrograph' for pre- and post-burn storms. Provisional examination of rainfall-runoff ratios precedes a fuller analysis in relation to the complete water balance in Chapter 6, which consolidates and develops the themes of the preceeding two chapters. Physical controls of the rainfall-runoff conversion are examined in more detail in this chapter, and a general water balance equation is derived for the moorland. Particular attention is paid to the evapotranspiration term of the equation, this component being used to determine the accuracy of values of evaporation predicted by the soil moisture models employed in Chapter 4.

The concluding chapter incorporates an interpretive summary of the findings of the dissertation and tentative conclusions are drawn regarding the implications which the results may have for future moorland management regimes. Suggestions for extension of the work are outlined in relation to the scope of the current investigation. In considering documented work, a schematic approach is applied throughout the report with the findings of this dissertation being discussed in conjunction with previous studies in the context of each aspect of the analysis. Existing literature is therefore reviewed in terms of its relevance to the subsystem under discussion, on a chapter by chapter basis.

CHAPTER 2

DESCRIPTION OF STUDY AREA

The selection of a field site for hydrological investigation requires appreciation of a combination of factors in the context of chosen experimental approach and hypotheses. Not only are physical and hydrological catchment characteristics important, especially in terms of the regional representativeness of the site, but considerations of site access, project expenditure, and land ownership and control must also be made. In general, 'the selection of the study area will normally reflect a compromise between the ideal and the expedient' (Ward, 1967a, p.498).

Data for the present study were collected from a 5.6 ha headwater tributary of Glaisdale Beck, draining Egton High Moor and referred to as Wintergill or Egton Catchment, in the central part of the North York Moors National Park (G.R. NZ 762015, Fig.2.1). The physical characteristics of the site, discussed in the present chapter, and its instrumentation, described in Chapter 3, are shown in Figure 2.2. The study area encompasses a section of open heather (Calluna) moorland with an adjacent strip of coniferous woodland, the latter known as Wintergill Plantation, planted approximately twenty-five years ago (Plate I). The Calluna moor is subdivided by a peat track into an area of relatively old heather (twenty-five years) and a smaller area of somewhat younger vegetation (eight years old) to the north of the track. Grazing activity comprises mainly grouse and occasional sheep. Total relief of the area is 15.95 m while measured slope angles range from 0.5^{o} to 11.5^{o} on the moorland, and 3.5^{o} to

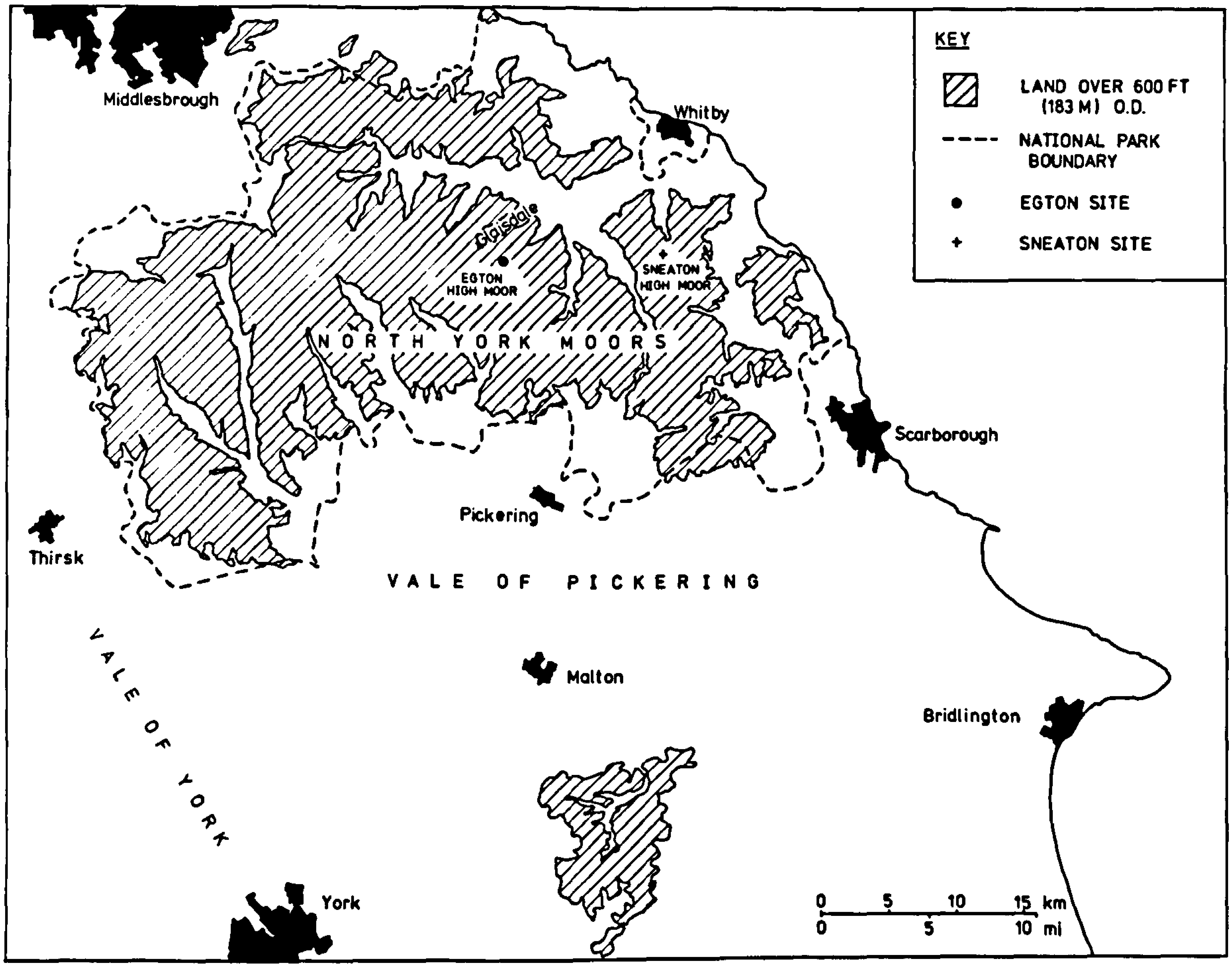


Figure 2.1 The North York Moors National Park

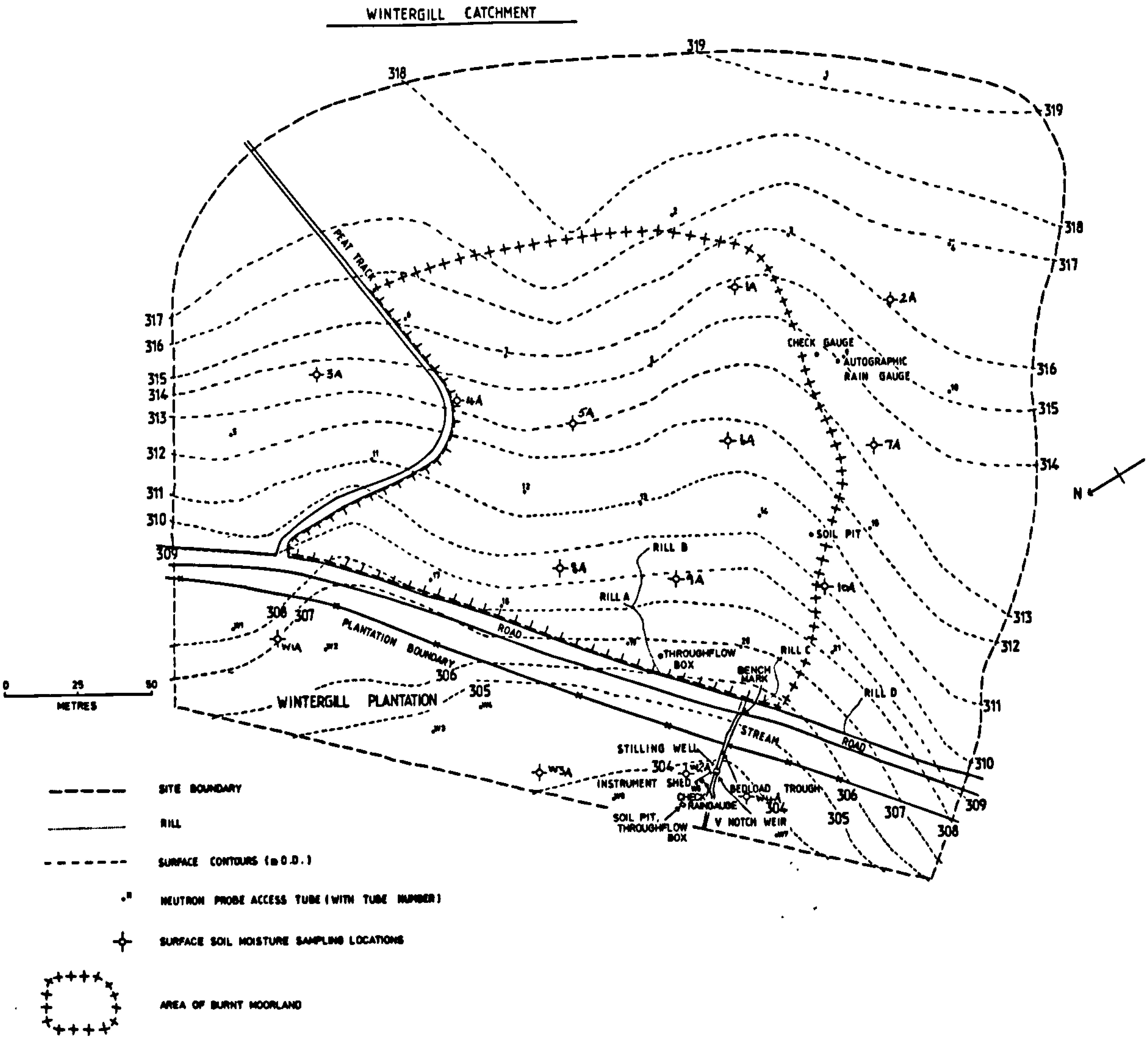


Figure 2.2 Relief and Instrumentation at the Egton Site (Wintergill)

Plate I Egton Catchment
Looking due west towards Wintergill Plantation

13.5° in the wood. A number of small rills feed the main stream channel which passes, via a culvert, under the boundary road. The rills sustain no surface flow during prolonged dry periods.

2.1 SITE SELECTION AND SCALE OF STUDY

This particular field site was chosen for study for a number of reasons, each of varying degrees of importance related to the aims of the dissertation previously described. The central purpose is to ascertain the impact of changing a surface vegetation cover on soil moisture status, evapotranspiration, and surface and subsurface runoff characteristics. Water balances are constructed largely to verify the accuracy of measured or calculated hydrological variables. No attempt is made to hypothesise about the implications of the results for wider moorland drainage basin activity and relationships, in hydrogeomorphological terms, but rather, fundamental responses to immediate land-use modification are evaluated.

Small plot-scale experiments would accommodate these intentions, but scope for wider application of the results and investigation of spatial interactions would be limited. Instead, an area more appropriate to practical scales of controlled heather burning was selected. This allows replication of monitoring sites over a manageable area, while also reducing problems of heterogeneity, found in even small basin studies. The Egton High Moor experimental area is homogeneous in terms of geology, soil type, and vegetation within each complex. Although hydrological relationships are not defined here on a drainage basin or regional scale, but rather over one part of a drainage system, the unit of study is large enough to render its wider representativeness a significant factor. In this respect, the physical features of the study site are typical of the North York Moors as a region. The specific objectives of the study therefore render a 'subcatchment' or stream-head scale more appropriate than either a small plot or a full catchment experiment. The former, plot-scale experiments, are more suited to highly controlled, agricultural investigations, while drainage basins are conducive to integrated studies of hydrological transitions and paths of movement. The present study is allied to a large lysimeter experiment, although differs from the latter in that it is confined only by a base seal. Comparative studies at this scale of operation include those of McCaig (1979) for a 4 ha pipeflow stream-head type; Finlayson (1977) for a 10 ha stream-head, carrying ephemeral flow in small depressions, on East Twin Brook, Somerset; and subsurface flow monitoring by Anderson and Burt (1977a, 1978) on hollow-spur hillslopes in the Bristol area.

More specifically, as an objective of the study is the construction of water budgets it is important to ensure that no water is being unaccountably lost from the system via subsurface routes. Catchment sealing is therefore a necessary prerequisite for study, and this is provided at the Egton site by an impermeable clay layer, discussed further in this context in Chapter 6. Determination of moisture fluxes into and through the clay horizon are best derived _in situ_ from a series of measurements of soil moisture potential, using for example, a bank of tensiometers, although for the purposes of the present investigation, analyses by the Soil Survey of England and Wales as reported in the literature (Carroll and Bendelow, 1981) are regarded as satisfactory.

An additional experimental requirement is the existence of perennially flowing surface water to enhance the possibility of measuring subsurface flow (Toebes and Ouryvaev, 1970), as separation of dry-weather baseflow from perennial surface water flow can be related to theories of subsurface movement. Although surface water discharges encountered in the study area are comparatively low, periods without streamflow were expected to be short, if not completely absent.

In practical terms, the Egton site was advantageous in that a programme of instrumentation had previously been initiated as part of a study by Fullen (1981). A natural siphon recording raingauge with associated Meteorological Office check gauge, a Munro water-level recorder and a streamwater sampler were already installed on-site. Problems of site access, data collection and site maintenance were anticipated as minimal for the site, these being important logistical considerations when an intensive sampling programme is implemented over a short study period of one to two years. Finally, co-operation of the land-owner, gamekeeper and National Park Authorities, and a desire on their part to burn the moorland vegetation in this area were positive influential factors during catchment selection.

2.2 CATCHMENT CHARACTERISTICS

2.2.1 GEOLOGY

The east/west trending Cleveland Anticline, an 'inversion structure', being initially a basin which was uplifted during Jurassic and Cretaceous periods and later folded (Kent, 1980), constitutes the structural basis of the North York Moors as a region. Lithologically the area is underlain largely by beds of the Great and Inferior Oolite Series of Middle Jurassic age, during which time, periods of marine deposition alternated with deltaic environments. These beds are surrounded by progressively older formations to the north and west, and younger to the south. The major valleys of the moors form a series of inliers comprising Lower, Middle and Upper Lias beds (Lower

Jurassic). The study area itself, as a headwater region of one of these valleys, is underlain by Middle Jurassic (Lower Oolite) beds consisting largely of sandstones and shales, these lithologies constituting the Estuarine Series (Kent, 1974), renamed the Deltaic Series by Hemingway (1949), and further altered in accordance with Hemingway and Knox (1973), as discussed below. Low angles of dip encountered in these beds partially account for the low slope angles of the study catchment. The area thus typifies the underlying geology of the Moors as a whole.

The base of the Middle Jurassic of the study area is marked by marine beds (Eller Beck Formation) and an unusual outcrop has been identified at 0.8 km (0.5 mi) north of the study area (Fox-Strangways et al., 1885; Hemingway, 1974). This section deviates from the normal sequence of shales-with-ironstone under sandstone, since it comprises calcareous sandstones, limestones, silts and thin mudstones beneath an arenaceous unit which includes an oolitic ironstone (Knox, 1970; Hemingway, 1974). Fox-Strangways et al. (1885) described an ironstone seam occurring at 30.5 m above the Eller Beck Bed as the possible representation of the Millepore Bed in the Wintergill area; this bed reflects a further period of marine deposition during which clays, silts and sandstones were deposited. The overlying Grey Limestone Series or Scarborough Formation is lithologically variable, but generally involves a narrow sequence of limestone, sandstone and shale which, although widening over Egton High Moor (Hemingway, 1958), is not included by the boundaries of the study site.

Although the central area of the North York Moors was generally unaffected by the last (Devensian) glaciation, valleys on the north-eastern edge were encroached upon by ice which reached westwards as far as the Leaholm moraine (G.R. NZ 7507) (Penny, 1974), about 5 km north of the study area. Egton Grange and Glaisdale contain the nearest drift deposits, while periglacial features are apparent in the south eastern area of the North York Moors (Dimbleby, 1952). Peat deposits provide probably the only evidence of post-glacial deposition in the Egton High Moor area, and the study site is incorporated in the belt of plateau bogs defined by Eyre (1973) as extending from Loose Howe (G.R. NZ 702008) to Three Howes (NZ 794012).

Catchment bedrock is overlain by a clay layer, at least 1 m thick in parts and, for which, ostensibly,there are several possible origins. Perhaps the most plausible explanation for its presence, however, is in situ weathering of the substrata. As the area lies just beyond the margins of the last glacial advance, and included erratic material is absent, a glacial origin is less probable, whilst a fluvial source would necessitate an extended drainage system, evidence for which is totally lacking.

2.2.2 STRUCTURE AND DRAINAGE

Large-scale drainage patterns of the North York Moors have been heavily influenced by the Cleveland Dome structure, from which all major river valleys except the Esk now flow radially. A number of theories have been proposed to explain present-day drainage, including renewed drainage with peneplain uplift (de Boer, 1974) and fluvial erosion of a series of surfaces originally formed by marine erosion (Palmer, 1973). As the area considered by this dissertation comprises a headwater region of one of these radiating river systems, Glaisdale Beck, the study is important with respect to both small-scale hillslope hydrology and denudation, and to processes operating downstream.

2.2.3 SOILS

Although the peat of this region has been termed 'hill-peat' by Fox-Strangways et al. (1885), its shallow depth, up to 25 cm, and lower acidity place the soil more appropriately, in the 'non-raw peat' category as described by the Soil Survey of England and Wales. Specifically the soils of this area are classified as the Onecote Series within the cambic stagnohumic gley subgroup (Carroll and Bendelow, 1981). All soils in this subgroup possess a humose or peaty topsoil up to 40 cm thick with reduced B horizon development. This soil type is the most common soil group throughout the Moors, covering 26% of the area studied by the Soil Survey. Much of the northern part of the Moors region is capped by raw peat soils, corresponding roughly to the outcrop of Calcareous Beds of the Grey Limestone Series, the study area being proximal to the boundary of this second major soil group. Soils of the primary moorland valley systems comprise non-calcareous pelosols on the upper slopes, with brown earths on the side slopes and stagnogleys on the valley floors (Carroll and Bendelow, 1981).

Typical soil profiles for the study site are shown in Table 2.1. The moorland description is based on a profile surveyed near the centre of the moorland area (G.R. NZ 761014), while the woodland profile represents the soil towards the boundary of the site (G.R. NZ 761015) (Fig.2.2). The results of physical and chemical soil analyses are summarised in Appendix I.

The accumulation of conifer needles under the woodland has led to the formation of a well developed litter layer since the time of plantation, and some litter decomposition is evident. Although differences in soil porosity may be apparent even between tree species

Depth (cm)		Horizon	Description
Moorland	Woodland		
0-2.5	0-3	L	Heather litter (moorland) Conifer needles (woodland)
2.5-25	3-11	Oh	Black, 10YR 2.5/1, peaty horizon. Mainly fine, fibrous roots; a few woody roots.
25-39	11-18	Eg	Dark brown, 10YR 3/3. Sandy texture. Medium angular blocky structure; slightly sticky, slightly plastic. Sharp, smooth boundary with Oh horizon. Fewer roots.
>39	>18	Bg	Yellowish brown, 10YR 5/4. Clayey texture. Medium-coarse angular blocky; slightly sticky, slightly plastic. Abrupt, smooth boundary. Evidence of gleying.

TABLE 2.1. Soil Profile Descriptions

within a period of twenty to forty years (Ovington, 1956), complete development of a forest soil profile may take several hundred years (Dickson and Crocker, 1954) and, apart from variations in depth of the Oh (peat) horizon over the study catchment, differences in field characteristics of each horizon under moorland and woodland are limited. Reduced peat depth under woodland may be a consequence of plantation preparation, although evidence of profile disturbance, for example in terms of horizon mixing, is generally minimal.

'Available water capacity', definable as the amount of water stored between 'field capacity' and 'permanent wilting point', and discussed further in Chapter 4, is described broadly by the Soil Survey for the Onecote Series as 'very large at the surface, and large below; retained water capacity is very large throughout' (Carroll and Bendelow, 1981, p.109). Hodgson (1976) included this soil type in wetness class VI, this being the wettest class in the range. More specific values given in the literature include a mean figure of 17% of soil volume for clays, and 30% for peat (Salter and Williams, 1965). Doorenbos and Pruitt (1977) quoted a maximum available capacity of 190 mm m^{-1} soil depth (19%) for a silty clay at field capacity. Land capability for the area is described by the Soil Survey as unimproved poor grazing, potential use mainly grass, with wetness a limitation (Carroll and Bendelow, 1981).

2.2.3.1 Peat Development

The history of peat evolution on the North York Moors is the subject of some debate. Palynological and archaeological evidence indicates that the start of the Atlantic period (7500 years B.P.) marked the earliest period of bog growth. A combination of a warmer, wetter oceanic climate and man's influence via burning of vegetation and grazing animals (Cundill, 1972; Dimbleby, 1962) probably resulted in more rapid runoff from slope areas. Forest clearance during the period 4000 to 5000 B.P. is one explanation for the initiation of peat formation at this time, although alternatively, peat could have expanded from local masses already in existence (Cundill, 1972). Further deforestation occurred after 3500 B.P. (Dimbleby, 1962) and by 2500 B.P. a wetter, cooler climate had accelerated leaching rates which have continued since that time. Present-day drier climates, animal grazing and heather burning have combined to halt peat accumulation or even induce peat loss over much of the moorland, subsequent erosional activity accentuating such removal (Arnett, 1978, 1980; North York Moors National Park, 1979). It is this sense in which the environmental and economic repercussions of man's activities are investigated by moorland research.

The nature of the substrata in this area lends itself particularly to hydrological studies. A peaty layer underlain by an impermeable soil horizon is especially significant in the context of surface and subsurface water flow. An increase in hydrostatic pressure, or reduction in soil resistance at such an interface may result in soil pipe initiation, although certain other conditions (a source of water in the profile, a means of outlet, sufficient hydraulic gradient, an erodible layer above the impeding horizon and soil cracking or swelling) may also be important (Jones, 1981). Differences due to semantics often make identification of such features questionable, and soil cracks, voids or macropores may be regarded as conduits for channel flow on a small scale. Further attention is given to subsurface runoff in the following chapters.

2.2.4 VEGETATION

Brief reference has already been made to long-term soil and vegetation changes of the North York Moors area. Present-day management practices tend to complicate further a situation which has resulted largely from anthropogenic and climatic influences. Effective sheep and grouse management entails a cycle of heather burning and recolonisation on the moors in order to maintain young, nutritious plant shoots, while much of the remainder of the moorland region has been extensively afforested during the last 60 years, so much so that few areas now remain available for planting. The Egton Moor catchment allows examination and close comparison of some of the effects of both of these major land-use changes.

Moorland vegetation communities have been the subject of a number of classification schemes such as those of Elgee (1914) and Gimingham (1960). However, management intervention, especially in moorland areas, results in decreased floral diversity and may also obscure environment/vegetation relationships (Bannister, 1976). Such prevailing factors render the placing of an observed community into a previously determined classification rather arbitrary, if not inexpedient. No rigorous attempt, therefore, is made to define the vegetation of the study area in this way. A background survey of species present in the area is, however, a prerequisite for a study of this type, the nature of the vegetation both influencing and being determined by such underlying factors as soil type and water status. The site supports a plant community typical of the dry moorland areas, with heather (Calluna vulgaris) as the dominant species. Calluna tolerates a wide range of temperatures, exposures and soil moisture levels, although development is reduced in waterlogged conditions, and rooting may be confined to the top few centimetres of very wet substrata; the species grows best in at least moderately-drained soil. At the study site used here, heather roots penetrate the subsoil (Table 2.1).

Calluna shows a four-phased developmental process (Watt, 1955). Plant establishment and early growth occurs during the first six to ten years and a height of up to 30 cm is reached during this 'pioneer' stage. Maximum cover is attained during the next, 'building' phase, the plant reaching 30 cm to 60 cm in height. Spreading of central branches leads to gap formation during the 'mature' phase by which time the plant is fourteen to twenty-five years old. Finally, when the plant reaches between twenty and thirty-three years, during the 'degenerate' stage, new pioneer plants colonise the enlarged gap and the cycle restarts. Management aims to achieve even-aged stands of heather, while the absence of burning results in the presence of all four growth stages (Barclay-Estrup and Gimingham, 1969). The latter situation was evident on the study site prior to the muirburn of April 1981, carried out as part of this research.

Moorland sub-dominant species include bilberry (Vaccinium myrtillus) and crowberry (Empetrum nigrum), while soft rush (Juncus effusus), heath rush (J. squarrosus) and common sedge (Carex nigra) colonise poorly-drained patches. Mosses (Hypnum cupressiforme, Plagiothecium undulatum and Sphagnum papillosum) are found especially on the area of younger heather and also, particularly in the case of Sphagnum, in wetter areas. Common cotton-grass (Eriophorum angustifolium) is a minor species at this site.

Sitka spruce (Picea sitchensis) and lodgepole pine (Pinus contorta) are the planted species of Wintergill Plantation. The former is well suited to peaty soils while Pinus contorta can withstand poor, exposed soils and has been extensively planted throughout north-east Yorkshire (Eyre, 1973). Few tree species are adapted to very high water table conditions (Armstrong, 1982) although P.contorta shows improved growth over Picea sitchensis under waterlogged soils because of the former species' ability to lower the level of the water table with rooting (Coutts and Philipson, 1978). Other tree species present in the plantation are occasional silver birch (Betula pendula) and rowan (Sorbus aucuparia), with a mixed ground flora of Calluna vulgaris, Vaccinium myrtillus, common bent grass (Agrostis tenuis), wavy hair-grass (Deschampsia flexuosa , ferns (Blechnum spicant and Dryopteris dilatata), mosses (Polytrichum commune) and Pellia sp. (liverwort). Bracken (Pteridium aquilinum) has also colonised the woodland, possibly as a result of improved drainage over this part of the catchment.

2.2.5 CLIMATIC BACKGROUND

Climatic characteristics vary quite markedly over the North York Moors, both in view of the region's altitudinal variation and its extension from the coast to a distance of almost 60 km inland. Long-term climatic data for the area (Smith, 1976) are summarised in Tables 2.2, 2.3 and 2.4. Values are based on an average areal height of 172 m so subsequent height adjustments have been made for the present study site. Data from one of a pair of Didcot Instrument automatic weather stations (Strangeways, 1972) on Sneaton High Moor (G.R. NZ 880017), sited at an altitude of 265 m, and at a distance of 11.7 km (7.3 mi) from the Egton site (Fig.2.1) are included in the tables. These stations were established by the Institute of Hydrology as part of an investigation into evaporation from heather, and were

MONTH	Long-term air temperature for North York Moors (°C)	Adjusted long-term temperature for Egton (311 m) (0.6°C/100 m)	Air temperature Egton (1981) * (°C)	Adjusted long-term temperature for Sneaton (265 m) (°C)	Air temperature Sneaton (1981) (°C)
JANUARY	2.5	1.7	2.7	1.9	4.1
FEBRUARY	2.8	2.0	0.9	2.2	2.8
MARCH	4.7	3.9	4.3	4.1	6.3
APRIL	7.3	6.5	5.7	6.7	6.5
MAY	9.7	8.9	8.3	9.1	9.6
JUNE	12.7	11.9	12.5	12.1	11.1
JULY	14.5	13.7	15.8	13.9	12.7
AUGUST	14.4	13.6	15.6	13.8	13.3
SEPTEMBER	12.7	11.9	15.6	12.1	12.0
OCTOBER	9.8	9.0	7.2	9.2	6.8
NOVEMBER	5.8	5.0	5.7	5.2	6.0
DECEMBER	3.7	2.9	-0.5	3.1	1.3

* Negretti and Zambra continuous temperature recorder at 1.22 m above ground level

Table 2.2 Temperature Data: Egton and Sneaton High Moors, and Long-term Averages for North York Moors (1941-1970)

MONTH	Long-term rainfall for North York Moors (mm)*	Egton rainfall 1981 (314 m) (mm)	Sneaton rainfall (1981) (mm)	Long-term average rainfall Farndale ** (mm)	Farndale rainfall (1981) (mm)	Long-term average rainfall Silpho Moor + (mm)	Silpho Moor rainfall (1981) (mm)
JANUARY	75	43.0	44.0	100	51	83	44.8
FEBRUARY	61	84.8	33.5	82	72	66	69.6
MARCH	50	150.2	(10days missing) 163.0	65	181	53	178.1
APRIL	52	41.0	50.0	67	74	56	59.2
MAY	59	83.4	54.5	73	92	65	48.4
JUNE	55	49.4	35.5	68	46	62	42.0
JULY	71	112.9	133.0	80	132	77	112.0
AUGUST	91	61.3	50.0	108	57	90	42.0
SEPTEMBER	74	116.8	113.5	84	136	71	94.6
OCTOBER	67	97.5	130.0	88	98	73	113.1
NOVEMBER	84	63.5	67.5	128	86	102	62.5
DECEMBER	69	10.9	14.5	102	92	83	116.2
		(1.12.81-9.12.81) gauge frozen	(9 days missing)				
	Σ808	Σ914.7	Σ889	Σ1045	Σ1117	Σ881	Σ982.5

* Rate of Increase of average annual rainfall: 0.315 Rm -119mm/100m, where Rm = areal average annual rainfall (mm) giving an adjusted total of 1000 mm for Egton and 934 mm for Sneaton

** Station altitude 322 m (G.R. SE 673995) 8.9 km (5.5 mi) from Egton catchment

\+ Station altitude 203 m (G.R. SE 957946)10.2 km (6.4 mi) from Sneaton automatic weather station

Table 2.3 Rainfall Data: Long-Term Averages (1941-1970) for North York Moors and Selected Stations

MONTH	Long-term Penman potential evaporation (PE) for North York Moors (mm)	Decrease of PE with height (mm/100 m)	Adjusted long-term PE for Egton (314 m) (mm)	Adjusted long-term PE for Sneaton (265 m) (mm)	Sneaton PE (1981) (mm)
JANUARY	0	1.5	0	0	7.0
FEBRUARY	8	2.0	5.2	6.1	11.3
MARCH	27	3.0	22.7	24.2	26.8
APRIL	48	1.5	45.9	46.6	50.8
MAY	71	1.0	69.6	70.1	94.5
JUNE	80	1.0	78.6	79.1	94.6
JULY	80	3.5	75.0	76.8	95.8
AUGUST	63	3.5	58.0	59.8	93.5
SEPTEMBER	39	3.0	34.7	36.2	59.7
OCTOBER	19	3.0	14.7	16.2	27.3
NOVEMBER	3	2.5	0	0.7	11.8
DECEMBER	0	2.0	0	0	6.0
	Σ 438		Σ 404.4	Σ 415.8	Σ 579.1

Table 2.4 Potential Evaporation Data: Sneaton High Moor and Long-Term Averages for North York Moors

used in the present study to provide rainfall and evaporation data. Their specific use is discussed further in subsequent chapters. Temperature and rainfall data have also been derived from continuously recording instruments established on the study catchment itself, and mean monthly rainfall data for neighbouring Yorkshire Water Authority stations are included for comparison in Table 2.3.

Air temperatures for 1981 at the Egton site follow the height-adjusted long-term average monthly trends although normal summer maxima are exceeded by 2^{o} to $3^{o}C$, and the recorded figure for December was below normal (Fig.2.3(a)). Measured temperatures for Sneaton High Moor deviate little from expected values (Fig.2.3(b)). Annual rainfall measured at Egton (915 mm) was below that derived from the conversion given by Smith (1976), although a period of missing data and instrument operation problems may account for some of the difference. A smaller margin separated long-term and measured annual values for Sneaton High Moor. Monthly totals at Egton reflect those measured at Farndale recording station and those for Sneaton correspond to Silpho Moor records for 1981 (Fig.2.4). Some deviation from long-term averages occurs at all sites in January, when snowfall was important during the early part of the month, and in July and August, although the main departure occurs in March, with two to three times the normal amount being recorded. Monthly Penman potential evaporation totals approximate to average values for Sneaton during the early and later months of the year, although summer maxima were maintained over a longer period than expected from long-term data (Fig.2.5). Local climatic factors (wind speed, net radiation and vapour pressure) explain this difference.

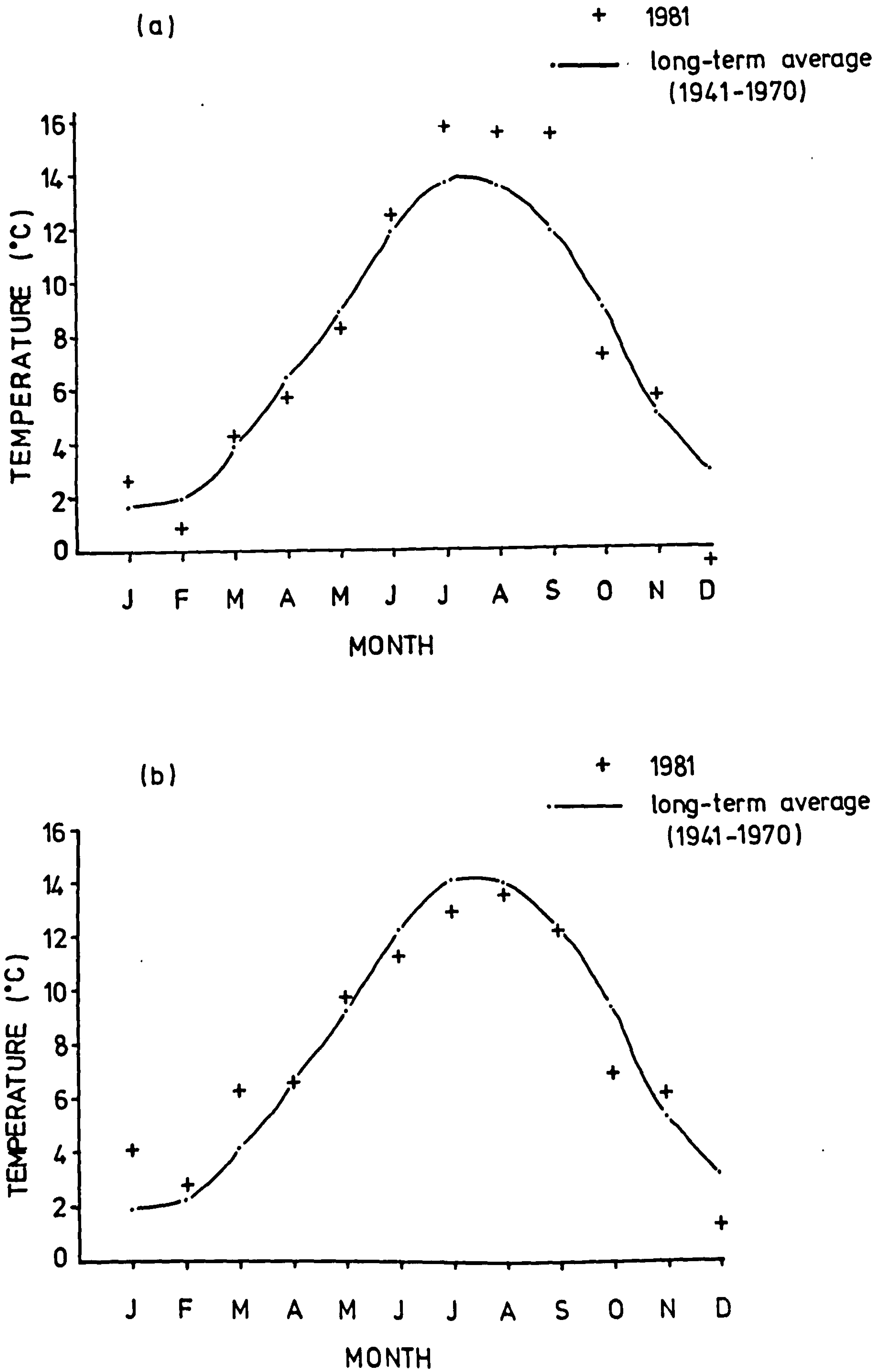
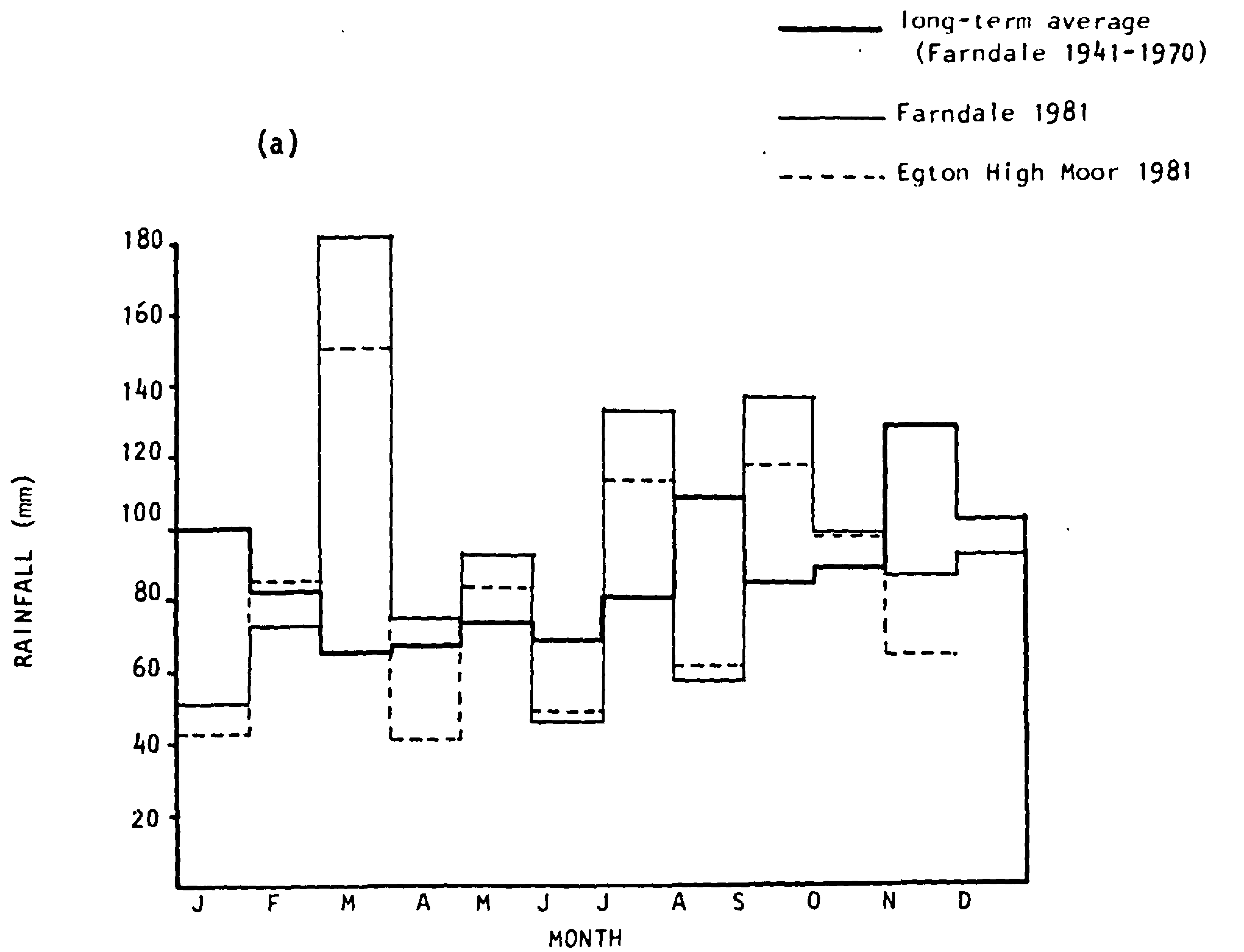


Figure 2.3 Mean Monthly Air Temperatures for 1981 in relation to long-term Averages for (a) Egton and (b) Sneaton High Moor Sites



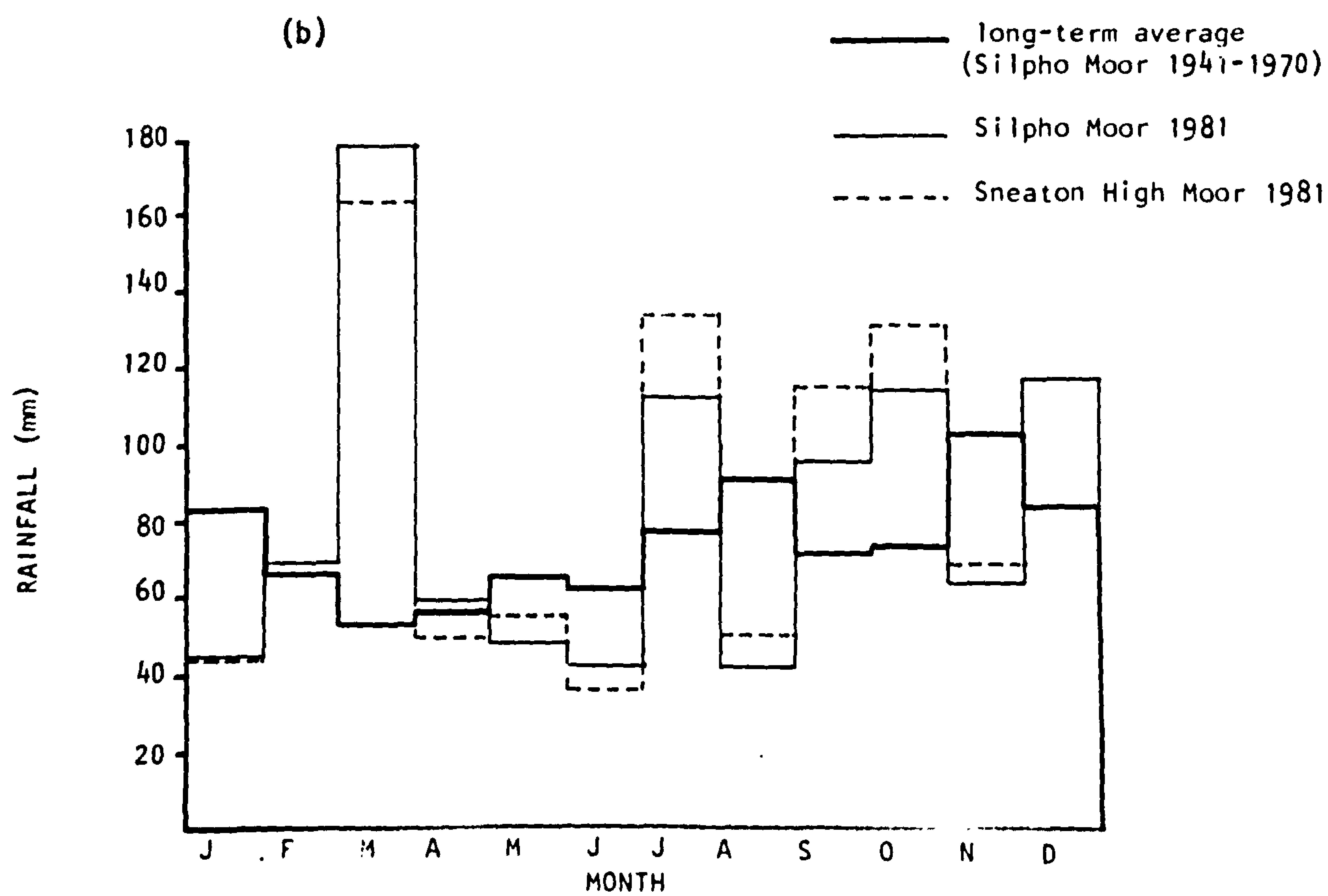


Figure 2.4 Monthly Rainfall Totals for 1981 on (a) Egton and (b) Sneaton High Moors in relation to comparable North York Moors Sites

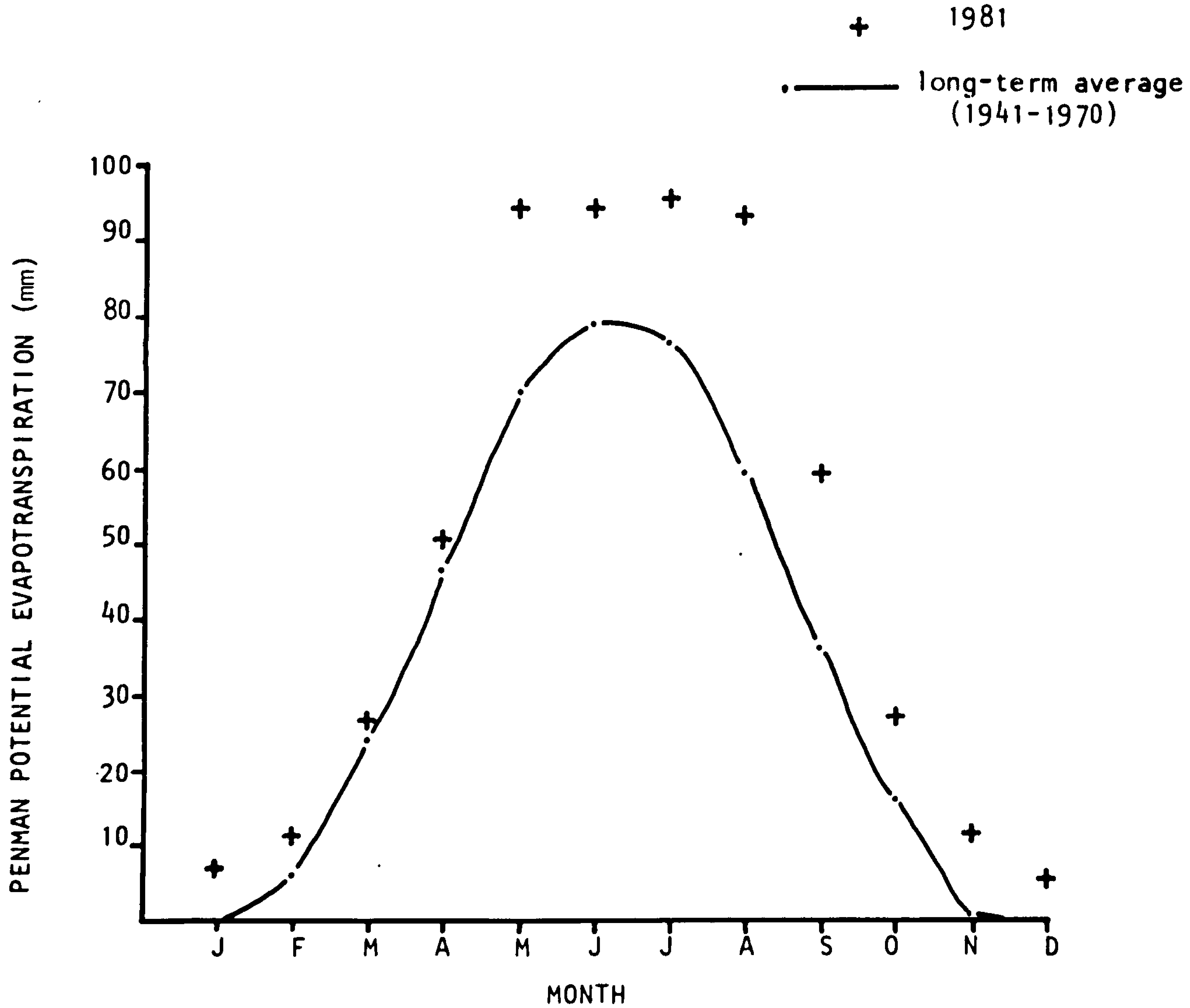


Figure 2.5 Monthly Penman Potential Evaporation Totals for 1981 Compared to Long-term Averages for Sneaton

2.3 SUMMARY

The area selected for study has potential for hydrological research not only in the context of changing land-use but also in its own right as a small peat headwater. Attention has been confined to a single unit of study with homogeneous environmental characteristics, the only major variation being that of vegetation type: a man-induced variation. The micro-catchment was chosen for its physical characteristics which are conducive to hydrological investigations; for its wider spatial representativeness; and because of the ability to examine the significance of major land-use changes within a small area. Interpreting the effects of these changes requires, at the outset, a reliable data base from which to deduce the nature of the hydrological system under operation. The means of acquiring these data are considered in the next chapter, along with reasons for choosing respective measurement techniques.

CHAPTER 3

DATA ACQUISITION AND MEASUREMENT TECHNIQUES

3.1 INTRODUCTION

Experimental sampling gives only a representation or estimate of reality and the most reliable results arise from a reproducible experiment in which the results are testable (John and Quenouille, 1977). The difficulty of obtaining precise and accurate experimental results was emphasised by Cox (1958), who warned that inherent variation within environmental systems severely limits the improvement in precision obtainable by using increasingly precise measuring equipment. It is therefore necessary to balance logistical limitations against acceptable levels of accuracy.

As indicated in the previous two chapters, field experimental conditions, in contrast to laboratory environments, are difficult to specify and control, particularly in the case of comparative studies. Experiments should therefore be designed with sufficient spatial and temporal resolution to enable identification of significant processes. Experimental timescales are particularly important for studies involving vegetation manipulation since a sufficient period is required for vegetation-induced effects to become apparent and representative, whilst in the case of vegetation removal, the effects of post-treatment recolonisation may need to be accounted for. The intention in this study is to assess the immediate, short-term effects of controlled heather burning, on a seasonal timescale, while the age of the adjoining woodland enables, more appropriately, the longer-term impacts of this vegetation cover to be determined. The experimental data collection period ran from July 1980 to March 1982, although much of the analysis and interpretation is focussed upon data acquired for 1981.

Precipitation and stream stage variables were monitored continuously and soil moisture status and subsurface storm flow volumes on a weekly basis. Supplementary surface moisture measurements were made monthly. Hourly data recorded by automatic weather stations on Sneaton High Moor were normally summarised as daily values.

3.2 PRECIPITATION

Precipitation input to the heather ecosystem was evaluated using a Casella natural-siphon autographic raingauge, supplemented by check totals from a standard Meteorological Office Mk. II gauge (Fig.2.2). The gauges, set at approximately canopy level, are unaffected by the aerodynamically induced errors associated with those sited above ground-level at grassland sites. Estimates of ground-level receipts under woodland were provided by a second Meteorological Office Mk. II gauge, although independent assessments of 'losses' to vegetation were incorporated in evapotranspiration calculations carried out during implementation of soil moisture estimation models (Chapter 4).

Gauge siting and installation followed recommendations of the Meteorological Office (1956), within the conditions of the study, with instrument checking and manual precipitation measurements being conducted on a weekly basis. Operational problems included occasional freezing of the recording gauge float, despite provision of insulation, as suggested by Kelway (1975), and loss of a few days' records due to distortion of the float chamber mechanism, again a consequence of sub-zero temperatures. Records were adequate for intermittent analyses, but measurements from the automatic weather station on Sneaton High Moor were used preferentially where continuous, daily data were required. Although these records were also used to represent gross precipitation input to the woodland plantation, a degree of error is expected to be incurred as a result of wind current effects at the woodland boundary (Penman, 1963).

Quantification of snow involves particular difficulties (underestimation by raingauges, wind effects, etc.) and numerous attempts have been made to overcome this problem, varying in complexity from the melting of snow collected by a raingauge (Rodda et al., 1976) to incorporation of remote sensing techniques (Bruce and Clark, 1966; Rodda et al., 1976). No separate measurement of snow coverage and depth was made in the present study, since light snowfalls were represented approximately by raingauge readings, whilst heavier falls tended to preclude site access and sometimes resulted in equipment malfunction, and loss of records. Adjacent monitoring stations suffered similar problems and data were lost from the automatic weather stations during late December 1981 and early January 1982.[1]

3.3 SOIL MOISTURE

Soil moisture content (wetness) may be determined either directly or indirectly, and continuously or on a non-continuous basis. Indirect measurement involves specification of an association between moisture content and a further moisture characteristic, usually soil moisture tension (potential). These relationships, however, demonstrate hysteresis, separate curves being applicable for drying and wetting phases. Measurements of both content and tension are often carried out on a non-continuous basis. Some of the more widely used soil moisture

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1 Data printouts from the automatic weather stations show several gaps in the record during this period, as a direct result of instrument failure due to freezing. Since the stations are checked fortnightly, instrument malfunction is rectified only at these intervals.

measurement techniques are outlined below, prior to a more detailed discussion of the methods adopted here.

3.3.1 INDIRECT METHODS OF ASSESSMENT

3.3.1.1 Tensiometers

These instruments are used to measure total soil moisture potential, and, on the premise that osmotic potential is insignificant in affecting liquid flow, can be used to determine matric potential. A tensiometer consists of a water-filled porous ceramic cup connected either directly to an electrical transducer, or via a water column to a vacuum gauge or simple manometer, each of which monitors pressure changes resulting from variations in soil water tension. Measurement relies on water in the porous cup reaching hydraulic equilibrium with the surrounding soil water, although, in some instances, hydraulic resistances in the soil/tensiometer system can lead to a lag time between change in moisture potential and reading on the gauge.

Instrument detection is restricted to low soil moisture tensions, in the range 0 bar to 0.8 bar, that is, below pF 2.9, pF being 'the common log. of the height of a water column in centimeters equivalent to the soil moisture tension' (Schmugge _et al_., 1980, p.965). At higher tensions, the air entry value of the ceramic cup is exceeded (Schmugge _et al_., 1980). Most of the anticipated range of moisture contents for the study site would therefore be measurable by this means. Certain types of tensiometer extend the normal tension range, and these include the pressure transducer type used by Cooper (1980) for measurement to a depth of 7 m, and that developed by Peck and Rabbidge (1969) which is claimed to be able to monitor potentials over the full range expected in agricultural and hydrological work. Theoretically, this instrument, which incorporates an aqueous solution of polyethylene glycol and a semi-permeable Visking membrane, is able to monitor either total or matric soil water potential.

Field use of tensiometers, however, involves a number of practical problems:

i) The instruments require frequent maintenance. Routine checks must be made for system leaks and disturbance, and air bubbles must be 'purged' on a regular basis.

ii) After installation in poorly drained conditions, water may tend to run into the hole created for the porous cup, although material such as powdered clay can be used as an infill seal to overcome this problem (Webster, 1966).

iii) Frosty conditions may cause the mercury column to break, while ground freezing can result in the development of cracks in the porous pot, rendering the instrument inoperable. Problems caused by sub-zero temperatures can be overcome, however, by using an ethylene glycol-water solution in the tensiometer system (McKim et al., 1976).

iv) A further disadvantage indicated by Towner (1981) may be the need, in swelling soils, to make additional measurements of the overburden pressure magnitude, in order to determine fully the in situ water content.

3.3.1.2 Electrical Resistance Methods (Porous Blocks)

Changes in soil moisture tension are measurable by concomitant tension variations in the water held by porous blocks buried at depth in the soil; measurements are conveyed to an electrical resistance meter via a pair of electrodes embedded in the block. Blocks made of gypsum are more sensitive in lower moisture content environments (0.4 bar to 19.7 bar) while porous nylon blocks are more responsive at the wetter end of the scale (Hillel, 1980a). Variations in the relationship between electrical resistance and moisture tension result from gradual dissolution of gypsum blocks, however, and this is especially prevalent in acidic layers and in soils with a high water table. Hysteretic effects both within these blocks and in the soil render them unsuitable for monitoring water content changes (Wellings _et al_., 1985), while blocks composed of inert materials, such as fibreglass, can show variations in resistance readings as a result of changes in surrounding soil solute concentrations. Further errors may result from slow responses to changes in soil water tension; an important consideration when interrelating moisture status with other hydrological variables. This lag, which may result from poor contact with the surrounding medium, may lead to gross errors for shrinking/swelling soils (King, 1968) and individual porous blocks at each depth need to undergo frequent calibration.

3.3.1.3 Thermocouple Psychrometers

These units measure total soil moisture potential by monitoring the equilibrium vapour pressure of soil water. They depend on differences between wet bulb and dry bulb temperatures creating an electromotive force, which is representative of a change in moisture potential. Measurements are restricted to those of high soil moisture tensions, generally in the range 2 bar to 50 bar, and on this basis alone, these instruments prove less suitable for the type of environment under study.

3.3.2 DIRECT MEASUREMENT OF SOIL MOISTURE CONTENT

The overall aims of the present investigation suggested that an assessment of moisture status based on water content or wetness (direct measurement) would be preferable to one based on measurements of soil water potential (indirect). Although measurements of potential would usefully complement moisture content data in a comparative land-use study, they are generally more beneficially applied in studies of plant water availability or in detailed specifications of lateral and vertical soil moisture flux patterns. The need for an extensive monitoring system and/or the nature of the soil or soil moisture properties themselves, also indicated that the indirect methods outlined above would be inappropriate. Tensiometers, particularly, require periodic maintenance, especially in an upland environment where the instruments are susceptible to frost damage. A further prerequisite both for these instruments and for porous blocks, would be their removal prior to vegetation burning, to prevent instrument damage, involving re-installation and therefore uncertainties about the effects of burning on soil moisture. Results of direct measurements are more amenable to subsequent data manipulation, both in soil moisture model implementation through calculation of soil moisture deficits, and in construction of water budgets. Two of the principal direct methods are therefore incorporated into this study; the thermogravimetric technique and neutron scattering.

3.3.2.1 Thermogravimetric Method

This technique, which acts as a means of calibration for other methods of soil moisture measurement, is incorporated into the experimental programme to investigate directly variations in moisture content in the immediate surface layers of the profile. The need for a representative and flexible sampling frame for these measurements, as well as those at depth, had to be balanced against the time required for network establishment and subsequent moisture measurement. A random design often necessitates a large number of measuring points, while stratified random sampling may be difficult to put into practice. Systematic sampling, on the other hand, is widely used (Petersen and

Calvin, 1965), leads to both better population coverage and a more representative mean value than does a random design, and, in most samples, involves only a limited and permissible degree of bias (Hammond and McCullagh, 1978).[1]

In the present study, a systematic network of surface and total profile sampling locations was established by aligning a series of surveying poles at measured distances. Fifty by fifty metre grid squares formed the basis for moisture measurement at depth (Section 3.3.2.2) while surface moisture conditions were assessed at set points within this framework (Figs.2.2, 3.1). The system allows any obvious small scale changes in physical, hydrological or vegetational factors to be accommodated into the design during establishment in the field, resulting, as far as possible, in a representative sampling network. It also enables subsequent selection of 'key' sites, should this be desirable. Precise surface monitoring locations were chosen on each sampling occasion by using a systematic rotation for each site; a spatially expandable system was thus devised, which reduced sampling variability. Referring specifically to sampling in the immediate surface layers, Hills and Reynolds (1969) set thirty to forty-five as an average sample size for areas larger than 961 m^2, while Reynolds (1970b) recommended that at least ten individual samples be collected to estimate mean moisture content to within $\pm 10\%$ at the 95% probability level.

1 Cliff (1973) denounced the idea of 'the sample' altogether, in that the statistical 'population' can itself be regarded as a sample, since it could consist of any number of values but particular circumstances lead to the evolution of a specific population.

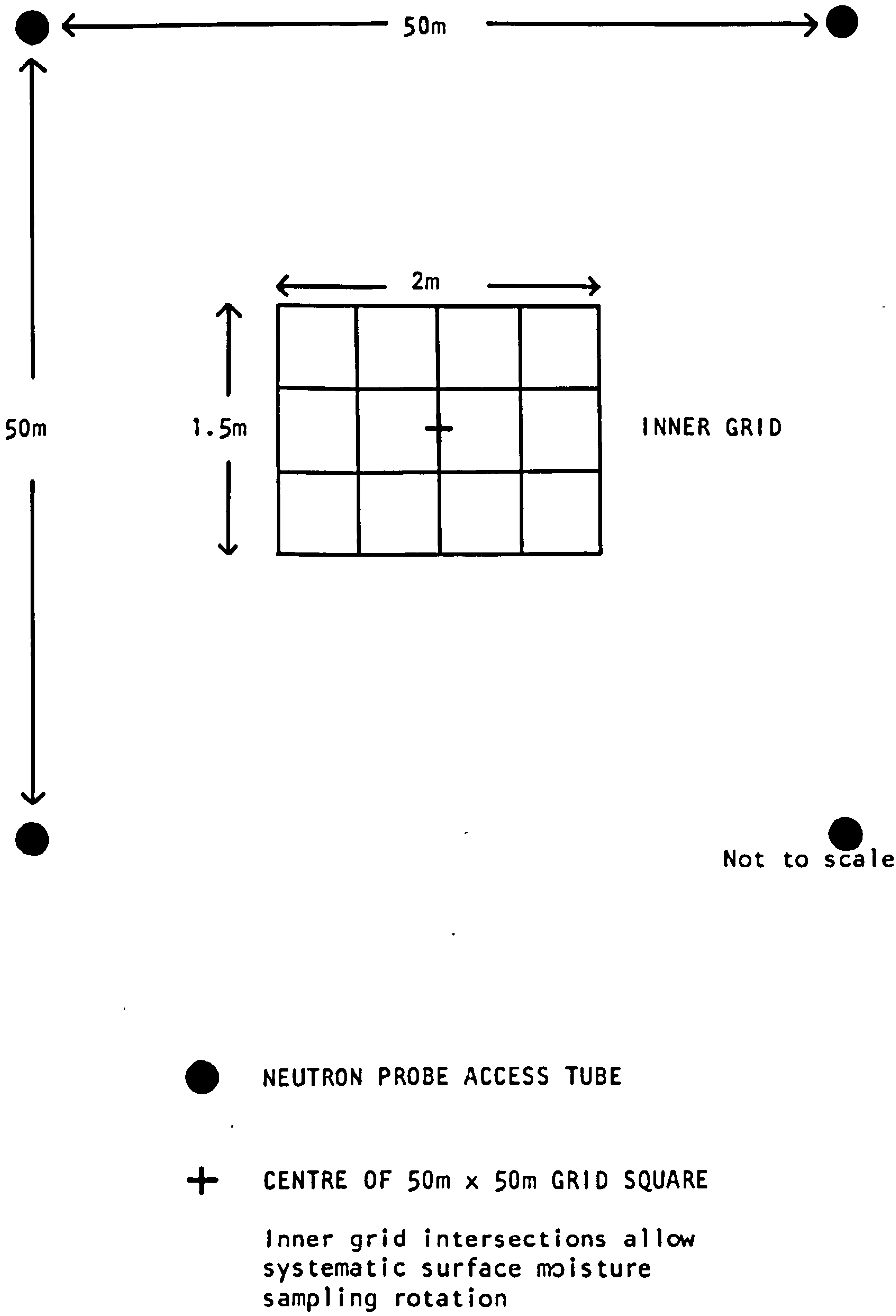


Figure 3.1 Framework for Soil Moisture Sampling

Limitations of time and resources restricted surface measurements at Egton to duplicate samples at each of two depths (0cm to 2.5 cm and 2.5cm to 7.5 cm) for fourteen sampling points over the study area (Fig.2.2).

As well as large spatial changes, variations in soil moisture are also evident at smaller scales. Sampling volumes must be large enough to reduce microscale variations and therefore surface samples of between 50 g and 100 g, and always at least 30 g, were removed for thermogravimetric analysis. A Dachnowski soil corer was employed, using a plunger to push out each sample, avoiding contact by hand as each core was pushed directly into a small aluminium container. These were then sealed in order to avoid moisture loss, and removed to the laboratory for weighing (to the nearest 0.2 g if the sample exceeded 100 g and to the nearest 0.1 g if less than this [Reynolds, 1970a]). Oven-drying was carried out at 105°C to constant weight, usually 24 hours. Samples were then cooled in a desiccator before re-weighing and subsequent calculation of moisture content. Some organic material oxidises at 50°C and certain clays may still contain structural water on drying at 105°C (W.H. Gardner, 1965). These are mainly the smectitic clay minerals, however, largely inapplicable to the present soil series (Carroll and Bendelow, 1981), while errors from these anomalies generally become important only when comparing different soil types.

Moisture content is calculated as a percentage of wet weight from the following expression:

$$\text{moisture content} = \frac{W_w - W_d}{W_w} \times 100\% \qquad \text{Eq. 3.1}$$

where:

W_w = wet (field) weight of soil (g)

W_d = dry weight of soil (g)

This index, unlike the more frequently used expression on a dry weight basis, has a specific range of values, 0% to 100%, and allows a constant increase in calculated moisture content percentage for a given increase in water content. This is not the case for the dry weight based index, for which soil moisture content displays a skewed frequency distribution, particularly for wet soils, and thus data transformation is required prior to statistical analysis. The wet weight index suffers from the disadvantage that water content is standardised by relating it to a soil characteristic (fresh weight) which itself depends on soil water content (Robinson, 1974). Expression as a percentage of soil volume, discussed in relation to neutron probe calibration in the following section, does not make any dependence on water content and is the most suitable mode of expression of absolute water content since plant roots occupy a certain volume, rather than weight of soil (Bannister, 1976). Variability in results is also lessened, and required sample size accordingly reduced, by this mode of expression. Boelter and Blake (1964) recommended this method for peats, in view of the varying bulk densities of these soils, although Bannister (1976) noted that this index may prove unreliable for shrinking/swelling soils.

In summary, the thermogravimetric method is inexpensive, simple to carry out and is widely acceptable. It is, however, time-consuming and destructive. Repetitive sampling may itself interfere with the local soil hydrology, placing constraints on sampling frequency, although these effects can be reduced by returning samples to their original locations. In the strict sense, this represents the only 'true' direct method of measuring moisture content, per se.

3.3.2.2 Neutron Scattering (Neutron Probe)

The neutron probe method, first proposed by Belcher et al. (1950) and Gardner and Kirkham (1952), and used here to evaluate total profile and selected depth moisture contents, has undergone increasing use with accompanying improvements in instrument design. All instruments operate on the same principle, however (Fig.3.2). A radioactive neutron source, in the present case 70 millicurie americium - 241 beryllium, emits fast neutrons into the soil at specified depths; collisions with soil atoms, largely hydrogen, slow down the neutrons which then form a 'cloud' around the detection point. Cloud density, representative of soil moisture content, is monitored by a slow neutron detector which transmits an amplified electrical signal which is displayed as a count rate, on a ratemeter or, for more accurate results, on a ratescaler as used here.

Soil elements other than hydrogen have the ability to slow down fast neutrons but, because of its low atomic weight (the hydrogen nucleus comprises only one proton with almost the same size of mass as a neutron [Milanov, 1969]) hydrogen is the only soil element that can do this effectively (Visvalingam and Tandy, 1972). The presence of hydrogen in the crystalline structure of clay minerals and in organic matter becomes important only when comparing results from different soil types, soils high in organic material tending to be high in moisture content in any case. The effect is further outweighed by soil constituents capturing thermal neutrons, causing a reduction in number

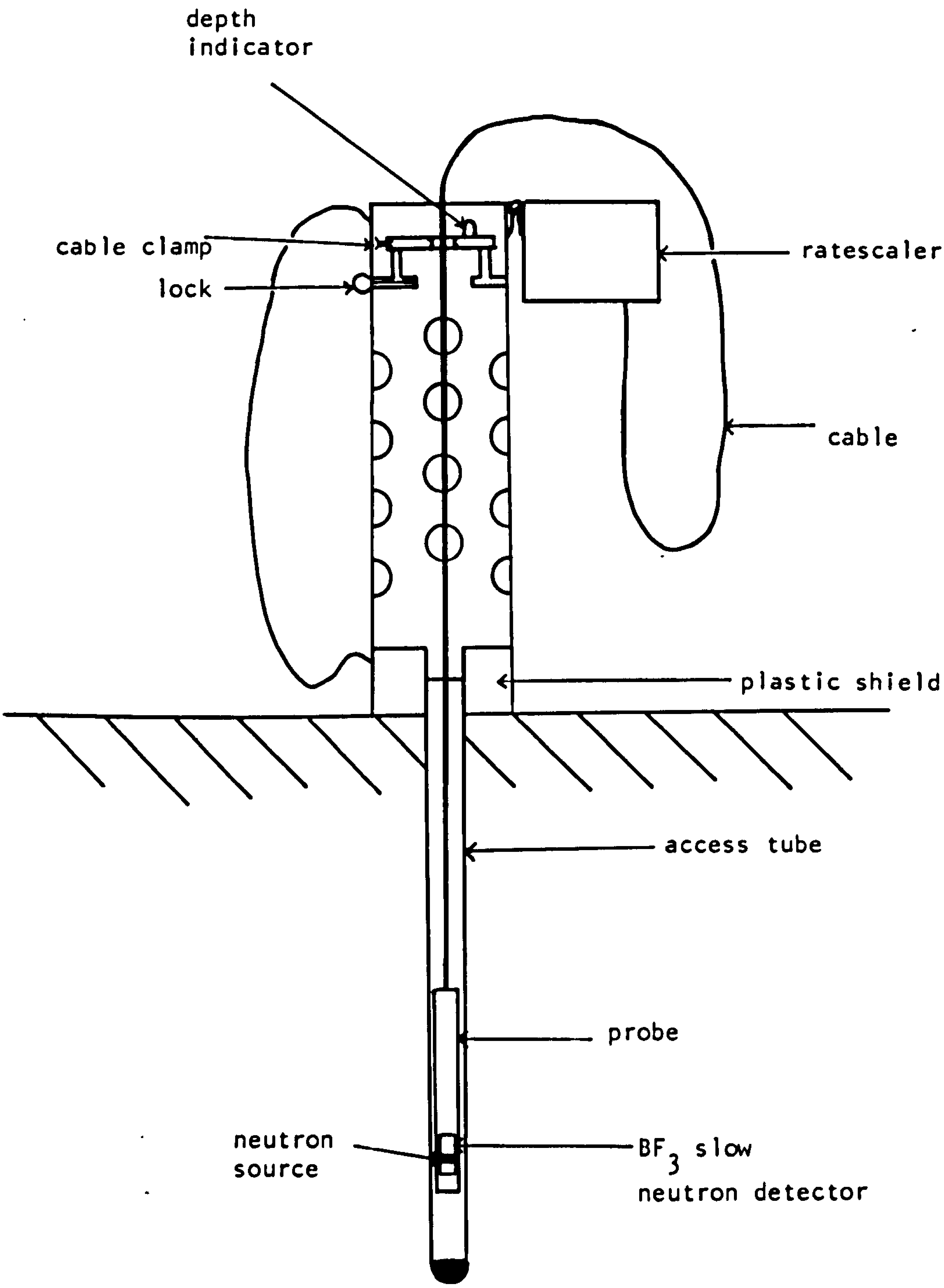


Figure 3.2 The Neutron Probe

counted (King, 1968).

Data quality and resolution are increased by the more recently introduced automatic neutron probe, or 'autoprobe', tested at the Institute of Hydrology. This instrument records at pre-determined depths and time intervals onto magnetic tape or solid-state type loggers (Roberts, 1981).

a) Access Tube Installation

Satisfactory establishment of the 50 m x 50 m sampling grid on the Egton site was followed by careful installation of a neutron probe access tube at each grid intersection. A total of twenty access tubes was sited on the moorland area, while visibility problems restricted the number in the woodland to seven (Fig.2.2). Bell et al. (1980) used the standard sample number formula for infinite, normally-distributed populations to calculate representative sample sizes:

$$n = 4\left(\frac{\sigma}{L}\right)^2 \qquad \text{Eq. 3.2}$$

where:

n = sample size

σ = standard deviation

L = level of accuracy required

(Snedecor and Cochran, 1967)

On this basis, the sampling scheme of the present study lies well within the recommendations of Bell et al., who suggested using sixteen to twenty-five sample points per 40 acre area.

Individual, bottom-sealed aluminium tubes were installed, generally to a depth of between 0.8 m and 1 m, initial excavation being accomplished by use of a soil auger slightly narrower than the access tube. Each tube was gently pushed in vertically to obtain a tight fit, while a rubber bung with attached silica gel bag provided a top seal and allowed the tube to remain dry inside. The upper 25 cm to 30 cm of the tube were left protruding above ground level. Problems of auger deflection by stones, as reported by Bell (1976), were not evident for this catchment, whilst careful augering prevented development of an enlarged hole at the ground surface, thus diminishing the possibility of water running down the side of the tube. Other potential sources of error during tube installation were itemized by McGowan and Williams (1980) as follows:

i) Anomalous moisture storage changes, resulting from cavity formation in the profile.

ii) Percolation around the tube, due to soil cracks.

iii) Loosening of originally compact soil, causing atypical rooting patterns.

iv) Compacting of soil, resulting in flow restrictions.

Great care was therefore taken during each stage of the installation process, since inadequacies in experimental skill and rigour can throw doubt on subsequent moisture readings (Bell, 1976).

The extensive spatial coverage provided by this sampling design limited probe count times to one 16-second reading for every 10 cm depth interval, readings commencing at 10 cm below the surface. Due to spatial soil moisture heterogeneity, however, this scheme is preferable to one involving long, high precision counts at only a few sites (Bell and McCulloch, 1966; Bell, 1976). Random and systematic errors arising from variability in plant water abstraction and net rainfall distribution (Calder, 1976) are also reduced by a wide sampling network. Measurements were taken weekly, and on an ordered basis over the study area in order to avoid variability due to small diurnal changes in soil moisture. Depth resolution is limited to between 10 cm and 15 cm since each reading represents the average moisture content for a sphere of between 10 cm and 30 cm radius, depending on moisture conditions and soil type. The neutron probe is therefore inappropriate for detection of detailed profile discontinuities, although total moisture contents are unaffected unless steep moisture gradients exist. Care must be taken during readings to reduce random errors caused by inadequate depth location of the probe, however.

b) Calibration

In order to derive soil moisture content values, the neutron probe method necessitates establishment of a relationship between probe count rate and moisture volume fraction (M.V.F.), the volume of water per unit volume of soil. The moisture/count rate relationship is a direct one and is not, therefore, subject to hysteresis (Visvalingam and Tandy, 1972). Variations in soil density and differing effects of soil elements make it desirable to obtain separate calibration curves for each soil type under consideration, although standard curves are also available (Bell and McCulloch, 1969). For the type of probe used in the present study, the Wallingford probe, peat and clay have been shown to be represented by a single calibration curve (Bell, 1976), and derivation of a similar curve was attempted through field calibration for the Egton site. Laboratory calibration was not undertaken because of anticipated difficulties in reproducing field conditions for these soil types, while theoretical calibration is unsuitable for empirical studies, being both time-consuming and expensive to carry out (Bell, 1976).

Calibration points were established by repeating the calibration procedure a number of times throughout the period November 1980 to November 1981, in order to incorporate the complete range of probable moisture conditions. On each occasion, a temporary access tube was installed close to tube number 13 (Fig.2.2) and ten 64-second counts were taken at each of a number of depths (10 cm, 20 cm, 30 cm, 40 cm, 60cm and 80 cm). A 10 cm 'Jarrett' auger, marked at appropriate depths, allowed excavation of soil close to the temporary tube, while six soil cores were removed at each depth by carefully hammering in a corer with removable cylinder liners (15 cm x 6 cm inside diameter). Use of a wooden duckboard prevented damage to the soil surface and vegetation during the field calibration process. On completion of sampling, the temporary tube was re-installed at a neighbouring location near tube 13, in preparation for the following sampling event. Soil cores were transported back to the laboratory in sealed plastic bags and, on return, core volume was calculated, and moisture content obtained thermogravimetrically; a calibration record form was completed for each depth sampled (Fig.3.3).

The neutron probe method suffers from the weakness of underestimating moisture content in the surface layers (approximately 20 cm), as a result of neutron escape into the surrounding atmosphere. The general calibration curve is not, therefore, directly applicable to surface measurements. Neutron reflectors, extension trays, specially designed surface-reading probes and correction factors have been employed in various circumstances as alternative means of estimation. However, neutron reflectors tend to give results merely representative of the back-scattering properties of the reflector, while surface extension trays, advocated by Bell (1976), are unsuitable for rough, moorland vegetation and for steeper slopes, and may involve difficulties of obtaining representative soil conditions. Development of a separate surface calibration between count reading and M.V.F. was attempted initially in the present case. Surface calibrations do not provide a total solution to the problem, however, since the relationship between

NEUTRON PROBE CALIBRATION RECORD FORM

SITE. 13.	DEPTH: 60cm.

	Weight in grams		1	2	3	4	5	6
a	Dish		199.44	194.20	212.45	199.63	208.17	253.5[illegible]
b	Wet core plus dish		529.87	909.10	1030.99	1020.01	974.74	1108.19
c	Dry core plus dish		425.03	730.80	815.02	781.46	781.21	876.19
d	Wet core	b-a	321.45	714.90	818.54	820.44	766.57	854.69
e	Dry core	c-a	225.59	536.60	602.51	581.83	573.04	642.67
f	Water expelled (vol. ccs)	d-e	103.86	178.30	215.97	230.61	193.53	212.00
g	Volume of core		258.05132	393.5733	473.8228	473.8225	401.5465	483.7821
h	Dry bulk density	e/g	[illegible]	1.34[illegible]	1.2717262	1.2448344	1.4059776	1.343422
i	Moisture Volume Fraction	f/g	0.4362814	0.4467408	0.4558	0.4867	0.4748	0.4337532

Mean Dry Bulk Density: 1.2551662 — Mean MVF: 0.4557631

PROBE COUNT RATES

	R	Rw
1	612	1045 (1092)
2	614	1032
3	608	1045
4	610	1043
5	612	1041
6	610	1038
7	608	1038
8	611	1042
9	606	1040
10	610	1046
Mean	610.1	1041
R/R_w	0.586071	

Total counting times

Soil (t) 640 sec

Water Standard (tw): 640 sec

Calculation of Random Counting

Error $\left(\sigma_{R/R_w}\right)$

$$\sigma_{R/R_w} = \frac{R}{R_w}\left(\frac{1}{Rt} + \frac{1}{R_w t_w}\right)^{\frac{1}{2}}$$

$$= 0.586071\left(\frac{1}{610.1 \times 640} + \frac{1}{1041 \times 640}\right)^{\frac{1}{2}}$$

$$= 1.18119 \times 10^{-3}$$

$$= 0.00118119$$

$$\simeq 0.00118$$

PROFILE DESCRIPTION

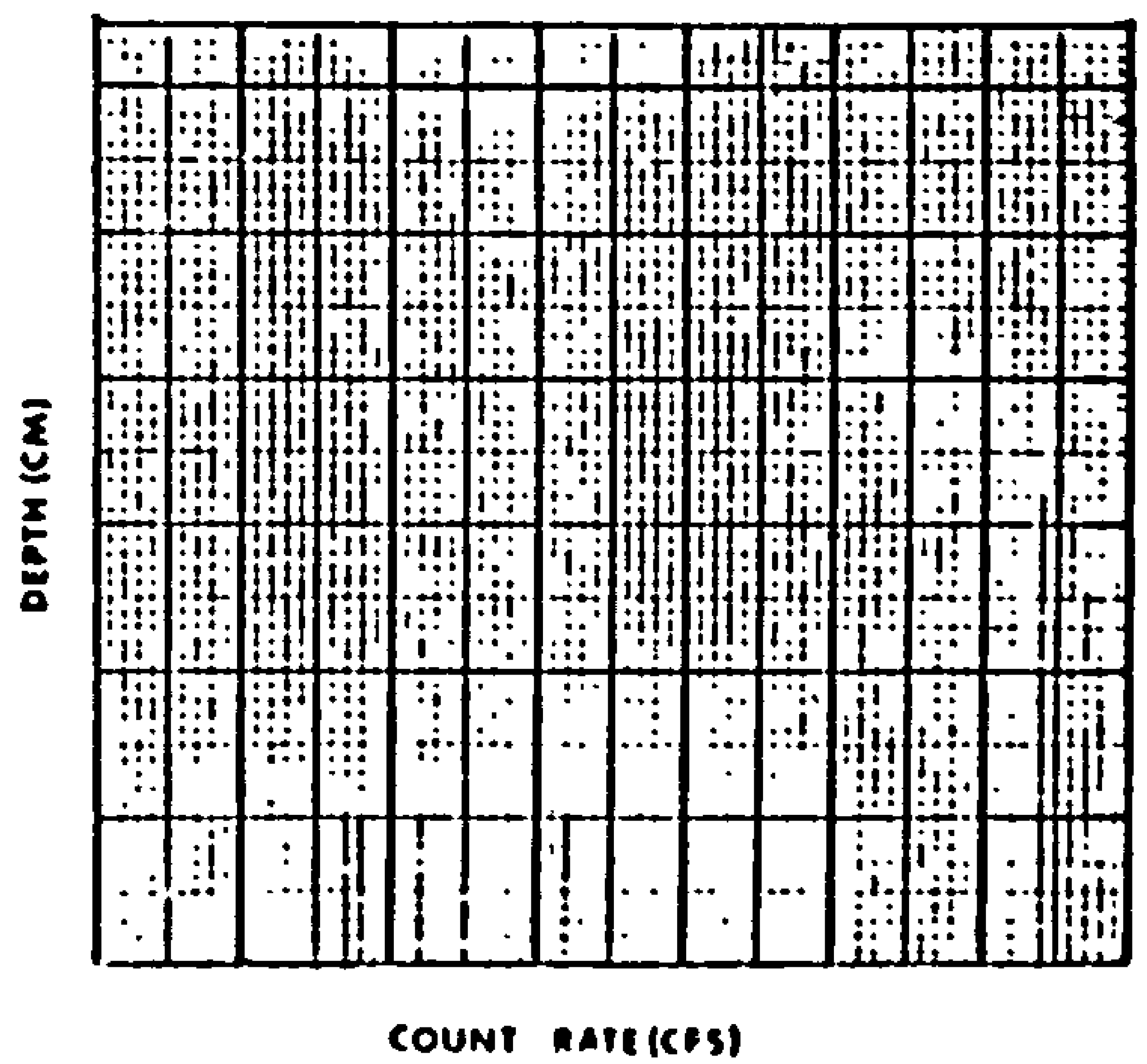
GENERAL INFORMATION

Probe No	
Meter No	
Date	26.5.81.
Observer	C E NNN

DETAILS OF COUNT RATE PROFILE

DEPTH	COUNT TIME	COUNT RATE

PROFILE GRAPH



Calibration Ref. No	Cal Curve No.	Plotting Symbol

Figure 3.3 Neutron Probe Field Calibration Record Form (water standard count rate represented by 'Rw')

count reading and moisture content may vary with changing moisture conditions in surface layers. Precision in depth location of the neutron source may also be of particular importance near the surface.

Calibration curves of count rate ratio (R/R_s) were used in each case, in preference to simple count rates (R). These ratios can be obtained by use of a laboratory or secondary standard, and in this study an access tube inserted vertically into a tank of water enabled weekly standard readings (R_s) to be taken using ten 64-second counts at the centre of the water profile. The procedure facilitates checks on possible drifting of readings and removes bias due to ageing of components and changes after instrument repairs (Bell, 1976). The use of laboratory standards also enabled detection of the need for a short (less than 32 seconds) 'warming-up' period, that is, the time taken for readings to become stable after initial instrument switch-on.

A random counting error, calculation of which is shown in Figure 3.3, results from the random process of radioactive decay (Bell, 1976). On no occasion was this larger than 0.002, in which case there was always a 95% probability that errors were less than 0.004 M.V.F. Inclusion of the R_s variable in calibration led to a larger random error than would have resulted from a simple count rate/M.V.F. relationship although the latter takes no account of drift, etc. (Bell and Eeles, 1967) and, unlike the calibration attempted here, is unable to eliminate the effects of source strength and counter efficiency.

Site calibration, nevertheless, eventually proved unreliable, as a result of both insufficient calibration points for the 'dry' zone of the curves and scatter about the curves (Figs.3.4, 3.5). Correlation between R/R_s and M.V.F. proved statistically insignificant for both surface and sub-surface values ($r = -0.041$, -0.115 for surface and sub-surface data sets, respectively, Pearson correlation). A number of

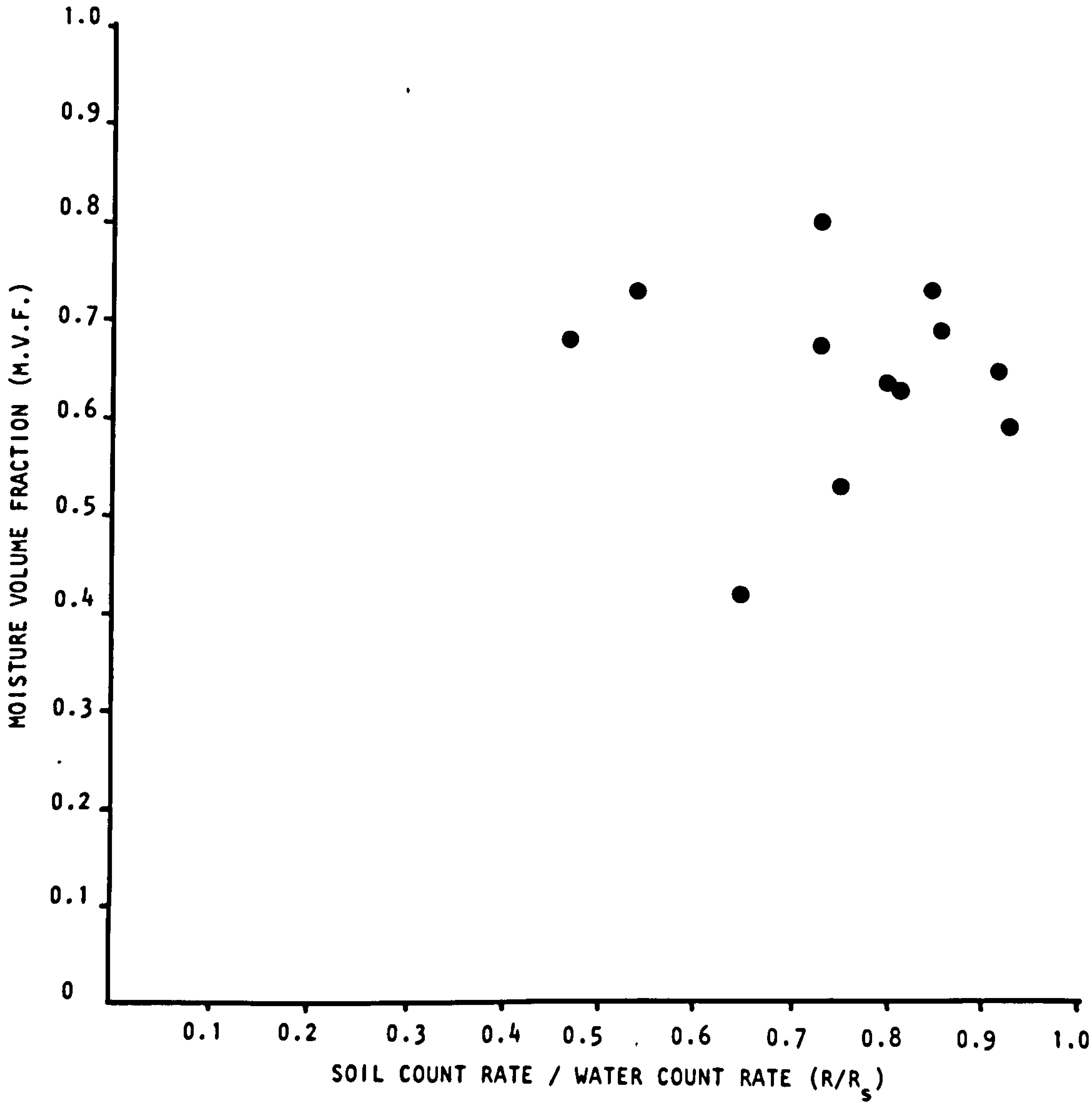


Figure 3.4 Calibration Points for Surface Depths (10 cm - 20 cm) at Egton

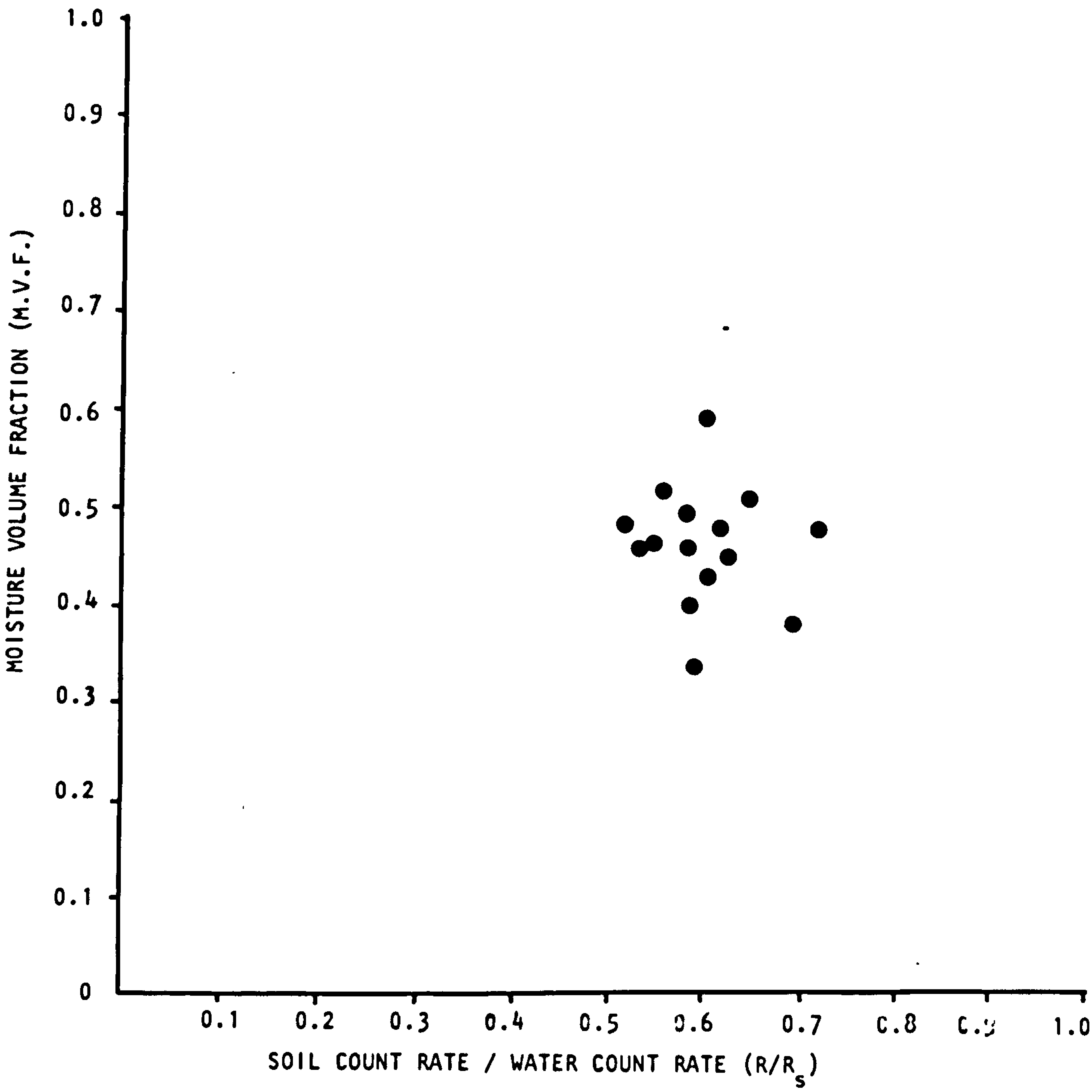


Figure 3.5 Calibration Points for Non-Surface Depths (below 20 cm) at Egton

factors can result in scatter, including soil heterogeneity, difficulties in obtaining undisturbed core samples and the fact that the count rate represents an average over a soil sphere (Bell and McCulloch, 1966). Calibration also depends on source and detector geometry and on access tube material.

The 'standard' calibration curve developed for clay and peat by the Institute of Hydrology was therefore adopted for count rate conversion for non-surface layers (Fig.3.6). Errors derived from calibration become less important in water balance studies (McGowan and Williams, 1980) and in other investigations requiring information on moisture content changes rather than on absolute moisture contents and only the slope of the calibration curve need be established for determining moisture changes. This remains generally constant for a particular soil group such as clay and peat, sand, and loams (Bell, 1976).

A separate correction factor was used to compute M.V.F. for the top layer of the profile. A number of authors have proposed the use of applied correction factors. Cole and Green (1969), for example, applied corrections using curves of probe response at air/soil interfaces. Similarly, experiments by the Institute of Hydrology on Sneaton High Moor include attempts to derive surface corrections using measurements taken at 10 cm above ground level and at ground level, as well as at 10 cm intervals below the surface (John Roberts, pers. comm.). Access tubes need to be well above ground level in this instance (at least 30 cm) to prevent interference from the radioactive shield. A series of correction factors for different depths was derived by Grant (1975) by artificially drying the soil to give a range of different moisture

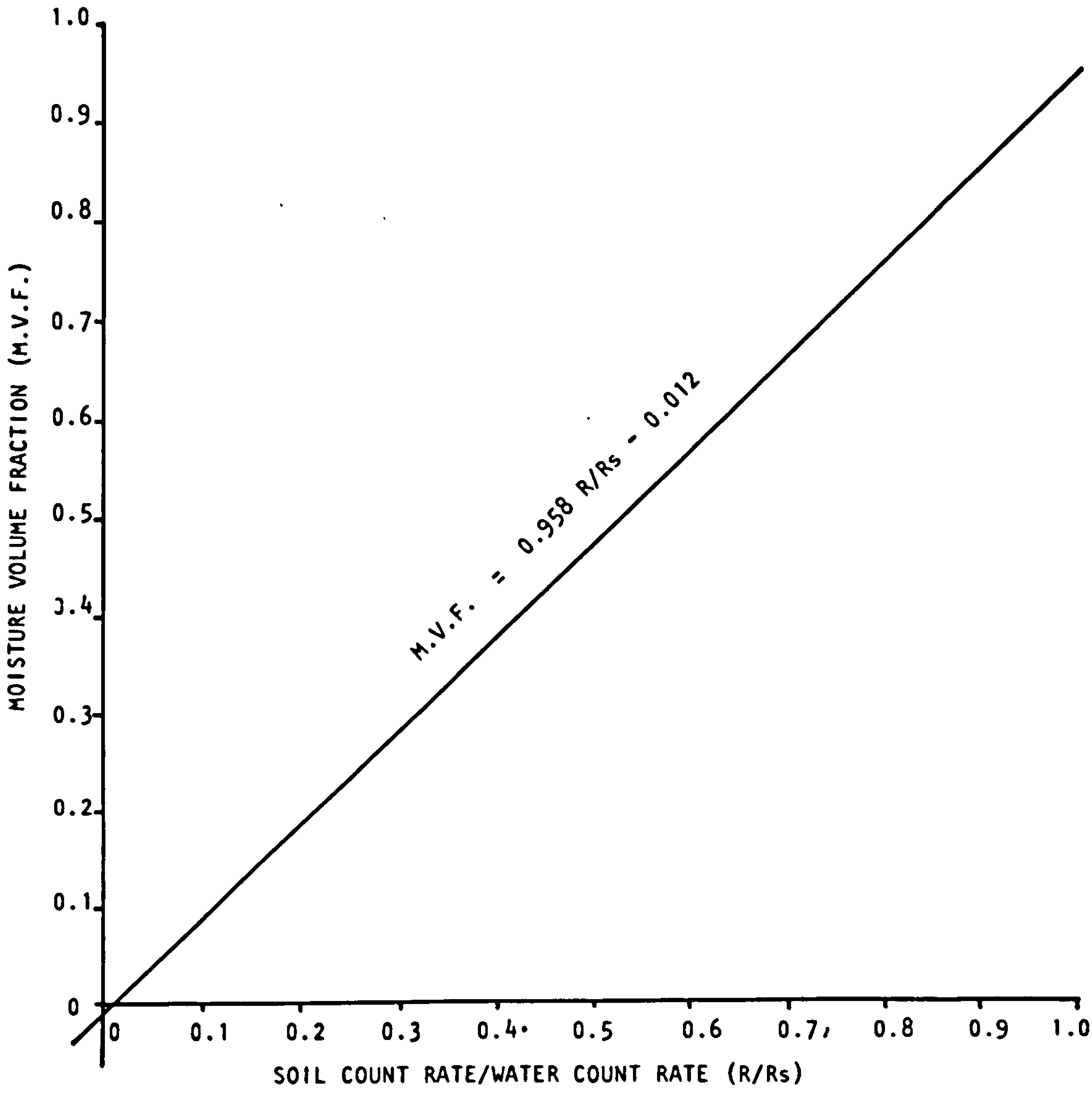


Figure 3.6 Standard Calibration Curve for Clay and Peat

conditions.

Surface reading correction in the present case was based on the techniques used by C.M.K. Gardner for a data bank of soil moisture measurements made throughout Great Britain (Gardner, 1981a). Although the non-surface calibration line may not be applicable for readings to 20 cm or even 30 cm, corrections were restricted to the 10 cm readings in the present case in the light of two factors:

i) In the study by Gardner, data for depths of less than 15 cm were substituted with the reading for the next available depth.

ii) Escape of neutrons at 20 cm or 30 cm is less severe for a wet site.

Moisture volume fractions for the 10 cm depth at each access tube site were directly substituted by the value for 20 cm for winter months (October to March, inclusive). Ten-centimetre depth readings taken during the period April to September were corrected individually for each tube by the following expression:

corrected M.V.F. at 10cm = Eq.3.3

$$\frac{\text{M.V.F. at 10 cm x mean winter count rate } (R/R_S) \text{ at 20 cm}}{\text{mean winter count rate } (R/R_S) \text{ at 10cm}}$$

where uncorrected M.V.F.'s for 10 cm are those obtained from the standard calibration curve. The conversion effectively assumes that the ratio of neutron probe count rate at 10 cm to that at 20 cm remains constant throughout the winter months, at each measuring site. The assumption is justified since, during winter, the soil profile is at field capacity and any differences in count between the two layers will be the result of changes in soil profile characteristics of which the winter ratio takes account. Further manipulation of both surface and subsurface moisture data is described fully in the following chapter.

In summary, the neutron probe method is a reliable and non-destructive technique, yielding precise measurements of soil moisture changes. Unlike most alternative means of measurement, the neutron probe provides an integrated expression of moisture content over a comparatively large volume of soil, and for the purposes of the present study, facilitated construction of accurate water balances and calculation of soil moisture deficits. The data also simplified separation of drainage and evapotranspiration in the soil moisture profile, enabling definition of maximum depths of moisture extraction under different land-uses.

3.3.2.3 Other Methods

Further methods of determining soil moisture status are available but many, such as remote sensing and time-domain reflectometry, are not yet widely used under field conditions. Lysimetric methods (Black *et al.*, 1969) normally involve estimating amounts of water percolating through an isolated soil column, sometimes using weighing techniques, but they are subject to limitations of spatial distribution and have the disadvantage of being isolated from upward moisture fluxes. Gamma-ray absorption techniques depend on the principle that radiation emitted from a source at depth and monitored a set distance away, varies only with soil moisture content (Hillel, 1980a). A high degree of spatial resolution is possible (for example, 2 mm) but the method depends on soil bulk density being either constant, or continually monitored and this, along with problems of accurate alignment of two parallel access tubes, has largely confined its use to controlled laboratory environments.

3.4 THROUGHFLOW

3.4.1 DEFINITION

For the purposes of this discussion, terminology largely follows that of Freeze (1972b) and Whipkey (1969). Subsurface runoff is defined as that part of the hydrological cycle comprising:

i) Subsurface storm flow, and

ii) Baseflow

Subsurface storm flow, also termed interflow or throughflow, comprises moisture flowing through the upper soil layers, either as saturated or unsaturated flow, towards a stream channel during and after a storm, without becoming part of the groundwater system. Rates and volumes of flow depend on rainfall rate and duration, and on soil hydraulic properties. Baseflow is derived from groundwater leakage at stream channel interfaces, and from unsaturated flow within the soil profile (Hewlett, 1961a).

Increasing interest in the phenomenon of subsurface moisture flow during the last twenty years has resulted in numerous attempts to assess its importance, using simulation or direct measurement techniques. The intention of the present study was to select a method of field measurement to indicate the effects of vegetation cover changes on at least the saturated flow component of subsurface stormflow: tenable measuring techniques are discussed below.

3.4.2 MEASUREMENT

In general, direct field measurement of subsurface flow may be achieved either by tracer experiments, or by physical interception and diversion of the flow:

3.4.2.1 Tracer Experiments

These methods provide an overall representation of subsurface flow on a large (catchment) basis. A tracer, such as a dye or radioactive element, is injected at a known point in a stream channel reach; when complete tracer mixing has been attained, sampling at a number of points downstream indicates the increase in stream discharge via detection of progressive tracer dilution. This yields an estimate of subsurface flow from adjacent hillslopes, once contributions from intervening tributaries have been taken into account. Similarly, weirs or flumes with accompanying water-level recorders, a set distance apart, can be used to indicate subsurface flow contributions from intervening slopes.

The restricted area under consideration and, therefore, inadequate tracer mixing times prevented employment of these techniques in the present study. Further, choice of tracer would have been limited, since many radioactive substances and chemical dyes become adsorbed onto clays and organic matter (Atkinson, 1978).

3.4.2.2 Flow Interception

Methods involving physical intervention of the flow normally entail insertion of a set of troughs or gutters at selected levels in the soil profile, effecting diversion of flow to a measuring device. Flow rates may be monitored manually, with a stop-watch and measuring cylinder (Weyman, 1973, 1974) or on a continuous basis to derive complete flow hydrographs. Continuous measuring systems have included a tipping bucket assembly connected to an electronic logger (Knapp, 1973); measurement of flow stage by means of weir slots (Dunne and Black, 1970a); and detection of accumulated throughflow in a collecting drum, using stage recorders.

3.4.2.3 The Study Method

In view of the preceding discussion, a design based on physical interception of lateral flow was deemed most appropriate for the purposes of the present investigation. Apparatus providing a continuous record of flow discharge, using a collecting vessel and rotating drum chart, was laboratory-tested but proved difficult to operate in the field. Application of a drain discharge meter, of the type used in agricultural tile drains, or of a drain outflow meter as described by van de Weerd (1977), was also considered. Eventually, however, an instrument was adopted which was simpler in design, easy to install and maintain, and yet one which was adaptable and accurate: the throughflow (interflow) box or trough designed by Arnett (1971), (Fig.3.7). Instrument locations were selected on the basis of prior field observations of subsurface flow. One instrument was therefore installed at the base of the moorland slope, close to the intermittent rills, and one near the experimental boundary of the woodland (Fig.2.2).

The measuring instrument, which basically consists of a two-layered galvanised steel plate box with the front open-ended, is inserted at 90° to the soil surface, the uppermost lip of the box being carefully pushed upslope below the litter layer. The top compartment fills with an undisturbed 0 cm to 15 cm layer and the lower section with the 15 cm to 30 cm layer, the complete box being 15 cm wide and 30 cm in length. It is important to ensure that each section is completely full of undisturbed soil in order to represent field conditions as accurately as possible. Since plastic tubing directs throughflow from enclosed pipes to collecting vessels, the design has the advantage that samples may be collected for subsequent water quality analyses. Topsoil (peat) depths of 14 cm and 11 cm in the moorland and woodland areas,

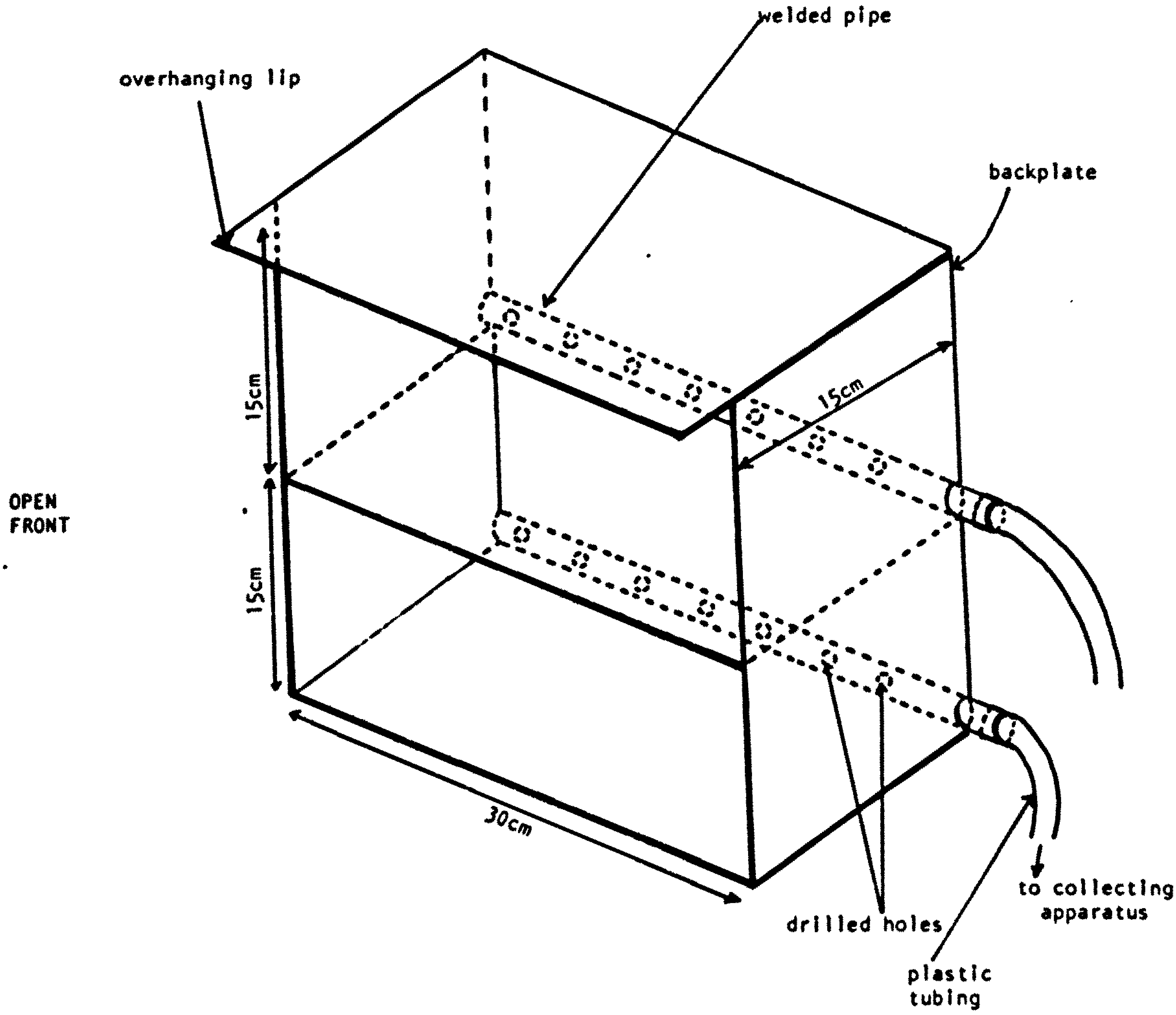


Figure 3.7 Throughflow Box [after Arnett (1971)]

respectively, enabled subsurface flow resulting from the impeding action of the clay layer to be received, in each case, by the upper collecting vessel. Increasingly large containers were used in a series of tests here, sample overflow being a problem during winter, and eventually a five-litre plastic vessel was maintained at each depth. These containers were inspected and throughflow volumes measured at weekly intervals. After instrument installation, soil should be carefully replaced behind the box, as close as possible to the original bulk density. This infilling attempts to overcome pressure differences at the air/soil interface. Instrument design and construction are more fully described by Arnett (1971).

If continued for further use for any length of time under these conditions, the design could usefully incorporate a drainage system to evacuate the collecting pit when necessary. For pit drainage, Knapp (1973) recommended excavation of a drainage channel lined with a rainwater drain, although the low slope angles of this area would probably warrant excavation to considerable depths and/or distances. An alternative possibility would be the installation of a pump (Knapp, 1973), although for the duration of the current study, occasional clearance of the pit proved adequate.

Several considerations of flow hydraulics need to be made in relation to the design and operation of a throughflow interception mechanism. Firstly, the throughflow boxes employed here capture only saturated flow. Opinions vary as to the significance of the unsaturated flow component for baseflow and for the total storm response. Some authors contend that streamflow is controlled by the saturated component (Weyman, 1970, 1973; Anderson and Burt, 1977b), and attention may be justifiably directed towards monitoring of saturated flow if this is deemed to be of greater significance than unsaturated flow (Whipkey and Kirkby, 1978). The latter constituent however, may comprise an important supply for stream baseflow in steep mountain watersheds (Hewlett, 1961b; Hewlett and Hibbert, 1963) while baseflow itself may form a substantial component of total runoff for some catchments, as reported for the eastern United States by Hewlett (1982). The part played by throughflow in generating storm runoff is discussed further in Chapter 5, although it may be noted at this point that in the final instance, it is the local physical conditions of a specific area which determine storm runoff generation processes (Rodda et al., 1976).

No distinction is made between matrix flow and pipeflow by the current design, both being collected together. Flow in discrete subsurface channels was seen only occasionally over the experimental area, however, and specific importance of this type of flow was considered to be limited. Identification itself may become subjective at this scale (less than 1 cm diameter) when based purely on size of feature and Atkinson (1978) more objectively defined pipeflow as being turbulent, and matrix flow as laminar. Removal of vegetation and the exposure of an erodible layer can increase the importance of piping, however (Jones, 1971, 1981).

Finally, a fundamental problem involved in pit excavation for throughflow measurement is the effect of an exposed soil face on soil hydrology itself. The unnatural face leads to a saturated wedge extending upslope, altering saturated and unsaturated flow conditions and distorting the net of hydraulic potential. The latter effect can have further consequences, in leading to an influx of water from areas other than those directly upslope of the pit (Atkinson, 1978). Such problems were avoided in an experiment by Dunne and Black (1970a) since, by monitoring throughflow from an entire hillslope, using an 84 m (275 ft) long trench containing tile drains, fluctuations in throughflow contributing area were contained within the measured zone. Ungauged sections at each end of the trench reduced edge effects. Mosley (1982), whilst agreeing that pit excavation results in flow line convergence and development of saturated wedges, proposed that this is relevant only in the case of matrix flow measurement, and not for flow along preferred lines or macropores. Sealing with mortar and artificial wetting beyond plot boundaries were techniques used by Whipkey (1965) in an attempt to eliminate unnatural flow paths, while Atkinson's suggestion (1978) of collecting flow from a natural soil face, although viable in the present instance for the moorland plot, is precluded under woodland due to the absence of exposures.

The complexities outlined above are, however, at least reduced by the apparatus used in the present case. Specific design and installation features of the instrument mean that profile disturbance is kept to a minimum: the box is completely filled with soil, while the backplate remains flush with the soil face; behind this, the soil is replaced to avoid exposure of a free face and the collecting pit is situated as distant as possible from the measuring pit, by directing the plastic tubing and then backfilling.

Subsurface moisture flow is difficult to monitor even under ideal conditions and the intention here was not to define it in great spatial or temporal detail, but rather to represent general variations due to land-use change. Sophisticated systems can be established, but are likely to involve time-consuming installation and may be best applied to longer-term projects.

3.5 STREAM DISCHARGE

Although measured 'at a point', stream discharge represents an integration of contributions from precipitation, overland flow, throughflow and groundwater flow, some arriving at different velocities, perhaps from different parts of the catchment and holding varying degrees of importance in different areas. Because measurements of stream discharge are not subjected to spatial extrapolation in an attempt to be representative of a larger area, they comprise probably the most accurate of all measurements of hydrological variables.

An instrumentation system to monitor stream stage had previously been established on the study site by Fullen (1981). The equipment, comprising a V-notch weir and water-level recorder, was set up as far as possible in accordance with the specifications outlined by the World Meteorological Organization (1974) and Toebes and Ouryvaev (1970), and was considered appropriate for the purposes of the present investigation.

In detail, a triangular 1/2 90 degree thin-plate V-notch weir was located downstream of the instrument shed which was constructed over the stream itself (Fig.2.2). This type of weir is a precise, sensitive instrument, particularly suitable for low minimum flow situations (British Standards Institution, 1981) and, apart from occasional sediment and debris blockage of the approach channel, no serious problems were encountered during operation of the weir. A good control structure will give discharge values to within ± 1% to 2% (Gregory and Walling, 1973), although Rothacher and Miner (1967) warned that the average field installation has an error of ± 2% to 10% or more, and that the best expected accuracy under field conditions is 3% to 5%. Detailed information on construction and installation of these instruments is given by the British Standards Institution (1981).

Weir structures may be calibrated by a number of different techniques, of varying reliability. Field calibration by current meter gauging, for example, is time-consuming, can be inaccurate and it may involve awaiting the occurrence of a complete range of flows to enable proper calibration. Laboratory methods are simple to perform, but can be subject to systematic error (Francis, 1966), and it may therefore be preferable to use a British Standards Institution rating equation and standard rating tables. The discharge equation applied to stage measurement in the present study was as follows:

$$Q = C \frac{8}{15} \tan \left(\frac{\alpha}{2}\right) \sqrt{2g_n}\, h^{5/2} \qquad \text{Eq. 3.4}$$

where:

Q = discharge ($m^3 s^{-1}$)

C = coefficient of discharge (from rating tables)

α = angle included between the sides of the notch ($53^{\circ}8'$)

g_n = acceleration due to gravity (9.80665 m s^{-2})

h = head referred to the vertex of the notch (stage, m)

A continuous record of stream stage was provided by a Munro water-level recorder with weekly drum chart mechanism housed in the instrument shed; the stilling well with float was placed directly in the stream, immediately upstream of the weir. Recorded stage height was regularly checked against stream level, but normally required adjustment only after heavy storms and on only one occasion (15 August 1980 to 16 August 1980) was the maximum recordable level exceeded. Whetstone and Grigoriev (1972) quoted float-stilling wells as being accurate to 2 mm.

3.6 AUTOMATIC DATA STORAGE, RETRIEVAL AND MANIPULATION

Data from one of the automatic weather stations established on Sneaton High Moor (p. 24) were used to supplement catchment measurements. The stations permit detailed monitoring of a range of meteorological variables and can operate for a period of up to four weeks unattended. The following attributes were recorded at five-minute intervals onto Microdata logger magnetic tape:

i) Solar radiation (W m^{-2})	- Kipp Solarimeter
ii) Net radiation (W m^{-2})	- Dirmhirn net radiometer
iii) Wet bulb depression (^{o}C)	- platinum resistance probes
iv) Temperature (^{o}C)	
v) Wind run (speed) (m s^{-1})	- anemometer cups, turning a potentiometer
vi) Wind direction (degrees)	- reed relay system
vii) Rainfall (mm)	- ground-level tipping bucket raingauge

Variables i) to vi) are monitored by sensors supported by a mast. Signals from each sensor are modified for input into the logger and tapes are read directly by computer. Daily summaries are produced after data translation, quality control, editing and conversion (Roberts, 1981). In addition to the seven recorded meteorological variables, data processing provides hourly values of specific humidity (g kg^{-1}) and specific humidity deficit (g kg^{-1}); hourly Penman-Monteith evaporation estimates for water, grass and forest (W m^{-2}); and daily totals of Penman E_O (open water evaporation) and Penman E_T (potential evapotranspiration) (MJ m^{-2}). Hourly values of aerodynamic resistance (r_a) are also included in the calculated variables. Hourly, daylight and 24 h totals or averages, along with maximum and minimum five-minute values of all recorded variables are provided by the summary sheets; frequency of the calculated values (humidities, evaporation and r_a) varies. An index of data quality helped when deciding to accept or reject suspect values.

In the course of the present analysis, daily records of selected variables were incorporated into soil moisture modelling and synthesis of water balances. Rainfall totals, Penman-Monteith estimates of evaporation from forest and Penman E_T values were drawn directly from the weather station output, with unit conversion where necessary. Penman-Monteith evaporation estimates for heather and bare ground were subsequently calculated using basic meteorological variables (temperature, rainfall, net radiation, wind run and wet bulb depression). The evaporation formulae were selected for their reliability and widespread use. Availability of input data and prior calculation of most of the required estimates rendered their use more appropriate than that of other, more empirical equations, based on temperature measurement, such as that of Thornthwaite (1948). Justification for establishing a more direct means of measurement, through field-based instrumentation at Egton, is equally questionable in view of the experimental difficulties involved (representativeness of plots, site requirements, degree of data resolution, etc.). The form of the equations used is outlined below.

3.6.1 PENMAN FORMULA FOR POTENTIAL EVAPOTRANSPIRATION

The Penman method of evaporation estimation amalgamates the energy budget and aerodynamic (vapour flow) approaches enabling evaporation estimation from readily available meteorological data:

i) The aerodynamic approach is based on Dalton's empirical equation:

$$E_O = (e_s - e_d)\, f(u) \qquad \text{Eq. 3.5}$$

where:

E_O = open water evaporation rate

e_s = vapour pressure at the evaporating surface

e_d = vapour pressure in the atmosphere above the surface

$f(u)$= a function of wind velocity

Penman's aerodynamic term may then be expressed as:

$$E_a = 0.35\ (1+0.01u)\ (e_a - e_d)\ \text{mm d}^{-1} \qquad \text{Eq. 3.6}$$

where:

E_a = drying term (the 'drying power' of the air)

u = wind speed at a height of 2 m (mi d^{-1})

e_a = saturation vapour pressure at mean air temperature (mm mercury)

e_d = actual vapour pressure at mean air temperature and humidity (mm mercury)

ii) The energy balance approach based on the partition of available energy:

Available energy comprises a sensible heat component, used in heating the atmosphere, and latent heat, used in evaporation; the balance between these two fluxes is termed the Bowen ratio. Available energy is apportioned between the net radiation income, a soil heat flux, a canopy/air heat flux and energy used in photosynthesis. Net radiation is the main component and may be either measured or estimated by equation:

$$H = (1-r)R_a(0.18+0.55n/N)-\sigma Ta^4(0.56-0.09\sqrt{e_d})(0.10+0.90n/N)\ \text{mm d}^{-1} \qquad \text{Eq. 3.7}$$

where:

H = net radiation

r = albedo (reflection coefficient of the surface)

R_a = theoretical radiation intensity at the surface (evaporation units)

n/N = ratio of actual/possible hours of sunshine

σTa^4 = theoretical black body radiation;

Ta = mean air temperature (^oK)

σ = Stefan's constant

e_d = as above

The basic Penman (1948) equation combines these two approaches to eliminate the measurement of temperature of the evaporating surface:

$$E_o = \left(\frac{\Delta}{\gamma} H + E_a\right) \Big/ \left(\frac{\Delta}{\gamma} + x\right) \text{ mm d}^{-1} \qquad \text{Eq. 3.8}$$

where:

E_o = open water evaporation

Δ = slope of saturation vapour-pressure curve at mean air temperature (mm mercury/^oF)

γ = constant of the wet- and dry-bulb psychrometer equation (0.27 mm mercury/^oF)

E_a = aerodynamic term (Eq. 3.6)

x = normally set to unity

Penman's potential evapotranspiration, E_T may be derived from calculation of open water evaporation, converted by a factor, f where:

$$f = E_T/E_o \qquad \text{Eq. 3.9}$$

where:

E_o = rate of evaporation from open water (mm d^{-1})

E_T = rate of evaporation from turf (short green cover) (mm d^{-1})

f = 0.8 for summer,
0.6 for winter

Penman (1956) later showed that this two-stage calculation is unnecessary if albedo and surface roughness terms for open water are substituted by appropriate values for vegetated surfaces.

3.6.2 PENMAN-MONTEITH EVAPOTRANSPIRATION

In a move towards integration of meteorological and plant physiological aspects of evaporation, Monteith (1965) introduced crop aerodynamic and stomatal resistances into the procedure to allow more realistic calculation of vapour transfer for several vegetation types, but especially for taller vegetation, as the Penman-Monteith combination equation:

$$\lambda E = \frac{\Delta H + \rho c\,(e_s(T) - e_d)/r_a}{\Delta + \gamma\,(1 + r_s/r_a)} \text{ mm d}^{-1} \qquad \text{Eq.3.10}$$

where:

λE = latent heat flux

ρ = air density (~ 1.20 kg m $^{-3}$)

c = specific heat of the air (1.01×10^3 J kg^{-1} $^{o}C^{-1}$)

$e_s(T)$ = saturation vapour pressure at air temperature T (mm mercury)

e_d = actual vapour pressure of air (mm mercury)

r_a = crop aerodynamic resistance (s cm^{-1})

r_s = surface resistance, in practice stomatal resistance (s cm^{-1})

Remaining terms as in previous equations.

Consideration of specific terms, and applications and limitations of each of these estimates are deferred until the following chapters, when evaluation of actual evapotranspiration is also discussed.

3.7 THE MUIRBURN

Controlled vegetation burning encourages the growth of young, nutritious shoots, removes ageing plant material and controls unwanted species. The objective of a controlled burning programme for grouse management is the achievement of a mix of both old heather, for nesting cover, and young plants, for feeding. Length of burning cycle depends on the time taken for the vegetation to reach the best conditions for burning, although it normally ranges from every seven years to every fifteen to twenty years (Muirburn Working Party, 1977). The legal permissible time for burning is 1 October to the end of April (England, Scotland and Wales). Small fires are preferred, while shape of burnt area is important for grouse moors; narrow strips up to 30 m wide are preferable to square or circular areas, thus maintaining adjacent protective cover (Gimingham, 1972; Muirburn Working Party, 1977). The area of moor burnt in the present study (Fig.2.2) was dictated by weather and vegetation conditions, and the requirements of the gamekeeper.

Several environmental variables were monitored during the muirburn, carried out on 10 April 1981 (Plate II). Wind direction was due south, changing to west, and average wind speed, which was recorded every 15 seconds by a hand-held anemometer at 1.5 m above ground level, was 6.9 m s^{-1} (n = 70, s.d. = 1.71). Fire temperature was measured by a chrome-alumel thermocouple at a height of 40 cm within the heather canopy and readings were taken from an electrical pyrometer connected to the thermocouple by a set of leads. A maximum temperature of 480°C was recorded during the burn (Fig.3.8), although temperatures up to 500°C to 800°C are attainable in a heather canopy and 250°C to 500°C at ground level (Muirburn Working Party, 1977).

Fire intensity, which relies not only on prevailing weather conditions, vegetation and soil surface characterstics, but also on burning technique, was estimated in the present instance from the

Plate II   Muirburn at Egton Catchment (10 April 1981)

Both photographs taken at approximately 1200h.

(a) Smoke - Photographer is standing 10 m from neutron probe access tube no. 2, and facing due west.
    Rainguage apparatus shown centre.

(b) Fire - Facing south east.

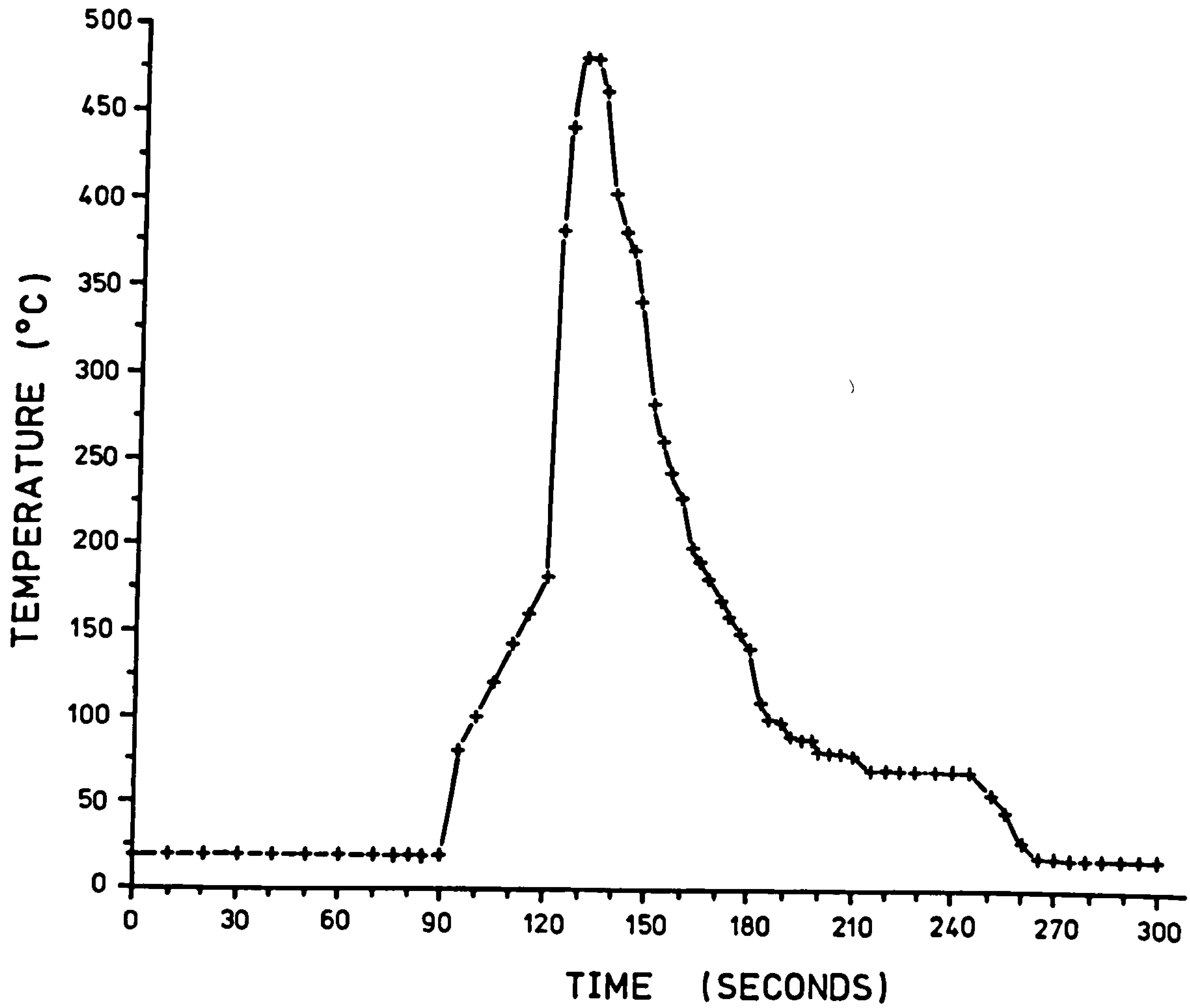


Figure 3.8 Fire Temperature within the Heather Canopy during the Muirburn

following expression of Bryam's (1959) formula, as used by van Wagner (1964) and Kayll (1966):

$$F.I. = t.v.m \qquad \text{Eq. 3.11}$$

where:

F.I. = Fire Intensity index (g cal s^{-1} cm^{-1})

t = heat of combustion (4800g cal g^{-1})

v = rate of fire advance (cm s^{-1})

m = amount of fuel consumed (g cm^{-2})

Speed of fire advance, 'v' in general varies from about 3 cm s^{-1} to 12 cm s^{-1} . In the present case mean rate of spread, derived from a series of timing measurements using marker poles, was 7 cm s^{-1} (n = 19, s.d. = 4.97). Random samples of vegetation collected prior to and immediately after the burn from five 0.5 m by 0.5 m quadrats, were used to determine the amount of fuel consumed, 'm'. From weights of vegetation before and after burning, 72% (13053.88 kg ha^{-1}) of the vegetation was found to be consumed by the fire, the pre-burn heather standing crop being 18051.64 kg ha^{-1} and the post-burn 4997.76 kg ha^{-1} (Aspinall, 1982). Fire intensity is therefore calculated as 4402 g cal s^{-1} cm^{-1} from Equation 3.11 (t = 4800 g cal g^{-1}, v = 7, m = 0.131 g cm-2), a value of a comparable order of magnitude to those determined by Fullen (1981) for Egton and Sneaton High Moors (4410g cal s^{-1} cm^{-1} and 1817 g cal s^{-1} cm^{-1}, respectively).

3.8 SUMMARY

The quality of experimental results depends ultimately on basic experimental design, and on rigour used in data collection. Effort therefore needs to be devoted to establishment of reliable monitoring systems, to satisfy the objectives and constraints of the project, without restricting the time allocated to subsequent data collection and analysis. In monitoring a series of variables in relation to the aims of this project the most appropriate measuring techniques were chosen where available. In this selection, a number of prevailing factors were considered. First, data resolution needed to be such that site visits were made at least weekly and thus automatic monitoring equipment was used where possible. Since the complete data base included a range of measuring frequencies (continuous, hourly, weekly and monthly) data-summarising techniques were used in the analysis of results. Secondly, sampling design needed to be representative. Siting of single monitoring instruments was therefore executed carefully, while variables requiring wider spatial coverage were monitored systematically on a regular, non-continuous, basis. Finally, instrument precision and accuracy are important in terms of reliability of results. Different variables are measured with different degrees of accuracy and mode of expression of results should reflect this.

Data processing, analysis and interpretation are covered by the following three chapters. Responses by the soil moisture variable are identified in Chapter 4, along with implications for water loss through evapotranspiration, while in Chapter 5 the nature of the rainfall-runoff conversion is assessed. Moorland water balances are calculated in Chapter 6 which also continues the evaluation of the interception and transpiration components introduced in the preceding two chapters.

CHAPTER 4

SUBSURFACE MOISTURE RESPONSES

4.1 INTRODUCTION

This chapter examines the consequences of changes in vegetation cover for subsurface moisture status. General soil moisture responses are well documented, but studies of the effects of multiple, conflicting land-use regimes over areas characterised by high water tables are more limited. It is necessary in the present study to determine whether vegetation effects are sufficiently important to prevail over meteorological influences in this context. Crop cover has both macroscopic and microscopic effects on soil moisture conditions. It determines amounts of effective rainfall through its interceptive effect, it facilitates infiltration and, through its effect on organic matter, modifies soil structure, density and porosity. Trees are particularly effective in determining soil water flow paths by means of soil channel generation by root systems.

Techniques used here to determine objectively the significance of heather burning and coniferous afforestation, consider the nature of the relationship between soil moisture status and evapotranspiration. A certain amount of controversy surrounds the specific role of plants in conducting water from soil to atmosphere. One school of thought (for example, Lee, 1967) states that plants are active in water transport control, for example, by stomatal opening and closing. Transpiration depends on potential gradients of the 'soil-plant-atmosphere continuum' (SPAC), a term derived by Philip (1966) for the integrated system of water transport through soil, plant and atmosphere, water flowing from regions of relatively high to relatively low potential energy. Flow rate is determined by the resistance of each part of the system and by the potential gradient (Hillel, 1980b), with water following the path of least resistance. If plants are active in controlling transpiration, it is still unclear which of stomatal or root resistances has ultimate control. The contrasting argument, now largely refuted, maintains that plants act as passive 'wicks' in water transport and that transpiration represents simply leakage from the plant (Penman, 1963; van Bavel, 1968).

Since daily fluctuations of transpiration in Calluna vulgaris do not follow exactly evaporation trends as indicated by the atmospheric saturation deficit, Bannister (1964b) suggested that stomatal control is an important influence over transpiration rates in this species, although atmospheric saturation deficit may not be an accurate representation of evaporation from plant or soil. Further evidence proposed in favour of stomatal control lies in the determination of the main component of plant water potential, pressure gradient, by evaporative demand, the result of large vapour pressure differences between atmosphere and leaves (Hillel, 1980b; Slatyer and Gardner, 1965). Sensitivity of transpiration response to changing soil water potential, however, varies from species to species (Jarvis and Jarvis, 1963).

The present investigation evaluates soil-plant relationships through soil moisture modelling for three land-use types, using the relationship between actual evapotranspiration and potential evaporative demand, as soil moisture content becomes limiting. While Chapter 5 considers spatial variations in catchment soil moisture, the present chapter concentrates on changes in moisture deficit throughout the year under different surface vegetation covers, and the accuracy with which soil moisture models may predict those conditions.

4.2 SOIL MOISTURE MODELLING

Variations in soil moisture status under different vegetation types may be quantified using average moisture content. In order to assess implications for other components of the hydrological cycle, however, a more objective means of interpreting soil moisture data is necessary. Soil moisture modelling techniques are used in the present instance to evaluate the effects of vegetation on soil moisture status and to examine implications for runoff and actual evapotranspiration. Such modelling of hydrological processes is of practical value to water authorities, farmers and water engineers, although acceptable degrees of accuracy may vary between users. Water authorities require detailed soil moisture deficit and drainage data as aids to flood warning and water resource management, whilst more general moisture deficit information may be sufficient for agriculturalists in estimating crop irrigation requirements. As soil moisture modelling is carried out on a number of different scales and for differing applications, several techniques have been developed, each of which may be generally classified into one of two categories, either physically-based or empirical models.

4.2.1 PHYSICALLY-BASED MODELS

These models rely on the Richards (1931) equation which is based on Darcy's law (Darcy, 1856). The latter states that for saturated conditions, any moisture flux is proportional to the hydraulic head gradient and hydraulic conductivity. Considering moisture flux in the vertical (z) dimension,

$$q = -k \frac{\delta H}{\delta z} \qquad \text{Eq. 4.1}$$

where:

q = moisture flux in the vertical (z) dimension (cm s^{-1})

k = saturated hydraulic conductivity in the vertical (z) dimension (cm s^{-1})

$-z$ = depth (cm)

H = total hydraulic head (cm) = $H_p + H_g$

H_p = hydrostatic pressure head (height of water resting on a point)

H_g = gravitational head (height above a fixed datum)

Moisture movement therefore relies on the establishment of a hydraulic gradient and is brought about by differences in potential energy between two points. Movement occurs from areas of high total hydraulic potential to areas of lower potential. Total hydraulic potential comprises the sum of gravitational, pressure (matric) and osmotic potentials, the latter generally being considered to be of least importance. Although the components of total potential may not be mutually independent, they do not all act in the same direction (Hillel, 1980a). Gravitational potential is due to the height of a particular point above a fixed, arbitrary reference datum. Pressure or matric potential is positive when the point of interest lies below the water table, and soil moisture is at a pressure greater than atmospheric. When the point is above the water table (unsaturated soil) and soil moisture remains below atmospheric pressure, suction or tension forces are operative, and matric potentials are negative. Osmotic potential is dependent on the presence of solutes in the soil water, which tend to lower the potential energy of soil water.

Richards (1931) adapted Darcy's law, for unsaturated flow by making conductivity a function of matric suction head (matric potential). In the vertical (z) dimension this is given by:

$$q = -k \; (\psi) \frac{\delta H}{\delta z} \qquad \text{Eq. 4.2}$$

where:

ψ = matric suction head

H = ψ + z (hydraulic head)

$k(\psi)$= unsaturated hydraulic conductivity

The applicability of Darcy's law to flow within peat soils is subject to some debate, although it appears to be most relevant in cases of low humification (Rycroft et al., 1975a,b). However, in practice the law may be applied in most cases of soil water movement (Hillel, 1980a).

The present discussion has thus far omitted to mention the role of plant roots in physically-based models. In this context, a dichotomy exists between macroscopic models which consider water uptake by the root zone as a whole and which include a sink term in the Richards equation to predict root extraction, and microscopic variants which assess flow to a single root. These two approaches are examined below.

4.2.1.1 Macroscopic Scale Approach

As the entire root system is seen to be one absorbing mass, macroscopic models ignore variations in potentials around individual roots. By combining a sink term, representing the root system, with the Richards (1931) equation, water uptake for one-dimensional (vertical) flow may be described by:

$$\frac{\delta\theta}{\delta t} = - \frac{\delta}{\delta z} \quad [k(\psi)\delta\Psi/\delta z] \; + \frac{\delta k}{\delta z} \quad (\psi) \quad -S \qquad \text{Eq. 4.3}$$

where:

θ = moisture content at depth z

t = time

$k(\psi)$= unsaturated hydraulic conductivity at depth z

ψ = matric potential at depth z

S = sink term

Numerous sink terms have been defined (Molz, 1981), but they are typically of a form specified by Hillel et al. (1976):

$$S = \frac{(\phi_{soil} - \phi_{plant})}{(R_{soil} + R_{roots})} \qquad \text{Eq. 4.4}$$

where:

ϕ_{soil} = total hydraulic head of the soil

ϕ_{plant} = hydraulic head in the plant

R_{soil} = soil resistance (a function of soil hydraulic conductivity and root density)

R_{roots} = root resistance

Most require a knowledge of root resistances and leaf potentials, quantities that are not easily assessed (Feddes et al., 1974), and in an attempt to overcome this problem, Feddes et al. (1976) presented an empirical sink term which is a function of soil water content. The sink term is employed by the authors in a 'finite-difference' model and their results are verified with field measurements of soil water content. They concluded that the empirical sink term compares favourably with a more physically-based formulation described by

Feddes et al. (1974).

Molz (1981) criticised the earlier model of Feddes et al. (1974) as having an extraction function operating successfully only within the context of the model. Similarly, Rowse et al. (1978) rejected the work of Feddes et al. (1974) on the basis that values of the empirical proportionality constant in their root extraction calculation were determined from field measurements of water extraction profiles, the very phenomena they were intended to predict. Simulated and measured water extraction patterns showed close agreement in the model of Rowse et al. (1978) and, indeed Molz (1981) described this model as one of the best in existence, in terms of its extraction function, the latter being one of only a few representations of water uptake which considers both soil and plant resistances. There is general disagreement over the relative importance of these resistances however, although Molz expressed the dominance of root resistance and thus criticised those extraction functions which ignore it.

Several other macrosopic models have been developed, for example, that of Feddes and Rijtema (1972) which incorporates an extraction function to calculate water uptake by red cabbage (*Brassica oleracea*). Soil-plant-atmosphere continuum relationships were examined in the model of Nimah and Hanks (1973a,b), while Herkelrath et al. (1977) developed a root extraction function which accounts for soil-root contact resistances to water uptake in a semi-empirical fashion.

4.2.1.2 Single Root or Microscopic Scale Approach

This procedure evaluates flow in the vicinity of a single root, of uniform water-absorbing properties and radius. Results are extrapolated to the whole root system, assuming roots are equally spaced. A sink term is not usually involved in the calculations, with rates of water uptake being determined by basic solution of the Richards equation. Gardner's model (1960) exemplifies this approach, wherein the root is approximated to an infinitely long cylinder, and water is assumed to move only in the radial direction. Results are multiplied by an 'average' root density in order that conclusions may be extended to the whole root zone. Difficulties in modelling flow to individual roots, in measuring root geometry and in assessing the comparative importance of root and soil resistances have severely hindered development of the microscopic approach.

In conclusion, although physically-based models are theoretically sound and provide accurate results, demands on input are large. Knowledge of root resistances, root geometry and leaf water potential is required, and unless this type of detailed information is available, empirical modelling techniques should prove more suitable.

4.2.2 EMPIRICAL MODELS

Empirical models generally represent soil moisture extraction in terms of one or more 'layers' from which water is removed according to a drying specification. Model 'layers' do not necessarily reflect physical horizons of the soil, but rather a temporal sequence portraying different components of the drying process. Actual extraction from each layer is determined according to potential extractive demand (normally potential evaporation) and remaining moisture content within the layer. The relationship between actual extraction and layer moisture content is usually described by a 'drying curve'. Runoff or drainage may occur only when the layers are filled to capacity although some models allow for 'direct recharge' without this prerequisite stipulation. Several 'drying curves' have been proposed to explain the relationship between actual:potential extraction ratio and layer moisture content, the exact form of the drying curve being a controversial issue. Most curves describe the drying process between the two end-points 'field capacity' and 'permanent wilting point', concepts themselves subject to much debate.

4.2.2.1 Field Capacity and Permanent Wilting Point

Field capacity, often used to calculate the upper limit of available water in the soil, was defined by Veihmeyer and Hendrickson (1949, p 75) as 'the amount of water held in the soil after excess water has drained away and the rate of downward movement of water has materially decreased, which usually takes place within 2 or 3 days after a rain or irrigation in pervious soils of uniform structure and texture.' It is sometimes equated with a soil water potential of 1/3 atm (0.34 bar, 34,000 Pa). The small range of soil moisture contents in which plants permanently wilt is denoted 'permanent wilting point' (PWP) and although it represents the lower limit of 'water availability', growth processes and transpiration can be inhibited before this point is attained, while under some circumstances, transpiration may continue beyond it (W.R. Gardner, 1965). Permanent wilting point may be defined as 'the root-zone soil wetness at which the wilted plant can no longer recover turgidity even when it is placed in a saturated atmosphere for 12hr' (Hillel, 1982, p.297). This point, nevertheless, remains arbitrary since plant water and soil water potentials may not reach equilibrium in this time. Permanent wilting is sometimes equated with a potential of 15 bar. Contrasting hypotheses concerning the relationship between available soil water and plant activity are discussed in the following section.

4.2.2.2 Evapotranspiration and Soil Moisture Content

Although numerous drying curve relationships have been suggested, they are broadly divisible into the following types (Fig. 4.1):

a) Actual Extraction Independent of Moisture Content

Water is equally available between 'field capacity' and 'permanent wilting point', and only when soil moisture is below the latter does extraction fall below the potential rate. This type of relationship was advocated, for example, by Veihmeyer and Hendrickson (1927, 1955) and Gardner and Ehlig (1963). The latter authors, using pot experiments on birdsfoot trefoil (*Lotus corniculatus*), cotton (*Gossypium hirsutum*) and pepper (*Capsicum frutescence*) concluded that little variation in transpiration rate occurs until the plants wilt, beyond which point, an almost linear relationship exists between water content and transpiration rate.

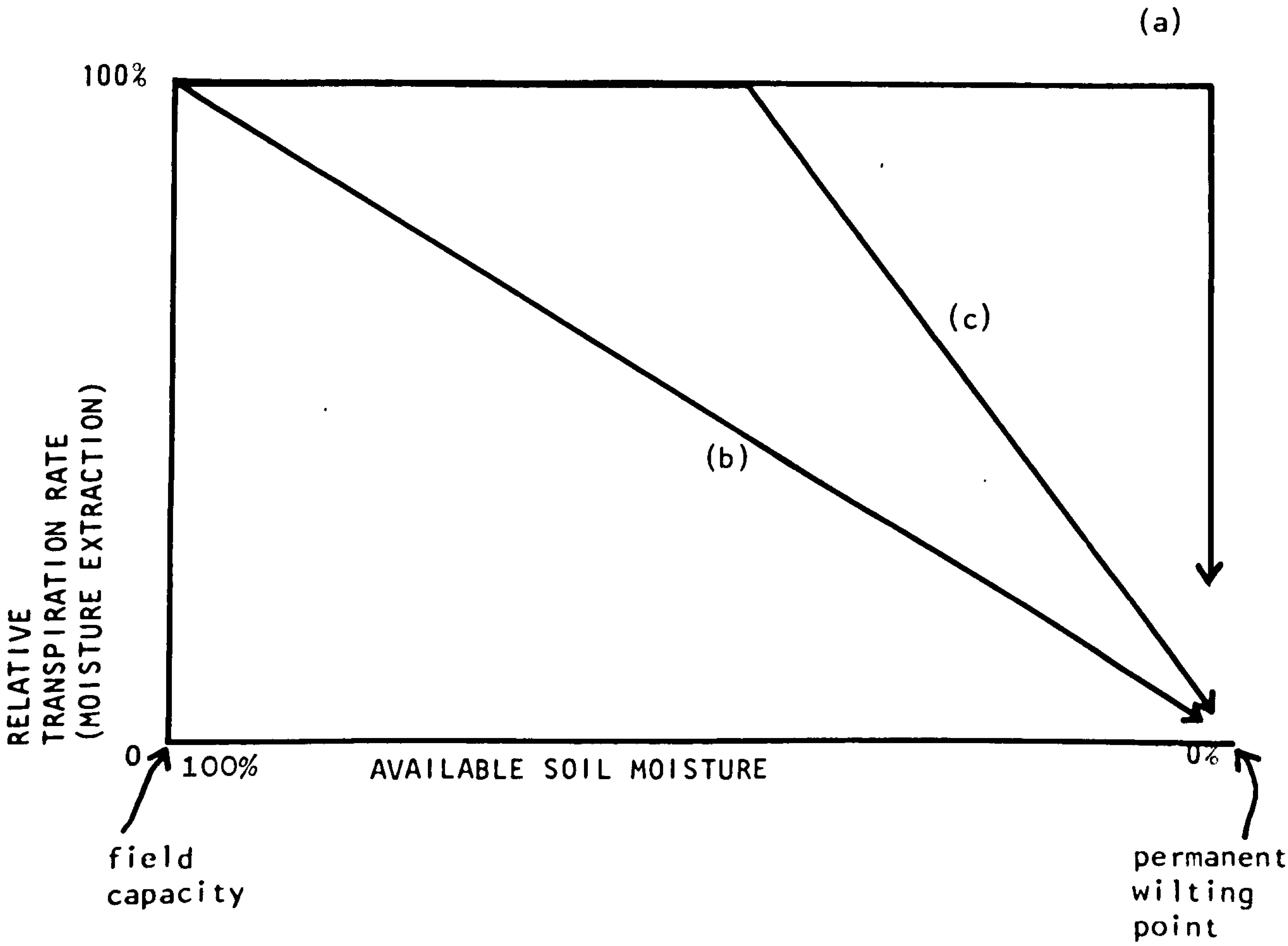
Veihmeyer and Hendrickson (1950, 1955) proposed three possible explanations for the rapid reduction in transpiration rate on reaching permanent wilting point:

i) A slight decrease in moisture content leads to a large increase in resistance to further water removal.

ii) The slowness with which water moves into dry soil around roots.

iii) Failure of roots to extend into areas holding moisture above permanent wilting point.

This type of drying curve may be applicable under conditions of low potential transpiration rate or under low soil moisture suctions (Denmead and Shaw, 1962).



(a) Soil moisture extraction independent of moisture content.

(b) Gradual reduction of extraction with decreasing moisture content.

(c) Intermediate drying curve (equal extraction from field capacity to a 'critical point', beyond which extraction decreases).

<u>Figure 4.1 Three Fundamental Hypotheses Regarding Soil Moisture Extraction by Plants in Relation to Available Moisture Content</u>

b) Reduction of Actual Extraction with Decreasing Moisture Content

Moisture stress may occur in the plant at any time before permanent wilting point is reached. Thornthwaite and Mather (1955) for example, proposed a linear decline in actual:potential evapotranspiration ratio (relative transpiration rate) with decrease in available soil water. Slatyer (1955) described results from plot experiments on grain sorghum, cotton and peanut, where limiting soil moisture resulted in a gradual reduction in transpiration in all cases. Transpiration was maintained for a longer period in sorghum than in the other two species, a fact which Slatyer explained in terms of sorghum's more extensive root system.

c) Intermediate Drying Curve

Soil water depletion proceeds at the potential demand rate until a critical point is reached, after which extraction is reduced as moisture content decreases. In certain instances the effects of varying environmental conditions have been built into the specified drying curve relationship, and Zahner (1967) proposed separate curves for differing soil textures. A roughly linear decline in actual:potential evapotranspiration ratios was evident for clay, while the curve for sand showed extended soil moisture depletion at the potential rate, followed by rapid alteration to curvilinearity towards wilting point, a result similar to that proposed by Penman (1949) for a range of soils. Rutter (1975) indicated, however, that in Zahner's work, textural differences may be partly compounded with variations in climate.

Similarly, Denmead and Shaw (1962) emphasised the effect of varying meteorological conditions on plant water availability. Under high potential transpiration conditions, for example, actual transpiration may fall below potential demand, even with adequate soil moisture content. Under low potential transpiration conditions however, actual and potential transpiration may be equal, down to very low soil moisture values.

The abundance of drying curves results from the considerable number of experiments carried out to evaluate the transpiration/soil moisture relationship, all conducted under different physical conditions, and with various plant species, soil properties and meteorological influences. No one curve is 'correct' but each may be applicable in particular circumstances. Penman (1963) remarked, for instance, that it may be that neither the curve of equal availability nor that of gradual decline in transpiration is precise at any time but that their inherent errors are probably less important than other sources of variability in their application.

More recently, Calder et al. (1983) reviewed the relative performances of empirical deficit models using increasingly accurate equations for the calculation of potential evaporation and increasingly sophisticated drying curves. For a grassland cover they concluded that site characteristics determine the choice of potential evaporation/regulating function combination. No particular drying function was nominated as being superior, although poor model fits resulted where actual evaporation was set permanently to potential.

4.2.3 MODEL SELECTION

The selection of a soil moisture model, theoretical or empirical, depends ultimately on the degree of accuracy required and the data available. Adoption of one of the simpler empirical models limits input information required, while empirical models in general are thought to be more practical and easier to use than the theoretical type. Further, empirical models may provide results comparable to those of some of the less complex theoretical models, where the latter fail to consider the importance of phenomena such as hysteresis. Application of an empirical or semi-empirical relationship between evapotranspiration and soil moisture status is generally preferred for field-based studies, while theoretical models are more suited to detailed assessments of plant water relations. In the light of these conclusions, empirical modelling techniques are employed for the present study.

One of the earliest empirical soil moisture models was that developed by Penman (1949) which, until recently, has been utilised by the Meteorological Office in an adapted form to predict soil moisture deficits on a nationwide scale (Grindley, 1960,1967,1970). Soil moisture deficit maps were prepared as a service to farmers and hydrological authorities (water authorities, water supply engineers, etc.) to predict irrigation needs, flood levels, groundwater recharge and reservoir replenishment. This 'Penman-Grindley' combination is a single-layer empirical model in which differences between rainfall and Penman estimates of potential evaporation are adjusted according to a specified drying relationship, and used to predict values of soil moisture deficit and actual evapotranspiration. Grindley and Singleton (1969, p.812) formally defined soil moisture deficit as 'the cumulative effect of evaporation minus rainfall with, as initial condition, the ground at field capacity and hence a zero deficit'. A soil moisture deficit is established when potential evaporation exceeds rainfall and soil moisture reserves are depleted (Grindley 1969).

The Meteorological Office Rainfall and Evaporation Scheme (MORECS) (Wales-Smith et al., 1976; Wales-Smith and Arnott, 1980) was introduced in 1978 to replace the Penman-Grindley model. It has a two-layered structure and is based on the premise that evapotranspiration causes abstraction of all the moisture from one layer before water is lost from the second layer. This represents a more realistic concept since uniform moisture distribution and abstraction are rare in natural soil profiles.

In 1959 Holmes and Robertson developed the 'modulated moisture budget'. This double-layered model accommodates both changing rooting depths and moisture stress during soil drying. All available moisture is initially evaporated from the top layer, at the potential rate. Depletion then proceeds from the second layer at a decreasing rate, depending on amount of remaining moisture and root distribution. This model was subsequently improved by Baier and Robertson (1966) with the introduction of their 'versatile budget' variation. This permits simultaneous withdrawal of moisture from several layers of varying capacities and also has the facility to involve any of a number of drying curves. The model includes estimates of runoff and drainage, the latter typically being assumed on attainment of a moisture excess. Shaw (1963) developed a multi-layered model to predict soil moisture under corn. Evapotranspiration is set to a constant rate for the early part of the year, but otherwise varies according to drying curves based on those of Denmead and Shaw (1962). As with other multi-layer models, however, filling of each layer is still performed in a sequential manner.

Stuff and Dale (1978) objected to those empirical models which take no account of capillary rise, which, in their opinion represents an important parameter in shallow water table areas. Consequently, their model introduced an assessment and prediction of this phenomenon, estimated from soil moisture deficits and water table depths. Capillary rise is also considered in the soil water balance model developed by Makkink and van Heemst (1974) using data from a polder in the Netherlands. This comprehensive empirical model represents the soil profile as a series of dynamic zones, between which water transfer is allowed. Both saturated and unsaturated zones are included, their capacities varying in accordance with water table depth. Unsaturated soil is further subdivided into a layer which is depleted by evaporation (the evaporation zone) and a zone from which water is extracted by plants, the transpiration zone. The capacity of the latter varies with crop development, while actual transpiration is calculated with reference to a reduction factor, which itself depends on the available moisture content of the transpiration zone. Normal percolation to the saturated soil zone is a function both of the amount of excess water in the unsaturated zone and of the height of this surplus (when 'collected' at the top of the profile) above the water table. Similarly, capillary rise is calculated from deficits in the unsaturated zone and the height of collected deficit above the water table. Although a detailed and representative empirical procedure, this model may be hindered in practice by the need for definition of several variables.

A semi-empirical model developed by Walley and Hussein (1982) employs a physically-based approach within an empirical framework. This model allows transfer of moisture between a total of four profile layers, fluctuations in water table level being accounted for. Root abstraction is calculated for each rooting zone layer, while evapotranspiration is evaluated for soil and plant factors separately, in relation to their relative areal coverage. Walley and Hussein advocated more extensive use of the model, application of 'typical' soil and plant parameters precluding the need for field calibration.

The aims of the present investigation demand a quantitative measure of vegetation effects using the hydrological data available. Ideally the selected model should allow optimisation of parameters such that 'best fits' may be chosen for prediction. As the Penman-Grindley or Estimated Soil Moisture Deficit model has undergone widespread application and been used for almost twenty years by the Meteorological Office, it is employed in the present study for reasons of reliability and compatability. The MORECS model is assessed both in its own right and in terms of potential improvement over Grindley estimates. Both models require only basic input data, provide objective land-use comparisons and moisture predictions, and enable parameter optimisation.

4.3 THE PENMAN-GRINDLEY MODEL

4.3.1 BACKGROUND AND APPLICATION

The Penman-Grindley model, first introduced for regional prediction in 1960 and refined in 1967 and 1970 (Grindley 1960, 1967, 1970), is conceptually simple, depending only on inputs of daily rainfall and potential evaporation, together with land-use data. It relies on soil drying processes proposed by Penman (1949) in the form of a single parameter drying curve (Fig. 4.2). In a state of zero moisture deficit, any additional rainfall is assumed to contribute to runoff. Soil moisture is extracted at the rate of atmospheric demand (potential evaporation) until the soil moisture deficit exceeds a specified, empirical 'root constant' value. This point marks 'a specified amount of soil moisture (expressed in mm equivalent depth) which can be extracted from the soil without difficulty by a given vegetation on a given soil' (Grindley, 1970, p 204) and which varies in magnitude between land-use types. Beyond this point, a further 25 mm (1 inch) of extraction occurs at the rate of potential evaporation, representing extraction from below the root zone, after which actual evaporation drops rapidly to 0.1 of the potential rate as the soil dries further. Tables of actual/potential evaporation relationships for a series of root constants, based on the Penman curve were given by Grindley (1969).

Penman (1949) calibrated the drying curve by predicting return to field capacity (as indicated by a resumption of flow in drain-gauges) on an experimental farm in Cambridge. He suggested that for general application under grass cover, the root constant is of the order of 75 mm, although he recognised that its value would vary depending on meteorological conditions, particularly in the early growing season. To account for this he proposed a relationship between root constant and spring rainfall:

$$c = 5.0 - 0.6\Sigma R \qquad \text{Eq. 4.5}$$

where:

c = root constant (inches)

ΣR = sum of April and May rainfalls

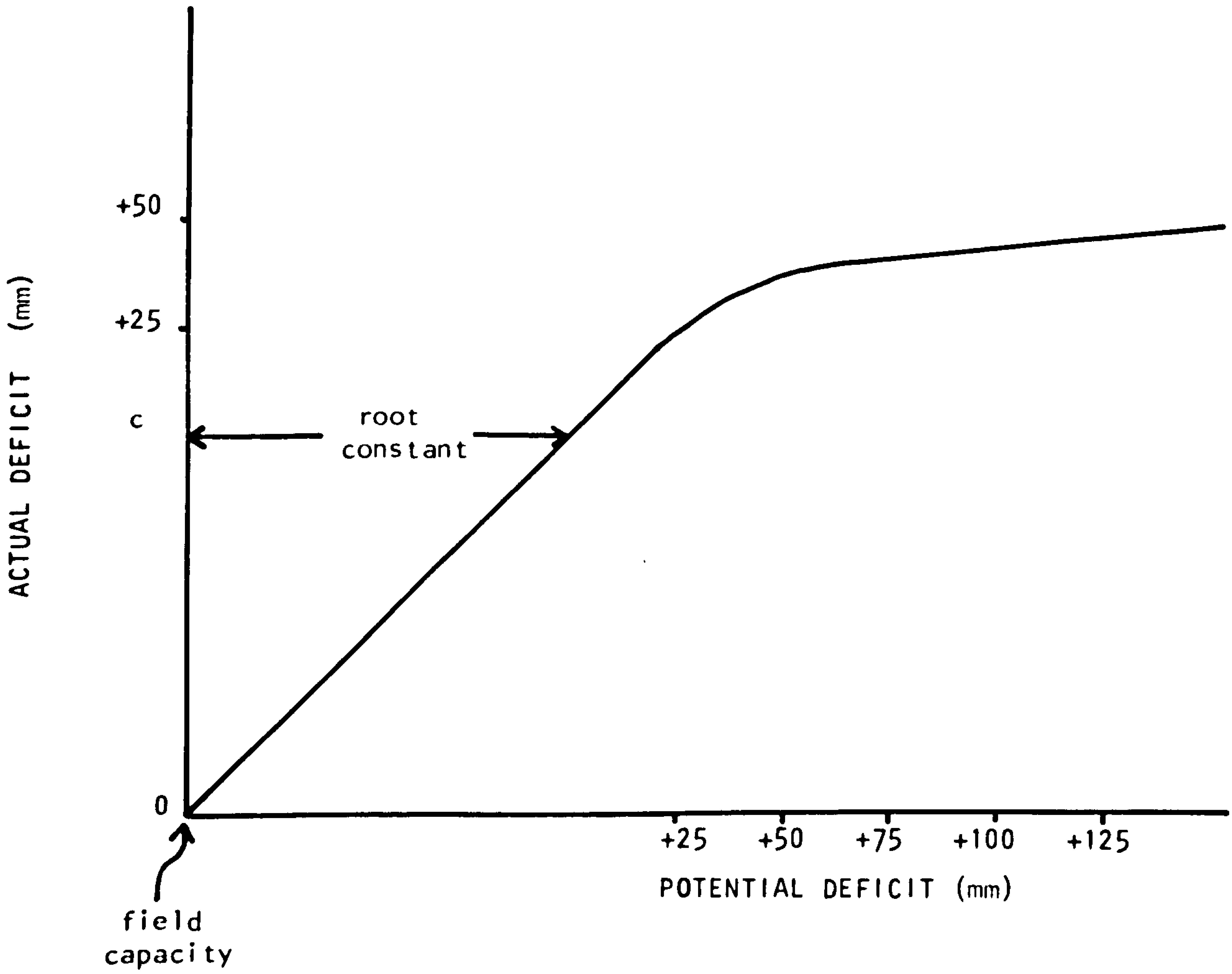


Figure 4.2 The Penman Drying Curve [after Penman (1949)]

4.3.2 APPLICATION TO STUDY AREA

For employment of the Grindley model at the catchment scale, a series of maximum actual soil moisture deficits and associated root constants have been devised for a variety of crops (Grindley, 1969, 1970). The root constant values take no specific account of variations in soil type except in so far as they reflect the nature of the vegetation, although root constants can be reduced accordingly for investigations involving a poor soil base (Grindley, 1969).

4.3.2.1 Preliminary Analysis

Before subjecting the data to soil moisture analysis, it was necessary to eliminate the possibility of soil moisture variation due to slope changes within the catchment, since significant differences in slope angles were found between woodland and moorland (vegetated and burnt areas combined) ($t = 1.863$, d.f. = 28, significant at 0.05 level, 1-tail test). Two-way analysis of variance (ANOVA) tests were executed on several runs of data comprising total profile moisture contents (0 cm to 80 cm), at randomly selected access tube sites, and calculated as described later, in Section 4.3.2.3. Some of the selected data sets displayed normal distributions (Fig 4.3) in accordance with more general findings (for example, Nielsen et al., 1973; Hills and Reynolds, 1969; Bell et al., 1980), while non-normal data were transformed by taking the square-root of each value, and substituting these for raw measurements. The analysis was restricted to soil moisture content, as the normality of soil moisture deficit data remains questionable. Slope angles were divided into categories of 'low' (0^{o} to 4.5^{o}), 'medium' (4.6^{o} to 9.0^{o}) and 'high' (9.1^{o} to 13.5^{o}), and vegetation was classified as moorland or woodland. For

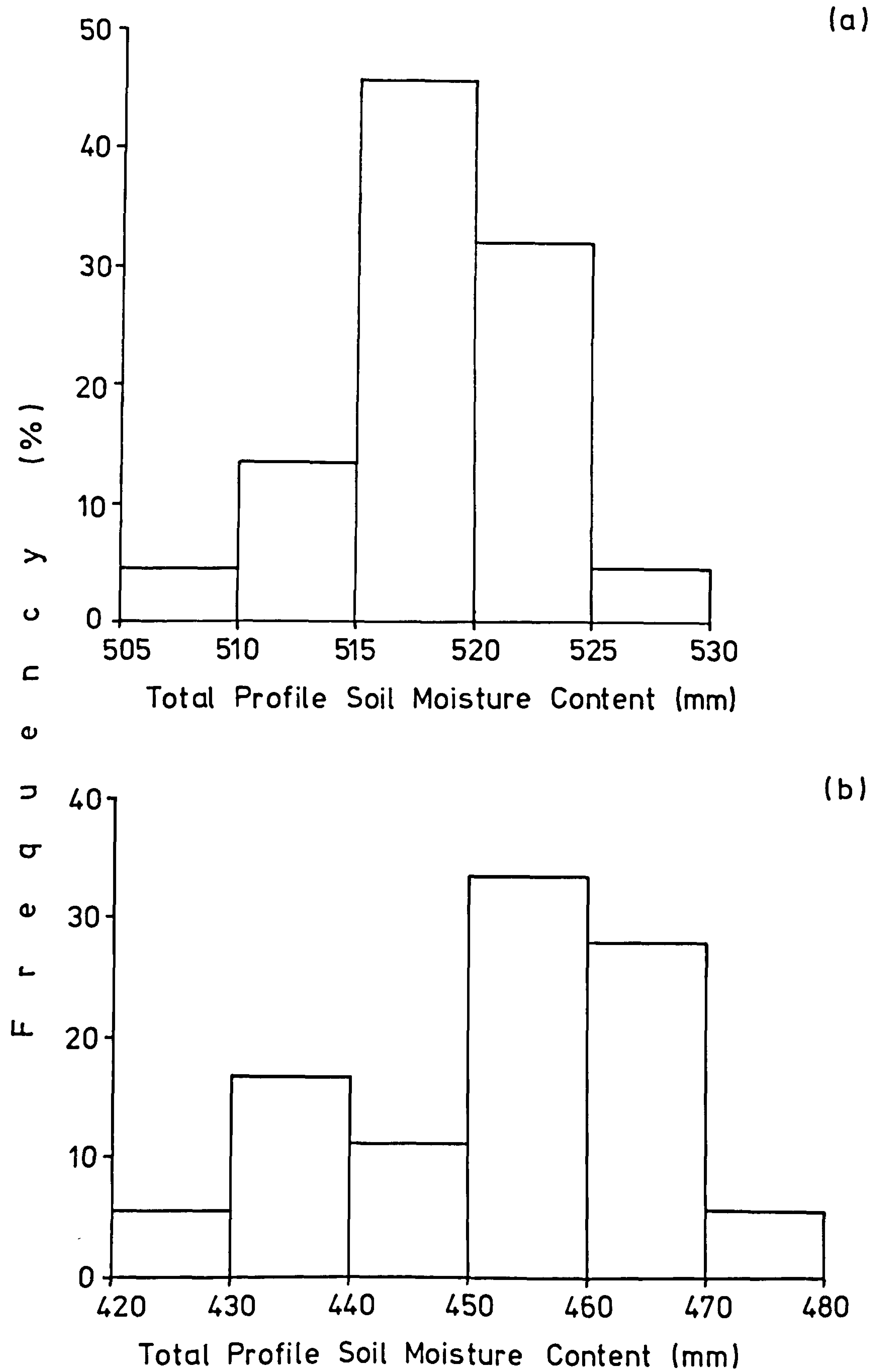


Figure 4.3 Selected Normal Frequency Distributions for Soil Moisture Content (a) Access Tube Site 19; (b) Site W4 using Data for Winter Months

each run, vegetation proved to have a more significant effect on soil moisture (typically $F = 53.1, 72.5$, d.f. in each case 1,103) than did slope angle. Further, slope means generally showed smaller deviations from the grand mean of all observations than did vegetation means (Table 4.1).

4.3.2.2 Model Input Data

The Grindley model version implemented in the present study operates on a 'plot' scale, in which each of the three vegetation types (heather moorland, burnt ground and woodland) is classified as one plot area. The drying curve, as specified by Penman, is supplied to the model as co-ordinate pairs of calculated potential soil moisture deficit and actual soil moisture deficit. Daily rainfall and Penman evaporation figures are based on records from the automatic weather station on Sneaton High Moor. The remaining input requirement is that of moisture deficit values.

4.3.2.3 Calculation of Soil Moisture Deficit

Soil moisture deficit equates to the amount of water lost from the soil by evaporation. In practice, the difference between soil moisture content and field capacity yields an estimate of profile soil moisture deficit. More realistic assessment of moisture deficit, on a soil layer basis is discussed in the following section. Summation of water content values for a series of soil profile depths yields total moisture content (mm) for the profile. Moisture content values are calculated by assuming that the moisture volume fraction, determined from the neutron probe calibration curve, represents moisture contained in a 100 mm depth of soil, the neutron probe count depth interval. A moisture volume fraction of 0.58, for example, represents 58 mm water per 100 mm soil. This assumption is applied to measurements made at 20 cm depth and below; readings at 10 cm below

Vegetation	Slope Category	Access Tube No.
Moorland	Low	15
	Medium	13
	High	3
Woodland	Low	W5
	Medium	W4
	High	W7

Grand Mean = 22.07

Slope:	Low	Medium	High
Mean Soil Moisture Content (mm)	22.94	21.86	21.39
Deviation from Grand Mean	0.87	-0.21	-0.68

Vegetation:	Moorland	Woodland
Mean Soil Moisture Content (mm)	23.51	20.62
Deviation from Grand Mean	1.44	-1.45

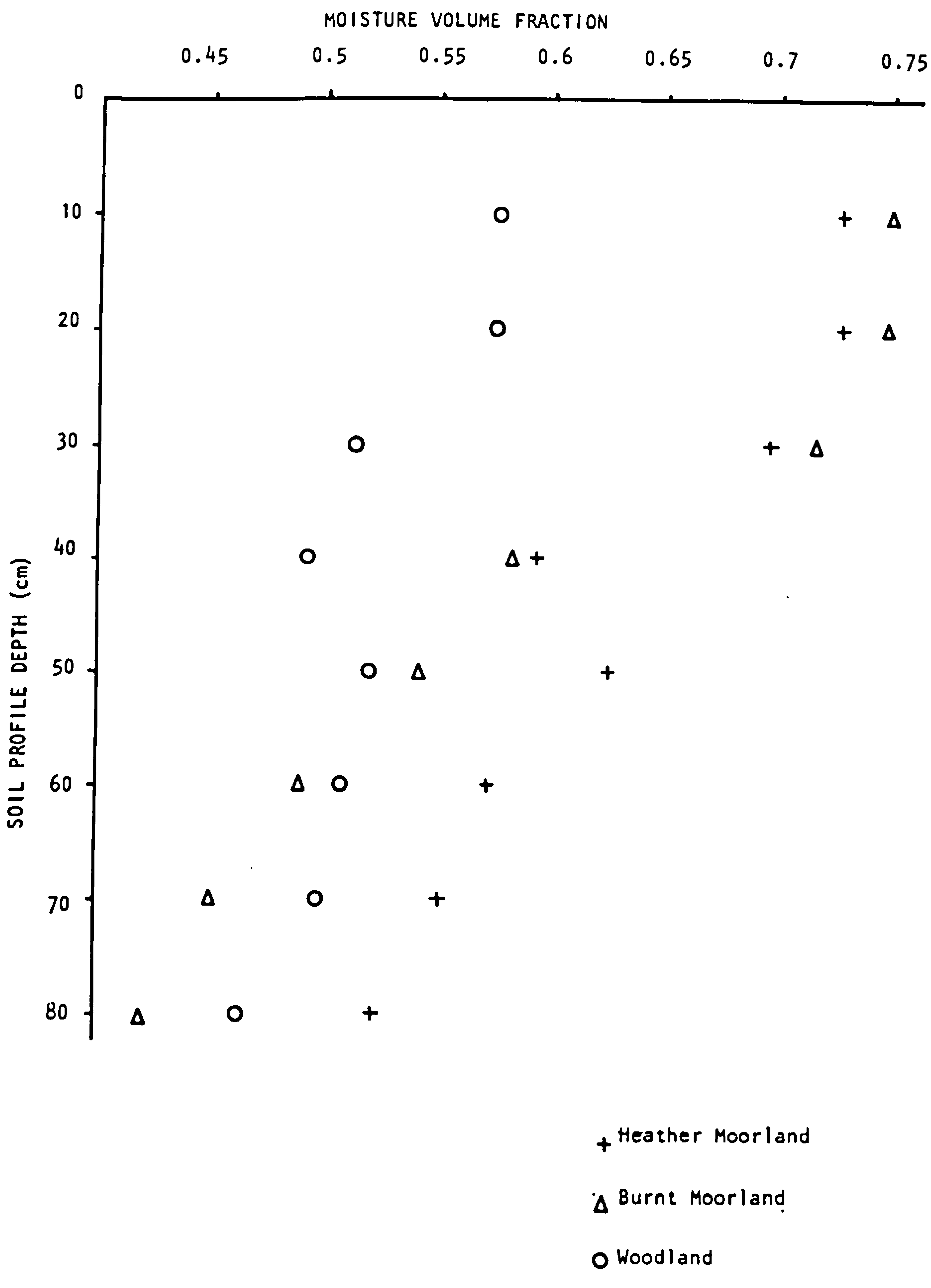
Table 4.1 ANOVA: Deviations from the Grand Mean of Soil Moisture Observations for a Sample Data Set (Transformed Data)

the surface are taken to represent the 0 cm to 15 cm layer and are therefore calculated as a fraction of 150 mm. Moisture volume fraction profiles (Fig. 4.4, Appendix II) may thus be converted to amounts of water held in the profile on each measuring occasion. Moisture content values are summed to yield total moisture content of part or all of each profile site monitored, and, through calculation of soil moisture deficit, enable objective evaluations of variations between land-use types.

4.3.2.4 Field Capacity Assessment

An estimate of field capacity needs to be established before determining moisture deficits for the area. A great deal of controversy surrounds this concept, and its theoretical basis has undergone much criticism. Field capacity as a soil water 'constant' has little physical meaning since the soil-plant-atmosphere continuum operates as a dynamic system. 'Available water', held between field capacity and permanent wilting point, is only one of several factors which affect plant water uptake and which are not described by the drying curve. Such factors centre upon the ability of roots to extract water and the ability of soil to supply it (Hillel, 1982). Field capacity remains, however, a useful concept for field applications and continues to be utilised for practical purposes. Only a brief outline of some of the arguments surrounding the use of this constant is therefore presented here.

Although field capacity is claimed to represent a state after excess water has drained away, this taking two to three days, it has been suggested that downward water movement may be appreciable for a considerably longer period of time. Hillel (1980b) contended that in medium- and fine-textured soils, redistribution can proceed at a significant rate for many days. This extended period of drainage was



<u>Figure 4.4 Typical Moisture Volume Fraction Profiles (average for October 1981-March 1982)</u>

alternatively considered by Veihmeyer and Hendrickson (1949) to be so slow in comparison with the rate of extraction of water by plants, that in practice it can be ignored. Similarly, although it is thought that significant drainage may occur after 48 hours from permeable soils with a deep water table, Hall and Heaven (1979) found losses after this time to be insignificant in sandy soils of the Newport Series, being within the variation in neutron probe count rate due to radioactive decay. Additional complications may arise in attempting to evaluate field capacity in shrinking/swelling soils. Delayed wetting may be a characteristic of a swelling clay which can hold large quantities of water on expansion (Hall and Heaven, 1979). This is not considered to be a significant problem for the clay at Egton, however, since it is largely non-montmorillonitic in composition (Carroll and Bendelow, 1981). Similar volumetric changes in the organic layer however, may induce errors in available water estimation. Although textural changes within the soil profile may affect the soil's storage capacity, no account is taken of this when determining field capacity, which may vary from around 9% (by volume) in sands to about 55% in peat (Rutter, 1975).

A further criticism is that field capacity results depend on the measuring technique used (Hillel, 1980b). Several means of arriving at a field capacity value exist, including laboratory estimations, water balancing from field measurements and optimisation procedures. Burrows and Kirkham (1958) suggested that laboratory determinations are unreliable where the soil profile contains extrinsic influences such as layers of different antecedent soil moisture, or small lenses of extraneous material. Specific determination was described by Hall and Heaven (1979) who selected individual depth field capacities as the mean of moisture content readings on days in winter and spring when little or no rain fell for at least three days before the date of measurement. In a consideration of the MORECS model, Gardner and Bell (1980) assumed field capacity to be that moisture content pertaining on the spring date after which the model specifies a continuous deficit. Calder *et al.* (1983) optimised field capacity, by minimising the sum of squares of differences between observed and predicted deficits. The procedure is restricted to winter data.

The present study employs neutron probe field measurements of soil moisture to ascertain field capacity values for both the complete soil profile and selected profile depths. Mean winter moisture content values are computed to represent a 'total profile' field capacity for each land-use category (535 mm for heather moorland, 507 mm for burnt moorland and 450 mm for woodland). Total profile moisture deficits are then assigned by subtracting moisture content from field capacity.

The method adopted here to calculate 'layer' field capacities and thus 'layer profile' deficits, represents an attempt to overcome the difficulty of separating drainage from evapotranspiration, an inherent problem associated with field capacity as a 'constant'. The procedure largely follows that of McGowan (1974) who has used the method to identify drainage in a variety of soils and crop types. The approach avoids the need to collect hydraulic conductivity and tension measurements, required by other methods of drainage estimation, and which may require calibration against soil moisture content measurements taken at a separate location. Layered soils and those subject to temporary waterlogging are difficult to assess theoretically, and may easily be subjected to McGowan's method of drainage separation.

Average soil moisture contents for each measured profile depth are plotted against time and inspected for a point of changing gradient (Fig. 4.5, 4.6 and 4.7). McGowan identified this point as marking the arrival of a drying front and associated it with root water extraction. The moisture content pertaining at this time is taken as field capacity for that particular layer, and layer moisture deficits are derived by subtraction of moisture contents from the resulting field capacity value. Additionally, times of deficit prior to and following the main deficit period are included, as identified from potential deficit values, discussed in Section 4.3.3.1, where these indicate existence of a deficit rather than drainage. In McGowan's study, the movement of the drying front through the profile as identified by this method, often corresponded with the movement of the 'zero flux plane', the depth at which soil water flux is zero, and, in cereals, it corresponded with maximum rooting depth. In this way, it is assumed that moisture above the effective depth is that which is extracted by roots, while that below comprises drainage. Water content changes identified above the zero flux plane therefore reflect upward fluxes (evapotranspiration) and those below, drainage. Potentially, therefore, the technique provides verification of soil water flow calculations.

Failure to assess simultaneous root extraction and downward drainage is a weakness of this and other methods of drainage estimation (for example, that using the zero flux plane). McGowan (1974), however, indicated that for the arable crops he examined, drying occurs from surface layers downwards and that the possibility of roots extracting water while the latter is undergoing downward movement represents only a small error. Drainage separation may become arbitrary when rates of root extraction and soil drainage are

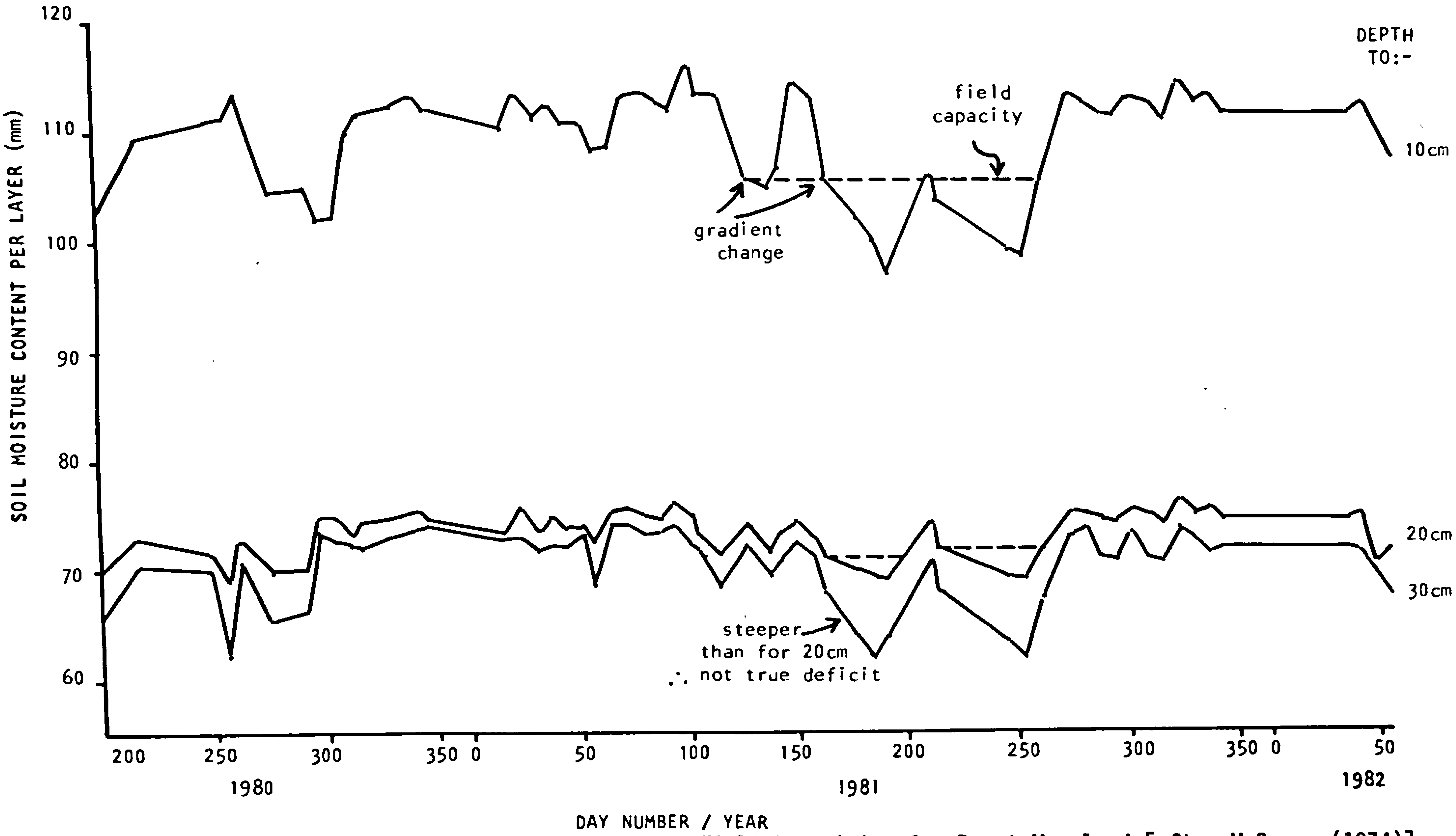


Figure 4.5 Drainage Separation to Identify Layer Field Capacities for Burnt Moorland [after McGowan (1974)]

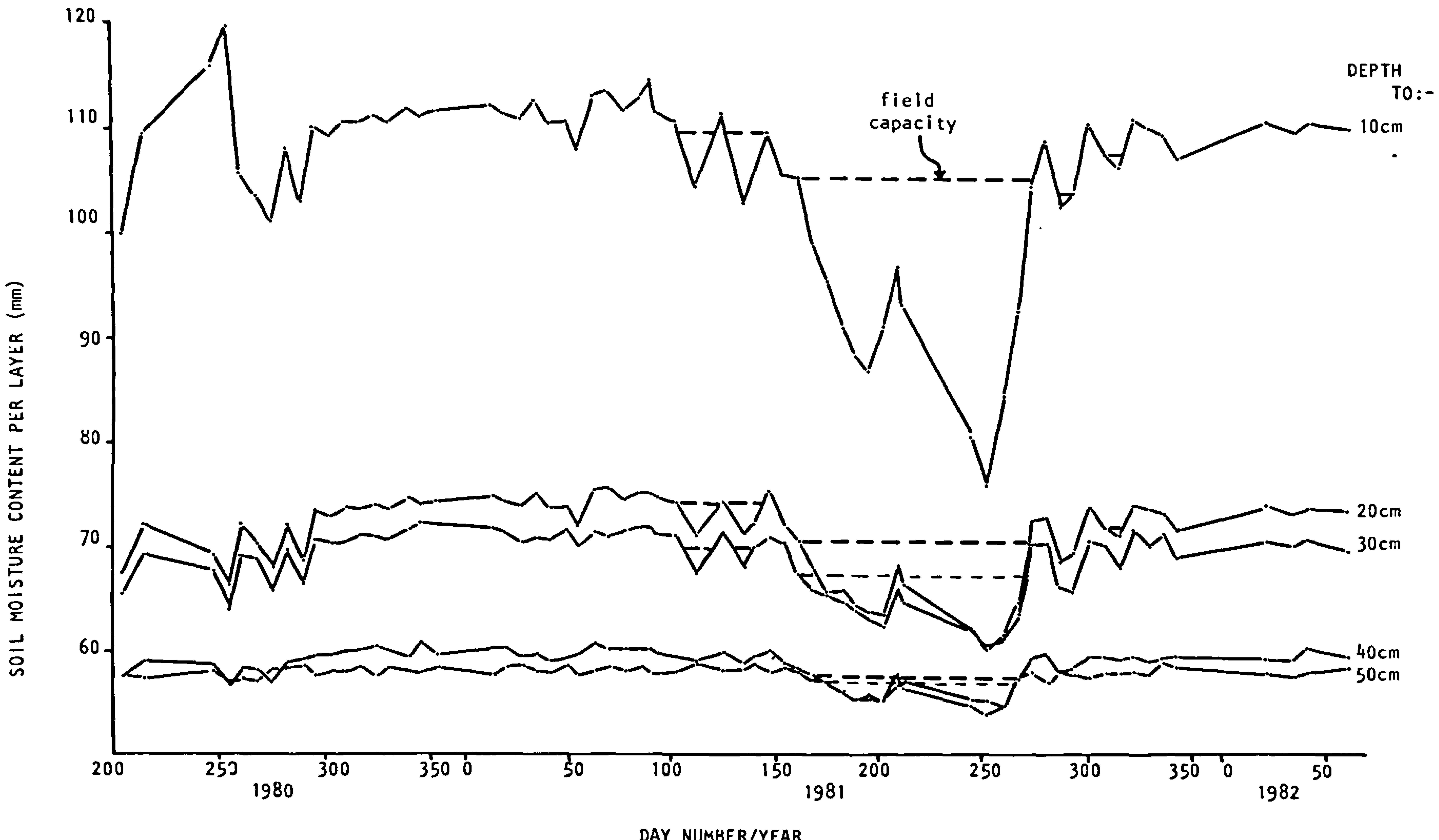


Figure 4.6 Drainage Separation to Identify Layer Field Capacities for Heather Moorland [after McGowan (1974)]

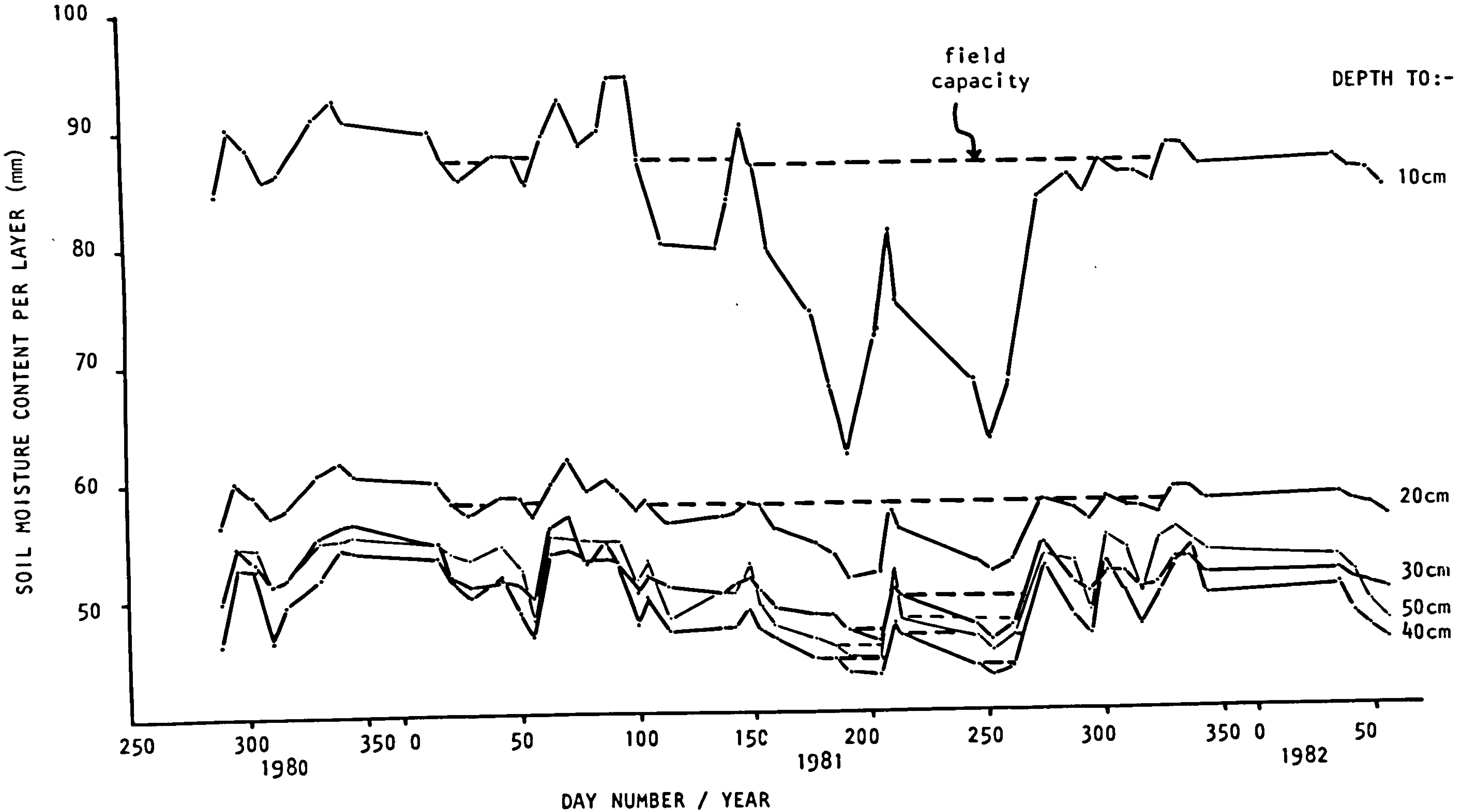


Figure 4.7 Drainage Separation to Identify Layer Field Capacities for Woodland [after McGowan (1974)]

similar, although McGowan reported this as occurring only where the time rate of change of water content is small (approximately 1% per month).

4.3.2.5 Extraction Depths

Total profile moisture deficits were derived from soil moisture measurements to 80 cm, under the assumption that all maximum rooting depths would be covered. An effective maximum depth, the depth to which roots extract measurable quantities of water, was subsequently identified for each land-use from Figures 4.5, 4.6 and 4.7 as the depth at which little change in moisture content occurred over time. Allocated values were as follows:

Land-Use	Effective Maximum Depth (cm)*
Heather moorland	50
Burnt moorland	20
Woodland	50

* from McGowan's (1974) method of drainage separation

These results are generally verified by site observations (Chapter 2) and to some extent by other rooting depth evidence. *Calluna vulgaris* is generally a shallow-rooting species, down to between 10 cm and 18 cm in peat (Boggie, 1956) with slight activity to 60 cm (Boggie *et al*., 1958). Gimingham (1960) specified *Calluna's* well-branched nature and depth as being determined largely by soil conditions, especially soil moisture. He maintained that rooting is generally restricted to organic layers of the soil and is inhibited under waterlogged conditions. Rutter (1955) noted the confinement of *Calluna* roots to tussocky areas in the wet heaths of south east

England but his results, in certain cases, showed the plant's distribution to be unrelated to water table depth and fluctuation. He concluded that geological and physiographic factors are most important in influencing water table levels. Calluna vulgaris seedlings were found by Bannister (1964a) to show reasonable root development in all except the wettest (above field capacity) moisture regimes tested. The species is characterised by winter dormancy of root growth (Bannister, 1976) and by its lack of root hairs (normally the zones of maximum water uptake). Under such circumstances, water is often absorbed over the whole root surface (Sutcliffe, 1968).

Of the two main tree species in Wintergill Plantation, Pinus contorta (lodgepole pine) and Picea sitchensis (sitka spruce), the former is the more tolerant of wet soil conditions. P.sitchensis may suffer windthrow in waterlogged peaty areas, as a result of its surface-rooting regime (Sanderson, 1977). Moisture extraction patterns, along with site inspections indicate that roots of both woodland species are generally confined to surface soil layers. In this way, conifers in general are able to adapt to a high water table and avoid the reducing conditions of lower horizons (Armstrong, 1982). P.contorta may, however, extend a few centimetres below the water table and, in so doing, abstract water and lower the water table as it roots; air entrapped in the root enables it to root more successfully than P.sitchensis (Coutts and Philipson, 1978). Similarly, Boggie (1972) found P.contorta roots to be confined generally to aerobic horizons above the water table, with rooting, which was predominantly lateral, showing improved growth with water table lowering. Surveys in north Wales and the English/Scottish border regions prompted Fraser and Gardiner (1967) to conclude that the overall mean rooting depth of Picea sitchensis on surface-water gleys and peaty clays with impeded drainage was 42 cm (16.5 inches) the most common rooting depths being between 30.5 cm and 45.7 cm (12 inches to 18 inches). This rooting pattern is also common in Pinus contorta (Coutts and Armstrong, 1976).

4.3.3 MODEL RESULTS

From model input information (rainfall, potential evaporation, actual weekly soil moisture deficits and drying curve co-ordinates) the Penman-Grindley model predicts daily values of soil moisture deficit using a specified or calculated root constant term for each land-use category. An error term, the root mean square error (RMSE) is included in model output, best model fits being those with the smallest error terms:

$$\mathrm{RMSE} = (F^2/n)^{0.5} \qquad \text{Eq. 4.6}$$

where:

n = number of observations

$$F^2 = \sum_{i=1}^{n} (\mathrm{SMD}_{oi} - \mathrm{SMD}_{pi})^2$$

SMD_{oi} = ith observed soil moisture deficit

SMD_{pi} = ith predicted soil moisture deficit

To define field capacity with any confidence a complete year of data is desirable. Similarly, since the model assumes a zero deficit at the start of calculation, it is inadvisable to commence model runs with data for summer months when actual deficit may not be zero. Results presented here, therefore, largely refer to information collected during 1981 only.

4.3.3.1 Actual and Potential Soil Moisture Deficits

An elementary need for soil moisture modelling for the study site is indicated by comparisons of actual and potential soil moisture deficits. The latter are calculated from the cumulative difference between daily Penman potential evaporation and rainfall figures. Should the difference become negative, potential deficit is reset to zero and calculation recommences. Figure 4.8 illustrates that potential evaporative demand is insufficient to assess soil moisture conditions at the site since actual, total profile deficits are significantly smaller than potential values, for all land-use categories. Maximum actual deficits are only 25% (burnt moorland) to 50% (heather moorland and woodland) of equivalent potential values. Profiles of potential deficits for 1980 are similar in magnitude to those of 1981, although the main deficit period begins and ends at earlier dates (Fig. 4.9).

Two main factors are responsible for the discrepancies between actual and potential deficits. Firstly, under dry conditions, because of high surface resistances, actual evaporation of heather and conifers remains below Penman potential. This disparity should be corrected by application of a drying curve model. Secondly, as Penman (1956) defined potential evapotranspiration assuming a 'short green crop ... of uniform height', transpiration rates from heather and conifers should be less, again because of higher surface resistance. Still greater deviation is shown between potential and actual deficits under a bare surface.

The extent to which actual and potential moisture deficits diverge depends largely on vegetation and soil conditions, to which no consideration is given in the calculation of potential deficits using the Penman equation. An improved correspondence between the two deficit types may therefore be provided by replacement of Penman with Penman-Monteith evapotranspiration data. Penman-Monteith estimates for woodland were drawn directly from computer printouts of automatic

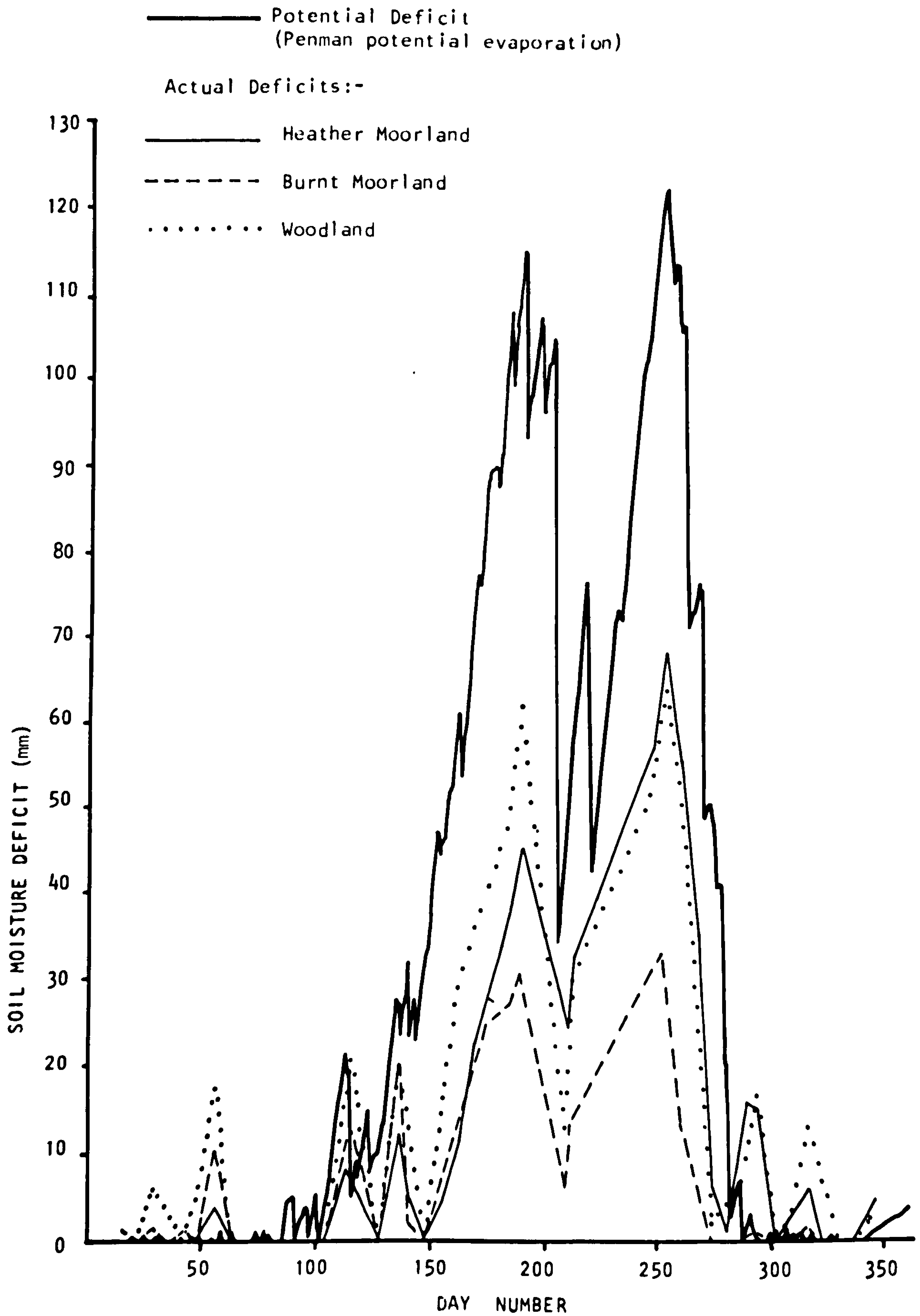


Figure 4.8 Comparison of Actual and Potential Soil Moisture Deficits (1981)

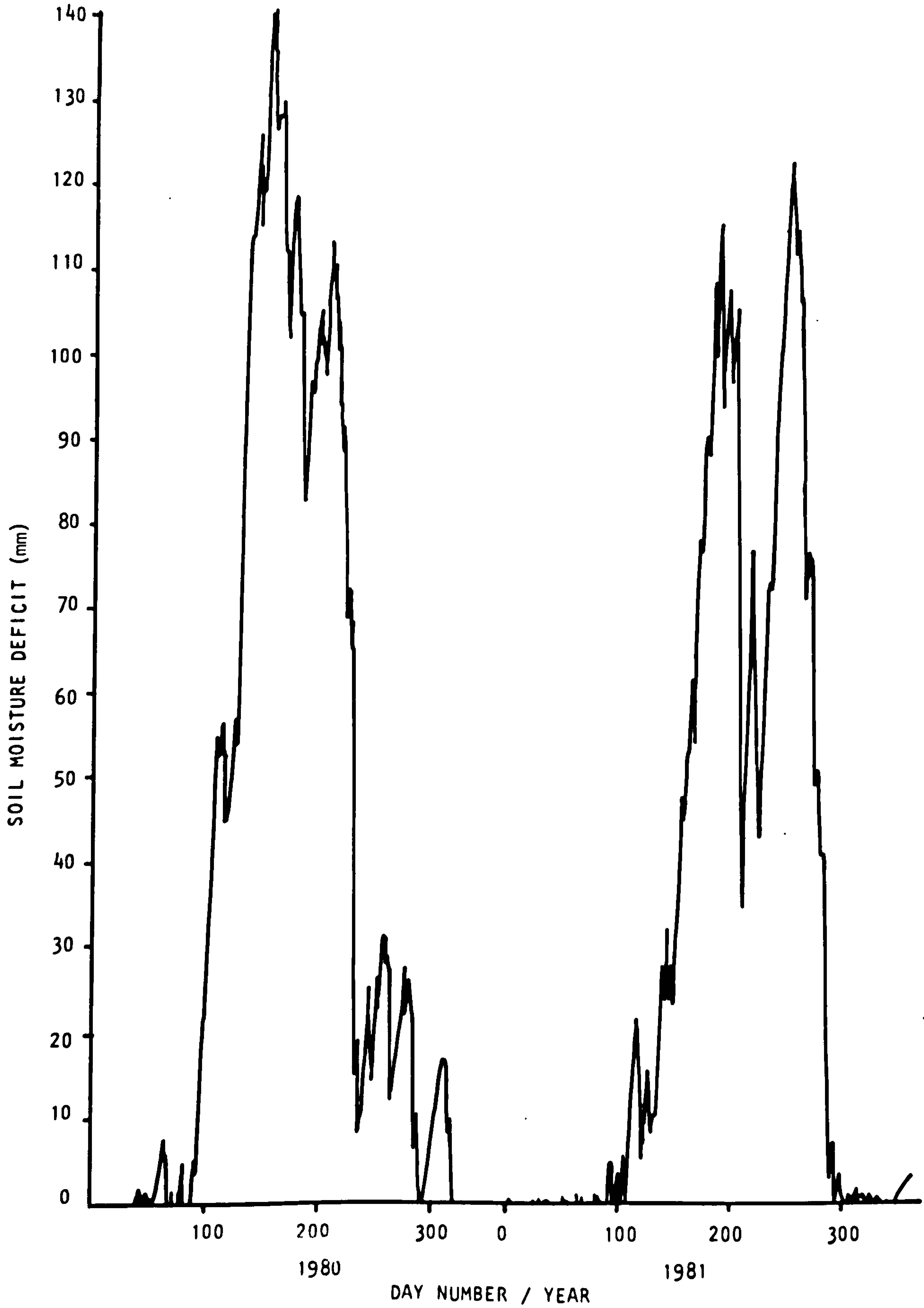


Figure 4.9 Potential Soil Moisture Deficits (1980-1981)

weather station data, calculated using measurements of the 'average' surface resistance of a Scots pine (<u>Pinus sylvestris</u>) forest in Thetford, Norfolk (pers. comm., J. Wallace) and are, strictly speaking, only applicable to forests with similar surface and aerodynamic resistances to those found in Thetford Forest (r_s = 120sm^{-1} to 150sm^{-1} ; r_a = 5sm^{-1} to 10sm^{-1} (Stewart and Thom, 1973; Gay and Stewart, 1974; Gash and Stewart, 1977). These estimates were, however, deemed quite adequate for the purposes of this dissertation. Evaporation of intercepted water is accounted for in the calculations by setting surface resistance to zero when vegetation is wet, that is, during rainfall and immediately after rain for as long as the atmospheric humidity deficit is below 1.0 g kg^{-1} (pers. comm. J. Wallace).

Values of evapotranspiration for heather and burnt moorland were not supplied directly and were therefore computed using the meteorological information recorded by the automatic weather station, along with appropriate resistance values. Estimates of surface resistance were derived from work by Thompson <u>et al</u>. (1981), and are shown in Table 4.2. Daily aerodynamic resistances were supplied by the automatic weather station for heather, while Monteith's calculation (1965) was used to determine daily values for burnt moorland, assuming a crop height of 0.05m (Thompson <u>et al</u>., 1981):

$$r_a = \frac{[\ln((z-d)/z_0)]^2}{k^2 u(z)} \qquad \text{Eq. 4.7}$$

where:

u = mean wind speed at height z (m s^{-1})

d = zero plane displacement,

= 0.6 x vegetation height = 0.03 m

z_0 = roughness length,

= 0.1 x vegetation height = 0.005 m

k = von Karman's constant,

= 0.41

Precise estimation of interception under heather is difficult because of the several developmental stages of this species and calculation here was based on a simple indicator described by Thompson et al. (1981), using leaf area index (LAI) to determine the proportion of rainfall, P intercepted:

$$P = 1 - (0.5)^{LAI} \qquad \text{Eq. 4.8}$$

where LAI is set to 3.5

Interception amount is allocated a daily maximum threshold of 0.2 LAI mm and is also corrected to allow for enhanced interception capacity due to multiple daily storm events. More rigorous estimates of interception are obtainable from the more complex models devised by Rutter et al. (1971, 1975) or Gash (1979), although this degree of detail was not considered appropriate for the present study, due to the empirical nature of the soil moisture estimation component of the models under investigation.

The results of the evapotranspiration index substitution are shown in Figures 4.10 to 4.12. A factor of 1.5 still separates actual and potential deficits during the summer period with actual deficits exceeding potential values under both heather and woodland. The underestimated potential deficits may be explained either by overestimated surface and/or aerodynamic resistance values in the Penman-Monteith equation, or by underestimation of interception. The latter possibility is quite plausible for this type of high rainfall environment.

	$r_s (sm^{-1})$
Upland	120 (Jan, Feb, Mar, Oct, Nov, Dec) 100 (Apr-Sep)
Bare Soil	100

(adapted from Thompson *et al*., 1981)

Table 4.2 Surface Resistance (r_s) Values for Heather (Upland) and Burnt Moorland (Bare Soil)

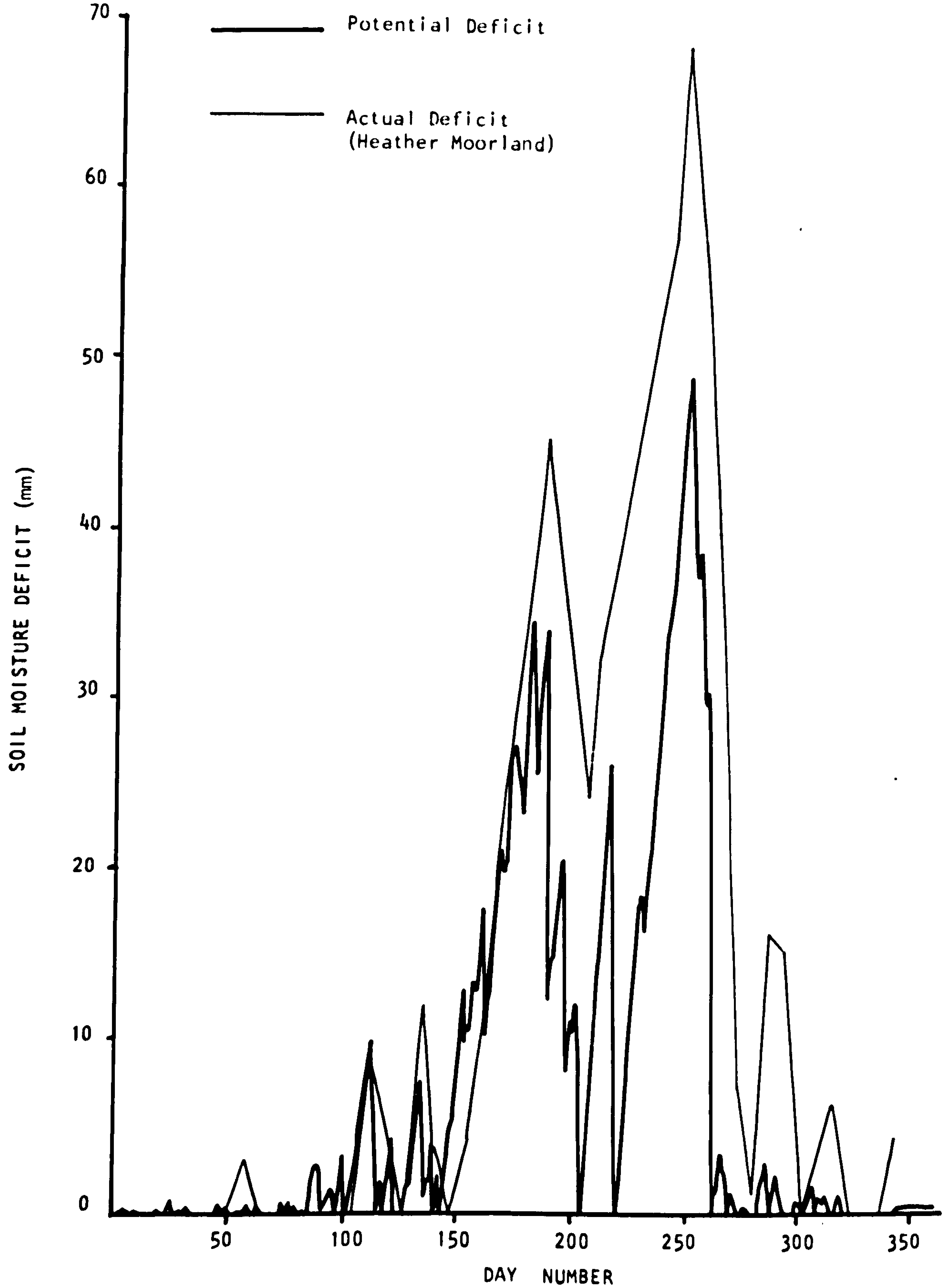


Figure 4.10 Comparison of Actual and Potential Soil Moisture Deficits for Heather Moorland using Penman-Monteith Potential Evapotranspiration (1981)

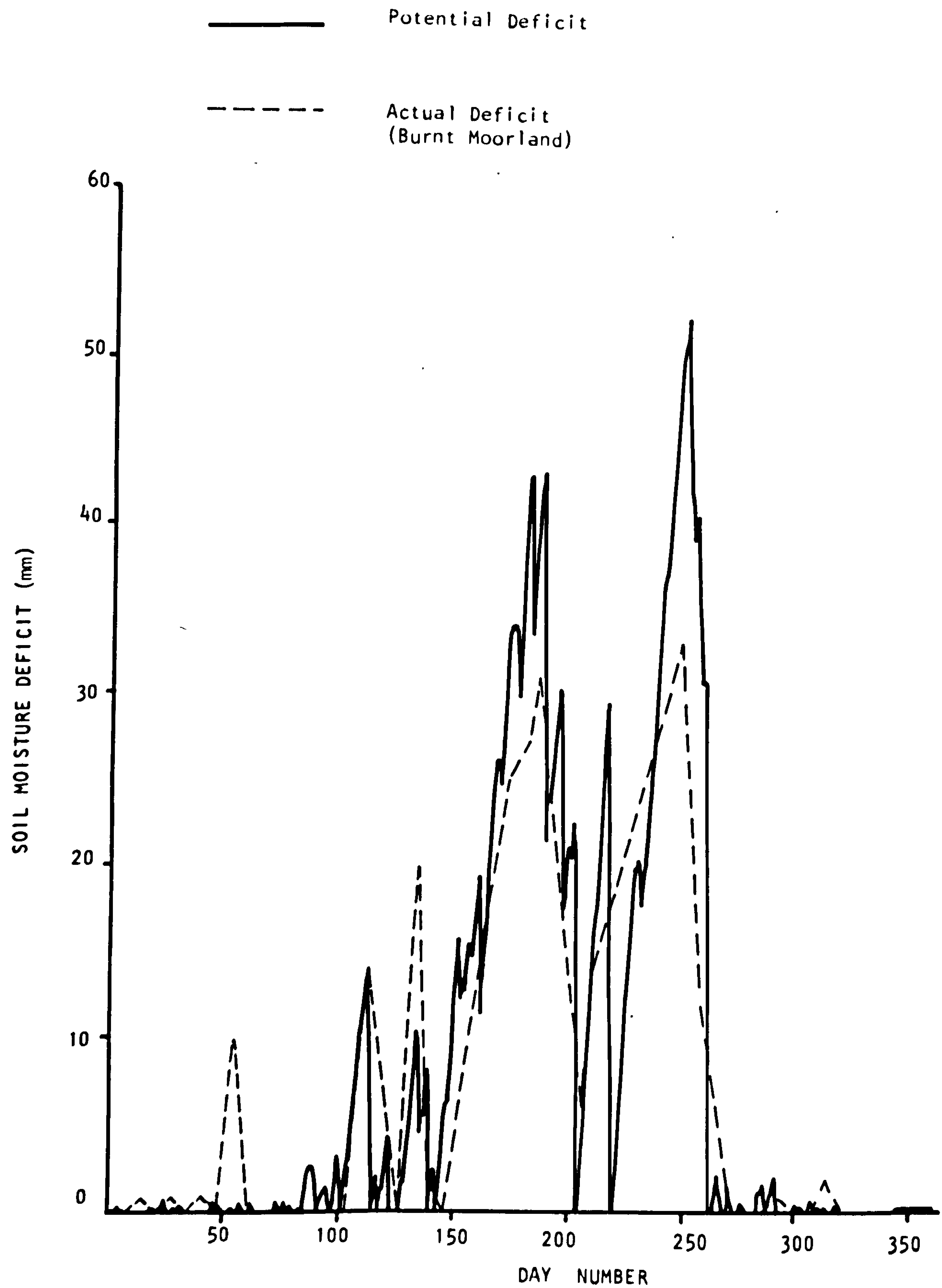


Figure 4.11 Comparison of Actual and Potential Soil Moisture Deficits for Burnt Moorland using Penman-Monteith Potential Evapotranspiration (1981)

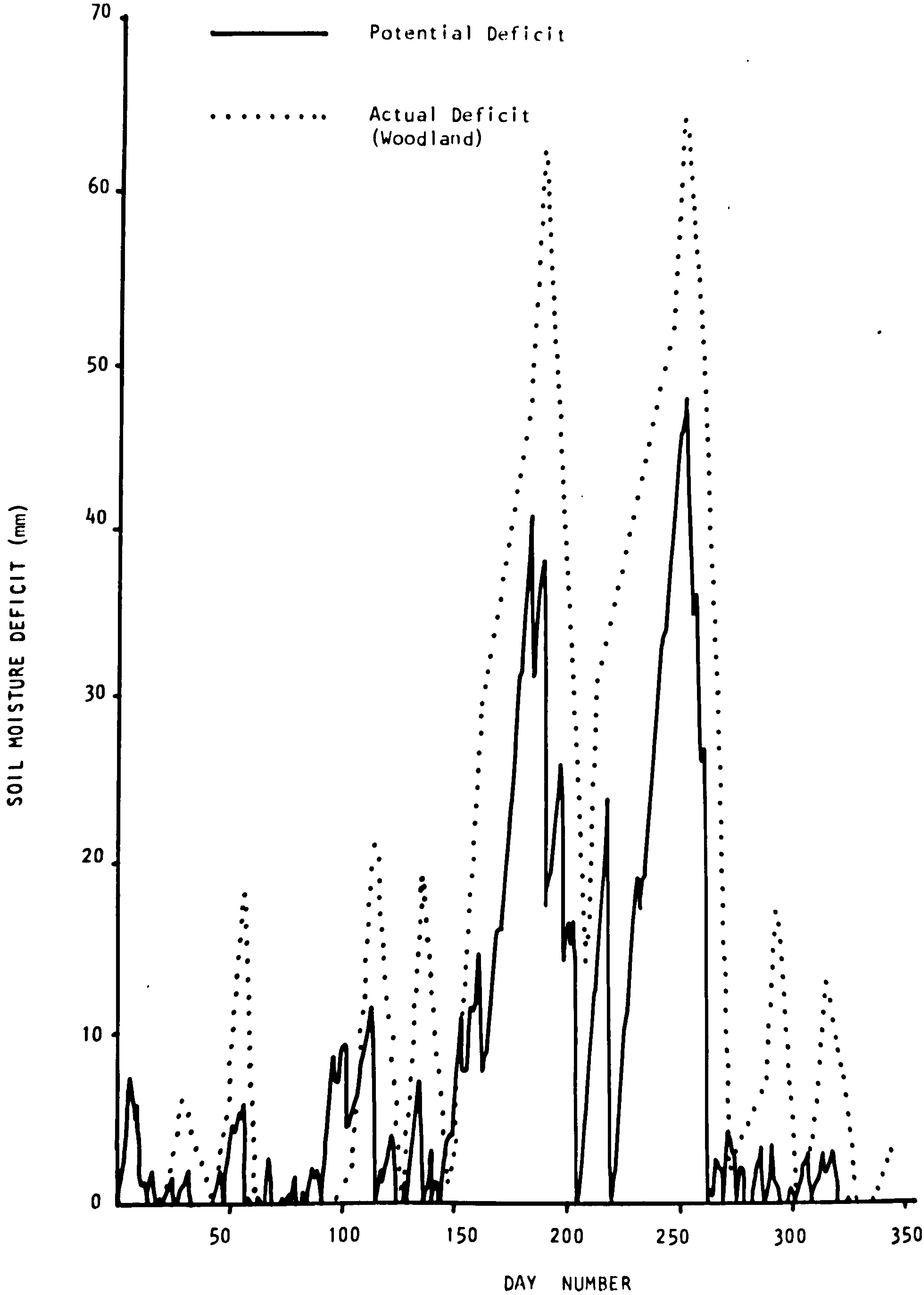


Figure 4.12 Comparison of Actual and Potential Soil Moisture Deficits for Woodland using Penman-Monteith Potential Evapotranspiration (1981)

4.3.3.2 The Drying Curve Model

Initial model runs were carried out using the Penman-Grindley model based on the root constant values recommended by Grindley (1969), these being 12 mm, 0 mm and 200 mm for heather moorland, burnt moorland and woodland, respectively. In addition, to investigate whether optimisation of the root constant value yields improved model simulation, a series of optimisations was implemented to determine the 'best' root constant for each vegetation type. Optimum root constants were selected as those associated with the smallest error terms from a series of trial root constant simulations. Both total profile deficits and those for maximum extraction depths were subjected to the complete procedure, all simulations relating to data for 1981.

4.3.3.3 Heather Moorland

a) Penman Evaporation

Using the recommended root constant of 12 mm and values of potential evaporation calculated from the Penman equation, the Grindley model simulates total soil moisture deficits for heather moorland with a root mean square error (RMSE) of 13.69 (Fig. 4.13). Root constant optimisation yields a best value of zero, which produces a marginally better simulation, with RMSE of 13.171 (Fig. 4.14). Both sets of predictions reflect the overall deficit pattern, distinguishing separate profile peaks although, in detail, demonstrate poor fits. Actual deficits are overestimated by the model throughout late spring and early summer and are underestimated later in the season. Several factors may explain these discrepancies which characterise a number of other simulations, discussed later. Firstly, the model assumes a complete crop cover, and reduced transpiration rates which result from limited plant development in the early season are disregarded. Allocation of an average, annual root constant

Figures 4.13 to 4.20:
Grindley Model Simulations for Heather Moorland

Figure 4.13 Simulation based on Penman Evaporation, Total Profile Deficits and Recommended Root Constant

Figure 4.14 Simulation based on Penman Evaporation, Total Profile Deficits and Optimised Root Constant

Figure 4.15 Simulation based on Penman Evaporation, Layer Deficits and Recommended Root Constant

Figure 4.16 Simulation based on Penman Evaporation, Layer Deficits and Optimised Root Constant

Figure 4.17 Simulation based on Penman-Monteith Evaporation, Total Profile Deficits and Recommended Root Constant

Figure 4.18 Simulation based on Penman-Monteith Evaporation, Total Profile Deficits and Optimised Root Constant

<u>Figure 4.19 Simulation based on Penman-Monteith Evaporation, Layer Deficits and Recommended Root Constant</u>

Figure 4.20 Simulation based on Penman-Monteith Evaporation, Layer Deficits and Optimised Root Constant

results in deficit overestimation in the early season, following root dormancy, and underestimation later in the year, with increased plant available water. Secondly, the drying curve shape which results in potential evaporation pertaining for a prolonged period may create exaggerated estimates in spring when evaporation should have been lowered below the potential rate. The model attempts to compensate for early-season deficit overestimation by underestimating during late summer and early autumn. Finally, a comparison of Figures 4.8 and 4.10 shows potential deficits based on Penman's formula to be similar during early and late season periods, while late season potential deficits derived using the Penman-Monteith formula are enhanced over equivalent early season estimates. Incorporation of Penman-Monteith evaporation data, discussed in a subsequent section, should therefore improve the model fit.

Return to a zero deficit state in autumn is represented by the model to within two to three weeks assuming end of year measured 'deficits' are, in reality, drainage. Identification of these spurious 'deficits' as drainage is supported by plots of actual deficit in the top 50 cm of the profile, the depth of maximum extraction for heather, which suggest that no significant deficit occurs after day 273 (end of September) (Figs. 4.15 and 4.16). The recommended root constant profile for total deficits predicts runoff during summer (days 204 to 205) while using the optimised root constant, return to zero deficit is additionally shown for day 220. Both predictions are supported by stream-level records, indicating that the heather plot was contributing to runoff at this stage. Day 204 marked the beginning of a large storm event, storm runoff continuing for some three to four days, while day 220 (8 August) was characterised by the recession of a storm which began two days previously. Missing soil moisture measurements restrict interpretation for the month of August (days 213 to 245) for these and all subsequent simulations. Soil moisture/runoff response relationships are considered further in Chapter 6.

With regard to layer deficits and a maximum extraction depth of 50 cm, a slightly poorer fit is apparent. For predictions using the recommended root constant, the RMSE increases to 15.769 (Fig. 4.15). This arises from the reduced moisture deficit values, the predicted plot being identical to that for the 0 cm to 80 cm soil moisture profile. Optimisation yields a negative root constant (-21 mm) only 4 mm moisture being evaporated at the potential rate, and a model fit error of 10.182. The generally uniform predictions throughout the year improve compatabilities for spring deficits (days 102 to 145) although rates of drying during early summer are still misrepresented and late season deficits again underestimated (Fig. 4.16).

b) Incorporation of Penman-Monteith Evaporation

In order to overcome some of the inadequacies of the Penman evaporation formula, the model was tentatively re-run substituting Penman-Monteith values of evaporation for those derived by the Penman formula. With regard to total profile deficits, incorporation of the recommended root constant for heather yields slightly improved simulations over those produced using Penman evaporation (RMSE = 11.105) (Fig. 4.17). These results should be treated with caution however, as the recommended root constants were not originally intended for use with Penman-Monteith evaporation. The predicted rate of rise in spring and early summer matches that observed and onset time for the main deficit period is indicated accurately. Whilst early season deficits show some improved representation therefore, deficits are still underestimated later in the year. This may be ascribed to an overall lowering of evaporation estimates by the Penman-Monteith formula over those generated by Penman's equation, by inclusion of a crop-specific surface resistance factor in the former equation. Further improvement in simulation fit results from root constant optimisation (RMSE = 10.575) (Fig. 4.18). Visually, the predicted profile is almost identical to the previous simulation incorporating the recommended root constant. Comparison with potential deficits (Fig. 4.10) demonstrates that optimisation is to potential deficits.

Applying the model to layer deficits of the top 50 cm produces the lowest root mean square errors of all simulations for heather moorland (6.404 for recommended root constant and 6.086 for optimised root constant). Deficit peaks are more accurately represented and underestimation in the late season is reduced in comparison to most of the previous plots (Figs. 4.19 and 4.20). Elimination of spurious deficits, in reality drainage, accounts for some of the improved fit over total profile simulations, while incorporation of Penman-Monteith evaporation estimates results in a more realistic simulation of overall temporal distribution of deficits.

Optimisation is again towards potential deficits, which may indicate that actual moisture deficits are, in fact, close to potential ones. Alternatively, actual and potential deficits may diverge, with actual values falling below potential, but exaggerated resistance values used in the Penman-Monteith calculations may have resulted in subdued evaporation and deficit estimates. Employment of collective values for r_s (Table 4.2) may have contributed to these errors, and inclusion of more frequent surface resistance estimates (daily, for example) would improve accuracy. The onset of spring drying is pre-empted by the model by about twenty days, although the prediction for total profile plots (Figs. 4.17 and 4.18) is more accurate. Errors in layer deficit calculation may therefore account for the difference.

4.3.3.4 Burnt Moorland

Since moorland vegetation was burnt early in 1981 (day 100) it was thought unnecessary to adjust data for the early season when deficits are typically low; each model plot therefore assumes a bare surface for the complete year. The effect of this on the resulting root constant is expected to be minimal since its optimisation relies largely on summer data.

a) Penman Evaporation

Grindley recommended a zero root constant for burnt (bare) ground but, for total profile deficits, as for equivalent heather moorland simulations, the model fit is poor (RMSE = 10.07) (Fig. 4.21). Deficits are in the main overestimated by the model throughout the summer months, while the onset of spring drying is predicted about six weeks in advance (layer deficit simulations (Figs. 4.22 and 4.23) indicate that early season 'deficits' are likely to represent drainage from the soil profile). Re-wetting is predicted more accurately, at about eleven days early. Optimisation yields a better model fit (RMSE = 4.671) for a negative root constant (-21 mm, that is, only 4 mm of evaporation at the potential rate) (Fig. 4.24). Deficit overestimation in the early season is reduced, but only at the expense of simulation later in the year when peak deficits are underestimated. Summer runoff is suggested by both fits for days 204 and 220 and the optimised plot accurately predicts runoff on two additional occasions in July (days 190 and 198).

Figures 4.21 to 4.28:
Grindley Model Simulations for Burnt Moorland

Figure 4.21 Simulation based on Penman Evaporation, Total Profile Deficits and Recommended Root Constant

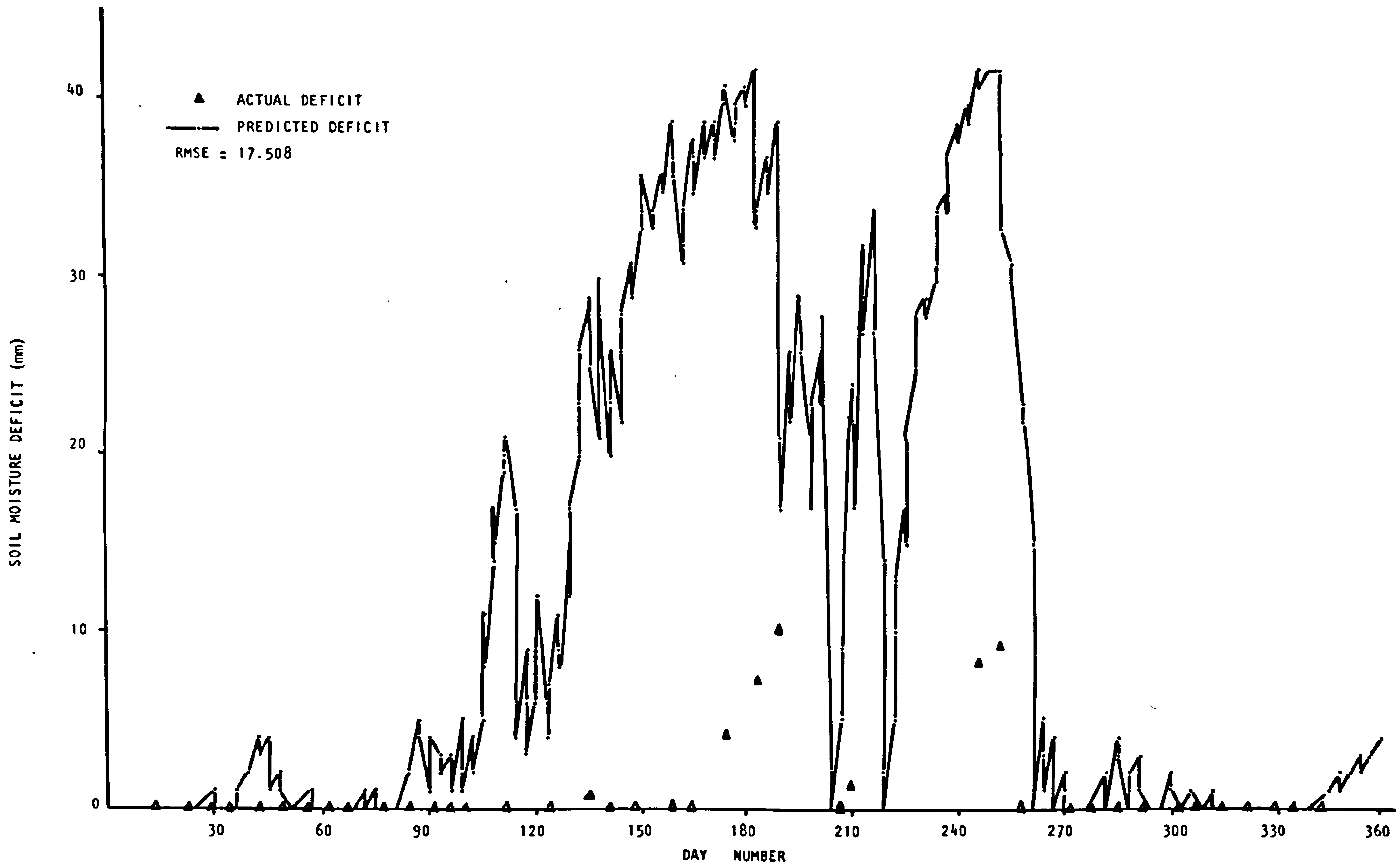


Figure 4.22 Simulation based on Penman Evaporation, Layer Deficits and Recommended Root Constant

Figure 4.23 Simulation based on Penman Evaporation, Layer Deficits and Optimised Root Constant

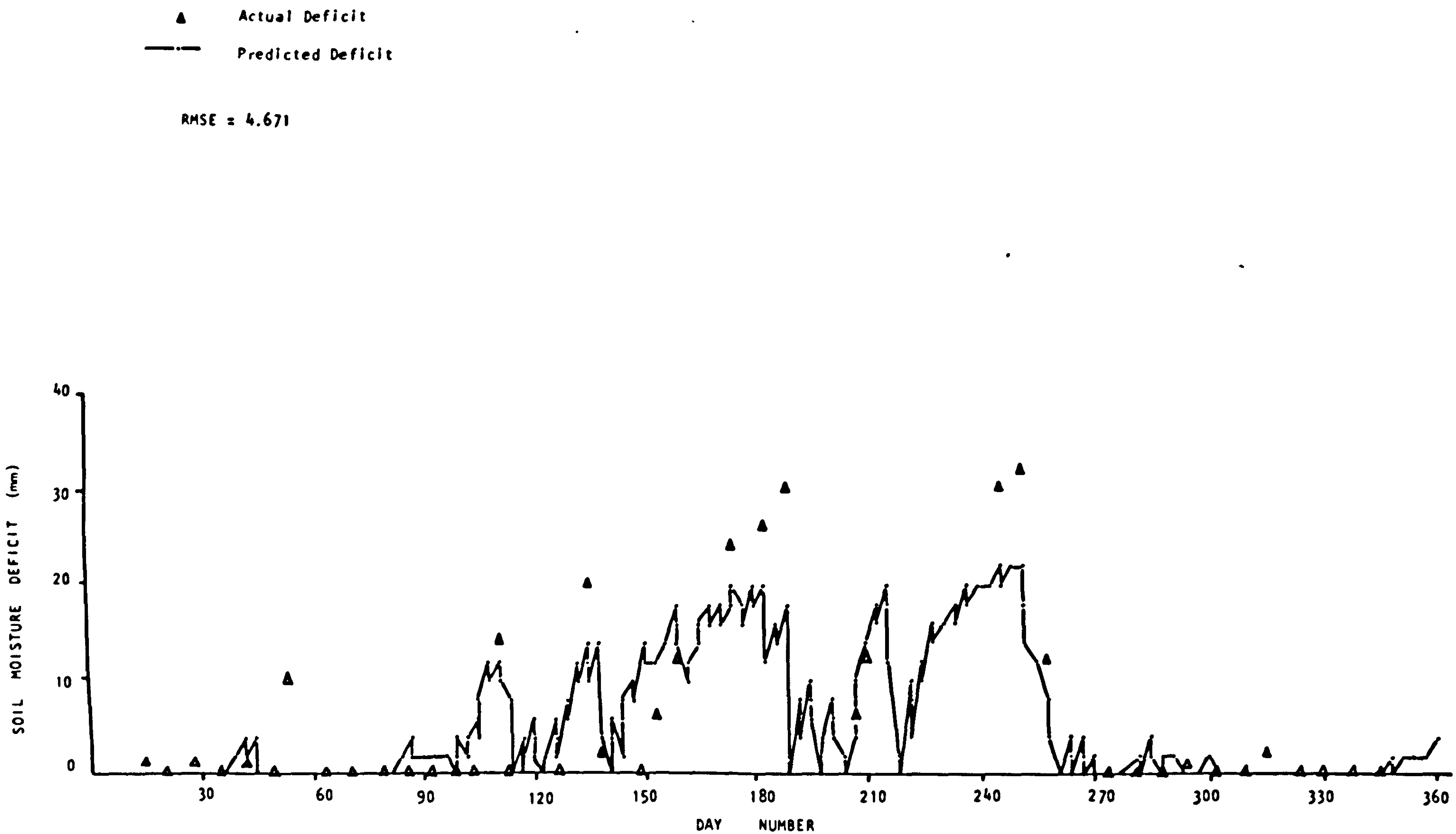


Figure 4.24 Simulation based on Penman Evaporation, Total Profile Deficits and Optimised Root Constant

Figure 4.25 Simulation based on Penman-Monteith Evaporation, Total Profile Deficits and Recommended Root Constant

Figure 4.26 Simulation based on Penman-Monteith Evaporation, Total Profile Deficits and Optimised Root Constant

Figure 4.27 Simulation based on Penman-Monteith Evaporation, Layer Deficits and Recommended Root Constant

Figure 4.28 Simulation based on Penman-Monteith Evaporation, Layer Deficits and Optimised Root Constant

With regard to layer deficits and a maximum extraction depth of 20 cm the small deficits of the top layer of bare soil inevitably lead to significant discrepancies between model simulation using the recommended root constant and actual values (RMSE = 17.508) (Fig. 4.22). Predicted deficits are reduced by the optimised plot (root constant = -25 mm) although misrepresentation of evaporation from this surface by Penman's formula still gives rise to deficit overestimation throughout the year (Fig. 4.23).

b) Penman-Monteith Evaporation

Simulations based on Penman-Monteith evaporation again prove more representative than those using Penman evaporation (Figs. 4.25 and 4.26). Root mean square error is reduced to almost one-third (3.709) of its value for the corresponding Penman plot using zero root constant, while optimisation yields a root constant close to that recommended (-3 mm). Much of the deficit overestimation is reduced for the zero root constant simulation, while peak deficits are more accurately predicted using an optimised value.

Plots for layer deficits show a worsening of fit over total profile simulations, for the recommended root constant, since actual deficits are restricted to those of the top 20 cm (Fig. 4.27). Optimisation is for a root constant of -25 mm, as for the equivalent Penman plot, although model fit is slightly improved over the latter. On applying Penman-Monteith evaporation, magnitude of predicted deficits is reduced for the optimised plot, although this also entails some underestimation during wetting up in July (days 189 to 209) (Fig. 4.28).

4.3.3.5 Woodland

a) Penman Evaporation

A poor model fit using Penman evaporation (RMSE = 31.841) is produced by incorporation of the Grindley recommended root constant of 200 mm for woodland (Fig. 4.29), largely because of the lower measured deficits. Examination of layer deficit plots (Figs. 4.30 and 4.31) reveals that apparent underestimation of total deficits at the beginning and end of the year relates to the inclusion of drainage in actual 'deficits'. The model optimises for a 9 mm root constant, reducing error of fit to 10.523 (Fig. 4.32). Patterns of model fit are similar to those discussed for several previous simulations using Penman evaporation, deficits being overestimated in spring and early summer and underestimated later in the year.

Summer runoff is predicted on only one occasion (day 204, optimised plot) implying enhanced water use under this cover, in terms of interception, root abstraction and transpiration. It is suggested that water yield is therefore lower. A Student's 't' test shows no significant difference between amounts of subsurface flow (throughflow) under woodland and heather moorland at 15 cm depth ($t=0.939$, $n=5$), although significantly larger volumes were found under heather at 30 cm ($t=2.834$, $n=4$). Greater volumes were also collected from the burnt plot, for which interception is much reduced, than from below woodland vegetation, at both 15 cm below the surface (significant at 0.05 level, $n = 24$, Wilcoxon test for paired samples) and 30 cm (significant at 0.05, $n = 23$, Wilcoxon test). During the summer period (days 154 to 273) throughflow was completely absent under woodland while times of subsurface flow under the burnt surface correspond to periods of surface storm runoff. The comparative period of subsurface flow between heather and woodland spans only from

Figures 4.29 to 4.36:
Grindley Model Simulations for Woodland

Figure 4.29 Simulation based on Penman Evaporation, Total Profile Deficits and Recommended Root Constant

Figure 4.30 Simulation based on Penman Evaporation, Layer Deficits and Recommended Root Constant

Figure 4.31 Simulation based on Penman Evaporation, Layer Deficits and Optimised Root Constant

Figure 4.32 Simulation based on Penman Evaporation, Total Profile Deficits and Optimised Root Constant

Figure 4.33 Simulation based on Penman-Monteith Evaporation, Total Profile Deficits and Recommended Root Constant

Figure 4.34 Simulation based on Penman-Monteith Evaporation, Total Profile Deficits and Optimised Root Constant

Figure 4.35 Simulation based on Penman-Monteith Evaporation, Layer Deficits and Recommended Root Constant

Figure 4.36 Simulation based on Penman-Monteith Evaporation, Layer Deficits and Optimised Root Constant

October 1980 to April 1981 and includes only a small number of valid measurements, whilst that between burnt ground and woodland incorporates a summer period and extends from April 1981 to March 1982 and thus facilitates more meaningful interpretation. Implications of vegetation cover changes for subsurface flow are discussed at greater length in a subsequent chapter.

Application of the recommended root constant to the smaller deficits held in the 0 cm to 50 cm layer inevitably promotes a worsening of fit (RMSE = 43.013) (Fig. 4.30). The root constant is optimised to -21 mm for these deficits and pattern of optimisation is comparable to the equivalent simulation for heather moorland. Some overestimation still pertains during the early season, followed by deficit underestimation in late summer, but rendering a smaller error of fit over that for the total deficit plot (RMSE = 7.187) (Fig. 4.31).

b) Penman-Monteith Evaporation

Figure 4.33 shows that despite a more accurate deficit simulation for the recommended root constant following inclusion of Penman-Monteith evaporation estimates, deficits are underestimated by about 20 mm throughout the year. Optimisation generates a slight increase in error of fit over the corresponding plot for Penman evaporation (RMSE = 11.511). The model optimises only to the highest potential deficit, leaving summer values underestimated and indicating that employment of a lower resistance value in evaporation calculations may, again, yield an improved fit (Fig. 4.34). Discrepancies during early and late 1981 again relate to included drainage in actual deficit estimates for these times.

Reduced errors of fit result from incorporation of Penman-Monteith data for layer deficits, in comparison with the two previous plots, yielding the best overall simulations for this land-use. Much of the improved simulation shown in Figure 4.35, however, accompanies modification of actual deficits against the same, predicted plot as that for the recommended root constant. Root constant is again optimised to a much lower figure (8 mm) than the recommended value might suggest and, because predicted layer deficits are not restricted to potential values, an improved model fit is observed in Figure 4.36 (RMSE = 4.68). The main deficiency involves enhanced deficits during early summer.

4.3.4 CONCLUSIONS

4.3.4.1 Observed Soil Moisture Deficits

Actual layer deficits for the three land-use plots are illustrated in Figure 4.37. Comparison with Figure 4.8 shows that total profile deficits display similar patterns throughout 1981, although in absolute terms, are less representative of actual moisture abstracted because of their included drainage. The extent of the main deficit period varies between land-uses since, although onset of spring drying occurs on similar dates, return to zero deficit in autumn varies from about day 260 (mid-September) for burnt moorland to day 280 (early October) for woodland. Both Figures 4.8 and 4.37 show deficits under burnt moorland to be reduced in comparison to other land-uses, transpiration and interception losses being greater from heather and conifers.

Greater wetting up is observed for the woodland profile in July (days 189 to 209) than for either moorland plot. This is explained in terms of runoff from the moorland, excess water from reduced

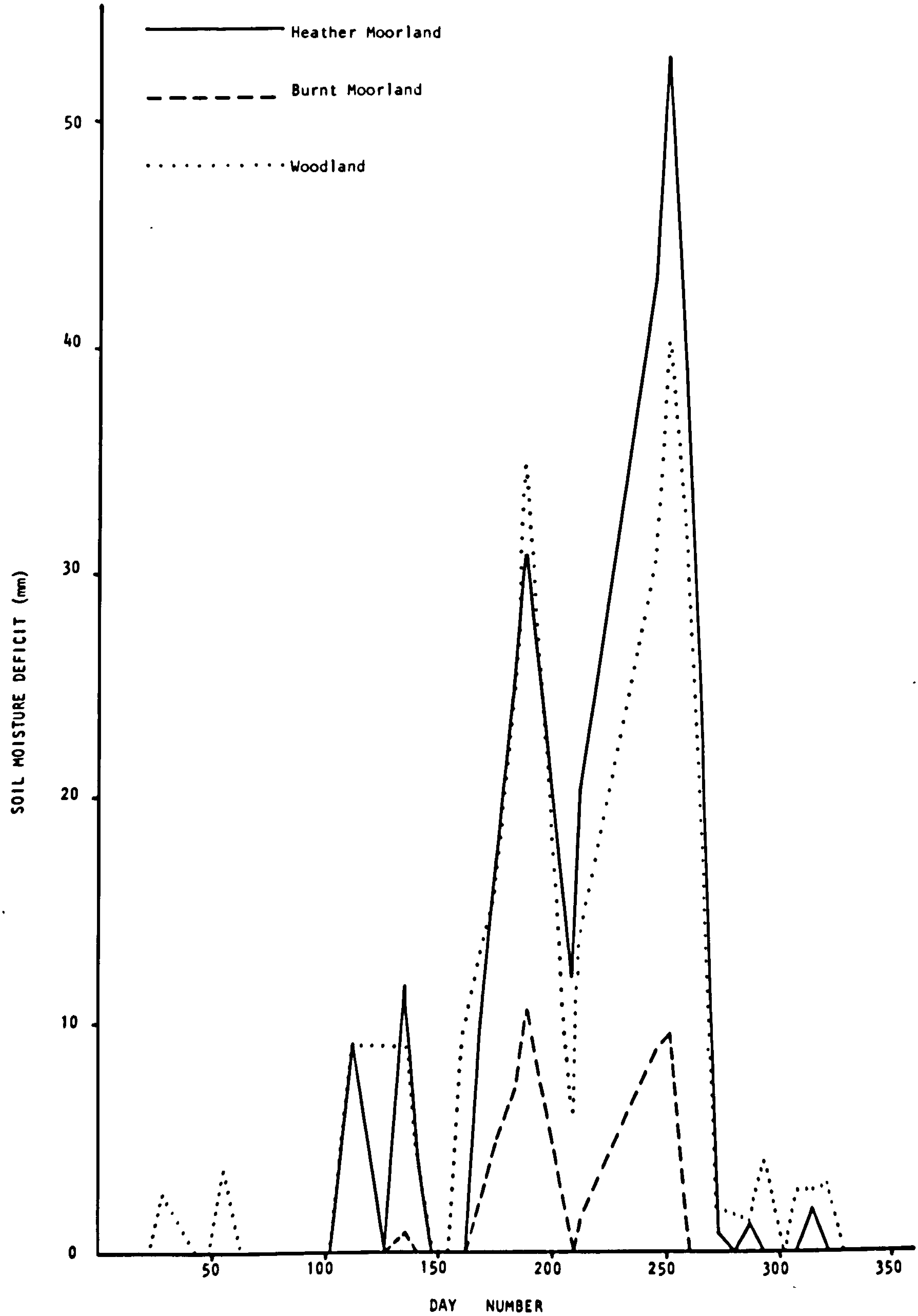


Figure 4.37 Actual 'Layer' Soil Moisture Deficits (1981)

interception and, for the burnt plot, reduced transpiration largely constituting surface and subsurface runoff, rather than soil moisture deficit reduction. The burnt moorland profile, characterised by smaller deficits, wets up completely to zero deficit during this period (Fig. 4.37).

4.3.4.2 Predicted Deficits

Soil moisture deficits are characteristically low in the type of environment under consideration here. Model simulations using the recommended root constants therefore generally produce poor fits, overestimating 'observed' deficits. Bearing in mind the influence of field capacity estimation on final values, a maximum observed deficit of 53 mm for heather (layer deficits) compares with 45 mm quoted for Sneaton High Moor for June 1980 by Wallace et al. (1982) who separated drainage and evaporation by identification of the zero flux plane. Calder et al. (1983) quoted a maximum of 100 mm deficit for peaty soils on Plynlimon, mid Wales during the 1976 drought.

Root constant optimisation subsequently generates more accurate predictions; root constants become reduced and, in certain cases, negative values produce the best model fits. Amounts of evaporation at the potential rate are therefore limited. Increased specification of species' characteristics by means of model runs with Penman-Monteith evaporation further reduces errors especially when combined with layer moisture deficits, which differentiate evaporation from drainage. Root constants and errors of fit are summarised in Table 4.3.

4.3.4.3 Deficiencies of Model Prediction

Some of the model errors reflect measurement or calculation problems while others relate to inherent model defects or

	PENMAN EVPORATION		PENMAN-MONTEITH EVAPORATION	
	TOTAL PROFILE DEFICITS	LAYER DEFICITS	TOTAL PROFILE DEFICITS	LAYER DEFICITS
HEATHER MOORLAND:				
Recommended	13.69	15.769	11.105	6.404
Root Constant	(12 mm)	(12 mm)	(12 mm)	(12 mm)
Optimised	13.171	10.182	10.575	6.086
Root Constant	(0 mm)	(-21 mm)	(mpd)	(mpd)
BURNT MOORLAND:				
Recommended	10.07	17.508	3.709	11.251
Root Constant	(0 mm)	(0 mm)	(0 mm)	(0 mm)
Optimised	4.671	5.301	3.405	3.283
Root Constant	(-21 mm)	(-25 mm)	(-3 mm)	(-25 mm)
WOODLAND:				
Recommended	31.841	43.013	11.511	4.723
Root Constant	(200 mm)	(200 mm)	(200 mm)	(200 mm)
Optimised	10.523	7.187	11.511	4.68
Root Constant	(9 mm)	(-21 mm)	(mpd)	(8 mm)

mpd = maximum potential deficit
(Figures in brackets refer to assigned root constants)

Table 4.3 Root Mean Square Errors of Grindley Model Predictions assumptions. Discrepancies may be introduced at an early stage if moisture volume fraction values are inaccurate as a result of errors incurred during neutron probe calibration. The incorporation of an established calibration curve and a correction factor for surface readings should help to eliminate such errors. Secondly, since the physical reality of the field capacity concept is questionable, 'observed' soil moisture deficits may be in error. Thirdly, divergence between predicted and observed profiles may result from Sneaton High Moor being climatically unrepresentative of the Egton site. This is unlikely, however, since both sites hold similar crops, lie on approximately the same line of latitude, are of similar heights and are separated by a distance of only 13 km. Any errors from this source are therefore expected to be insignificant; values of evapotranspiration in particular, remain fairly uniform over a wide area (Hall and Heaven, 1979). Disagreements between rainfall amounts recorded by the automatic weather station sited on open moorland and those actually received under woodland, however, may arise from eddy currents at the plantation's boundary.

Model inefficiencies may result from the fact that no direct account is taken of soil type except when using root constant optimisation, and that the Penman drying curve may be in doubt. Alteration of the drying curve shape may improve predictions in that an earlier reduction in the actual:potential evaporation ratio would reduce the overestimation of early season deficits. In addition, criticism may be aimed particularly at the root constants suggested by Grindley, since for a given crop the value may vary with, for example, geology or soil type and the findings from the present study lend support to the idea suggested by a number of other investigations, albeit in a different context with chosen field capacity promoting drainage underestimation, that the model overestimates actual evapotranspiration and that root constants should accordingly be reduced. Thus, Headworth (1970), in a study of root constants in chalk areas, related rainfall to both infiltration and actual evapotranspiration using different root constants and, from a correlation with river flow data, concluded that 25 mm was the most suitable root constant for the short-rooted vegetation of the area, as opposed to Grindley's value of 75 mm. Similarly, from discrepancies between measured groundwater recharge[1] and that calculated using Meteorological Office estimates, Kitching et al. (1977) concluded that differences could be due to overestimation of actual evapotranspiration by the model, and of the concomitant root constant. Relating generated root constants with calculated recharge led them to suggest a value of 35 mm as the most suitable for short-rooted vegetation, although comparisons of actual and calculated soil moisture deficits yielded an optimum value of 50 mm. A tendancy for the Grindley model to underestimate recharge in chalk as noted by Rushton and Ward (1979) led them to postulate an allowance for direct recharge, equating it to 15% of actual precipitation greater than 5 mm, plus 15% of effective precipitation.

Penman (1949) in his formula for root constant evaluation discussed earlier (p. 95) implied that different values be implemented each year, maintaining that, for example, a dry growing season would lead to deeper rooting. An annually varying value may be justified if the crop's ability to maintain stomatal opening varies from year to

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1 'That amount of surface water which reaches the permanent water table either by direct contact in the riparian zone or by downward percolation through the overlying zone of aeration' (Rushton and Ward, 1979, p 345).

year. Also, different aspects of plant physiology have varying sensitivities to plant water deficits, some being significant only during certain phases of development (Ritchie, 1981). In conclusion, therefore, definition of a yearly root constant should improve model predictions: this is, in effect, an optimised root constant.

Runoff, by model definition, is not allowed unless field capacity has been attained. Where direct runoff occurs when a deficit exists, moisture deficits are underestimated since the model assumes precipitation is evaporated or used to reduce the deficit. Finally, the model is limited by its inherent representation of the soil as a single layer. If greater sophistication is desired, a more realistic, multi-layer model should be implemented, perhaps allowing enhanced evaporation of incoming rainfall. To test the reliability of such a specification, the more recently developed two-layered MORECS model was run against the study data. In its more widespread use throughout Great Britain, the model has begun to produce improved results despite some preliminary problems.

4.4 THE MORECS MODEL

The Meteorological Office Rainfall and Evaporation Calculation Scheme (MORECS) was introduced for practical use at the Meteorological Office in April 1978 and has now largely replaced the Grindley model for purposes of nationwide soil moisture deficit prediction. A series of modifications was instituted during the three years following 1978 and a revised model version was completed in 1981. The original model (1978), represented diagrammatically by Figure 4.38, is implemented in the present study since this has received the most rigorous testing. In addition, the revised model retains some of the inadequacies of the earlier versions (Gardner and Field, 1983) and relies not on a drying

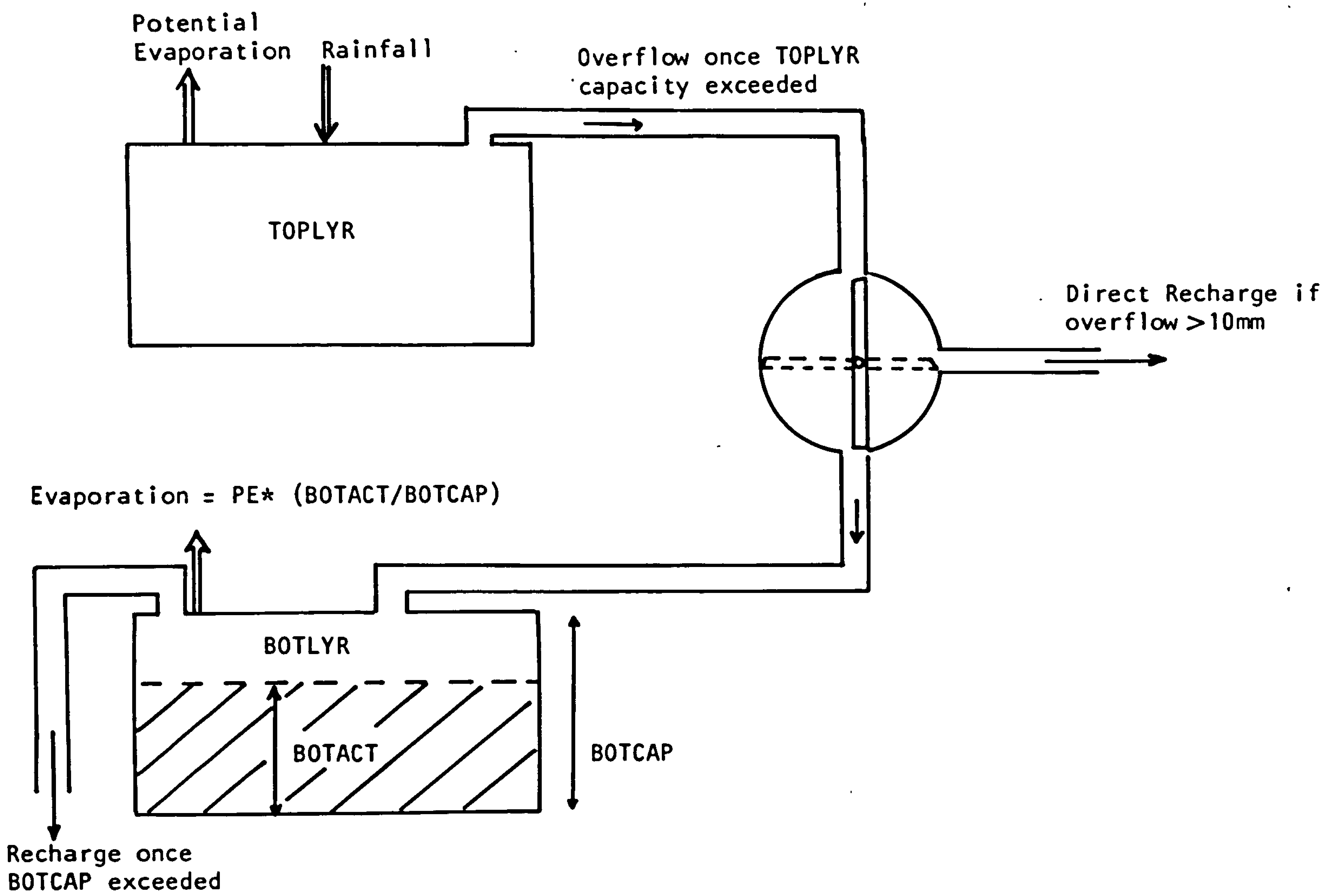


Figure 4.38 Conceptual Structure of the MORECS Model

curve but on the relationship between surface resistance and soil moisture deficit. The more recent model also makes no allowance for drainage or runoff, discussed below. In the model considered here, daily rainfall is added initially to the top layer (TOPLYR). When TOPLYR is full and rainfall in excess of 10 mm is added, excess moisture is sent to a general drainage/runoff component of the model. This may include direct recharge, that is recharge to groundwater in the presence of a soil moisture deficit, since the state of the bottom layer (BOTLYR) is unimportant. 'Drainage' may therefore occur when a deficit exists. In the more recent version of the model, no such allowance is made. Evapotranspiration, at potential demand, creates water loss initially from TOPLYR until this layer is empty, whence loss occurs from BOTLYR. This loss is calculated as a function of potential evaporation, and actual moisture content (BOTACT) in relation to the maximum moisture capacity of the layer (BOTCAP). Continued evapotranspiration results in the maximum deficit being obtained, daily soil moisture deficits being calculated as the sum of deficits in the two layers. The revised version of MORECS accepts low, medium and high available water capacity soils, with 'medium' as standard issue. The standard issue for the original model is for soils with high available water capacity (Table 4.4).

The basic Penman-Monteith equation is used as part of model computations to calculate potential evaporation from each layer. Evaporation calculations for the present study are similar to those implemented by the Meteorological Office for MORECS and as described by Thompson <u>et al</u>. (1981) but do not correspond in detail. Aerodynamic and surface resistances all remain as described earlier (Section 4.3.3), although a value of 70 sm^{-1} was adopted as r_s for woodland (Thompson <u>et al</u>., 1981). The model takes account of

	Original Model ((High AWC (>180 mm m^{-1})) (mm)	Current Model (Medium* AWC (100-180 mm m^{-1}, * high=+25%, low=-25%)) (mm)
Bare Soil	15	20
Conifers	500	175
Uplands	50	50

AWC = available water capacity

(based on Thompson, 1981)

Table 4.4 Defined Maximum Soil Moisture Deficits for Two Versions of the MORECS Model increasing leaf areas where relevant, and a crop resistance response to changing temperature and vapour pressure deficit is included for conifers.

The soil moisture extraction function of the model again relies on the drying curve (Fig. 4.39) and field capacity concepts. Field capacity exists when both layers of the model are full. In the original model 40% of available water is held in TOPLYR and is denoted MAX. This is assumed to be extracted at the potential rate, with surface resistance constant and while incident radiation is fully intercepted by the (assumed) dense crop (or bare soil). MAX is simliar in concept to the root constant and 2.5MAX defines maximum soil moisture deficit, as shown in Table 4.4. Once the top layer is exhausted of moisture, the remaining 60% in BOTLYR is subsequently extracted at a linearly decreasing rate as the soil dries (Wales-Smith and Arnott, 1980).

4.4.1 GENERAL APPLICATION

The Meteorological Office produces a weekly series of nationwide maps giving meteorological conditions, soil moisture deficit estimates and water balance calculations. Daily meteorological data (sunshine, temperature, vapour pressure, wind speed and rainfall) collected from synoptic stations throughout Great Britain, are interpolated objectively to yield 40 km x 40 km grid square averages of these five variables. Using the Penman-Monteith equation, daily estimates of potential evaporation are subsequently calculated on the same grid square basis. Daily water balances are calculated both for individual land-use categories in each grid square and for a weighted average land-use. The Meteorological Office's version of the model accounts for condensation (negative night-time

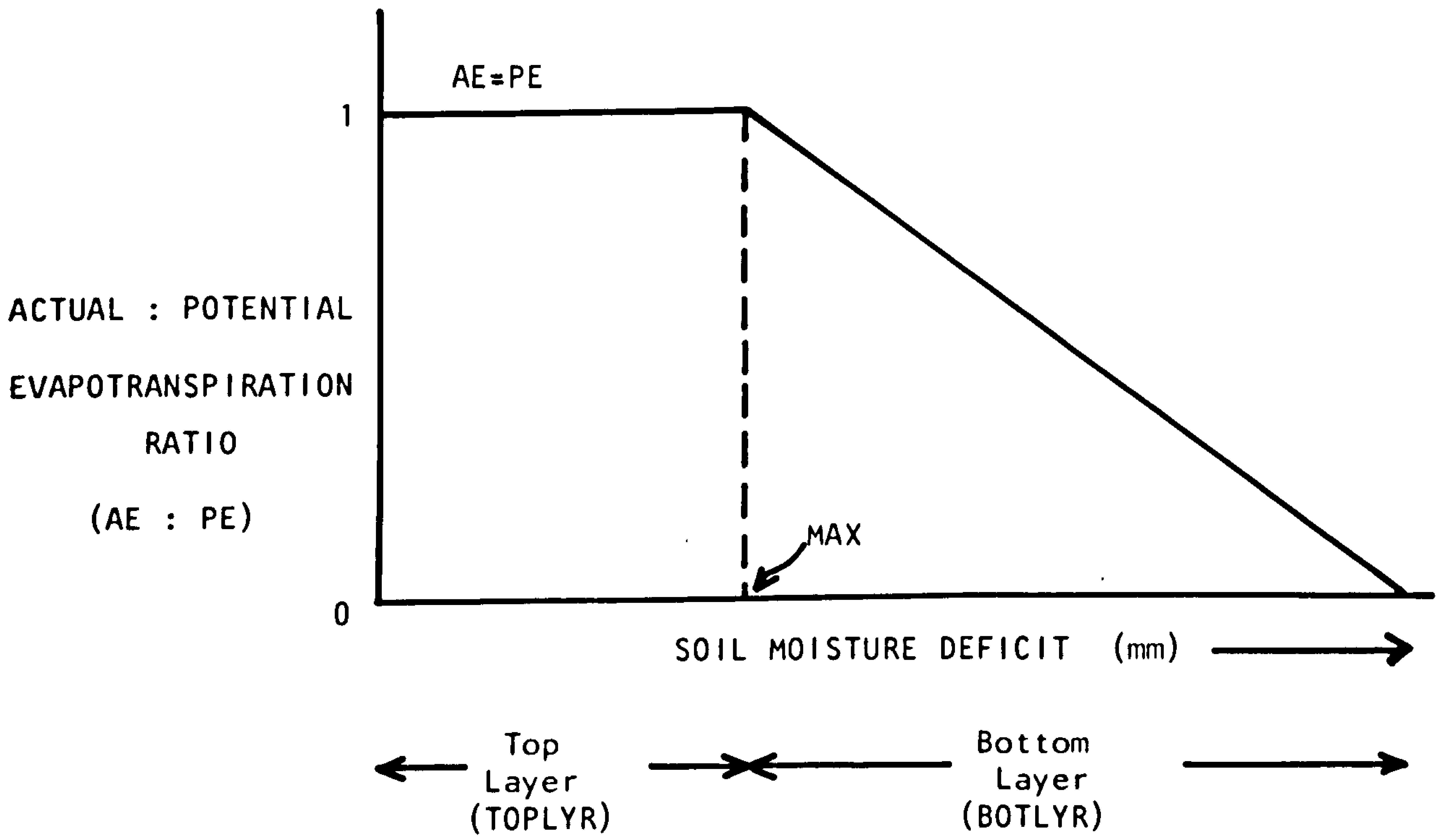


Figure 4.39 The MORECS Model Drying Curve

evaporation) which, along with daily rainfall input, constitutes effective daily rainfall. A proportion of this is 'lost' to interception, the latter varying on a seasonal basis, being dependent on leaf area index. Allowance is made for multiple daily rainfall events and the possibility of evaporation of intercepted rainfall during a storm. After definition of interception, any remaining rainfall is assumed to be diverted to soil moisture, evapotranspiration, percolation or runoff components.

Weekly soil moisture deficit estimates are produced for observed land-use types in each grid and as a weighted grid square average, produced by weighting individual land-use deficits according to land-use distribution. Although eventual output format depends on the user's requirements and a certain number of 'output options' are available, final maps generally include representations of meteorological variables, potential evapotranspiration, actual evapotranspiration, soil moisture deficit and excess rainfall (runoff and groundwater recharge) for both grass and actual land-use (Thompson et al., 1981).

4.4.2 APPLICATION TO STUDY AREA

Model runs were performed for the three land-use types at Egton using both Penman and Penman-Monteith daily evaporation estimates, and optimising the ratio of water held between TOPLYR and BOTLYR (Fig. 4.39) the initial drying curve shape being maintained. The ratio of the moisture capacities of the two layers is defined by the expression:

$$\frac{\text{Moisture capacity TOPLYR}}{\text{Total moisture capacity (TOPLYR+BOTLYR)}} \times 100\%$$

Calibration of the model was again confined to data collected during 1981, data input format resembling that of the Grindley model, and comprising soil moisture deficits for both total and layered profiles and averaged for each land-use plot, daily estimates of potential evaporation and daily values of rainfall. The model returns daily actual evapotranspiration and predicted soil moisture deficits, along with magnitudes of model parameters and error terms. Sizes of TOPLYR and TOPLYR+BOTLYR are given in millimetres as well as by the ratio TOPLYR:TOPLYR+BOTLYR. As in the previous model, the root mean square error of fit is calculated.

4.4.3 MODEL RESULTS

Table 4.5 summarises the complete set of results. Overall, goodness of model fit shows only marginal increases in accuracy over corresponding results from the Grindley model using optimised root constants. Further, this improvement is restricted to predictions based on Penman evaporation. All simulations optimise towards a lower TOPLYR:TOPLYR+BOTLYR ratio than that recommended for the original model (40%) (Wales-Smith and Arnott, 1980), indicating that evaporation at the potential rate is restricted. Model predictions are discussed in the following sections for each land-use plot.

4.4.3.1 Heather Moorland

For simulations based on Penman evaporation, the optimum ratio of moisture contained in TOPLYR:TOPLYR+BOTLYR is low (5%) both when using total and layer moisture deficits. Use of Penman-Monteith data suggests that model parameters could not be determined as the model optimises to potential deficits. Figure 4.40 illustrates that in mid to late summer observed exceed predicted deficits, again suggesting that resistance values in the Penman-Monteith calculation are too large. Overestimation of early season deficits is a characteristic feature of the remaining plots (Figs. 4.41 to 4.43) succeeded by, in

	PENMAN EVAPORATION				PENMAN-MONTEITH EVAPORATION			
	TOPLYR (mm)	TOPLYR+ BOTLYR (mm)	TOPLYR: TOPLYR+ BOTLYR(%)	RMSE	TOPLYR (mm)	TOPLYR+ BOTLYR (mm)	TOPLYR: TOPLYR+ BOTLYR(%)	RMSE
HEATHER MOORLAND:								
Total profile	6	120	5	11.332	48 (mpd)	320	15	10.575
Layered profile	2	40	5	9.421	47 (mpd)	188	25	6.086
BURNT MOORLAND:								
Total profile	2	40	5	3.979	10	40	25	3.867
Layered profile	1	4	25	2.332	1	4	25	2.206
WOODLAND:								
Total profile	7	140	5	8.25	51 (mpd)	340	15	11.511
Layered profile	2	40	5	7.366	27	77.143	35	4.635

mpd = maximum potential deficit

Table 4.5 MORECS Parameters and Errors for Optimised TOPLYR:TOPLYR+BOTLYR Ratios

Figures 4.40 to 4.43:
MORECS Model Simulations for Heather Moorland using Optimised Ratios between TOPLYR and TOPLYR + BOTLYR

Figure 4.40 Simulation based on Penman-Monteith Evaporation and Total Profile Deficits

Figure 4.41 Simulation based on Penman Evaporation and Total Profile Deficits

Figure 4.42 Simulation based on Penman Evaporation and Layer Deficits

Figure 4.43 Simulation based on Penman-Monteith Evaporation and Layer Deficits

the case of Penman-based predictions, compensatory underestimation of late season values. Model fit may be improved in this case by allowing TOPLYR:TOPLYR+BOTLYR ratios to vary throughout the year. Peak deficits are timed with reasonable accuracy, especially when using Penman-Monteith evaporation, while estimated extent of the main deficit period shows most improvement over that obtained by Grindley's model for Penman-based predictions. Occurrences of zero deficit in summer are similarly comparable to those of the previous model and each is corroborated by evidence from stream-level records.

4.4.3.2 Burnt Moorland

Lowest overall root mean square errors apply to this land-use, substitution of Penman with Penman-Monteith evaporation data resulting in only negligible improvement in fit (Table 4.5). Each model run generates an optimised ratio below the 40% recommended constant. In general, evaporation is checked at a smaller absolute deficit than under heather, indicating a smaller available water capacity for the burnt plot. In comparison to Grindley model predictions, better reproduction of peak total profile deficits is shown in terms of both timing and magnitude (Figs. 4.44 and 4.45) particularly when using Penman evaporation. Simulation of layer deficits again proves superior to total profile equivalents, however, with root mean square errors of 2.332 (Penman) and 2.206 (Penman-Monteith) (Figs. 4.46 and 4.47). Nevertheless, because layer deficits lie at or close to zero for most of the year, MORECS weights predicted values towards the zero axis and thus underrates observed peak summer deficits. Small magnitude deficits enable frequent predictions of summer runoff, only about half of which, the larger storms, are validated by hydrograph records.

Figures 4.44 to 4.47:
MORECS Model Simulations for Burnt Moorland using Optimised Ratios between TOPLYR and TOPLYR + BOTLYR

Figure 4.44 Simulation based on Penman Evaporation and Total Profile Deficits

Figure 4.45 Simulation based on Penman-Monteith Evaporation and Total Profile Deficits

Figure 4.46 Simulation based on Penman Evaporation and Layer Deficits

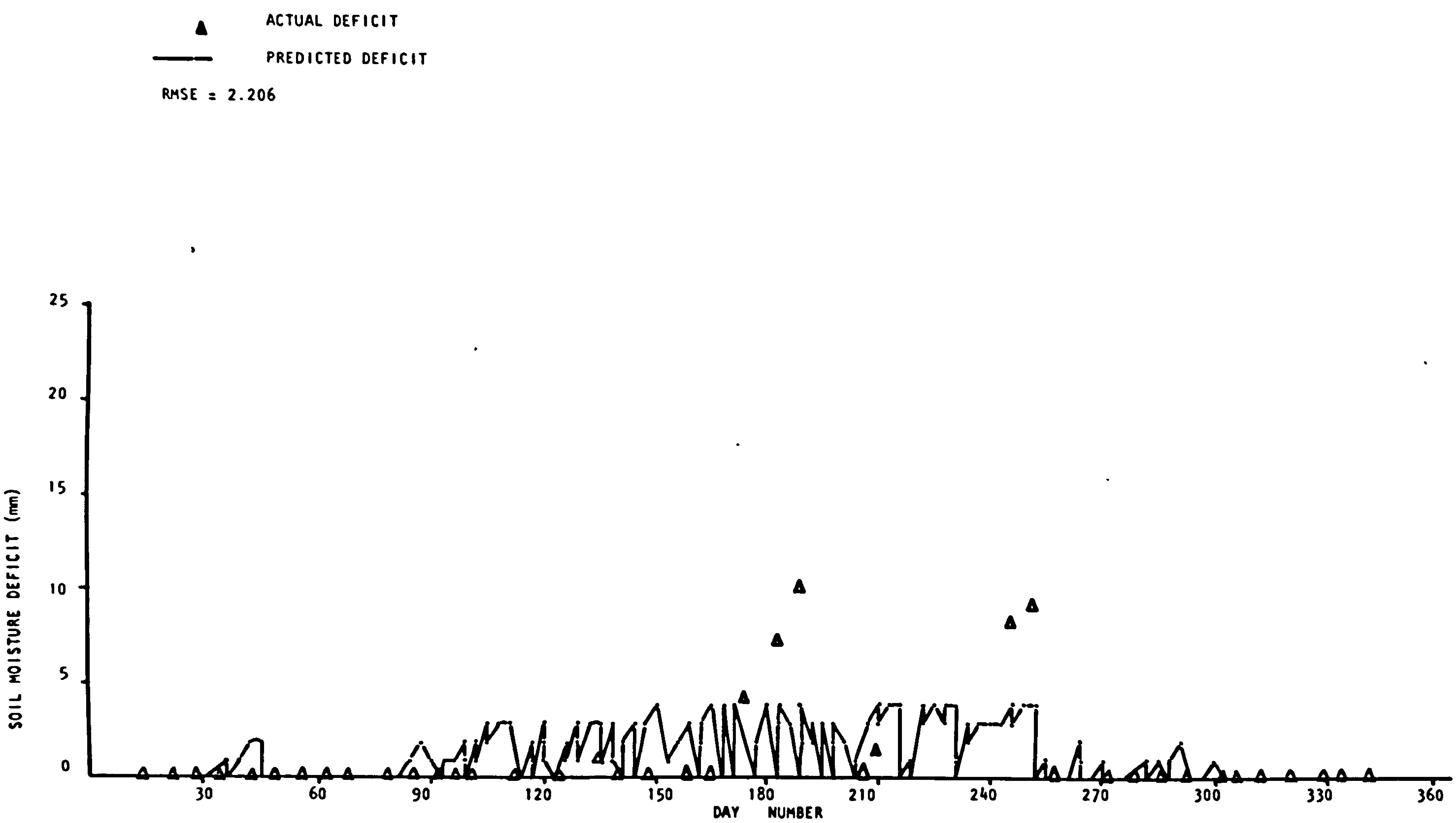


Figure 4.47 Simulation based on Penman-Monteith Evaporation and Layer Deficits

4.4.3.3 Woodland

MORECS simulates woodland data with an accuracy similar to that for the vegetated moorland. The Penman-Monteith evaporation/layer deficits combination provides the best overall fit, with 35% of total evapotranspiration occurring at potential demand. Some improvement in prediction over Grindley results is shown for total profile deficits (Penman evaporation); early season overestimation in particular, is largely eliminated (Fig 4.48). Using Penman-Monteith evaporation, however, MORECS fails to rectify the consistent underestimation of total deficits observed for the first model (Fig. 4.49). TOPLYR moisture is optimised close to maximum potential deficit, again implying that the allocated resistance values may be too high.

Plots of layer deficits return similar root mean square errors to analogous Grindley predictions while visually, some improvement is observed in MORECS' simulation of late summer values, using Penman evaporation (Fig. 4.50). Both this plot and its Penman-Monteith equivalent (Fig. 4.51) demonstrate some deficit overestimation earlier in the season, however, although alteration of the drying curve for this part of the year, by invoking an earlier reduction in actual:potential evapotranspiration ratio, would partially correct this deficiency. Reducing the value of MAX, the amount of water held in TOPLYR of the model, could, theoretically, also diminish the degree of overestimation. This would lower the maximum allowable deficit in the soil, but would also, however, involve its attainment arising earlier in the year (Gardner, 1981b) thus, in the present case, worsening any representation of peak deficits. In general, no improvement is observed for timing of initial spring drying and autumn re-wetting, although Figure 4.48 displays a more accurate assessment of the onset of the main deficit period, its Grindley counterpart (Fig 4.32) being affected by included drainage in early 'deficits'.

Figures 4.48 to 4.51:
MORECS Model Simulations for Woodland using Optimised Ratios between TOPLYR and TOPLYR + BOTLYR

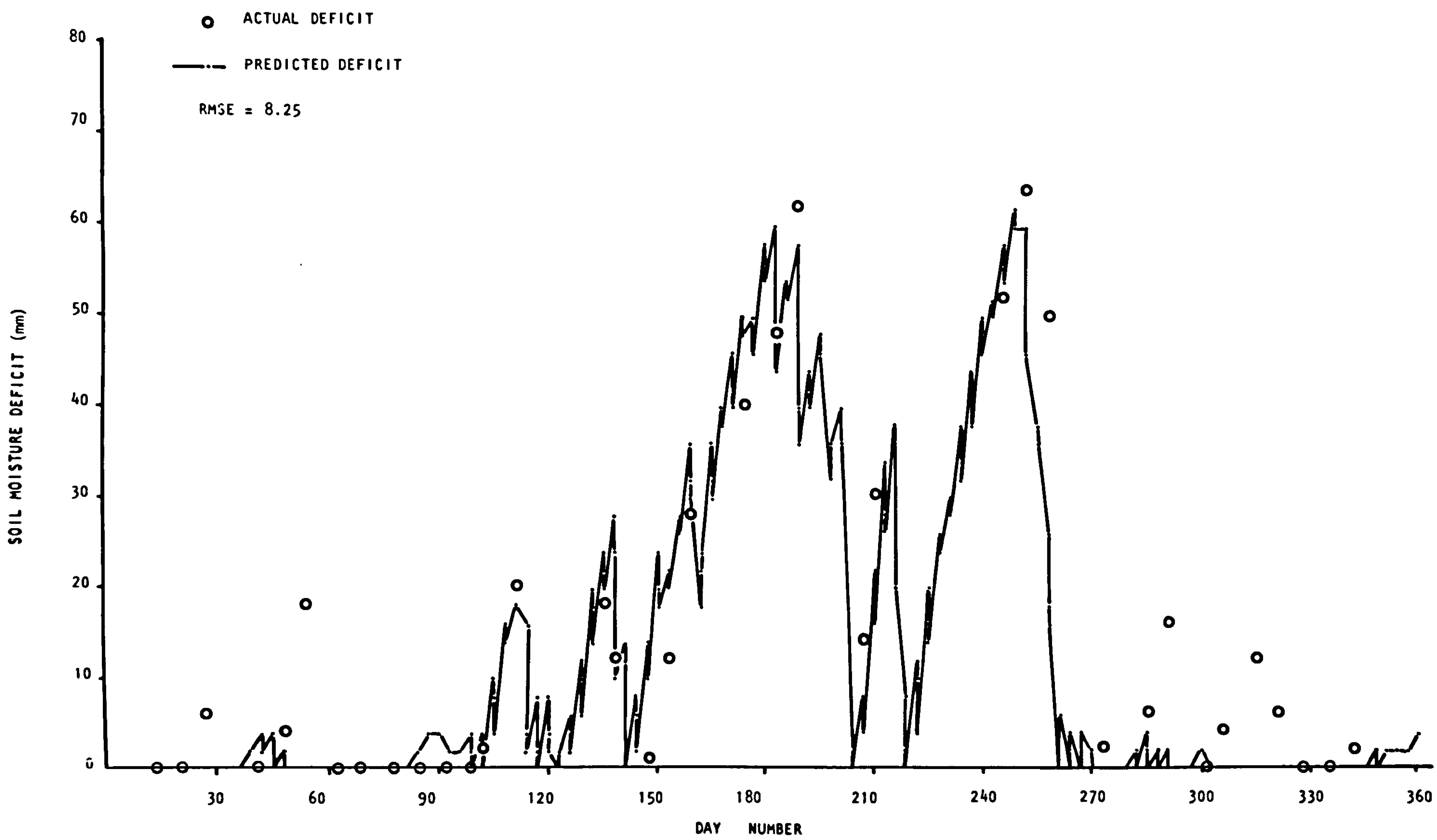


Figure 4.48 Simulation based on Penman Evaporation and Total Profile Deficits

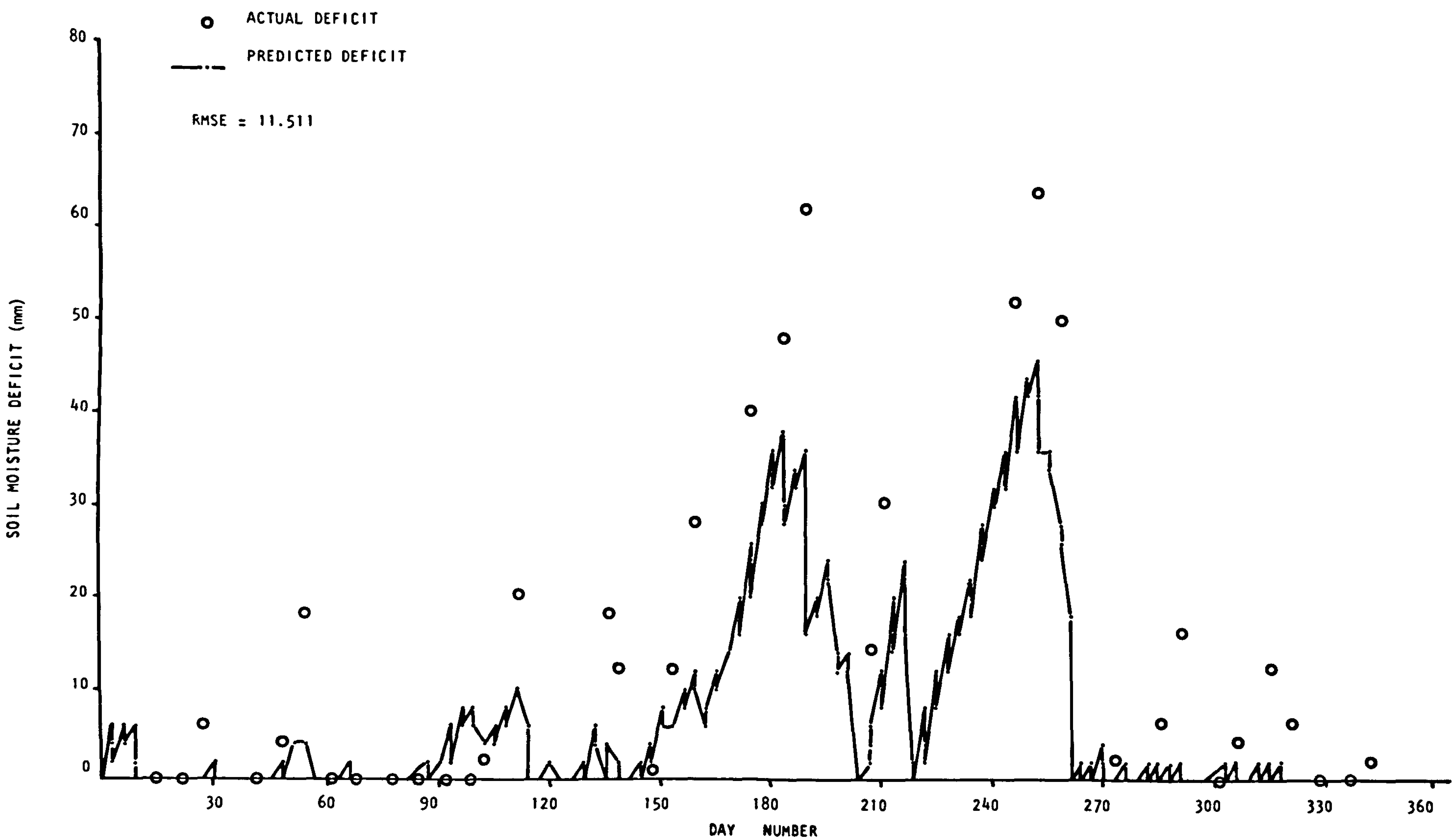


Figure 4.49 Simulation based on Penman-Monteith Evaporation and Total Profile Deficits

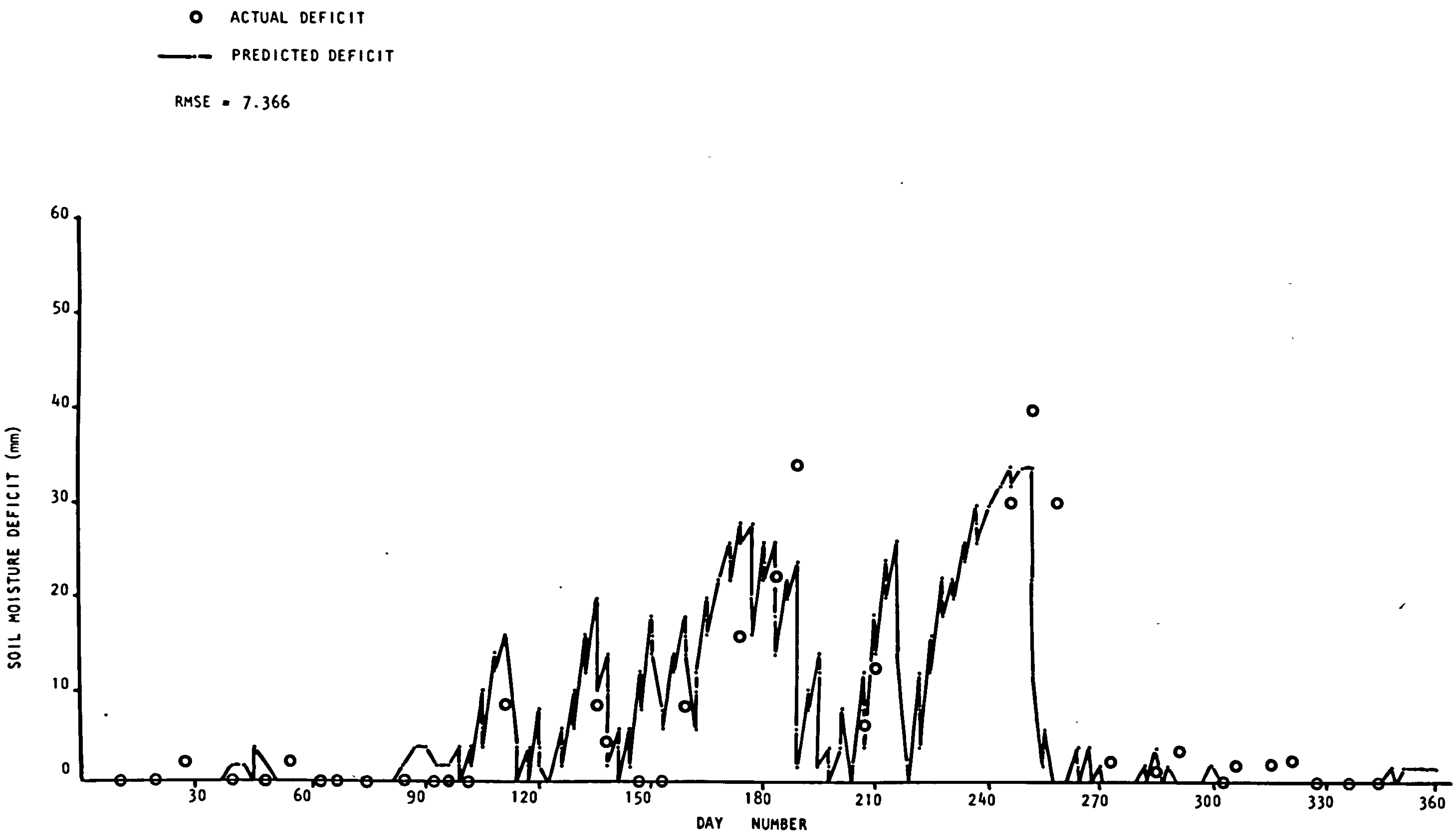


Figure 4.50 Simulation based on Penman Evaporation and Layer Deficits

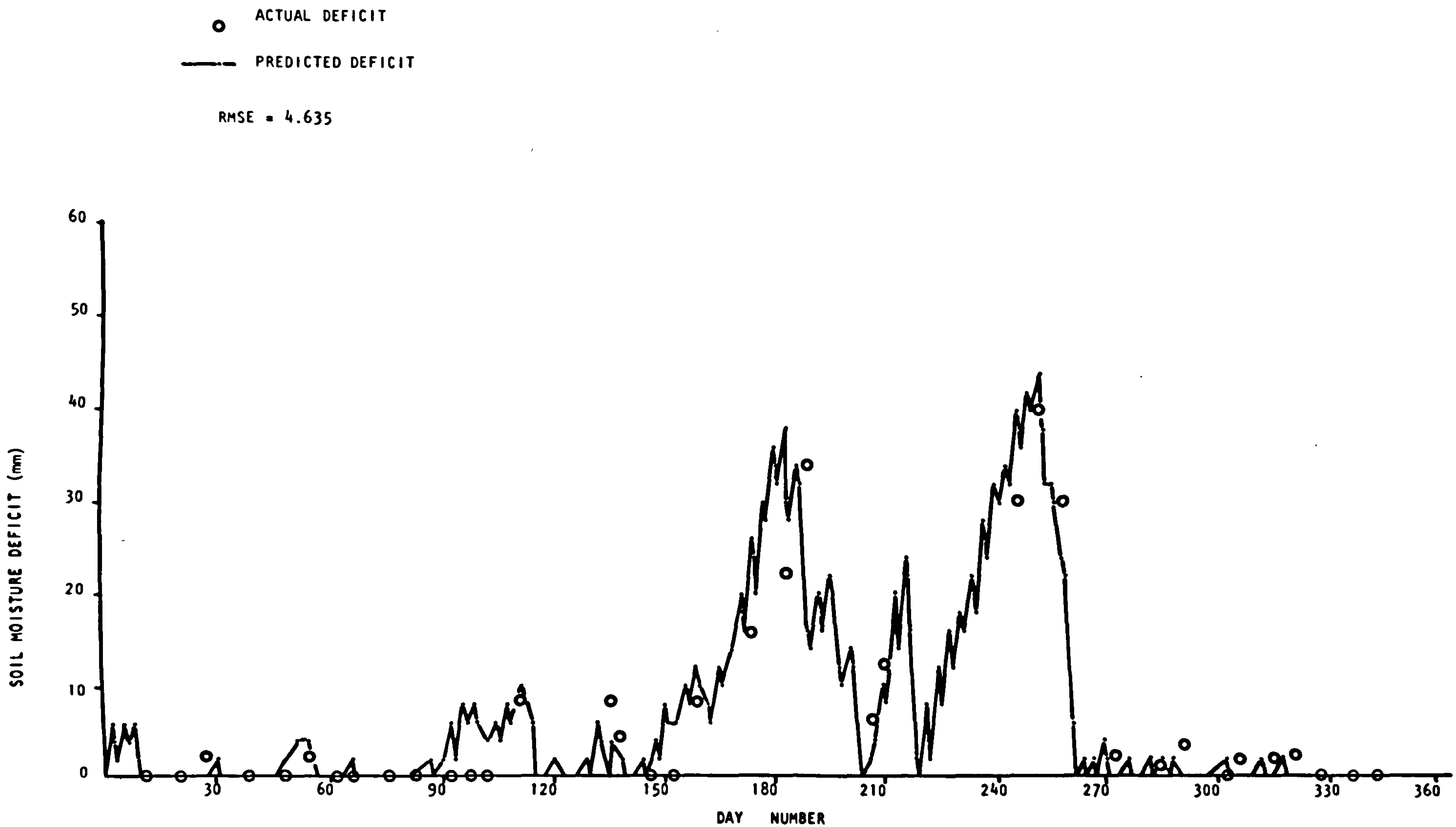


Figure 4.51 Simulation based on Penman-Monteith Evaporation and Layer Deficits

4.5 A COMPARISON OF THE TWO MODELS

Temporal changes in soil moisture deficit under the three land-use types at Egton are simulated with reasonable accuracy by the Grindley and MORECS models when parameter optimisation is introduced. This modification generally indicates that less evapotranspiration takes place from each plot at the potential rate than that 'recommended'. In common with a number of other investigations, therefore, it is suggested that actual evapotranspiration is overestimated by the Grindley and MORECS models and that drying specification parameters should consequently be reduced in magnitude. In some cases, however, recommended root constants have been used successfully (for example, Wheater and Weaver, 1980).

Despite virtual replacement of the Grindley model with MORECS for nationwide deficit prediction by the Meteorological Office, the single-layer model with optimisation proved adequate for deficit simulation in the present analysis. Improved fits are observed following substitution of the recommended Penman evaporation by Penman-Monteith estimates, especially for layer deficit predictions. In contrast, evaporation formula proves less important for the two-layered model. Error of fit for both models may be further reduced by introduction of seasonally or annually varying root constants or TOPLYR:TOPLYR+BOTLYR ratios.

Surprisingly few other published comparisons exist between measured deficits and those predicted by the Grindley model. Hall and Heaven (1979) applied the model to data collected from the fen and Chalk Wolds areas of Lincolnshire and found model deficits to be fair representations of reality. Deviations were related, for example, to crop senescence and variations in crop cover. Working in the Gloucester area, on agricultural land of poor natural drainage, Wheater and Weaver (1980) discovered variations in Grindley model performance, notably in the representation of peak deficits, between years and under different crops.

Evaluation of MORECS' performance is limited still further. Gardner (1981b) found changes in moisture deficits at grassland sites on a variety of soil types in southern England to be represented reasonably well by the original version of the model although in general absolute deficits were overestimated. Apart from alteration of inherent model characteristics, other means of improving model fit suggested by Gardner included increasing the soil profile depth under consideration and retrospective adjustment of field capacity. Thompson et al. (1981) concluded that the newly developed version of MORECS gives acceptable results for a range of soil and crop types.

For the present study site, results from both models support the idea of diminished amounts of potential evapotranspiration for 1981 following heather burning. The Grindley model shows that, on average, a total of 25 mm moisture evaporated at Penman potential demand from heather moorland (total profile deficits) compared to only 4 mm from the burnt plot. A total of 34 mm applies under woodland. Comparative figures for layer deficits are 4mm (heather moorland), 0 mm (burnt moorland) and 4 mm (woodland). Similarly, Table 4.5 illustrates that as the soil dries evapotranspiration is checked at the lowest deficits under burnt ground. The evaluation of changing soil moisture status continues in the next chapter, through its significance for storm runoff generation, not only in terms of its modification by land-use, but also in relation to spatial concentrations.

CHAPTER 5

IMPLICATIONS FOR SURFACE AND SUBSURFACE RUNOFF

The analysis of runoff responses to rainfall inputs represents one of the most important continuing problems in hydrology. Factors determining rainfall-runoff relationships, such as physical characteristics of the drainage basin, climatic influences and interference by man, are integrated in the flood hydrograph, a 'single empirical curve' defining the complexities of basin characteristics (Chow, 1964). Ward (1975) noted the difficulty with which catchment effects on runoff, especially those of vegetation, are assessed, while the variable source area concept implies that the effects of particular catchment characteristics on runoff will vary spatially as well as temporally.

Quantitative rainfall-runoff relations have been developed through hydrological systems investigations, and, ideally, a 'full synthesis' approach is needed in order to understand these relationships (Amorocho and Hart, 1964). A range of approaches is discussed in the present chapter, the main form of analysis being a type of partial-system synthesis. Although the development of rainfall-runoff models has been the subject of much research, rather less has been accomplished in terms of application and objective testing of models on catchments other than those for which the models are developed (Weeks and Hebbert, 1980).

The aims of the present chapter are the quantification and explanation of differences in runoff characteristics between heather moorland and burnt sites. Surface runoff responses are examined for pre- and post-burn moorland cases, while the importance of subsurface flow is considered for the woodland in addition. Conclusions relating to the implications of vegetation change for the stream hydrograph are drawn against a background of hydrological processes operative in the catchment. Emphasis is placed upon the implications for hydrograph form as opposed to changes in storm runoff volumes, since the latter are already well documented, although still generally in the context of forest land-use change.

5.1 DEVELOPMENT OF IDEAS

5.1.1 RUNOFF GENERATION

5.1.1.1 Variable Source Areas

Following strong criticism of the applicability of the Horton (1933) runoff model, alternative ideas were proposed to explain runoff response, the most important investigations culminating in the 1960s with the work of Hewlett (Hewlett, 1961a, b; Hewlett and Hibbert, 1967) and his advocation of partial or variable source area runoff generation. This concept, which gained support throughout the 1960s and early 1970s with both British and American workers introducing evidence to corroborate the theory (Tennessee Valley Authority, 1965; Weyman, 1973; Kirkby and Weyman, 1973), maintains that the area contributing to streamflow varies over time. Kirkby and Chorley (1967) designated the concept of a varied response to rainfall, and that of Horton's overland flow, as two opposing end-members of a series of possible infiltration processes, the importance of any one of which may vary spatially over a hillslope. Contrary to Horton's idea of runoff control by infiltration capacity, the variable source area concept specifies subsurface flow as important, both in sustaining baseflow and as a major component of stormflow. Bernier (1985) recently extended the application of the concept to arid and semi-arid areas, for which Hortonian runoff generation is normally thought appropriate.

Inputs to stream channels from throughflow and infiltration result in expansion of the channel network, in the variable source hypothesis. The size of this 'contributing area' controls the resulting stream hydrograph shape and runoff volume. The expanded network returns again to a lower density as water inputs cease, although expansion can continue after cessation of rainfall if throughflow continues to feed the system (Hewlett and Nutter, 1970). Source areas also fluctuate seasonally as well as on an individual storm basis. Hewlett and Hibbert (1967) explained channel expansion in terms of the inability of the soil to transmit subsurface flow, which then intersects the ground surface to become 'saturated overland flow'. The capacity for subsurface flow to contribute to storm runoff was additionally justified by these authors through the concept of 'translatory flow', by which rain falling on the upper regions of a slope travels downslope below the surface by gradually displacing moisture in the lower slope regions. The role of unsaturated flow in producing soil moisture gradients with slope had been previously demonstrated by Hewlett (1961a, b) who found increasing moisture contents downslope, with upslope zones draining to lower slope positions. Rain entering the soil profile in locations proximal to a stream therefore travels faster than that entering further upslope. During storms in headwater regions moisture deficits are thus satisfied first in areas at the slope base, these zones subsequently making the largest contribution to the storm hydrograph. As the storm continues, the area contributing to stream channel flow increases, as the saturated 'wedge' expands, satisfying both areas further upslope, and the upper layers of the soil profile (Weyman, 1973). While both saturated and unsaturated subsurface flow may be important especially in headwater areas (Ward, 1984), it is likely that only saturated flow contributes significantly to stormflow (Weyman, 1970; Anderson and Burt, 1977b) although unsaturated flow may supply baseflow or the recession limb of the hydrograph (Hewlett, 1961b; Hewlett and Hibbert, 1963; Weyman, 1973).

Despite the proposal by Amerman (1965) that runoff-contributing areas of saturation are located randomly on ridgetops, valley slopes and valley bottoms, it is more generally thought that contributing areas are limited to certain zones of a catchment. These have been variously defined, either as specific physical locations, such as the slope base, or in relation to particular catchment characteristics which are in turn related to physical or climatic features. In general, topography, antecedent moisture and rainfall characteristics may have a direct or indirect influence on the disposition of subsurface flow and stormflow contributing zones (Ragan, 1968; Betson and Marius, 1969; Whipkey and Kirkby, 1978). Soil characteristics are especially significant, with areas of shallow or compacted soil (Hewlett and Nutter, 1969; Betson and Marius, 1969), fine textured material (Whipkey, 1969) and poorly drained soil (Dunne and Black, 1970b) being potential runoff-generating zones. Heathland vegetation composition, through its relationship with soil moisture, has also been used to identify and map runoff contributing zones (Gurnell et al., 1985). Four, now widely known areas of moisture concentration were identified by Kirkby and Chorley (1967) as follows:

i) Base of slopes, adjacent to stream channels

ii) Hollows

iii) Slope-profile concavities

iv) Areas of thin or less permeable soils

Hollows are associated with slope concavities, both of which are generally found at slope bases, and only zone 'iv' may be initially unconnected to the stream channel. Flow convergence also arises in the soil profile as a result of reduced hydraulic conductivity with profile depth (Ward, 1984).

The significance of concave areas, points of change of slope to a lower angle, was emphasised by Zaslavsky and Sinai (1981). It was postulated that water is concentrated at these junctures since the incoming horizontal flux exceeds that leaving, the horizontal flow component being proportional to slope. This lateral flow is independent of rainfall amount and can occur in unsaturated soil conditions. In contradiction to traditional ideas of partial area contribution, the authors proposed that these waterlogged zones lead to increased groundwater recharge. Later, Abdul and Gillham (1984) highlighted the role of the capillary fringe in rainfall-runoff relations in humid areas, through its enhancement of groundwater discharge to stream runoff.

Variations in absolute size of contributing area between catchments is exemplified in work by Betson (1964) who concluded that the effective runoff-generating area is somewhat less than the whole watershed, but comprises a relatively consistent area. From a number of different North American watersheds, he quoted figures of 4.6%, 40% and 85.8% of the area as contributing to runoff. Definition of contributing area size may be complicated by non-linearities in the rainfall-runoff process due to source area expansion and contraction, and only minimum area can be evaluated, although even this may vary within and between regions. Minimum contributing areas for a number of North American watersheds were shown by Dickinson and Whiteley (1973) to be generally less than 10%, and by Weyman (1974) for the East Twin Brook catchment to be 0.7% to 2% and 0% to 73% of the basin area for surface runoff and throughflow, respectively, expressing flow as a proportion of storm rainfall. Contributing area may be underestimated in the presence of subsurface pipes, however (Jones, 1979). These features generate flow without the prerequisite of surface saturation, and contribute to stream channel flow either directly, or indirectly by feeding contributing areas or connecting isolated source areas to a stream channel, thus extending the source network. Pipeflow as a producer of quick storm responses has been stressed for headwater areas by McCaig (1983) and more generally by Jones (1971, 1981) in agreement with earlier work such as that of Whipkey (1969) who advocated the importance of biological and structural channels in conveying subsurface stormflow in forested catchments; subsurface flow was shown to be a major component of flood flows. As discussed in Chapter 3 (p. 63), the significance of pipeflow for the Egton catchment is uncertain.

5.1.1.2 Subsurface Flow

Although the importance of subsurface flow as a concept has been increasingly appraised over the last twenty years, and numerous empirical and mathematical attempts have been made to characterise its features and origins, some of the earliest work dates from the 1930s. Hursh (1936) for example explained 'subsurface-stormflow' in terms of an impervious soil horizon underlying an 'absorptive' layer. Hydrograph separation techniques were later used by Hursh and Brater (1941) to demonstrate the existence of 'underground storm-flow' in a Coweeta Forest watershed in North Carolina, and water moving in the immediate surface layers was proposed as a possible contributor to the storm hydrograph. Some workers have dismissed the idea of subsurface flow altogether, while others have questioned the possibility of its contributing to the storm hydrograph. Ragan (1968) for instance, interpreted ungauged lateral inflow to a stream channel in a forested Vermont watershed as rapid groundwater response. Despite favourable conditions in the Sleepers River watershed, Vermont, Dunne and Black (1970a) concluded that subsurface stormflow was insignificant in augmenting storm runoff and that overland flow from a restricted hillslope area made the only important contribution to channel runoff.

Freeze (1972b), while supporting the idea of subsurface flow, used mathematical simulations to show that its significance in supplementing storm runoff is limited. He was later criticised by Hewlett (1974) both for inadequate representation of natural basin conditions since Freeze confined his attention to a fixed catchment segment, and for ignoring the possibility of expansion and contraction of the stream channel system during storms. Weyman (1973), maintaining that throughflow provides stormflow only under high rainfall intensities or if organic horizons are saturated, emphasised the importance of lateral flow contributions to the hydrograph recession and to baseflow.

General disagreement over definition and interpretation of the components of runoff underlies much of the divergence of opinions on the significance of throughflow (Ward, 1975), and there has long been a need for more universal specifications of surface and subsurface water movement (Hewlett, 1974). Generalisations concerning runoff processes over wide areas are not always possible, since runoff-generating mechanisms may vary spatially, laterally and vertically even on a small scale and, furthermore, runoff generation may not be confined to a single process in one area (Pilgrim et al., 1978). Topography, soil properties, rainfall and vegetation characteristics determine specific runoff-generating processes, both regionally and within a basin. Contrasting experimental results are therefore inevitable, and the various runoff models complement rather than contradict each other (Dunne, 1978).

5.1.2 LAND-USE EFFECTS ON RUNOFF GENERATION

The complexities of analysing runoff processes and characteristics discussed above, are intensified by the effects of catchment land-use change. Experimental results must therefore be regarded firstly in the context of specific areas before broad comparisons are drawn between wider regions. Experiments on vegetation influences, which include alteration, replacement and removal of vegetation, date from the later part of the nineteenth century. More important early work was conducted during the first decades of the present century, however, with experiments at Wagon Wheel Gap, Colorado, commencing in 1909 (Bates and Henry, 1928), in Switzerland by Engler (1919) and at the Coweeta Experimental Forest in the southern Appalachians by Hoover (1944). Most studies concentrate on the effects on water yield, stressing the importance of the evapotranspiration component which includes transpiration and interception. Reduced evapotranspiration rates from devegetated zones or areas of low-growing crops result in lower soil moisture deficits and consequently greater runoff volumes than found under vegetated, especially forested areas.

In Great Britain, experimental work in this field reached a decisive change in the 1950s with the work of Law (1957a, 1957b). Large disparities in runoff between forested and grass-covered areas were exposed in controversial work at Stocks Reservoir in Yorkshire. From a lysimeter experiment in a sitka spruce (Picea sitchensis) plantation, Law concluded that 38% of rainfall was intercepted and evaporated over a one-year period. Comparisons made with results from the remainder of the largely grass-covered watershed indicated that trees yielded additional losses of 290 mm. The proposed extent of the loss generated a great deal of criticism, Law's work being attacked mainly on grounds of the small size of his study area (0.1 acre), introducing complexities created by edge effects, increased wind ventilation and enhanced radiation, together creating higher evaporation rates than would be expected from a large forest.

Research commencing in the 1960s by the Institute of Hydrology, on forested and grass-covered catchments on Plynlimon in central Wales supports the magnitude of losses expressed earlier by Law, however (Institute of Hydrology, 1976; Clarke and Newson, 1978; Calder and Newson, 1979; Clarke and McCulloch, 1979). Reduced water yields from the forested Severn catchment (mainly sitka spruce) were again explained largely in terms of the evaporation of intercepted water. Water balance calculations for the period 1972 to 1975 showed that mean annual evapotranspiration (mean annual precipitation minus mean annual streamflow) is approximately 281 mm greater from the Severn catchment than that from the adjacent grass-covered Wye catchment.

Changes in storage capacity may directly determine runoff response. The greater sensitivity of a burnt peat area was explained in these terms by Conway and Millar (1960) for the upper Tees in northern England. The devegetated surface, being drained by both natural and artificial channels, yielded earlier and higher flood peaks than a comparative Sphagnum-covered catchment. Delayed runoff response and prolonged recessions of the vegetated catchment were explained in terms of the larger storage capacity of the loose-textured _Sphagnum_ surface. Opposing work, however, reviewed by Wilcock (1979), indicates a moderated runoff response after peatland drainage, the latter being said to cause increased storage capacity above the water table.

Although there is general agreement on the fundamental causes of variation in hydrological regimes with vegetation modification, and extensive support has been lent to the idea of enhanced water use consequent upon afforestation and to one of greater runoff volumes following vegetation removal, reported magnitudes of change vary (Hoyt and Troxell, 1932; Hewlett and Helvey, 1969; Lewis, 1968; Pierce _et al._, 1973). Each experiment is governed by a set of contributing factors, including physical catchment features and climatic variables, which moderate vegetative influences on hydrology and in this context some evidence, discussed below, has been presented to contravene general opinion (Sodemann and Tysinger, 1967; Gash and Stewart, 1977). Thus, it may be necessary to identify the significance of other variables, such as underlying geology, rainfall regime and snow storage effects and to isolate distinctive hydrological cycle components (Hewlett and Hibbert, 1961).

Runoff responses from forested catchments may be modified according to the stage of forestry operations. An immediate effect, especially during heavy or prolonged storms, is the reduction in response time due to drainage and ploughing preparations, which effectively increase drainage densities. Thus, although mature forests promote reduction of peak runoff and of runoff volumes, under heavy rainfall flow may be augmented as a consequence of drain construction (Calder and Newson, 1979). Cyclical hydrological changes may signify times of thinning while clear felling is reflected in runoff increases (Binns, 1979). Using paired watershed experiments at the Coweeta Hydrologic Laboratory, Hewlett and Hibbert (1961) monitored the effects of different types of forest removal treatments, reporting increases in annual water yield of up to 41 cm (16 inches) over pre-treatment flow, the highest increases, as expected, pertaining to complete reduction of the forest stand.

Changing magnitudes of runoff following vegetation removal can depend on seasonal influences of catchment storage. Significant increases may be evident during the growing season, for example, while land-use remains unimportant in winter (Reinhart and Eschner, 1962; Rothacher, 1965). Conversely, in the years immediately following grass seeding of a Coweeta catchment, the early part of the growing season was marked by a greater water use than under the original hardwood forest cover, while significant increases in stream discharge were identified later in the year (Hibbert, 1969). Streamflow increase during the conversion however, was much less than that found by Hewlett and Hibbert (1961) immediately after other treatments at Coweeta.

Such apparently conflicting results may be explained in terms of further prevailing factors. Interception losses, for example, are modified by rainfall intensity and duration. Infrequent, high intensity or duration storms result in surface saturation and loss of protection by vegetation (Penman, 1963; Clarke and Newson, 1978), while under low rainfall regimes, losses from forest and grass may be similar when the higher interception losses found under forest fail to compensate for relatively low transpiration rates in summer (Gash and Stewart, 1977; Calder, 1979).

The presence of vegetation may have a limited effect on runoff from areas of coarse, permeable soils. Sodemann and Tysinger (1967), for example, reported that afforestation in east Tennessee instigated no significant change in water yield, since the shallow, well-drained soils overlying limestone bedrock facilitated infiltration of water which eventually emerged as springs near stream channels and was correspondingly unavailable for evapotranspiration. Magnitude of stormflow and of basin response may play important roles in determining flood volume from certain catchments, through sensitivity to rainfall intensity. Hewlett and Bosch (1985) found for South African catchments that rainfall intensity, and therefore overland flow, becomes important in flood runoff only as storm flow and basin response reduce in magnitude. As the proportional area of the catchment contributing to storm runoff increases, with more responsive basins, storm flow becomes less sensitive to small-scale variations in rainfall intensity. Land-use change (afforestation and grass veld burning) was therefore demonstrated to be of little importance to storm flow. Other important runoff-influencing factors include antecedent soil moisture conditions, topography, drainage density and the presence or absence of snow.

These conditions notwithstanding, mainstream thought remains in support of the directions of response emphasised previously. Thus, Bosch and Hewlett (1982), who updated and expanded Hibbert's (1967) earlier review of the effects of forest vegetation removal to include other vegetation types, generally corroborated Hibbert's conclusions of increased water yield following reduction in cover, increased streamflow diminishing in proportion to rate of vegetation recovery. The position of cleared zones in relation to runoff source areas was also identified as important in determining the results of only partially cutting a forested catchment.

The large body of existing literature on forest hydrology, as well as difficulties of control at the Egton study sites, designate the present dissertation as an assessment of the significance for surface runoff of removing medium-height moorland vegetation by controlled burning, a land-use change which has been subjected to little previous documentation; experiments on the implications of vegetation burning have largely been confined to semi-arid areas, involving forest, scrub or chaparral (e.g. Krammes and Rice, 1963; Pase and Ingebo, 1965; Wright et al., 1976). Subsurface flow modifications are examined for all three land-use types, however. Consequences for runoff are investigated both in terms of general rainfall-runoff relationships and through changes in stream hydrograph properties, while preliminary comparative predictions for runoff events are also proposed. Physical controls of water use are interpreted with particular reference to the importance of both soil moisture deficit and evapotranspiration, although discussion of the importance of these two variables in a wider context, in terms of the hydrological balance, is deferred until the next chapter. Subsurface response patterns are examined specifically, in the first section of the present chapter in relation to runoff generation theory. The second, and major part of the chapter is devoted to analysis of flood hydrograph responses.

5.2 SUBSURFACE FLOW RESPONSES

A comprehensive definition of subsurface stormflow was given earlier (Chapter 3, p.58). This lateral movement of water is subject to a number of modifying influences, affecting its speed of arrival at a stream channel, volumes of flow and flow characteristics. The importance of surface vegetation cover is illustrated in this section.

5.2.1 CONCEPTS OF SUBSURFACE MOISTURE MOVEMENT

Amounts and velocities of subsurface flow are basically attributable to changes in soil hydraulic conductivity, this generally decreasing with profile depth. Following rainfall, concentration of moisture in surface layers leads to a dominantly lateral movement of water downslope; this is due to greater lateral hydraulic conductivity (near the surface) than vertical profile conductivity. Hydraulic conductivity in turn depends on total soil porosity and on pore size. Thus, a saturated clay, consisting of tightly compacted particles, is less conductive than a highly porous layer. Under unsaturated conditions, however, a soil with small pores is more conductive than one characterised by larger voids, since small pores more easily retain water even at high suctions. Throughflow is further encouraged by a sloping surface, saturated soil, the presence of a sharp or gradual discontinuity in the profile, usually a less permeable soil horizon or iron pan, or by the existence of large soil voids or pipes. Assuming an average profile depth of 2 m, Tomlinson (1979) suggested the following vertical classification of flow in peat:

Profile Depth	
0 - 10 or 20 cm	Lateral flow (active layer)
Below 10-20 cm	Vertical flow

Vegetation controls throughflow magnitudes by its effect on infiltration and through the influence of organic matter on soil structure (Whipkey and Kirkby, 1978). Forested areas may be most conducive to subsurface flow, since soil permeability is maintained by forest litter while structural channels (old root holes, animal burrows and roots) may be more abundant under this cover (Whipkey, 1965).

Flow of water through a soil may be represented formally by combination of the Richards equation (an adaptation of Darcy's law, p. 80) with the Continuity Equation. Darcy's law is strictly valid only for low flow velocities within the laminar range (Chorley, 1978) although Whipkey (1967) maintained that it may become inapplicable at a point within the laminar range. Although the law is equally relevant to both saturated and unsaturated soil conditions, in the unsaturated state the porous body must be uniform and velocity of flow needs to be sufficiently small before the law can justifiably be applied (Childs, 1969).

The Continuity Equation is 'a statement of the conservation of mass during fluid flow through an elemental volume of the porous media' (Freeze, 1978, p.183). Net inflow of moisture to a volume of soil is the difference between inflow and outflow discharges: Continuity Equation for three dimensions:

$$\frac{\delta\theta}{\delta t} = -\left(\frac{\delta q_x}{x} + \frac{\delta q_y}{y} + \frac{\delta q_z}{z}\right) = -\nabla.q \qquad \text{Eq. 5.1}$$

where:

θ = volumetric moisture content

t = time

q_x, q_y, q_z = moisture fluxes in the x, y and z directions, respectively

∇ represents the spatial gradient of the flux, q

(Hillel, 1980a)

Knapp (1974) considered there to be a dichotomy between investigations which attempt to solve the equation of flow and those which confine themselves to the field situation, disregarding the equation.

General flow of water, represented as the Richards and Continuity Equations combined, is as follows:

$$\frac{\delta\theta}{\delta t} = -\frac{\delta}{\delta x}\left(k\frac{\delta\psi}{\delta x}\right) - \frac{\delta}{\delta y}\left(k\frac{\delta\psi}{\delta y}\right) - \frac{\delta}{\delta z}\left(k\frac{\delta\psi}{\delta z}\right) + \frac{\delta k}{\delta z} \qquad \text{Eq. 5.2}$$

where:

k = unsaturated hydraulic conductivity

ψ = matric suction head

It is possible to classify flow response to sufficiently heavy rainfall into dispersed matrix flow (flow through inter-granular pores and small structural voids) and concentrated pipeflow, the latter of which is not described by Darcy's law (Atkinson, 1978; Whipkey and Kirkby, 1978). These reflect two end-members of a whole sequence of different types of void. Weyman (1973) related unsaturated flow to capillary pores, and saturated to non-capillary pores. Matrix flow may be either saturated or unsaturated and, if laminar, is subject to Darcy's law, as discussed earlier in the context of theoretical modelling of soil moisture. Pipeflow, which is characterised by its rapid arrival at stream channels, is concentrated into subsurface conduits, usually ranging from between 1 cm and 2 cm up to several metres in diameter and often circular in cross-section (Atkinson, 1978). Whipkey (1967) concluded that a field equation is necessary to describe 'pipe-type' flow, with variable physiographic conditions incorporated and he described a number of different equations which have been proposed to illustrate turbulent flow in porous media.

5.2.2 ANALYSIS OF THROUGHFLOW AT EGTON

5.2.2.1 Results

Implications of land-use change for throughflow regimes were examined briefly in a previous section (4.3.3.5). The present discussion attempts to account for the variations in weekly throughflow volumes under different surface covers at the Egton site, using parametric and non-parametric statistical tests, as appropriate. Most factors which promote subsurface flow are apparent for the Egton catchment and combine to produce observed throughflow under forested, moorland and burnt surfaces.

Throughflow movement was almost continuous throughout winter under all three plots, declining fairly rapidly in spring to produce, with water table lowering, almost completely dry conditions in summer, notably in the surface layers. No significant difference was apparent when comparing amounts under heather moorland and woodland for the complete study period, at 15 cm depth; at 30 cm significantly larger volumes occurred under moorland than woodland (p. 143). Following heather burning, significantly greater amounts were found below the burnt plot than under woodland at both measuring depths suggesting that variations in vegetation cover have overriding effects over any due to slope, in the surficial soil layers at least. Comparison of results under heather with the same plot after burning illustrates no statistically significant difference for either measuring depth ($t = -1.058$, $n = 6$ for 15 cm depth and $t = -0.517$, $n = 6$ at 30 cm). However, the results must be viewed in a relative rather than an absolute sense, since significantly smaller volumes of throughflow were found to occur under woodland during the post-burn period than in a comparable period before vegetation removal (at 15 cm, $t = 1.755$, $n = 10$ (pre-burn), 24 (post-burn) significant at 0.05 level for 1 tail test; at 30 cm, $U = 50.5$ (Mann-Whitney U test), $n = 7, 24$, significant at 0.01 level for 1 tail test). Thus, since a constant vegetation background demonstrates a decrease in throughflow amounts between pre- and post-burn periods, then a stable situation for the moorland (no significant differences) suggests a relative increase in throughflow following vegetation removal by burning.

The results are explicable largely in terms of interception and transpiration, which are of greater significance under heather and woodland plots than for burnt moorland, leaving less water to reach the ground surface under vegetated sites. As discussed later (p.211), rather than enhancing soil permeability, the litter layer under heather and woodland may act as a further moisture interceptor, whilst under the burnt plot, dieback of old heather roots may provide small concentrated pathways for subsurface moisture movement. Soil compaction under bare ground, leading to formation of an increasingly impermeable surface and reduced infiltration is therefore proposed as an insignificant effect for the moorland during the period immediately following muirburn.

5.2.2.2 Sources of Error

The results are viewed in terms of potential errors both in their calculation and in the light of spatial variability. During heavy rainfall, points of throughflow on a slope are fed by flow from further upslope so that depth of saturation eventually increases towards the slope base. On saturation, lateral flow is promoted by lines of equal moisture potential lying orthogonal to the slope (Weyman, 1973). The relative increase in throughflow found after vegetation burning may therefore be underestimated to some extent, upslope regions of the site remaining vegetated (Fig. 2.2). Subsurface flow to the lower slopes is thus reduced in comparison to expected volumes, had the complete moorland slope area been burnt.

Spatial variability is expressed on different scales. As hillslopes generally show increasing wetness towards the base, measuring sites were selected to hold comparable positions in relation to the rest of the slope area. It is difficult, in this respect, to assess the influence of the intervening road, roughly marking the woodland/moorland boundary. It may act to divide moisture conditions, enhancing moisture accumulation at the moorland/road junction. It is suggested however, that interference by the structure is moderated since moisture is transferred from the moorland area via a culvert to the stream channel, although re-direction of water along the ditch on the upslope side of the road marks an artificial flowline for moisture received by the wooded part of the slope.

Spatial variation of throughflow status within each land-use plot strictly demands extensive instrumentation for complete assessment (Chapter 3). Throughflow patterns may vary over small distances through the effect of topographical contours on convergence and divergence of flow. Spatial variability, which can itself be altered between storms, may be enhanced under forest where stemflow can lead to localised concentrations of soil saturation and thus to areas of high hydraulic pressure (Whipkey, 1967). General spatial patterns of moisture concentration are considered in detail in the following section (5.2.3).

In certain instances, further errors may result from overestimation of the differences in flow between wet and dry periods. This is directly due to the effect of a measuring pit on moisture flow lines; in wet periods flow lines from an area larger than that of the pit width converge on the pit, whilst under dry conditions, upslope soil water tensions divert flow movement away from the measuring area. These two effects are to a certain extent counteracted in the land-use comparisons undertaken here and this, along with reduced flow line alteration with the present instrumentation, lessens the possibility of errors from this source.

5.2.3 IDENTIFICATION OF MOISTURE CONCENTRATION ZONES

Spatial variations in soil moisture content, both within and between vegetation zones are revealed for the Egton catchment by plotting and interpolating surface and total profile moisture values, the latter being considered here in terms of potential runoff source areas. Eschner (1967) remarked that 'extreme variation is the rule rather than the exception in both interception and soil moisture distribution', and it is therefore surprising that although lateral variation in physical properties of soils is generally recognised, implications for water regimes of confined areas, such as agricultural fields, have not been evaluated (Hillel and Hornberger, 1979). Although the extent of soil moisture variation differs with the degree of scale resolution, moisture generally varies to a significantly greater extent under bare plots than in vegetated areas, and with increasing proximity to surface layers of the soil. Factors influencing soil moisture variability are divisible into two groups; these are not necessarily mutually exclusive (Reynolds, 1970c):

i) Static factors, for example, slope aspect and certain soil properties.

ii) Dynamic factors, including vegetation cover and antecedent moisture conditions.

Variations in factors such as soil type, vegetation and topography were classified as mesoscale variations by Hawley et al. (1983) while macroscale differences are the result of variations in meteorological conditions. The effects of vegetation and topography (slope angle, aspect and position on slope) are reflected through infiltration, runoff and evapotranspiration characteristics, although size of sampling area determines which of these factors is important.

To represent spatial soil moisture variations in the present study, a series of maps was produced using the computer package 'SYMAP'. Neutron probe access tubes provided data points for total profile (0 cm to 80 cm) moisture content, represented on an average monthly basis, while surface moisture sampling sites acted similarly as reference points to produce maps for 0 cm to 2.5 cm and 2.5 cm to 7.5 cm depths for each sampling occasion (Appendix III). Contour interpolation between data points provides the final map version (Figs. 5.1 to 5.6).

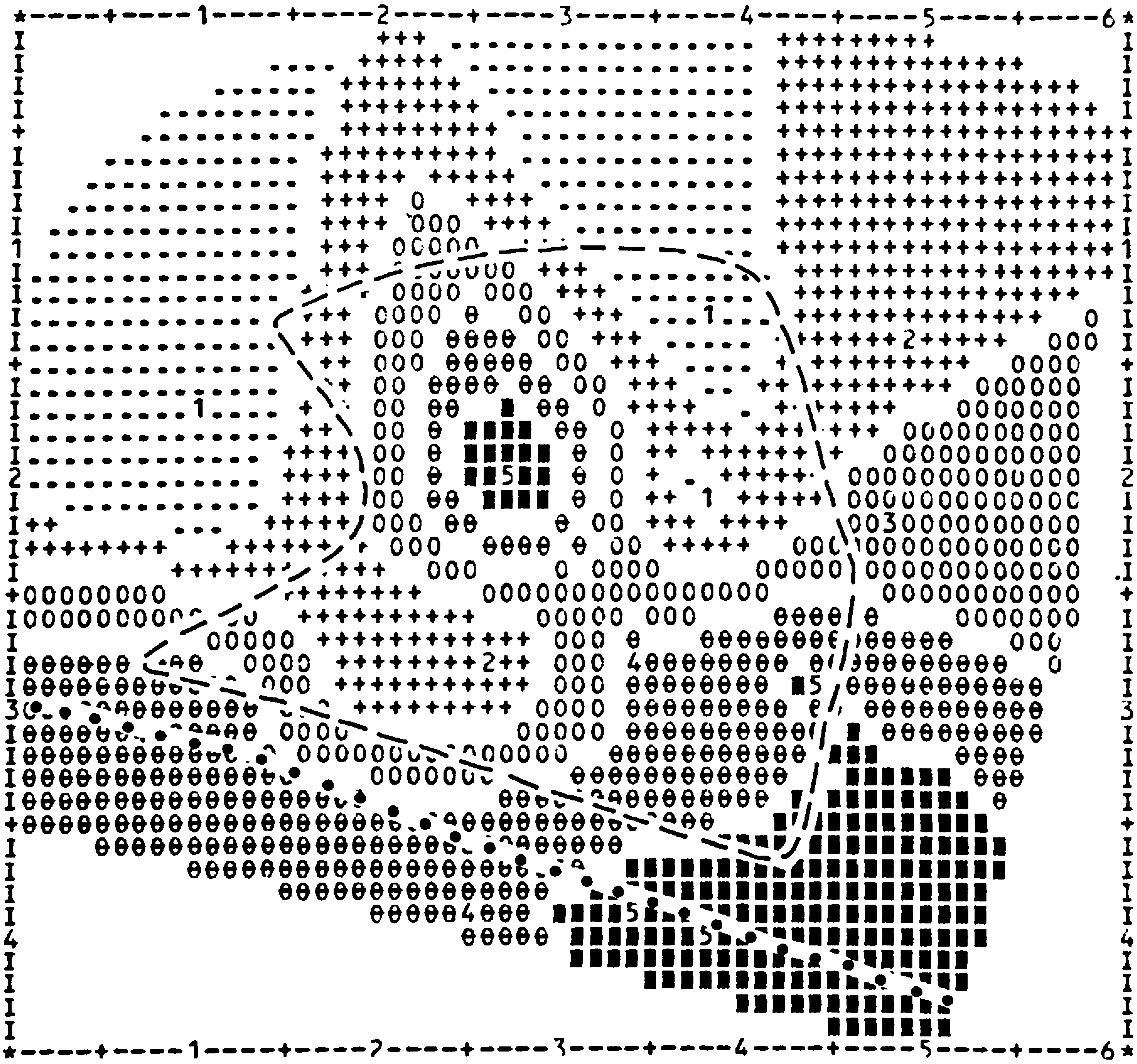
5.2.3.1 Surface Moisture Variations

Soil samples were collected from the immediate surface (litter) layer (0 cm to 2.5 cm) and the underlying 5 cm as part of a monthly programme to determine moisture variations within part of the surface layer, the zone in which the neutron probe becomes less accurate. Maps produced from these measurements are less representative than the complete profile maps since the former display conditions pertaining at one specific sampling time only. Thus, rather than interpreting relationships between general processes and temporally integrated distributions, discussion is confined here to the short-term influences on spatial patterns of surface moisture, for which variations due to antecedent moisture conditions or water table fluctuations are exposed. Moisture patterns are therefore examined below for specific cases in terms of prevailing topography, land-use, season and antecedent moisture conditions. Although the general effects of antecedent moisture may vary with soil type (Hawley _et al._, 1983) soil moisture variability is probably minimal after a dry period, when both the effects of soil heterogeneity on infiltration, and soil moisture holding capacity are at a minimum. In comparison, standard deviation is greatly enlarged following a rainfall event

Figures 5.1 to 5.6:
Spatial Variations in Soil Moisture Content over the Egton Catchment, derived using 'SYMAP'
Values in Figures 5.1 to 5.5 represent percentage moisture of wet weight of soil; those in Figure 5.6 are mm. moisture in the soil profile.

● ● ● ● Moorland/Woodland Boundary

Area of Burnt Moorland



ABSOLUTE VALUE RANGE APPLYING TO EACH LEVEL
('MAXIMUM' INCLUDED IN HIGHEST LEVEL ONLY)

MINIMUM	43.95	50.50	57.06	63.62	70.17
MAXIMUM	50.50	57.06	63.62	70.17	76.73

```
FREQUENCY DISTRIBUTION OF DATA POINT VALUES IN EACH LEVEL
 LEVEL          1          2          3          4          5
          ===================================================
          .........  +++++++++  000000000  θθθθθθθθθ  █████████
          .........  +++++++++  000000000  θθθθθθθθθ  █████████
 SYMBOLS  ....1....  ++++2++++  000030000  θθθθ4θθθθ  ████5████
          .........  +++++++++  000000000  θθθθθθθθθ  █████████
          .........  +++++++++  000000000  θθθθθθθθθ  █████████
          ===================================================
 FREQ.        3          3          1          3          4
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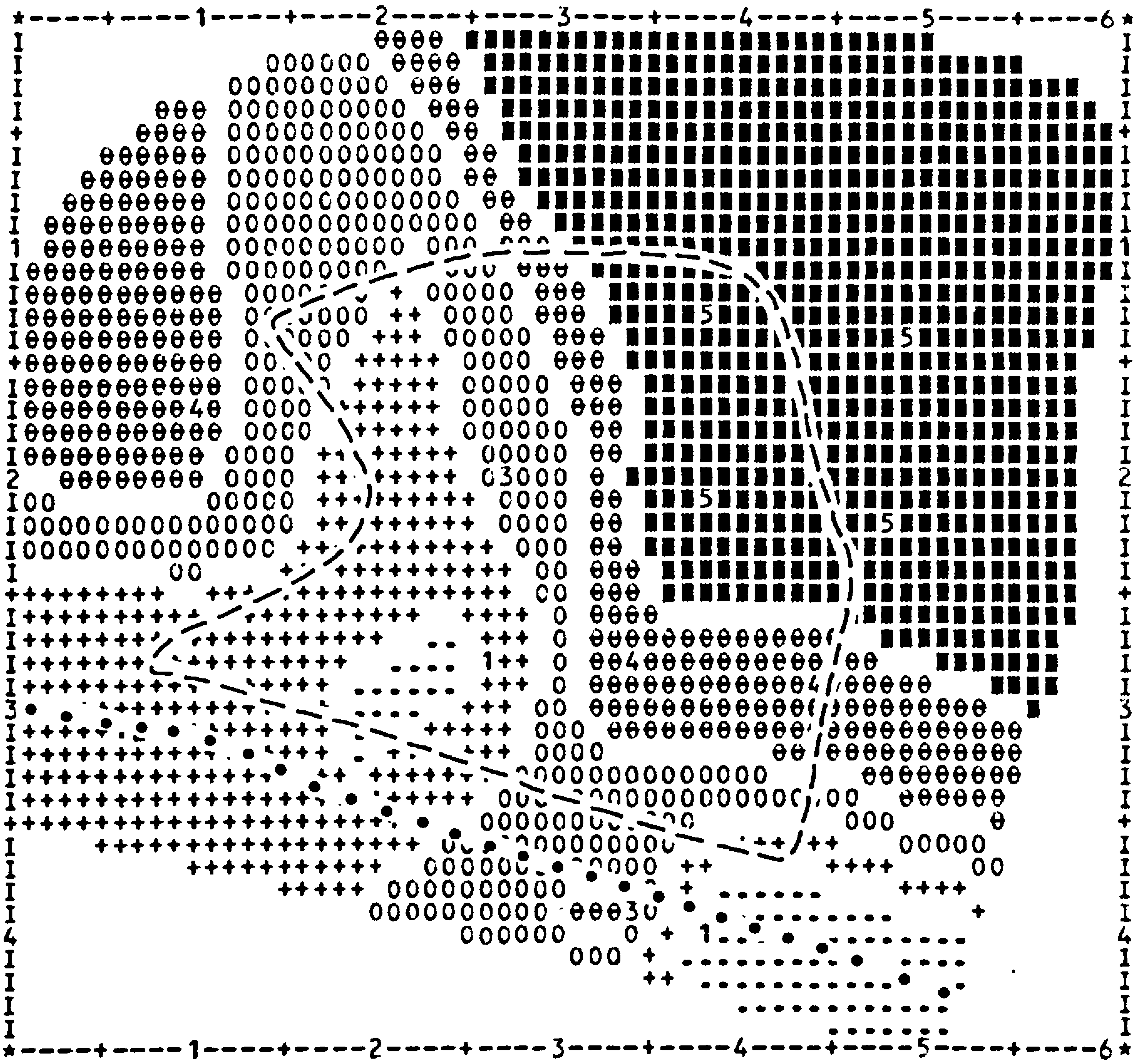
       2   I..1..I    I++2++I               Iθθ4θθI    I██5██I
       3   I..1..I    I++2++I               Iθθ4θθI    I██5██I
       4                                               I██5██I
```

Figure 5.1 Litter Layer (0 cm - 2.5 cm) Variations, 14 May 1981



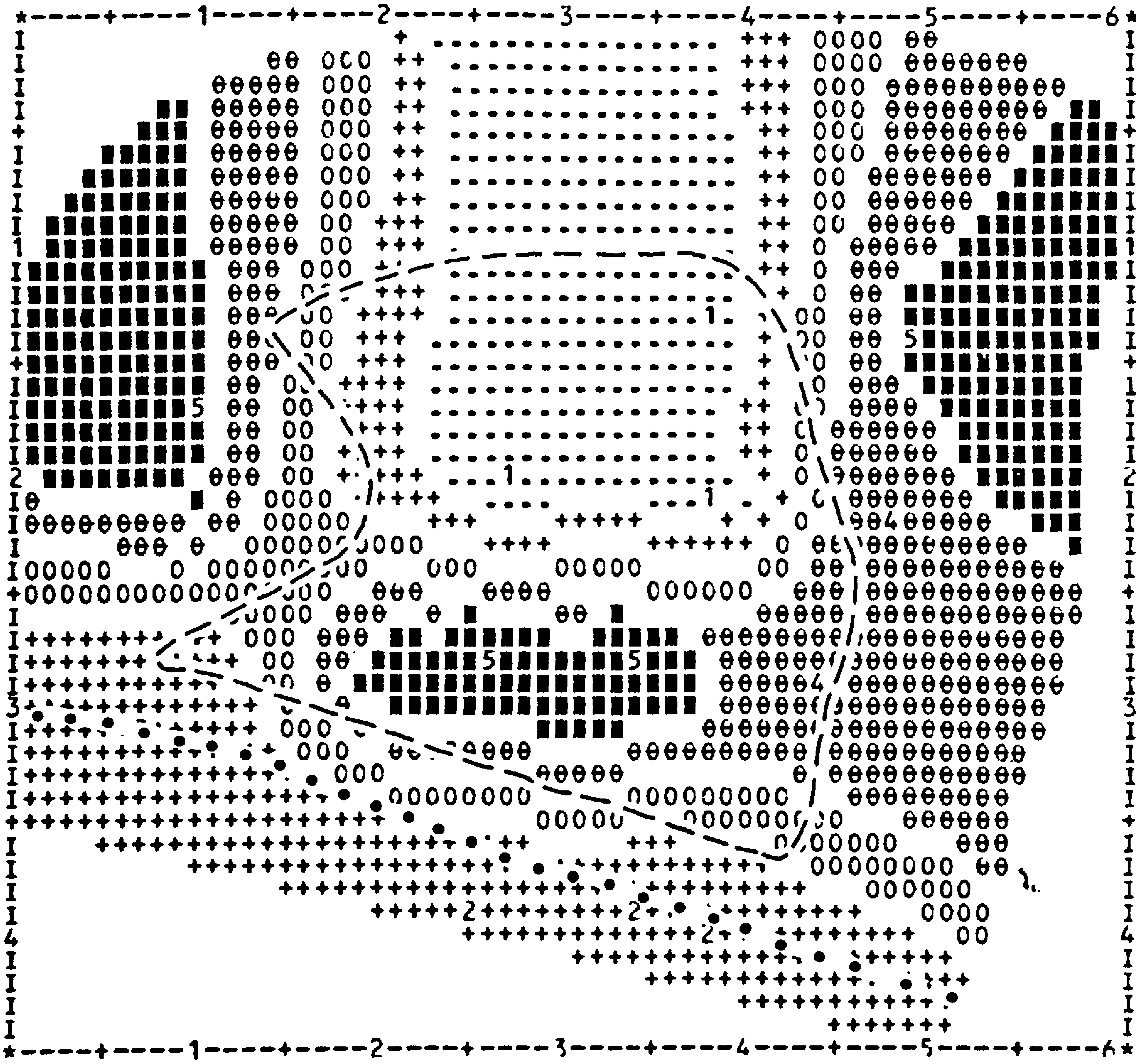
ABSOLUTE VALUE RANGE APPLYING TO EACH LEVEL
('MAXIMUM' INCLUDED IN HIGHEST LEVEL ONLY)

MINIMUM	26.85	37.29	47.72	58.16	68.59
MAXIMUM	37.29	47.72	58.16	68.59	79.02

FREQUENCY DISTRIBUTION OF DATA POINT VALUES IN EACH LEVEL

LEVEL	1	2	3	4	5
	....	+++++++++	000000000	ΘΘΘΘΘΘΘΘΘ	■■■■■■■■■
	....	+++++++++	000000000	ΘΘΘΘΘΘΘΘΘ	■■■■■■■■■
SYMBOLS	....1....	++++2++++	000030000	ΘΘΘΘ4ΘΘΘΘ	■■■■5■■■■
	....	+++++++++	000000000	ΘΘΘΘΘΘΘΘΘ	■■■■■■■■■
	....	+++++++++	000000000	ΘΘΘΘΘΘΘΘΘ	■■■■■■■■■
FREQ.	2	2	2	3	4
1	I..1..I	I++2++I	I00300I	IΘΘ4ΘΘI	I■■5■■I
2	I..1..I	I++2++I	I00300I	IΘΘ4ΘΘI	I■■5■■I
3				IΘΘ4ΘΘI	I■■5■■I
4					I■■5■■I

Figure 5.2 Litter Layer (0 cm - 2.5 cm) Variations, 18 November 1980



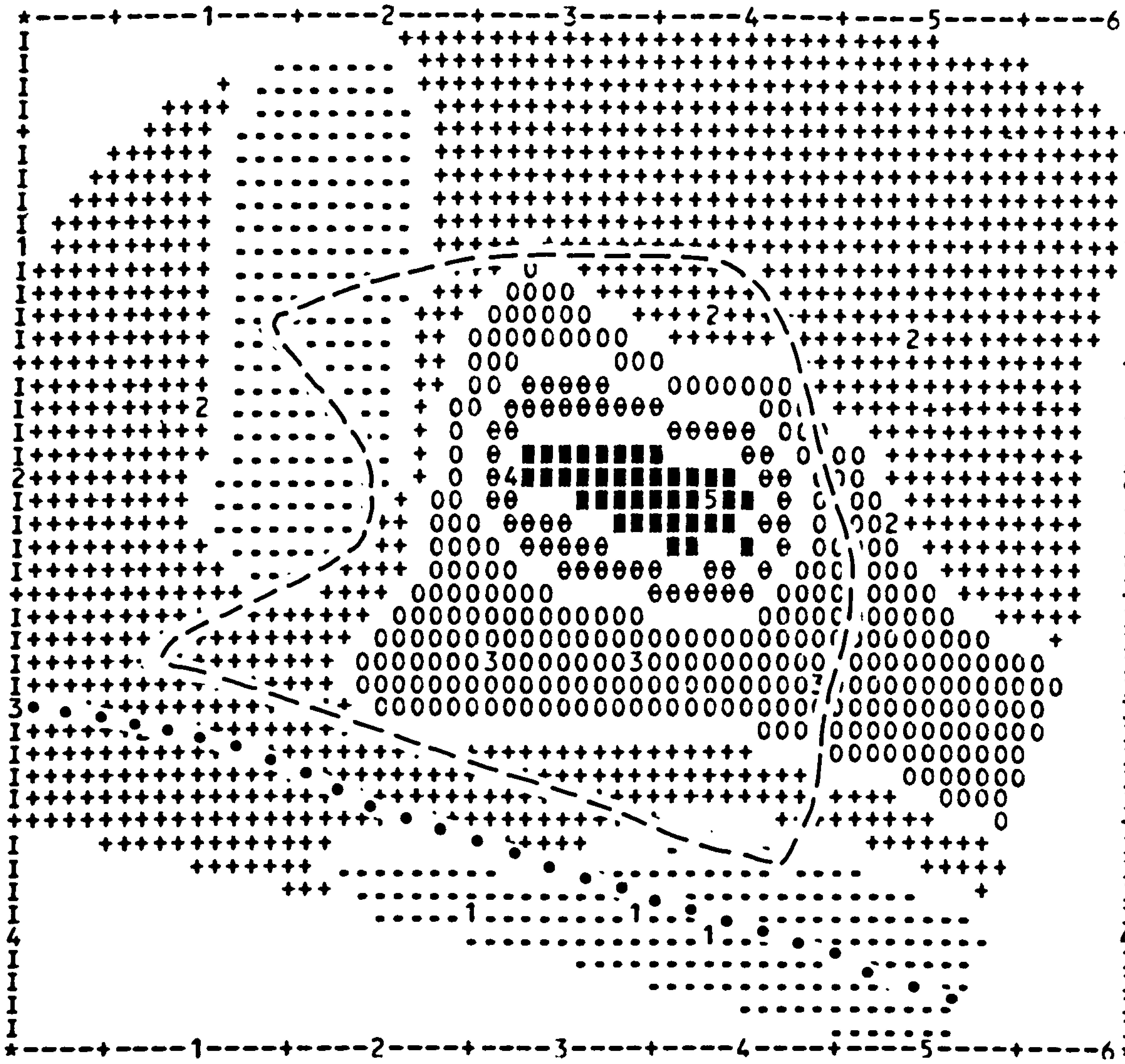
ABSOLUTE VALUE RANGE APPLYING TO EACH LEVEL
('MAXIMUM' INCLUDED IN HIGHEST LEVEL ONLY)

MINIMUM	14.92	25.31	35.69	46.08	56.46
MAXIMUM	25.31	35.69	46.08	56.46	66.85

FREQUENCY DISTRIBUTION OF DATA POINT VALUES IN EACH LEVEL

LEVEL	1	2	3	4	5
	....	+++++++++	000000000	θθθθθθθθθ	■■■■■■■■■
	....	+++++++++	000000000	θθθθθθθθθ	■■■■■■■■■
SYMBOLS	....1....	++++2++++	000030000	θθθθ4θθθθ	■■■■5■■■■
	....	+++++++++	000000000	θθθθθθθθθ	■■■■■■■■■
	....	+++++++++	000000000	θθθθθθθθθ	■■■■■■■■■
FREQ.	3	5	0	2	4
1	I..1..I	I++2++I		IθθθθθI	I■■5■■I
2	I..1..I	I++2++I		IθθθθθI	I■■5■■I
3	I..1..I	I++2++I			I■■5■■I
4		I++2++I			I■■5■■I
5		I++2++I			

<u>Figure 5.3 Typical Litter Layer (0 cm - 2.5 cm) Variations during Post-Burn, Summer Months</u>



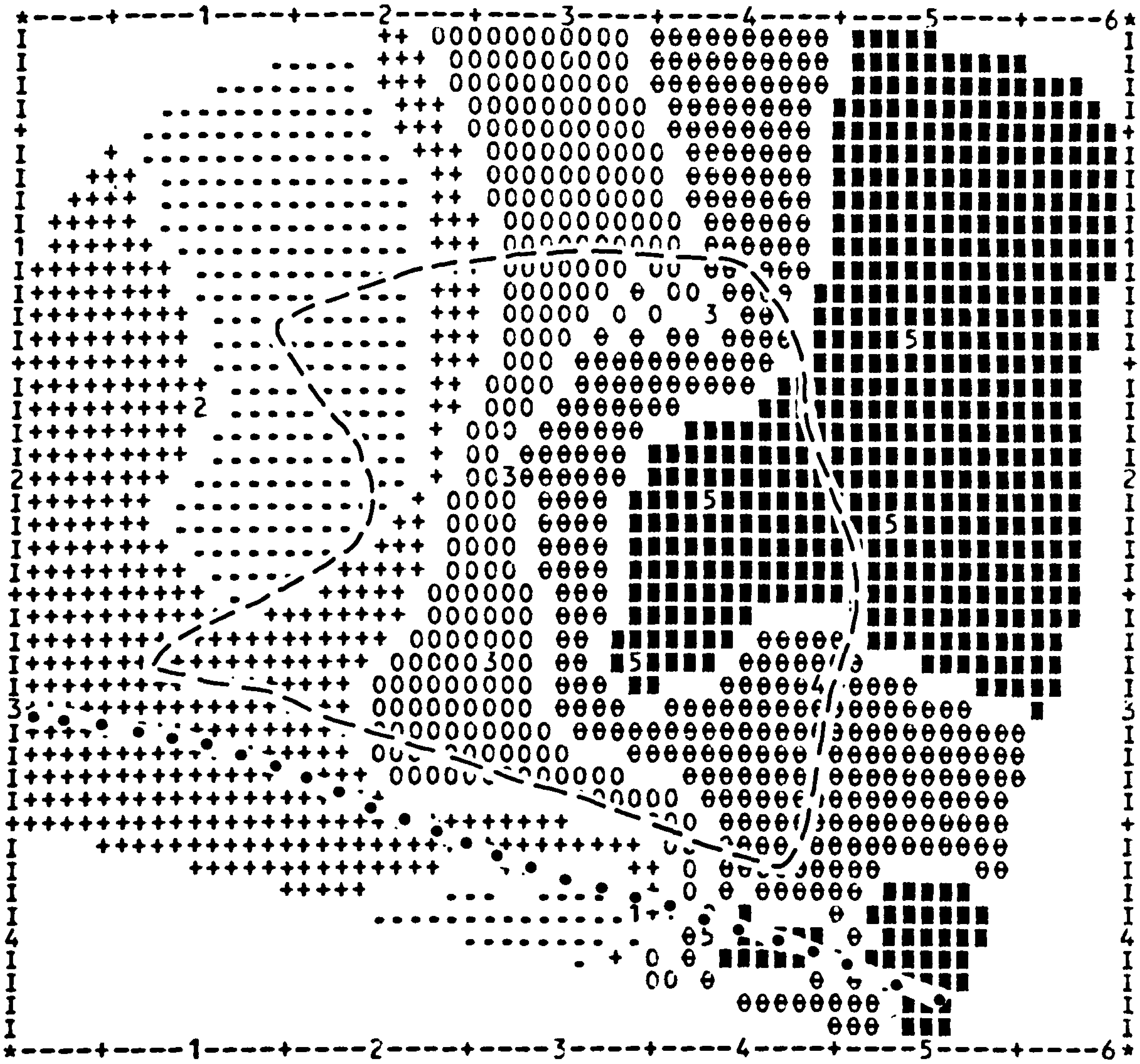
ABSOLUTE VALUE RANGE APPLYING TO EACH LEVEL
('MAXIMUM' INCLUDED IN HIGHEST LEVEL ONLY)

MINIMUM	60.24	67.76	75.27	82.78	90.30
MAXIMUM	67.76	75.27	82.78	90.30	97.81

FREQUENCY DISTRIBUTION OF DATA POINT VALUES IN EACH LEVEL

LEVEL	1	2	3	4	5
	....	+++++++++	000000000	ØØØØØØØØØ	■■■■■■■■■
	....	+++++++++	000000000	ØØØØØØØØØ	■■■■■■■■■
SYMBOLS	....1....	++++2++++	000030000	ØØØØ4ØØØØ	■■■■5■■■■
	....	+++++++++	000000000	ØØØØØØØØØ	■■■■■■■■■
	....	+++++++++	000000000	ØØØØØØØØØ	■■■■■■■■■
FREQ.	4	5	3	1	1
1	I..1..I	I++2++I	I00300I	IØØ4ØØI	I■■5■■I
2	I..1..I	I++2++I	I00300I		
3	I..1..I	I++2++I	I00300I		
4	I..1..I	I++2++I			
5		I++2++I			

Figure 5.4 Typical Litter Layer (0 cm - 2.5 cm) Variations during Spring 1981



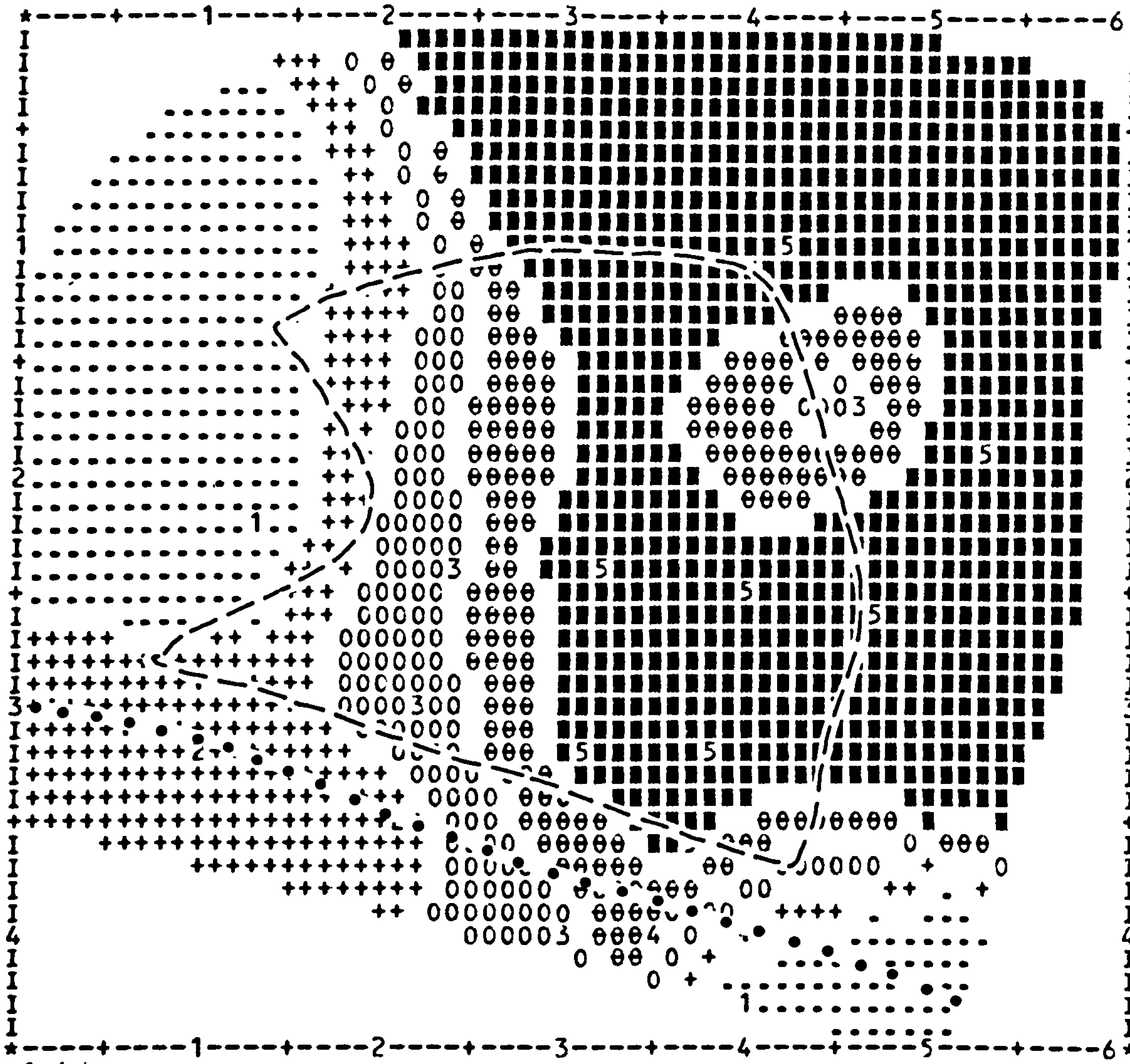
ABSOLUTE VALUE RANGE APPLYING TO EACH LEVEL
('MAXIMUM' INCLUDED IN HIGHEST LEVEL ONLY)

MINIMUM	27.46	37.65	47.84	58.03	68.22
MAXIMUM	37.65	47.84	58.03	68.22	78.41

FREQUENCY DISTRIBUTION OF DATA POINT VALUES IN EACH LEVEL

LEVEL	1	2	3	4	5
SYMBOLS	....	+++++++++	000000000	θθθθθθθθθ	■■■■■■■■■
	....	+++++++++	000000000	θθθθθθθθθ	■■■■■■■■■
	....1....	++++2++++	000030000	θθθθ4θθθθ	■■■■5■■■■
	....	+++++++++	000000000	θθθθθθθθθ	■■■■■■■■■
	....	+++++++++	000000000	θθθθθθθθθ	■■■■■■■■■
FREQ.	2	2	3	1	5
1	I..1..I	I++2++I	I00300I	IθθθθI	I■■5■■I
2	I..1..I	I++2++I	I00300I		I■■5■■I
3			I00300I		I■■5■■I
4					I■■5■■I
5					I■■5■■I

Figure 5.5 Typical Variations for 2.5 cm - 7.5 cm Depth during Winter Months



ABSOLUTE VALUE RANGE APPLYING TO EACH LEVEL
('MAXIMUM' INCLUDED IN HIGHEST LEVEL ONLY)

MINIMUM	385.10	462.36	539.62	616.88	694.14
MAXIMUM	462.36	539.62	616.88	694.14	771.40

FREQUENCY DISTRIBUTION OF DATA POINT VALUES IN EACH LEVEL

```
LEVEL          1           2           3           4           5
          ==================================================================
          .........   +++++++++   000000000   θθθθθθθθθ   ■■■■■■■■■
          .........   +++++++++   000000000   θθθθθθθθθ   ■■■■■■■■■
SYMBOLS   ....1....   ++++2++++   000030000   θθθθ4θθθθ   ■■■■5■■■■
          .........   +++++++++   000000000   θθθθθθθθθ   ■■■■■■■■■
          .........   +++++++++   000000000   θθθθθθθθθ   ■■■■■■■■■
          ==================================================================
FREQ.          2           2           4           1           8
     1    I..1..I     I++2++I     I003001I    Iθθ4θθI     I■■5■■I
     2    I..1..I     I++2++I     I003001I                I■■5■■I
     3                            I003001I                I■■5■■I
     4                            I003001I                I■■5■■I
     5                                                    I■■5■■I
     6                                                    I■■5■■I
     7                                                    I■■5■■I
     8                                                    I■■5■■I
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Figure 5.6 Typical Variations for 0 cm - 80 cm Depth shown throughout the Period July 1980 - March 1982

(Reynolds, 1970c; Bell et al., 1980). After a particularly heavy and prolonged storm, either moisture uniformity prevails due to saturation, or variation is maximised on a microscale due to pore size differences.

Consideration of factors other than vegetational changes needs to be included during map interpretation. Although, as has been previously stressed, physical homogeneity played an important part in catchment selection here, a certain degree of variability in factors such as topography and soil properties is inevitable in an empirical study. On a large scale this diversity is acceptable, but may be used on a small scale to explain spatial differences, as here, in soil moisture.

a) Litter Layer Variations

Sampling on 14 May 1981 (Fig. 5.1) showed surficial soil layers under woodland to have increased moisture contents over those found in the moorland zone. The preceding period 22 April to 12 May was characterised by several short, light bursts of rain each of which enabled moisture to be retained by woodland litter, with insufficient time for drainage before onset of further rain. The effect is pronounced under woodland vegetation since greater accumulation of litter under tree crowns prevents 'loss' to the soil profile (Eschner, 1967). Stemflow and throughfall are particularly important in inducing localised moisture differences under trees and their significance in this respect varies in extent with stem diameter and form, and crown position. Soil moisture generally increases with increasing distance from the centre of the tree crown, although variation can occur both within certain species and as a result of differences in soil properties and micro-relief (Eschner, 1967).

Occurrence of two large rainfall events, 12 mm and 8 mm, in the four days preceding a pre-burn November sampling date helps to explain the higher degree of wetness over the upslope regions of the catchment at this time (Fig. 5.2). Reduced slope gradients aid water retention towards catchment boundaries, a certain lag occurring in drainage to lower slope levels. Following heather burning this pattern was generally typical during summer months when the central moorland zone was characterised by the driest layers, and catchment divides and slope base were wettest (Fig. 5.3). Woodland soils showed distinctly reduced moisture contents at this time. During spring 1981, both before and after vegetation removal, the reverse, drying towards catchment boundaries, was evident (Fig. 5.4). Surface layers of burnt ground appear to be subject to extremes of moisture content since increased moisture in a zone covering most of this plot was a marked feature of winter 1981/1982.

b) Depth 2.5 cm to 7.5 cm

Early season (spring) months, both before and following heather burning, were marked by increased wetness over the moorland slope base and throughout the central moorland zone at this depth. The lowest moisture contents evident close to the stream in the woodland plot at both depths of measurement, although potentially explicable in terms of small-scale variations in soil characteristics, are more probably explained by the interpolating qualities of the mapping process and the use of only specific sampling dates. A denser concentration of sampling points would help to clarify surface moisture patterns for this zone. Winter patterns at the lower depth (1980/1981 and 1981/1982) (Fig. 5.5) begin to reflect those of total profile moisture content, which are interpreted below in terms of topography and 'source areas'.

5.2.3.2 Total Profile Variations

Similar patterns of moisture pertain throughout the complete study period (July 1980 to March 1982) (Fig. 5.6), the general pattern being one of increasing moisture content in a southerly direction across the catchment area, with a drier 'island' towards the south eastern edge. The driest moorland region is usually that beneath the youngest heather, to the north. Increased wetness towards the stream zone is evident throughout the year.

The zone of high moisture content observed over a section of the moorland slope base shows physical characteristics coincident with the first three of Kirkby and Chorley's (1967) zones of moisture concentration (p. 187). Delimitation of this kind of physiography as an area of flow convergence is supported by the work of Anderson and Burt (1977a, 1978) who showed the importance of topographical hollows in generating a throughflow response and in maintaining baseflow during hydrograph recession. The driest moorland profiles beneath younger heather may be similarly explained in terms of topography. Contours here form a spur at the slope base, with an area of parallel contours (straight slope) further upslope (Fig. 2.2) and drainage in this area is therefore good.

On shallow slopes as here, however, contributing areas may not always correspond to slope concavities and hollows, but may move during the course of a storm to include hillslope spurs. The important influence of soil water potentials on total potential explains this variation (Anderson and Kneale, 1980) and several such changes in the topography/soil water potential relationship may occur throughout drainage (Anderson, 1982). High total moisture concentrations towards the remaining catchment divides at Egton may be partly the result of lower slope angles and the inability of soil to transmit water. Plant species characteristic of poorly-drained areas (Juncus effusus, J.squarrosus, Carex nigra and Sphagnum papillosum) begin to colonise towards the top of the slope.

No obvious large-scale feature is evident to explain the small area of low moisture content towards the south eastern edge of the catchment and minor variations in soil properties, through their effect on infiltration capacity, may account for its presence. Surface cracks, for example, may promote the by-passing of surface layers by infiltrating water which is then absorbed three-dimensionally, in contrast to the normal one-dimensional process (Hillel, 1980b). Alternatively, soil storage capacity may differ over this part of the catchment. Thus, limited organic matter facilitates rapid attainment of saturation but retains less water than a deeper layer so that different stages in the infiltration process can be reached by different parts of a catchment at separate times, and wetting and drying may occur simultaneously in different parts of the profile.

5.2.4 SUMMARY

The influence of land-use on subsurface water has been examined both in terms of throughflow volumes, using the woodland area as a control against which to determine variations on the moorland, and in relation to potential source areas for runoff. Catchment characteristics proved conducive to subsurface flow, the process being observed under all three plots, although, despite suitable conditions (Jones, 1981), substantive evidence of flow through subsurface channels, or pipes, was limited. Both positive identification and further investigation into the causes of these features are required before definitive conclusions are drawn regarding this form of subsurface flow in the area studied.

Throughflow is enhanced after vegetation removal by burning and shows clear correspondence with periods of surface runoff. Increased flow is explained in terms of reduced losses to transpiration and evaporation of intercepted water, while remains of heather vegetation, stems and roots, enhance infiltration and subsurface flow processes, respectively. Topographic variations and areas of saturation help to explain spatial patterns of moisture content; the slope base is proposed as an area of flow convergence and is thus a potential contributing area, while the driest moorland area is identified with a spur feature.

5.3 STORM RUNOFF IMPLICATIONS

5.3.1 APPROACHES TO RUNOFF ANALYSIS

All comparative investigations require objective analysis and in this connection, several procedures for assessing storm runoff variations are considered here, prior to a more detailed discussion of the methodology adopted for the present study. Documented flood estimation and analysis have been approached through several different techniques involving varying degrees of sophistication. A complete evaluation of streamflow changes necessitates detailed description of hydrograph shape and runoff volume, while more limited assessments may suffice where data are lacking or basin requirements are not fulfilled. Runoff analysis procedures may be classified in a number of ways, different methods estimating separate flood features. Nash (1958), for example, distinguished three elements of rainfall-runoff analysis, as follows:

i) The relationship between rainfall volume and resulting storm runoff volume, defined as 'quick response' or 'direct runoff'.

ii) The distribution of storm runoff through time, incorporating the effect of the catchment itself.

iii) The relationship between the frequency of rainfall and that of discharge, requiring previous solution of the preceding two elements (i and ii).

The second component is the main consideration of the present chapter. Specific runoff events are evaluated through application of a proven technique, in order to define land-use controls on the hydrograph. The relationship between gross rainfall and total runoff is also briefly examined, in that proportionate allocations of rainfall to measured stream runoff are calculated on both an annual and a seasonal basis in this chapter, prior to a more specific interpretation of relationships within the water balance in Chapter 6. A range of available runoff estimation techniques, along with some practical applications, is considered in the following section.

5.3.2 METHODS OF RUNOFF EVALUATION

Approaches to runoff forecasting and prediction may be classified in a number of ways. Useful divisions may be made between statistical analyses, empirical and analytical techniques, and catchment models. Each of these approaches is discussed in the ensuing paragraphs and its potential for inclusion in the present study is considered.

5.3.2.1. Statistical Approaches

Apart from data summarising techniques, such as runoff percentages and maximum and minimum recorded flow, considered in a later section, statistical analysis is generally used to determine recurrence intervals of peak flow discharges or a given rainfall amount, and is most suitable for application to long-term records. Interpreting floods as random events, subject to chance, a probability distribution may be fitted to a series of flood events, and the recurrence interval of a given flood magnitude predicted from the distribution's parameters. Hydrological records may be extended synthetically, in order to facilitate flood prediction, by means of stochastic methods. The simulated values have the same stochastic properties as the recorded data and are normally derived by techniques such as Monte Carlo analyses (Kleijnen, 1974, 1975) or Markov chain processes (Bharucha-Reid, 1960; Dynkin, 1982).

5.3.2.2. Empirical and Analytical Procedures

Most important early rainfall-runoff relationships began to be constructed in the late nineteenth and early twentieth centuries, and although numerous empirical formulae were developed to calculate runoff volume or peak discharge, results were generally poor. Rodda (1969), for example, enumerated several formulae derived to identify flood peak discharges for certain recurrence intervals and commented on their predictive inaccuracies. The simpler relationships involve only one variable, usually catchment area, as exemplified below (Eq. 5.3), while further terms are incorporated in the more complex equations.

$$Q = C \cdot A^{0.6} \qquad \text{Eq. 5.3}$$

where:

Q = peak discharge

C = a constant

A = catchment area

The 'rational formula', often attributed to Lloyd-Davies (1906) represented an attempt to introduce rainfall intensity into calculation of peak discharge:

$$Q = C\ I\ A \qquad \text{Eq. 5.4}$$

where:

Q = peak discharge ($m^3\ s^{-1}$)

C = rational coefficient of runoff, depending on soil type, topography, surface roughness, vegetation and land-use

I = rainfall intensity ($mm\ h^{-1}$)

A = catchment area (km^2)

This formula relies on simplified assumptions, such as spatially uniform rainfall intensity and Hortonian runoff generation, which render it justifiable only for very small, relatively impervious drainage basins, and it returns generally unreliable results for large heterogeneous catchments. Peak discharge, Q is said to occur when the time of concentration is reached, that is, the period required for water received by the hydraulically most distant part of the catchment to reach the measuring outlet; this condition is taken to represent achievement of a steady state. Use of the relationship is therefore questionable for areas of variable source contributions. Nash (1958) showed the formula to assume a rectangular 'instantaneous unit hydrograph' (p. 232) having a uniform runoff ordinate for the entire time of concentration.

Analytical procedures for assessing runoff vary in their reliability and include Horton's (1933) now largely discredited approach to runoff volume estimation, based on infiltration capacity and rainfall intensity. An early useful method of estimating flood volume was coaxial graphical correlation developed by Linsley et al. (1949). Direct runoff amounts are estimated using a number of independent variables, usually an antecedent precipitation index and week of the year which, when combined provide an index of antecedent soil conditions, together with rainfall duration and amount, read off from a coaxial diagram comprising a series of curves. The variables are considered in sequence, initial moisture conditions being assessed first and storm variables introduced last. Net rainfall is adjusted until it is directly related to runoff volume (Weyman, 1975). The overriding importance of soil moisture deficit over storm duration and time of year was emphasised by Nash (1966) in his criticisms of the technique. He also maintained that the graphical correlation is difficult to apply and accuracy of prediction may not be assessed correctly. Further, statistical significance of the independent variables cannot be evaluated objectively. Correlation with basin characteristics is difficult because of the nature of the included parameters and the method is, therefore, to a large extent empirical (Weyman, 1975).

Information on the complete hydrograph, rather than on only a single characteristic such as peak discharge, is required by engineers and planners for reservoir design, in their assessment of modification of the flood form by flood-detention structures and in calculation of flood duration (Dunne and Leopold, 1978). The 'unit hydrograph', the hydrograph of direct surface runoff resulting from excess rainfall of unit volume and duration, is usually applied in the description of the time distribution of runoff. The method involves specification of the hydrological, geometrical and land-use relationships of a catchment (Overton and Meadows, 1976) and more detailed discussion of this method is delayed for later sections.

Other methods of calculating the form of the flood hydrograph include the application of linear reservoir transformation and the isochrone method, as well as combinations of different concepts. Reservoir transformation attempts to model catchment response by routing flow through a series of simulated reservoirs, a technique used by Nash (1957) to develop an equation for the instantaneous unit hydrograph. The isochrone method, involving division of the catchment into a number of sub-areas, not necessarily corresponding to internal watershed boundaries, relies on flood discharge being proportional to the areas from which water simultaneously reaches the catchment outlet (Ward, 1978). Thus, isochrones are lines which originate and terminate on the catchment boundary, and each of which joins a series of points which are separated from the catchment outlet by the same flow translation time, the time taken for water to reach the outlet (Dooge, 1959). Catchment sub-division in this way can then be used to generate hydrographs by using runoff routing procedures (Laurenson, 1964).

Traditional analytical methods such as coaxial graphical correlation and the unit hydrograph are disadvantaged by a dependence on the arbitrary separation of hydrograph components which have limited physical meaning. Reliable results can be obtained for individual catchments, however (Nash and Sutcliffe, 1970), while later methods of hydrograph analysis have concentrated, more realistically, on the speed of arrival of flow to the stream channel (Ward, 1975). Under certain circumstances, however, for example in making predictions for ungauged catchments, a catchment model of the rainfall-runoff relationship, involving interpretation of physical controls, may prove more reliable.

5.3.2.3. Catchment Models

Several catchment models, varying widely both in degree of complexity and in their application, have been developed as a further approach to flow forecasting. Complete basin, conceptual models, often using the idea of 'stores' in different parts of the catchment, are distinguished from the 'black box' classification of models, to which the unit hydrograph belongs, which ignore many of the physical processes and temporal and spatial non-linearities involved in the rainfall-runoff conversion. Catchment models aim to simulate all processes and components of basin hydrology and are generally designed in order that they may be subsequently applied to ungauged basins. They suffer from the potential disadvantage of over-complexity, although Kirkby (1975) suggested that between five and ten parameters is adequate for a model predicting runoff hydrographs from rainfall. The problem of autocorrelation, or parameter interdependency is also reduced with the number of parameters. The greater complexity of some catchment models was considered redundant by Betson and Ardis (1978) since, in large catchments at least, input variations become dampened at the output stage by the effects of channel characteristics, and land-use influences become less important. These authors suggested modelling separate sub-catchments, recombined by channel routing. General model requirements as proposed by Nash and Sutcliffe (1970) are summarised as follows:

i) A means of representing model significance.

ii) Simplicity, without excessive deviation from physical laws.

iii) A minimal number of parameters.

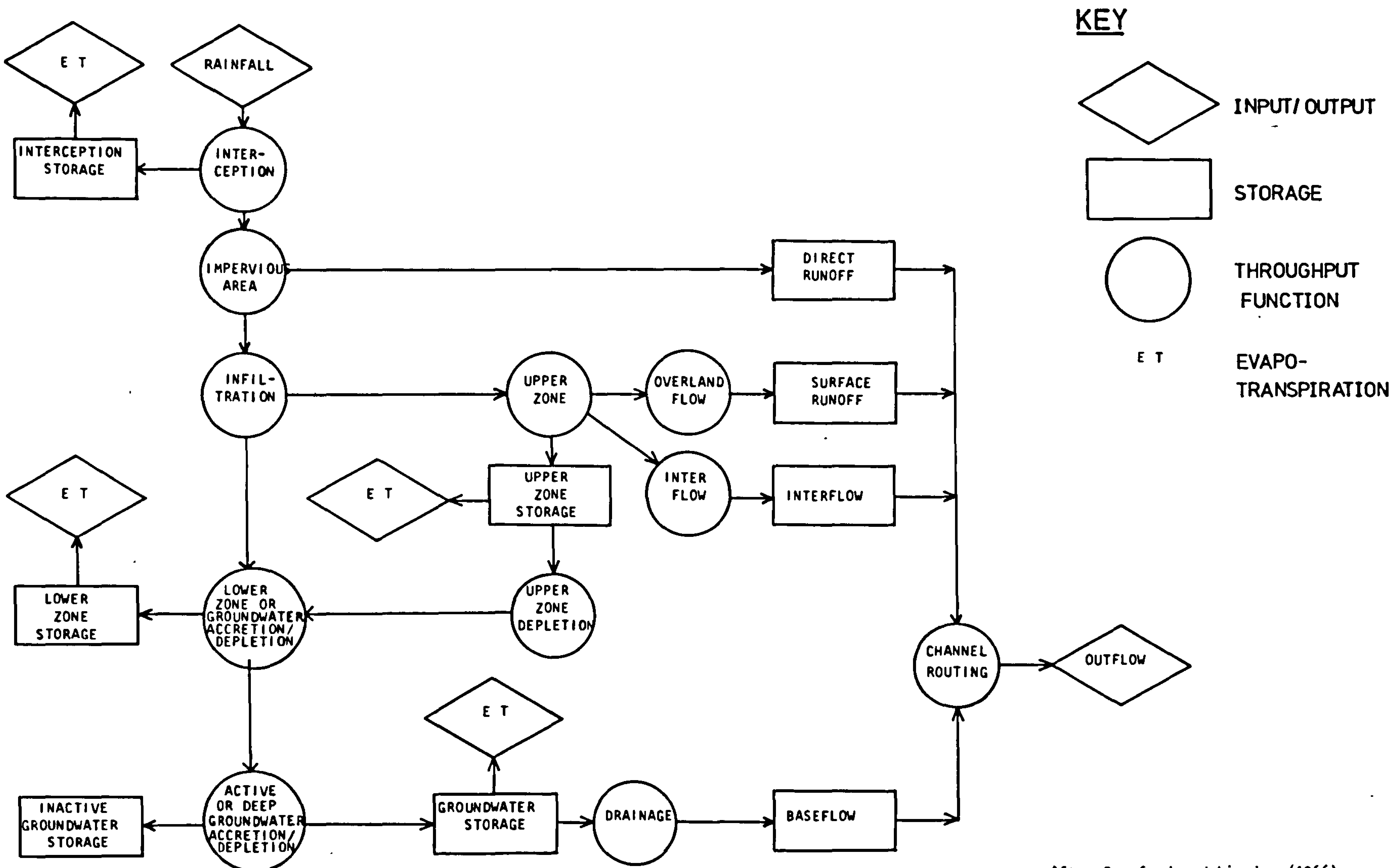
iv) Versatility of application.

One of the most widely known catchment models and one of the first to require computer usage is the Stanford Watershed Model (Crawford and Linsley, 1963, 1964, 1966), representing a major advance in hydrological analysis in the 1960s. The model has undergone considerable modification and improvement, including a variety of adaptations for climate and basin size, as well as increased resolution by means of reduced time increments. All facets of the hydrological cycle are incorporated in this model in order to generate runoff hydrographs by a water balance accounting procedure and, once calibrated, the model generates high quality simulations (Weeks and Hebbert, 1980). The forerunner of the model was a simple moisture accounting procedure to estimate mean daily surface runoff, using daily rainfall and potential evapotranspiration data (Linsley and Crawford, 1960). The method was employed on a trial basis in California and the predicted total daily streamflow, which comprises direct runoff, previous channel storage and groundwater flow, indicated general agreement with measured values. The amount of water held in a series of basin storage zones is calculated using inputs and outputs, any excess water contributing to runoff; direct runoff comprises precipitation in excess of infiltration capacity and runoff from impervious areas. Infiltration and percolation to groundwater are assumed to vary in a linear fashion with soil moisture storage, which is represented on two levels: soil moisture is assumed to be depleted at the potential rate from the upper level, while depletion of lower level storage is regarded as less important.

The Stanford Watershed Model per se was developed as an expansion of the earlier work to enable more detailed monitoring of hydrograph shape from short records (Fig.5.7). The model incorporates a partial area runoff component, infiltration being interpreted as spatially variable. Surface and subsurface runoff are routed downstream and, together with the groundwater component, comprise the total hydrograph, with throughflow assumed to be proportional to local infiltration capacity (Crawford and Linsley, 1966). Evapotranspiration occurs at the potential rate from the upper soil moisture zone until depletion, whence it occurs from the lower zone at a rate determined by moisture availability.

The model compares predicted and observed runoff and groundwater flows, altering those basin parameters which control runoff volume until reasonable estimates are obtained. Particular components of the model can be modified to enable simulation of the effects of different combinations of input variables. Assessment of the influence of catchment alteration on the water balance, for example, is achieved through variation of the parameters related to catchment characteristics. The Stanford Watershed Model has been criticised on the grounds of complexity, expense and difficulties in discerning parameter interrelationships (Overton and Meadows, 1976; Weeks and Hebbert, 1980), although a simpler version of the model is represented by the United States Geological Survey parametric model developed by Lichty et al. (1969) and concentrating on simulation of peak flows.

Numerous other conceptual models have been presented and include those developed in a series of papers by Nash and Sutcliffe (1970), O'Connell et al. (1970) and Mandeville et al. (1970). These



After Crawford and Linsley (1966)

Figure 5.7 Flow Diagram of Stanford Watershed Model IV authors developed a runoff forecasting model which uses a soil moisture accounting procedure with a limited number of parameters. The damping effect of basin storages on rainfall-runoff volumes is also included. 'HYSIM', a model developed by Manley (1978), has a potential range of uses and employs parameters based on catchment characteristics to predict flow in ungauged basins. A further simulation model, 'SHOLSIM', was developed by Aston and Dunin (1980) to predict the hydrological consequences of land-use change for a catchment in New South Wales. Meteorological data, along with several soil, vegetation and land form parameters are used as input. Land-use changes in selected 'hydrologically homogeneous' zones are simulated by altering the assigned vegetation type in the model.

Employment of a catchment model in the present study is rejected on fundamental grounds. The relatively short period of records hinders development of a valid catchment-specific model, while implementation of an existing model entails the need to select a system appropriate to the physical and hydrometeorological conditions of the study site. Since many models are designed specifically for larger catchments, application of a simpler rainfall-runoff conversion is perhaps more suitable for a small headwater area for which data are restricted, and for which many of the more sophisticated models, some requiring solution of complex equations or a large number of parameters, would be redundant. Thus, the intricate functions and demands for several calibrations of some models may be unwarranted where input data fail to reach the same degree of accuracy, and when the model inevitably falls short of its full potential. Rainfall-runoff evaluation and the analytical procedures selected, along with reasons for their applicability, are discussed in the remainder of this chapter.

5.3.3 RAINFALL-RUNOFF VOLUME RELATIONSHIPS FOR THE EGTON CATCHMENT

General catchment response is determined initially by examining relationships between rainfall and runoff volumes in terms of percentages of gross rainfall and total measured stream runoff. Relationships are assessed over a one-year period and are interpreted for the moorland area as a whole. Specific effects of heather burning on the hydrograph are evaluated through application of the 'unit hydrograph' method in relation to storm and catchment features in Section 5.3.4.

The proportion of rainfall which contributes to runoff varies with a number of factors, including rainfall intensity and duration, catchment geology, vegetation and soil charateristics, size and topography of the catchment and the size of the subsurface catchment. Complexities introduced by such factors as rainfall intensity and antecedent conditions are less important for the derivation of monthly or annual rainfall-runoff relationships, when storms are averaged over a period of time, than for those calculated on an individual storm basis (Linsley, 1967). Ratios of rainfall:runoff are particularly variable for headwater regions, where the percentage contribution of precipitation to quickflow (direct runoff) varies from storm to storm (Ward, 1984). In this context, therefore, simple rainfall:runoff ratios are derived for annual and monthly periods, and more objective analysis of the effects of land-use on storm hydrographs, considering antecedent conditions and rainfall 'losses', is left to the application of the 'unit hydrograph' technique to pre- and post-burn storm events. Indeed, over a sample of thirteen storm events, the latter type of analysis demonstrates that the proportion of total rainfall yielding quick response runoff, that is, after baseflow separation, varies from less than 1% to 33%.

Numerous rainfall:runoff ratios have been published and only a few general examples are quoted here, before discussion of results for the Egton catchment. In Great Britain as a whole, the ratio of annual runoff to rainfall generally varies from 18% to 86% (Francis, 1973). Calder and Newson (1979), in demonstrating enhanced water use by forested catchments, reported that an average of 83% of gross precipitation runs off from the grassland Wye catchment, compared to only 62% from the tree-covered Severn. Maxima of 25% to 50% of gross rainfall were quoted by Hewlett and Hibbert (1967) for forested catchments in the eastern United States, while Hewlett (1961b) found that for the Coweeta Hydrologic Laboratory specifically, the percentage of rainfall contributing to stormflow after separation of baseflow, seldom exceeded 35% and magnitudes of between 10% and 15% were more usual. An approximate, linear relationship between stormflow as a proportion of rainfall, and area of watershed generating stormflow was found to yield a maximum contributing area of approximately 45% of the watershed. Sodemann and Tysinger (1967) discovered that basic rainfall-runoff relationships have remained the same for the White Hollow watershed, Tennessee, over a long-term (thirty-year) period, with average annual runoff comprising roughly 40% of the average annual rainfall.

Percentages are calculated for the Egton site on an annual basis (17 December 1980 to 8 December 1981) and for individual monthly periods within the year. These specific periods are determined by data quality and availability and are chosen to coincide with periods for which all water balance data were attainable. Data collected during the initial monitoring period (July to November 1980) are omitted due to a lack of continuity of measurements during this time.

Runoff volumes are those measured on-site as gross stream discharge and converted to equivalent depth values in millimetres using the following expression:

$$\text{Runoff } (\text{mmh}^{-1}) = \frac{Q(\text{m}^3\text{s}^{-1}) \times 3600\text{s}}{\text{Basin Area } (\text{km}^2) \times 1000} \qquad \text{Eq. 5.5}$$

where:

Q = stream discharge

Daily precipitation values monitored by the Sneaton automatic weather station are used to determine total precipitation for each time period. This data source is selected in preference to measurements made on-site because the continuous, high quality data which this type of progressive analysis demands are more readily obtained from the Sneaton measurements. Allowances made for local storms, along with more specific details of data collection are described in the following chapter in relation to annual and seasonal water balances.

Over the total annual period, 57% of gross rainfall is accounted for by stream runoff, while seasonal variations in rainfall-runoff relationships are as illustrated in Figure 5.8. Stream runoff reaches its highest percentages during December and the early months of the year when, under saturated or nearly saturated soil conditions, runoff response is rapid, with over 70% of rainfall contributing to stream discharge during the winter season. Runoff exceeds rainfall input during two fortnightly periods at the end of 1981 and beginning of 1982. This apparently anomalous feature is explained by the occurrence of snow in preceding periods, followed by a discharge response only after snowmelt during these later intervals. An average of 35% of rainfall is shown to contribute to runoff during the summer months April to September, being reduced to only 9% in June and July. Rainfall-runoff relationships are elaborated further, with reference to the general water balance, in

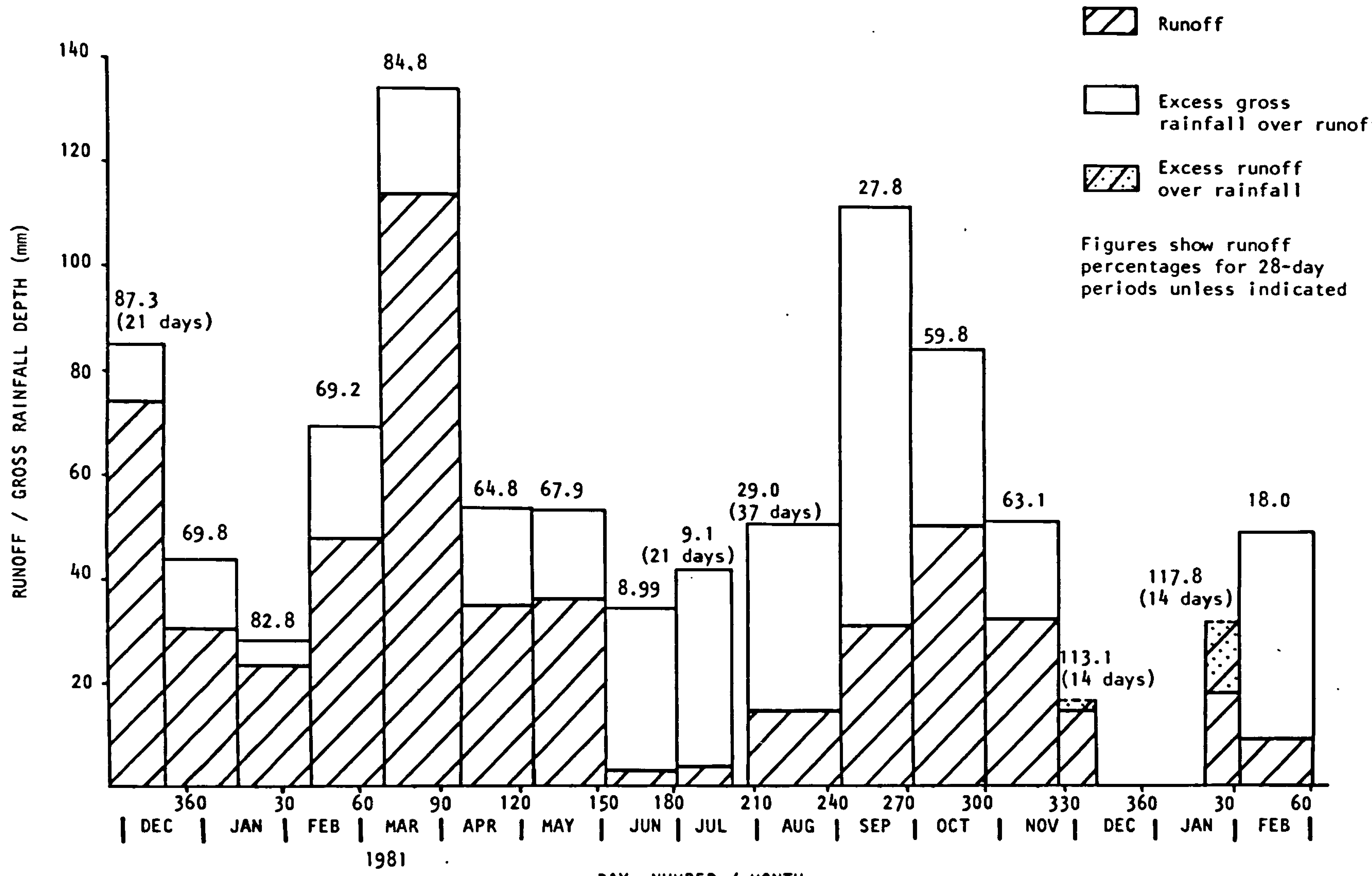


Figure 5.8 Seasonal Rainfall-Runoff Relationships

the next chapter.

5.3.4 THE DISTRIBUTION OF STORM RUNOFF IN TIME:THE UNIT HYDROGRAPH

The unit hydrograph model is adopted as the main runoff analysis procedure in the present study. This deterministic, part empirical, part theoretical method is chosen since it fulfills the objectives of the study and, indeed facilitates quantitative, comparative analyses of catchment response, by providing direct illustration of the effect of a unit amount of rainfall. In the light of its reliability and flexibility, the approach has met with extensive usage for runoff estimation from both gauged and ungauged catchments and, more specifically, in the evaluation of peak flows in engineering applications. The technique has been used particularly in demonstrating the effects of catchment urbanisation (Hall, 1974, 1977a, 1981; Hollis, 1974). Changes in flood hydrograph shape can be quantified easily, and hydrograph properties compared with physical catchment features. A single unit hydrograph can be derived and subsequently used to predict the effects of storms of differing size. The method is selected also for ease of application, since it yields effective results for practical purposes, yet requires only basic input data and involves fewer variables than would a catchment model, without the need for parameter optimisation. Diskin (1979, p.199) described the introduction of unit hydrograph and instantaneous unit hydrograph techniques as 'very important steps in the development of modern hydrology'. Methods of derivation are described in a later section (p.239).

5.3.4.1 Model Development

Introduction of the unit hydrograph concept is generally attributed to Sherman (1932), although a number of revised and improved versions have evolved since this procedure was first presented. Early modifications to the original technique include that by Bernard (1935) who described a 'distribution graph', determined by catchment characteristics and relating rainfall and runoff, with ordinates of the graph being expressed as a proportion of total catchment flow. Horner and Flynt (1936) used unit hydrograph equations to derive a 100% runoff graph which assumes no absorption or evaporation over an impervious area, while some of the early uses of the technique were discussed by Hoyt (1936). A suggestion for improved allocation of subsurface and surface storm flow was made by Langbein (1938) and in the following year, Brater (1939) showed the applicability of the unit hydrograph principle to small watersheds in particular. Despite early association between the unit hydrograph theory of runoff and Horton's (1933) runoff model, the technique may in fact be more applicable at present, for large catchments at least, in view of succeeding work on variable source areas (Hewlett and Hibbert, 1967). Quickflow-producing areas remain relatively constant and produce similar runoff volumes under similar rainfall conditions (Ward, 1975). For small catchments, where stream channel expansion and contraction occur rapidly and become more important, the unit hydrograph may yield more erratic results than envisaged by Sherman, as the direct runoff generating areas vary in extent (Hewlett and Hibbert, 1967). It is, nevertheless, often the smaller areas from which the best general results are obtained, since spatial variation in rainfall is more limited.

5.3.4.2 Concepts and Principles

The fundamental unit hydrograph principle is based on the conversion of rainfall to surface response runoff, a certain proportion of rainfall being 'lost' via interception, evaporation, transpiration, infiltration and seepage. The unit hydrograph itself is derived from the remaining quick response runoff hydrograph and is the result of a unit volume of net or effective rainfall, usually 1 mm, 1 cm or 1 inch falling uniformly over the catchment area at a uniform rate and in a time, 'T'; this is referred to as 'unit rain'. Runoff from rainfall of any duration or intensity can then be defined (Sherman, 1932). The unit hydrograph was later summarised by Sherman (1942, p. 514) as 'the hydrograph of surface runoff (not including ground-water runoff) on a given basin, due to an effective rain falling for a unit of time', the latter usually being less than 24 hours and always less than the time of concentration. For any catchment, similar responses, in the form of hydrographs, are said to be produced for similar storms and antecedent conditions, since physical basin characteristics, such as slope and size, remain constant or are specified (Sherman, 1932).

5.3.4.3 The Instantaneous Unit Hydrograph

If time of effective rainfall is synthetically reduced to zero, that is, a fall of unit net rain occurs instantly, the 'instantaneous unit hydrograph' or 'instantaneous response function' is the resulting unit hydrograph. This artificial concept is used in theoretical studies of rainfall-runoff relationships and has the advantage that rainfall duration is eliminated as a variable (Chow, 1964). Nash (1960) related the statistical moments of the instantaneous unit hydrograph to catchment characteristics and developed a general instantaneous, and finite period, unit hydrograph equation:

$$u = \frac{1}{K\Gamma(n)} (t/K)^{n-1} \exp^{-t/K} \qquad \text{Eq. 5.6}$$

where:

u = ordinate at time 't'

K = time constant in a first-order linear system

τ = a gamma function

n = a numerical parameter

Dooge (1959) proposed a general theoretical basis for the unit hydrograph method, developing an equation for the instantaneous unit hydrograph as a special case of a general unit hydrograph formula. Making the assumption that the reservoir action in a catchment can be separated from translation, and lumped in a number of reservoirs, he presented the following equation:

$$\frac{uT}{V_0} = \int_0^t P(m,n-1)\,\omega(\tau')\,dm \qquad \text{Eq. 5.7}$$

where:

u = ordinate of the instantaneous unit hydrograph

T = maximum translation time

V_0 = volume of rainfall excess

t = time since occurrence of rainfall excess

$P(m,n-1)$= Poisson probability function

m = dimensionless time factor, $(t-\tau)/K$

τ = translation time

K = size of linear reservoirs (all equal)

$n(\tau)$= number of linear reservoirs downstream of τ

$\omega(\tau')$= dimensionless time-area-concentration curve[1] (p. 234) adjusted for variation in rainfall intensity

In an applied sense, however, the instantaneous unit hydrograph is difficult to derive from discrete interval data, and has no real practical advantages over the finite period unit hydrograph (N.E.R.C., 1975).

5.3.4.4 Model Assumptions and Limitations

A number of fundamental assumptions underlie the unit hydrograph concept, some of which may not always be satisfied, although on the premise that they are met as closely as possible, valid results are obtained. Firstly, net rainfall must be uniformly distributed, both spatially over the catchment and temporally throughout a storm. Ideally, short duration storms giving a uniform, intense fall of rain should be selected for analysis, since these induce well-defined hydrographs with a short time base (Chow, 1964). One of the most common misapplications of the unit hydrograph technique arises from violation of this assumption (N.E.R.C., 1975). The spatial distribution of rainfall is most irregular over large catchments and under these circumstances, areal distributions must be carefully calculated using, for example, a network of raingauges and appropriate analytical techniques such as the Thiessen polygon method. The effects of this simplifying assumption may also be suppressed through representation of the catchment as a series of sub-basins, in conjunction with streamflow routing (Amorocho and Hart, 1964). Certain authors, however, for example, Rogers (1972), have reported that limited spatial variation in rainfall characteristics may occur throughout a storm without seriously affecting surface runoff hydrograph features. In small study areas such as that under consideration here, storm uniformity and stationarity can be expected;

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1 A time-area-concentration curve represents the relationship between the time of flow and area of the catchment 'enclosed' by a time contour (line of equal time of flow to the gauged outlet). With increasing time, the time contour eventually encompasses the whole catchment, that is, when the time of concentration is reached (Nash, 1966).

indeed for this reason the Flood Studies team (N.E.R.C., 1975) limited their application of the unit hydrograph to areas of under 500 km^2. Wilson (1974) and Linsley et al. (1982) cited 5000 km^2 as an arbitrary cut-off point, while Wisler and Brater (1959) quoted 8000 km^2. The requirement for uniform rainfall may also be relaxed when a single catchment is under examination; storms should be merely characteristic of the area since non-uniformity is incorporated into the unit hydrograph (N.E.R.C., 1975).

Net rainfalls of equal duration but different intensities are assumed to produce hydrographs of equal time duration. Hydrograph time base depends on the chosen method of baseflow separation and this introduces the importance of administering a consistent separation technique. Němec (1964) argued that the time base of a hydrograph, especially on the rising limb, is influenced by the volume of net rain, while theoretically, the hydrograph recession limb may be regarded as having an infinite time base (Chow, 1964). Dooge (1959) stated that it is not essential to satisfy this assumption, however, and that it is physically realistic only in catchments of evenly-distributed storage zones.

The principle of superposition, which can be used to alter unit hydrograph duration, asserts that a number of unit hydrographs may be combined or added together, so that, for example, the hydrograph resulting from three separate storms is the sum of the three separate hydrographs (Wilson, 1974). This encompasses a further principle, that of proportionality or linearity of catchment response, which states that direct runoff hydrograph ordinates are directly proportional to rainfall volumes. The unit hydrograph model therefore represents the system linearly, the same unit response resulting from given storm and catchment characteristics. In order to accommodate the linearity assumption, evaluation of the rainfall-runoff relationship demands that rainfall losses and baseflow be eliminated before unit hydrograph derivation (Chiu and Bittler, 1969). Doubt was cast on the assumed linearity of hydrological systems in the 1950s and 1960s (for example, Minshall, 1960; Amorocho, 1963; Machmeier and Larson, 1968). Nash (1958), in showing certain methods of solving the rainfall-runoff relationship to be specific cases of the unit hydrograph, each technique relying on the assumption of linearity, exposed some of the inherent weaknesses of these methods. Non-uniform distributions of rainfall and losses, complex overland flows and spatially-varied unsteady channel flow result in runoff non-linearity, although irregularities can be smoothed out by the basin's integrating effect (Singh, 1964). Storage and outflow factors are also related in a non-linear fashion, involving complexities of friction and velocity gradients (Amorocho, 1961). Calver et al. (1972) summarised possible causes of temporal and spatial non-linearities in rainfall-runoff processes as follows:

i) The presence of saturated throughflow at different depths in the soil profile, resulting from variation in soil properties with depth.

ii) The delay between rainfall occurrence and lateral flow caused by the time needed for saturation levels to approach the surface.

iii) Spatial variations in antecedent moisture content having the effect that only zones of moisture concentration contribute directly to runoff.

In comparison, the results presented by the Flood Studies Report (N.E.R.C., 1975) failed to produce evidence in support of non-linear systems and the variable contributing area concept was proposed as an argument to support linearity, on the premise that as the contributing area expands, average travel time increases. However, as initial source area grows, the proportion of rainfall contributing to storm runoff does not remain constant, but increases with the size of the storm (Weyman, 1975).

Simple linear relationships ignore the fact that surface runoff velocity varies with discharge, since greater volumes may travel at higher velocities than lesser volumes (Kirkby and Chorley, 1967; N.E.R.C., 1975), although these arguments are less applicable where subsurface flow dominates. Rastogi and Jones (1971), in using a time-invariant, non-linear mathematical model, found rainfall excess intensity to be important in influencing lag time, time to peak and peak flow rates by its effect on flow velocity, although the authors acknowledged the fact that field verification of their model was not possible due to the absence of suitable data and that it should be used in a theoretical context only. Boyd et al. (1979) similarly observed that lag time does not necessarily remain constant for a catchment, varying in relation to flood magnitude.

The linearity principle is based on the need for simplicity, however, and although the complexity of the runoff generation process may militate against a simple linear system, the assumption represents an approximation which is usually found acceptable. This is particularly so for small and medium-sized catchments, if storm selection is executed with care, since unit hydrographs from different storms show least similarity on larger areas (Henderson, 1963; Francis, 1973). The importance of satisfying the linearity assumption may also vary with runoff characteristics. Difficulties of incorporating a linear operation are manifested for example during 'catastrophic' floods when overland flow is evident, and problems are aggravated in hilly areas with large acquifer outcrops where lag time decreases rapidly as rainfall amount increases (N.E.R.C., 1975). Kirkby and Chorley (1967) maintained that non-linearities are significant only for the hydrograph recession, since this is more sensitive than the rising limb or peak (especially) to variations in contributory slope hydrographs. Furthermore, over moderately high flow conditions, stream velocity is generally constant, itself a requirement for linearity. In using the unit hydrograph procedure to testify the validity of runoff routing, Laurenson (1964) discovered a high degree of success with the unit hydrograph approach in a catchment characterised by non-linear catchment storage effects. In general, although great care must be exercised in interpreting results based on an assumption of linearity, the degree of approximation caused by the concept depends on the extent of the system's non-linearity (Amorocho and Hart, 1964). The latter authors concluded that the unit hydrograph method tends generally to underestimate large floods and overestimate small events. The arguments surrounding linearity were summarised by Freeze (1972a) who questioned the physical explanations for any consistent response, linear or otherwise.

The unit hydrograph theory also assumes a time-invariant system, maintaining that the direct runoff hydrograph from a specific basin remains the same for a given pattern of effective rainfall, irrespective of recurrence times. No temporal change is considered in the hydrological system relative to its past behaviour. This assumption may be valid over geological time scales, but response changes are to be expected on a monthly, storm by storm, or even daily basis as a result of, for example, changes in urban cover, or the introduction of a reservoir. Several authors have therefore considered alternative, time-varying and/or non-linear models (for example, Prasad, 1967; Chiu and Bittler, 1969; Chiu and Huang, 1970; Datta and Lettenmaier, 1985).

5.3.4.5 Unit Hydrograph Derivation

A number of different approaches, of varying complexity, have been adopted for formulating the unit hydrograph. Wilson (1974), for example, described a limited technique involving division of the runoff hydrograph ordinates by net rainfall, but this method is unsuitable for application to the complex type of storms found in Great Britain, where multi-peaked hydrographs occur regularly (N.E.R.C., 1975). Three fundamental modes of approach are available for derivation of the unit hydrograph for these more typical situations:

i) Trial and error, or iterative methods.

ii) Direct analytical methods.

iii) Methods based on knowledge of the functional form of the unit hydrograph.

The least accurate is the first approach. Trial and error methods include that described by Linsley et al. (1949), involving calculation of the unit hydrograph ordinates by successive solution of a series of equations for each ordinate. Collins (1939) also developed a simple iterative method, incorporating the use of distribution percentages for the drainage area, defined as proportions of total flow occurring in unit periods. This distribution graph is applied to runoff for all periods of rain except the largest. Resulting discharges are subtracted from the actual discharge hydrograph and the residual should represent the distribution of discharge from the largest runoff amount. Further trials are conducted until the residual graph is in close correspondence with the assumed distribution graph.

In the Flood Studies Report (N.E.R.C., 1975) two direct analytical methods were compared with Nash's (1960) method, the latter using moments of the instantaneous unit hydrograph as parameters to express the relationship between net rainfall and storm runoff. Hydrograph analysis by matrix inversion, developed by Snyder (1955) was, with the addition of 'smoothing', found preferable to the harmonic analysis method as produced by O'Donnell (1966) and modified by Hall (1977a). The harmonic analysis approach involves definition of rainfall excess volumes as well as surface runoff hydrograph and unit hydrograph ordinates in terms of harmonic series. In comparing instantaneous unit hydrograph results from harmonic analysis with those calculated from an exponential equation given by Nash (1957), in which catchment response is represented as a series of reservoirs, O'Donnell (1960) commended the former approach. The technique is hindered by the fact that it may lead to oscillations in the resulting unit hydrographs, however, although a method of controlling this, involving truncation of the number of terms in the harmonic series, has been discussed by Hall (1977b).

Matrix inversion with smoothing is adopted for unit hydrograph derivation in the present study and procedures are based largely on those given by the Flood Studies Report, assumptions and requirements of the model being met as closely as possible. Site characteristics at Egton match catchment recommendations as proposed in the report, although deviation from these criteria may be permissible in certain instances. The following features were proposed:

i) Catchment area less than 500 km^2.

ii) Reliable gauging station rating curve.

iii) Presence of at least one autographic raingauge.

iv) Catchment displaying some evidence of short-term runoff response.

Hydrograph derivation for Egton is discussed in detail in the following sections.

a) Storm Selection

Where possible isolated rainfall and runoff events are selected for analysis, the most suitable discharge hydrographs having a well-defined peak, a smooth rising curve and an uninterrupted recession limb. Storms should be separated by a sufficient period of time such that they may be identified as individual events (Hall, 1977a; Wheater et al., 1978, 1982). Should only complex hydrographs be available, these may be separated into several simple hydrographs (Chow, 1964). Selection of short-duration storms renders rainfall intensity variations of less importance, while large floods yield unit hydrographs which are representative of flood conditions for the area under investigation (Nash, 1966). The total number of storms is limited in the present study, since all those selected for analysis were chosen to meet the recommendations outlined above, resulting in a total of thirteen events. Table 5.1 summarises the characteristics of the accepted storms, for which small to moderate rainfall catches apply, with generally low intensities (less than 4 mm h^{-1}).

b) Data Collation

Site records of rainfall and stream stage are used to abstract data for each storm and its preceding period. Rainfall and stage values are read directly from chart records, following corrections for timing and other errors. Stream stage is converted to discharge by means of the standard rating curve appropriate for the type of V-notch weir used in this study and as given by the British Standards

STORM EVENT NUMBER	STORM DATE	GROSS RAINFALL (mm)	$\log_{10}$	RAINFALL DURATION (h)	$\log_{10}$	AVERAGE RAINFALL INTENSITY (mm h^{-1})	$\log_{10}$	SMD (mm)	$\log_{10}$ (1+SMD/100)
PRE-BURN:									
1	7.8.80 - 9.8.80	14.4	(1.158)	11.0	(1.041)	1.31	(0.117)	7.5	(0.031)
2	27.7.80	10.4	(1.017)	3.5	(0.544)	2.97	(0.473)	26.0	(0.1)
3	30.7.80 - 31.7.80	10.1	(1.004)	4.0	(0.602)	2.53	(0.403)	20.5	(0.081)
4	1.8.80 - 2.8.80	15.2	(1.182)	5.0	(0.699)	3.04	(0.483)	13.6	(0.055)
5	14.11.80 - 16.11.80	12.8	(1.107)	11.0	(1.041)	1.16	(0.065)	0.0	(0.0)
6	17.11.80 - 19.11.80	8.6	(0.935)	4.5	(0.653)	1.91	(0.281)	0.0	0.0)
7	25.11.80 - 26.11.80	12.7	(1.104)	5.5	(0.74)	2.31	(0.364)	0.0	0.0)
8	2.12.80 - 5.12.80	13.5	(1.13)	30.0	(1.477)	0.45	(-0.347)	0.0	(0.0)
POST-BURN:									
1A	11.9.81	8.1	(0.909)	4.0	(0.602)	2.03	(0.308)	58.1	(0.199)
2A	26.9.81 - 27.9.81	19.4	(1.288)	5.5	(0.74)	3.53	(0.548)	15.0	(0.061)
3A	28.10.81 - 31.10.81	5.6	(0.748)	3.5	(0.544)	1.6	(0.204)	0.0	(0.0)
4A	18.11.81 - 20.11.81	8.5	(0.929)	11.0	(1.041)	0.77	(-0.114)	0.0	(0.0)
5A	4.12.81 - 6.12.81	10.0	(1.0)	9.0	(0.954)	1.11	(0.045)	0.0	(0.0)

Table 5.1 Storm and Catchment Characteristics of Events Selected for Unit Hydrograph Analysis

Institution (1981) (Equation 3.4).

Net rainfall calculation requires information on soil moisture deficits, for which spatially averaged values are calculated from previously derived estimates (Chapter 4). Amendments are made to deficit figures for intermediate evaporation and rainfall between the time of soil moisture measurement and that of storm occurrence. Deficiencies in the procedure are immediate, but this approach was deemed the most suitable in the absence of detailed information on infiltration and redistribution rates.

c) Baseflow Separation

Following visual checks on the rainfall-runoff relationships of the selected storm events, hydrograph analysis proper was initiated. Unit hydrograph analysis confines itself to a simple division between baseflow and quick response runoff for each flood hydrograph. Nash (1958) noted the difficulty with which both hydrograph and rainfall separation are carried out for natural as opposed to urban catchments. Numerous different techniques are available by which baseflow separation may be achieved and, according to Nash (1960), almost all lack physical justification and are completely arbitrary. Nash failed to develop a non-arbitrary system but recommended that at least a consistent approach be maintained.

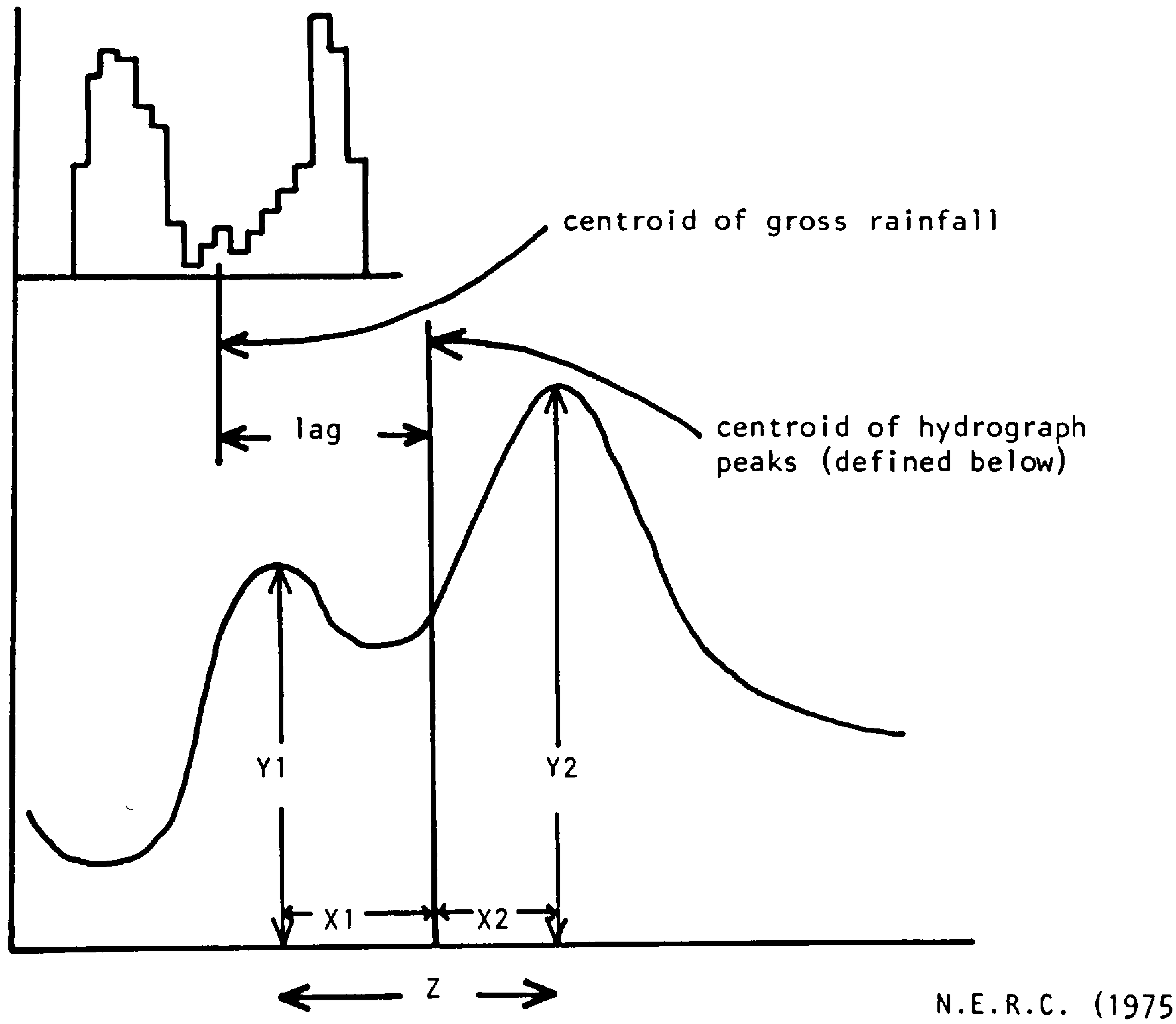
Wheater et al. (1978, 1982) however, found no great change in hydrograph shape with small changes in location of the baseflow separation point on the recession limb, and the method selected becomes less important as flood response increases in size in relation to preceeding flow (N.E.R.C., 1975). Chow (1964) agreed that baseflow separation procedures are arbitrary and, because baseflow generally constitutes only a small percentage of most peak flows, resulting errors are small and therefore simple methods of straight line separation can be used. The inherent difficulty of spatially and temporally defining separate runoff components was used by Hewlett and Hibbert (1967) to explain the subjectivity of hydrograph separation techniques. After examining records from small forested watersheds in the Appalachian-Piedmont region, the authors developed a simple technique separating quickflow, discharge running rapidly from the watershed, from delayed flow. A line is extended from the start of hydrograph rise to the recession limb with a constant slope of 0.033 $m^3\ min^{-1}\ km^{-2}\ h^{-1}$.

The master depletion curve method of baseflow separation requires plotting of recession limbs (log discharge) against time, and the fitting of a straight line to the lower parts of the curves. The point of deviation of the hydrograph from this tangent marks the end of surface runoff, and baseflow separation is then represented as a straight line joining this point to the beginning of hydrograph rise. This method is most suitable for use with continuous discharge records of a few years' duration. Alternatively, the baseflow separation point is identifiable by locating the point of greatest curvature on the recession limb, using ratios of discharge at a point to that at a fixed interval later. This makes fewer demands on the data base than the foregoing technique (Wilson, 1974).

The method adopted in this study is that described by the Flood Studies Report and is similar to that used by Nash (1960). It centres around the concept of lag time, in this context the period required for the effect of rainfall to reach the catchment outlet, and defined specifically as the time from the centroid of gross rainfall to the peak flow or centroid of hydrograph peaks, as explained diagrammatically for a double-peaked hydrograph in Figure 5.9. The point at which hydrograph flow begins to increase ('X', Fig.5.10) is projected to meet point 'A', which represents the centroid of peaks. A sketched extension of the previous recession limb suffices for this purpose, particularly if flow is low or constant. Point 'A' is then joined by a straight line to 'B', the point where the time from end of rainfall equals four times the lag period. Total response runoff is that represented by the area under the hydrograph curve and above the baseflow separation line. Discharge is converted to equivalent depth in millimetres using Equation 5.5. Poor prediction of total response runoff may result from errors in the rating curve, in assessment of catchment area or from channel storage effects. On one occasion a small negative lag resulted in the corresponding event being rejected from the analysis (Fig.5.11). Rain storms ending after the flood hydrograph peak in this way may be storms of diminishing intensity (Chow, 1964).

d) Rainfall Separation

Having defined each response runoff hydrograph it is necessary to derive a series of net rainfall increments for each storm event, in order that these two data sets can be presented to a matrix inversion routine. Again, several methods are available to separate net or effective rainfall from that which is 'lost', although the method chosen becomes less important with storms of almost steady rainfall (N.E.R.C., 1975) and the use of different loss methods may result in only small differences in unit hydrograph shape (Wheater *et al.*,1978). A simple horizontal line drawn across the hydrograph was used by Chow (1964) such that the area above the line represents the volume of direct runoff and that below, losses. Four separation techniques, shown in Figure 5.12, were compared in the Gloucester



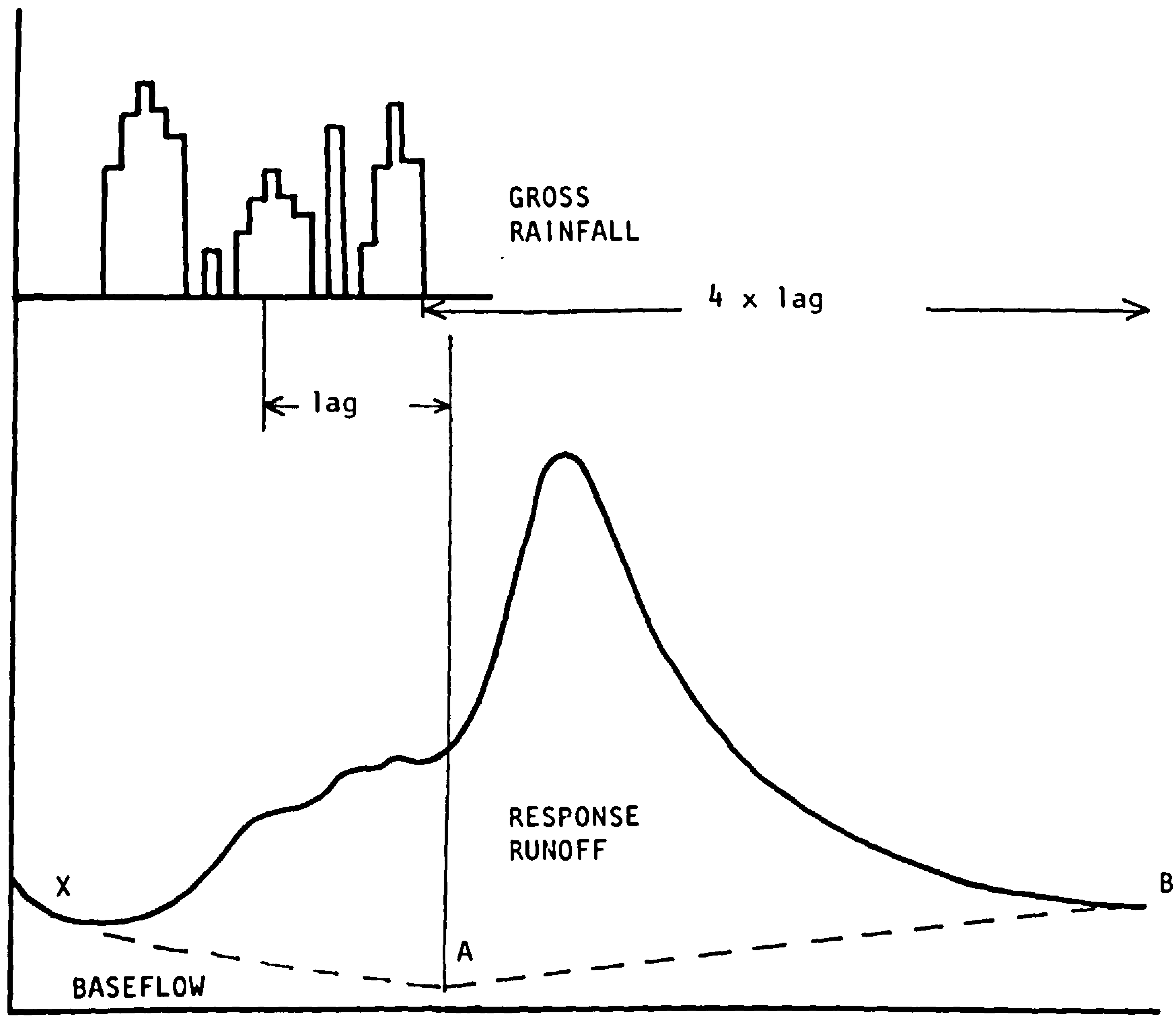
centroid of hydrograph peaks defined by

ratio X1:X2 :-

$$X1 = \frac{Z}{(Y1+Y2)} \times Y1$$

$$X2 = \frac{Z}{(Y1+Y2)} \times Y2$$

<u>Figure 5.9 Definition of Centroid of Hydrograph Peaks and Lag Time for Baseflow Separation</u>



N.E.R.C. (1975)

Figure 5.10 Hydrograph Baseflow Separation

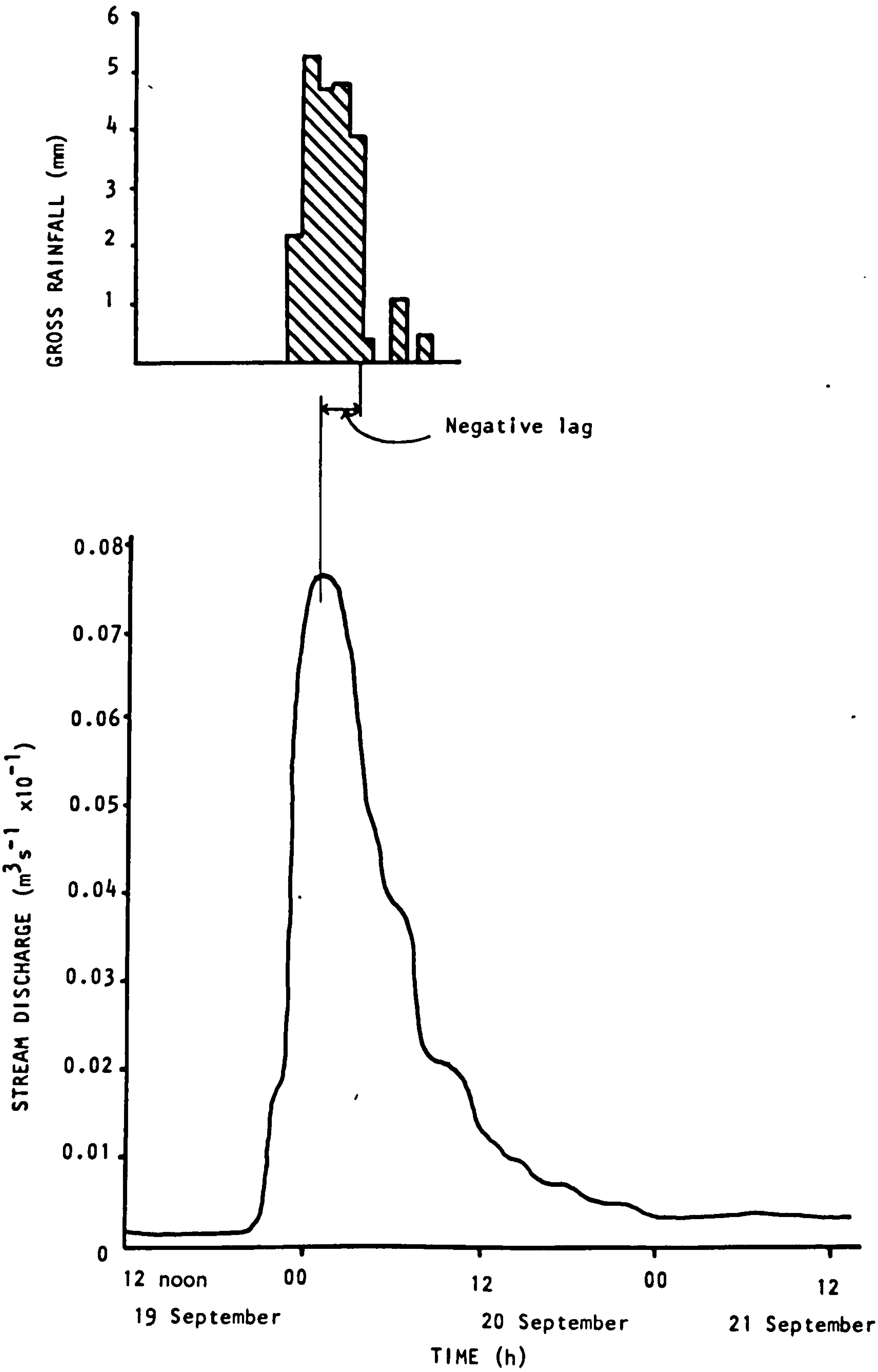


Figure 5.11 Negative Lag Time Derived for a Post-burn Storm Event

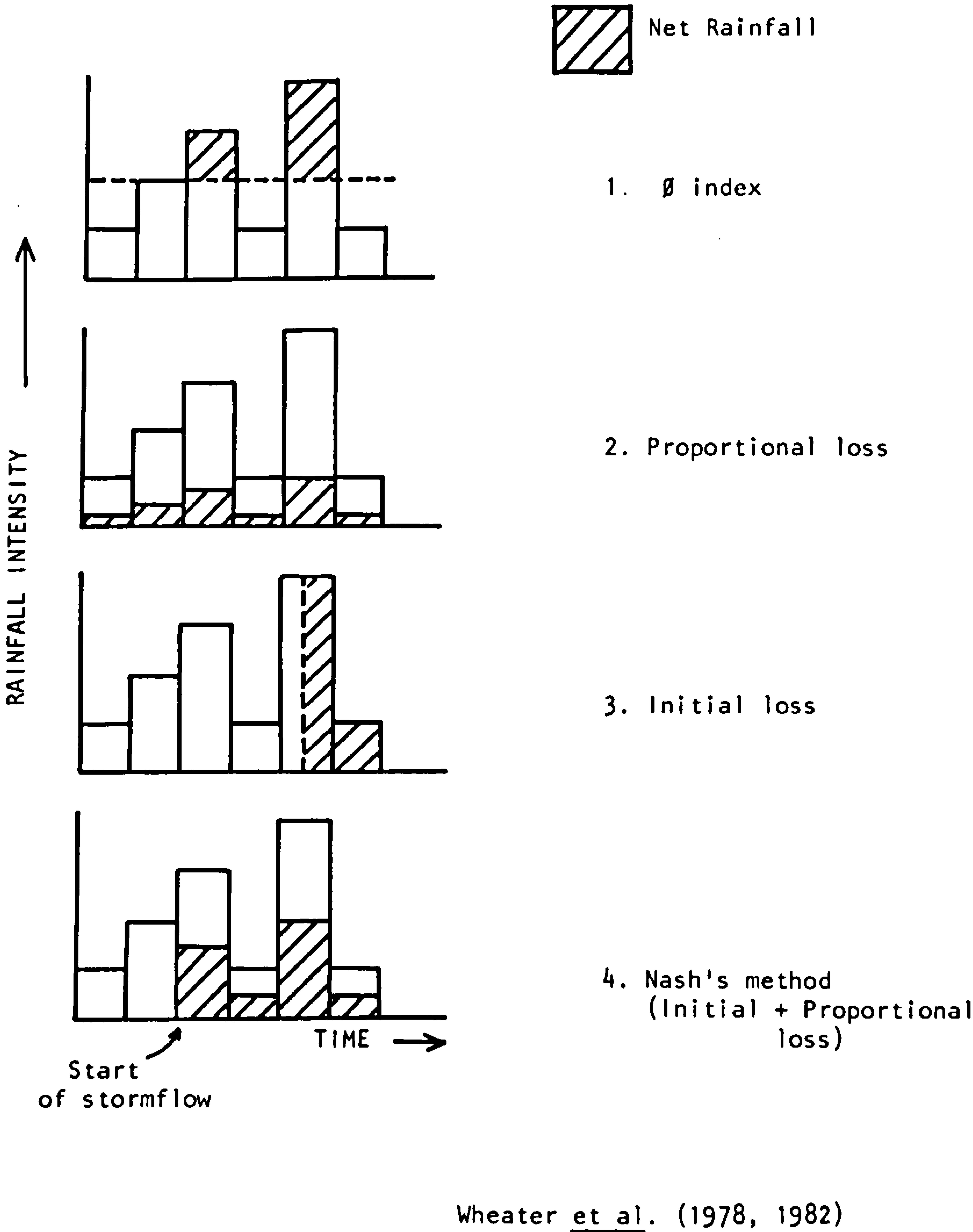


Figure 5.12 Methods of Defining Net Rainfall

study by Wheater et al. (1978, 1982):

i) Ø index - constant loss rate.

ii) Proportional loss - constant proportion of total rainfall lost.

iii) Initial loss - all losses deducted at the start of rainfall.

iv) Nash's method - all rainfall before stormflow lost (initial loss), followed by losses distributed proportionally or using the Ø index.

A constant loss rate may be defined from those rainfall events which fail to generate storm runoff (Dickinson and Whiteley, 1973). It may be found, however, that such 'uniform' indices vary according to factors such as antecedent conditions, and the time interval used in the rainfall hyetogram (Linsley, 1967). This type of index is also insensitive to any significant spatial variability in runoff generation (Clark, 1980). The proportional loss method was selected by Wheater et al. since it avoids the timing difficulties of the first and third methods of allocating losses and yielded fewer unrealistic hydrographs. Volume of initial rainfall was found difficult to predict while Nash's method incorporates, in addition, an extra parameter.

The Flood Studies team recommended a percentage-based method as being most appropriate and computations for this technique are summarised by the example shown in Table 5.2. In detail, the method involves firstly dividing the storm into a series of time intervals; recommended time periods as presented by Sherman (1942) are shown in Table 5.3. In the present study, an interval ('T') of 0.5 h is chosen to define most storms, although a few are adequately described by setting T equal to 1 h. This complies with N.E.R.C.'s recommendation

Time Interval (h)	Total Rain (mm)	Soil Moisture Deficit (mm)	A.P.I.5	C.W.I.	Rain x C.W.I.	% Runoff	Net Rain (mm)
0	0.3	7.5	3.3	120.8	36.2	16.7	0.05
1	0.3	7.2	3.6	121.4	36.4	16.7	0.05
2	1.5	6.9	3.9	122.0	183.0	13.3	0.2
3	2.3	5.4	5.4	125.0	287.5	17.4	0.4
4	2.7	3.1	7.7	129.6	349.9	18.5	0.5
5	1.7	0.4	10.4	135.0	229.5	17.7	0.3
6	1.3	0	12.1	137.1	178.2	15.4	0.2
7	1.3	0	13.4	138.4	179.9	15.4	0.2
8	0.3	0	14.7	139.7	41.9	16.7	0.05
9	1.5	0	15.0	140.0	210.0	20.0	0.3
10	1.2	0	16.5	141.5	169.8	16.7	0.2
	Σ 14.4				Σ1902.3		

$$F = 2.5/1902.3 = 1.314 \times 10^{-3}$$

Table 5.2 Computation of Net Rainfall Increments for a Summer Storm, prior to Vegetation Removal (Event 1)

Catchment Area (sq mi)	Time Unit (h)
>1000	12
100 - 1000	6,8 or 12
20	2
<20	∼1/3 or 1/4 of concentration time of basin

<u>Table 5.3 Recommended Unit Hydrograph Time Periods</u>

(Sherman, 1942)

of assigning T to yield at least five ordinates on the hydrograph rising limb. Total rain is then determined for each interval, and any unrelated rainfall occurring before or after the main event is excluded. The 'total profile' soil moisture deficit is derived for the first time interval from the most recent soil moisture measurements, any intermediate rain being subtracted from the deficit and evaporation added. Deficits for all subsequent time intervals are determined similarly by subtracting the rain for the preceding time interval. The value is reset to zero once a negative estimate is obtained.

An antecedent precipitation index (A.P.I.) is then calculated for each time interval. Several A.P.I. equations are available for this purpose, each based on the idea of a decreasing antecedent rainfall effect over time. The index for the first time interval is determined here from the following equation:

$$A.P.I.5_d = 0.5^{\frac{1}{4}} \; (P_{d-1} + 0.5P_{d-2} + (0.5)^2 P_{d-3} + (0.5)^3 P_{d-4} + (0.5)^4 P_{d-5}) \qquad \text{Eq. 5.8}$$

where:

$A.P.I.5_d$ = antecedent precipitation index for 5 days preceding the storm

$P_{d-1,2}\ldots$ = total rainfall for 1st, 2nd ... day preceding the storm

Succeeding A.P.I.'s are then calculated as follows:

$$A.P.I.5_t = P_{t-1} \times 0.5^{T/48} + A.P.I.5_{t-1} \times 0.5^{T/24} \qquad \text{Eq. 5.9}$$

where:

P_{t-1} = rainfall during preceding time interval

T = unit hydrograph time base, for example, 1 h

$A.P.I.5_{t-1}$ = A.P.I. for preceding time interval

$0.5^{T/24}$ = decay constant; replaced by 1 where T $\leqslant$ 1 h

Each rainfall value is then multiplied by a corresponding catchment wetness index (C.W.I.) derived from the expression:

$$C.W.I. = 125 + A.P.I.5 - \text{soil moisture deficit} \qquad \text{Eq. 5.10}$$

Total response runoff (mm) divided by the sum of these values yields a factor, F, which converts rainfall to runoff on the basis of catchment wetness for each time interval. The product (F x C.W.I. x Total Rain) yields a series of net rain increments, each expressed as a percentage of the total rain for the interval and normally reduced to one decimal place in the last column of the table. The quantities of rainfall excess amount (net rain) and response runoff should be equal and either may need slight adjustment to satisfy this requirement.

c) Hydrograph Computation

Definition of each unit hydrograph requires response runoff hydrograph ordinates and the net rain sequence for each event to be presented as data to a matrix inversion routine. The general aim is to solve for $|u|$ in the following equation:

$$|p| \cdot |u| = |q| \qquad \text{Eq. 5.11}$$

where:

$|\ |$ represents a matrix of values

p = net rainfall

u = unit hydrograph ordinates

q = response runoff ordinates

The response function, $|u|$ is a one column vector comprising N ordinates, where

N = m-n+1

m = number of response runoff ordinates

n = number of net rainfall ordinates.

Multiply through by the transpose matrix of p, $|p^T|$:

$$|p^T| \cdot |p| \cdot |u| = |p^T| \cdot |q| \qquad \text{Eq. 5.12}$$

and divide both sides by the product $\{|p^T| \cdot |p|\}$:

$$|u| = \{|p^T| \cdot |p|\}^{-1} \cdot |p^T| \cdot |q| \qquad \text{Eq. 5.13}$$

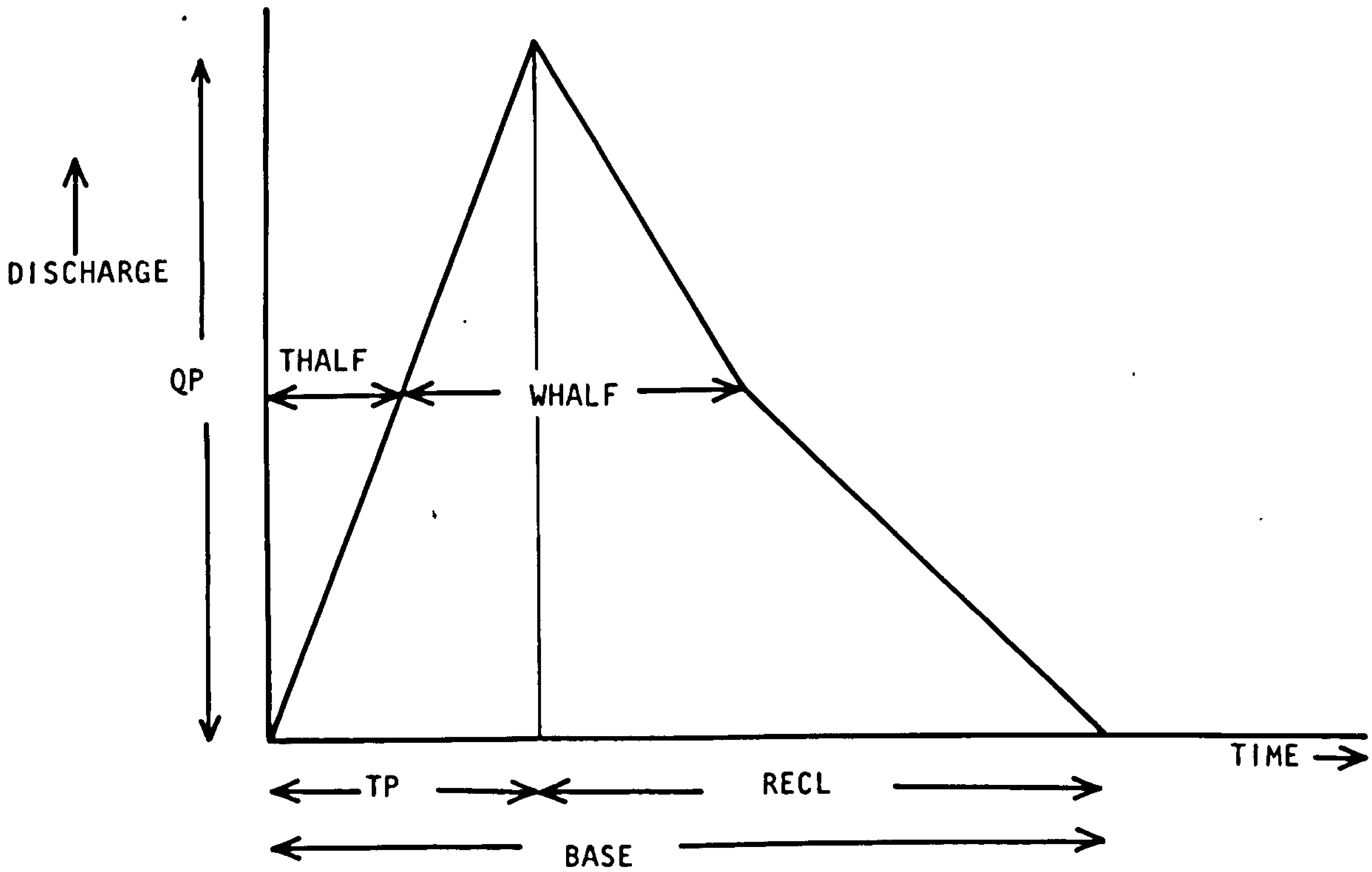
where $\{|p^T| \cdot |p|\}^{-1}$ is the inverse matrix of $\{|p^T| \cdot |p|\}$.

This is accomplished by firstly, constructing the rainfall matrix, $|p|$ by filling out the column of net rain increments with zeros, as indicated in Appendix IV for the example shown in Table 5.2 (p.251). This matrix is then made square by pre-multiplying by its transpose, $|p^T|$ formed by interchanging matrix rows and columns. The result is inverted and used as a multiplier on the one-column matrix product $|p^T| \cdot |q|$ (Eq. 5.13). The resulting values, $|u|$ comprise the ordinates of the 'least squares' unit hydrograph, for which the sum of squares of differences between the observed hydrograph ordinates and those of the reconstituted hydrograph is at a minimum. Matrix manipulations were performed in the present case through computer library subroutines incorporated into purpose-written programs.

An oscillating, unstable unit hydrograph can arise particularly when the number of unit hydrograph ordinates is large, from a lack of restriction on the relative values of the unit hydrograph ordinates (Boorman and Reed, 1981). To reduce instability, moving average smoothing was employed. This entails calculation of the average of a point and its two neighbours. The averaging procedure is carried out twice in succession and final average (smoothed) points are re-plotted. Smoothing facilitates objective identification of the unit hydrograph peak, and providing the data interval is sufficient to describe the hydrograph, undesirable shape dampening should be kept minimal (N.E.R.C., 1975). Occasionally, further manual smoothing is required to remove 'noise' in the unit hydrograph, in which case the area beneath the unit hydrograph should not be altered.

Unit hydrograph ordinates may be expressed on the basis of a standard area, although since the present study is confined to examination of a single catchment, values are maintained in their original units, and flow rate is simply scaled as $m^3\ s^{-1}\ mm^{-1}$. Diskin (1979) specifically drew attention to expression of unit hydrograph ordinates, nominating three different types of units, and recommending expression in terms of discharge per unit depth or discharge per unit area per unit depth, in accordance with runoff measurement as discharge or discharge per unit area, respectively. Hall (1981) derived a dimensionless unit hydrograph, expressing time as a proportion of lag time and multiplying the ordinate axis by lag.

Finally, the smoothed curves of the hydrograph are represented by straight lines to facilitate shape description. These are fitted by eye to the rising and upper half of the recession limb, the intersection of these two lines forming the approximating hydrograph peak. A third line is fitted to the lower half of the recession, meeting the first recession line at a flow equivalent to half the peak. This gives a more accurate hydrograph definition than would a simple triangular shape, although in a few cases the latter was found to describe the hydrograph adequately. The procedure is intended to produce a 'best fit' straight line approximation, providing good overall representation of the observed unit hydrograph, and inexact specification of the observed peak is therefore acceptable (N.E.R.C., 1975). Figures 5.13 to 5.21 illustrate unit hydrograph derivation procedures for pre- and post-burn events,



QP = Peak discharge

THALF = time to half-peak discharge

WHALF = width at half-peak

BASE = base width

TP = time to peak

RECL = recession curve length

LQP = $\log_{10}$ QP, LTHALF = $\log_{10}$THALF, etc.

Figure 5.13 Unit Hydrograph Straight Line Approximation and Parameter Definition

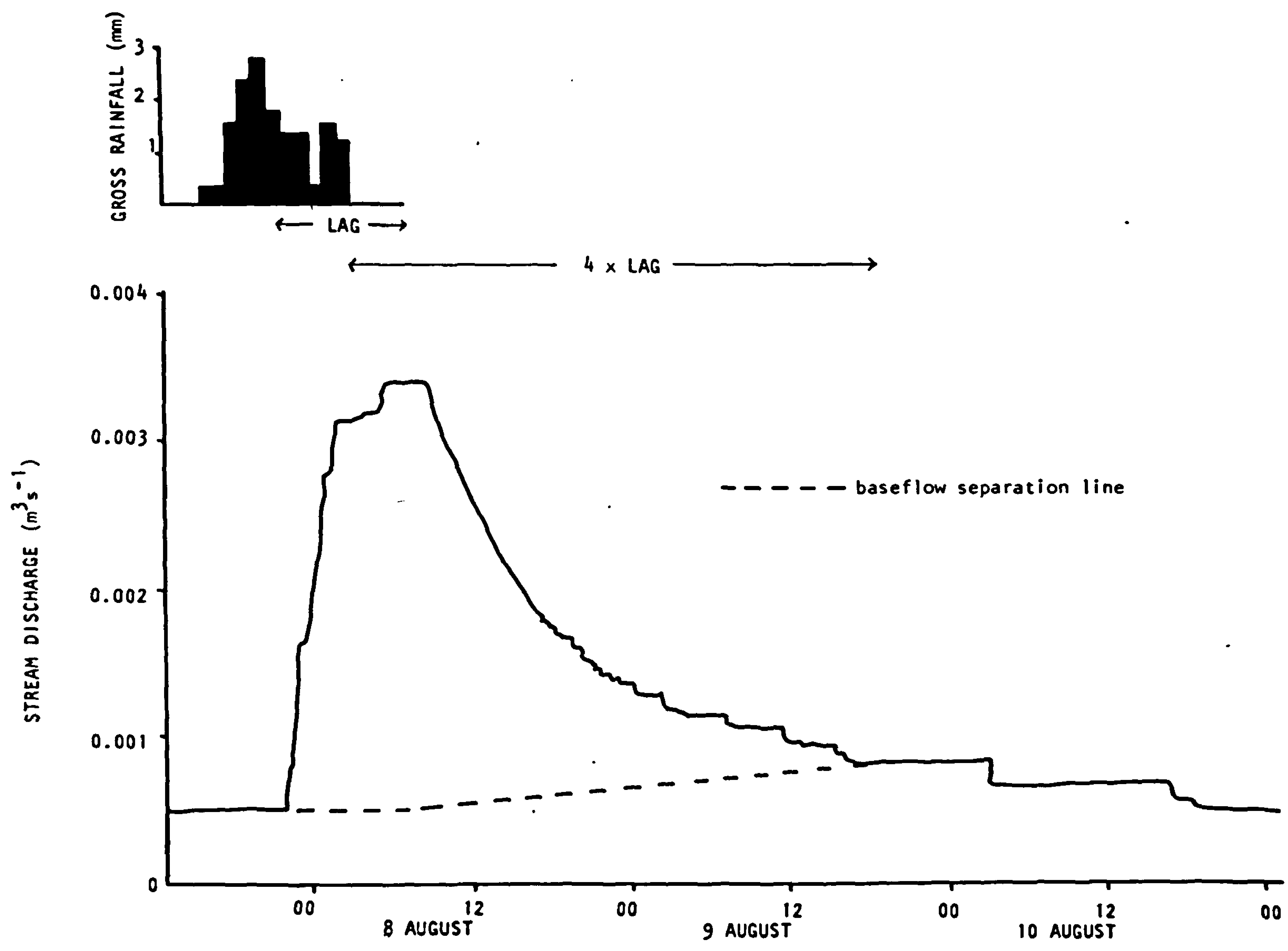
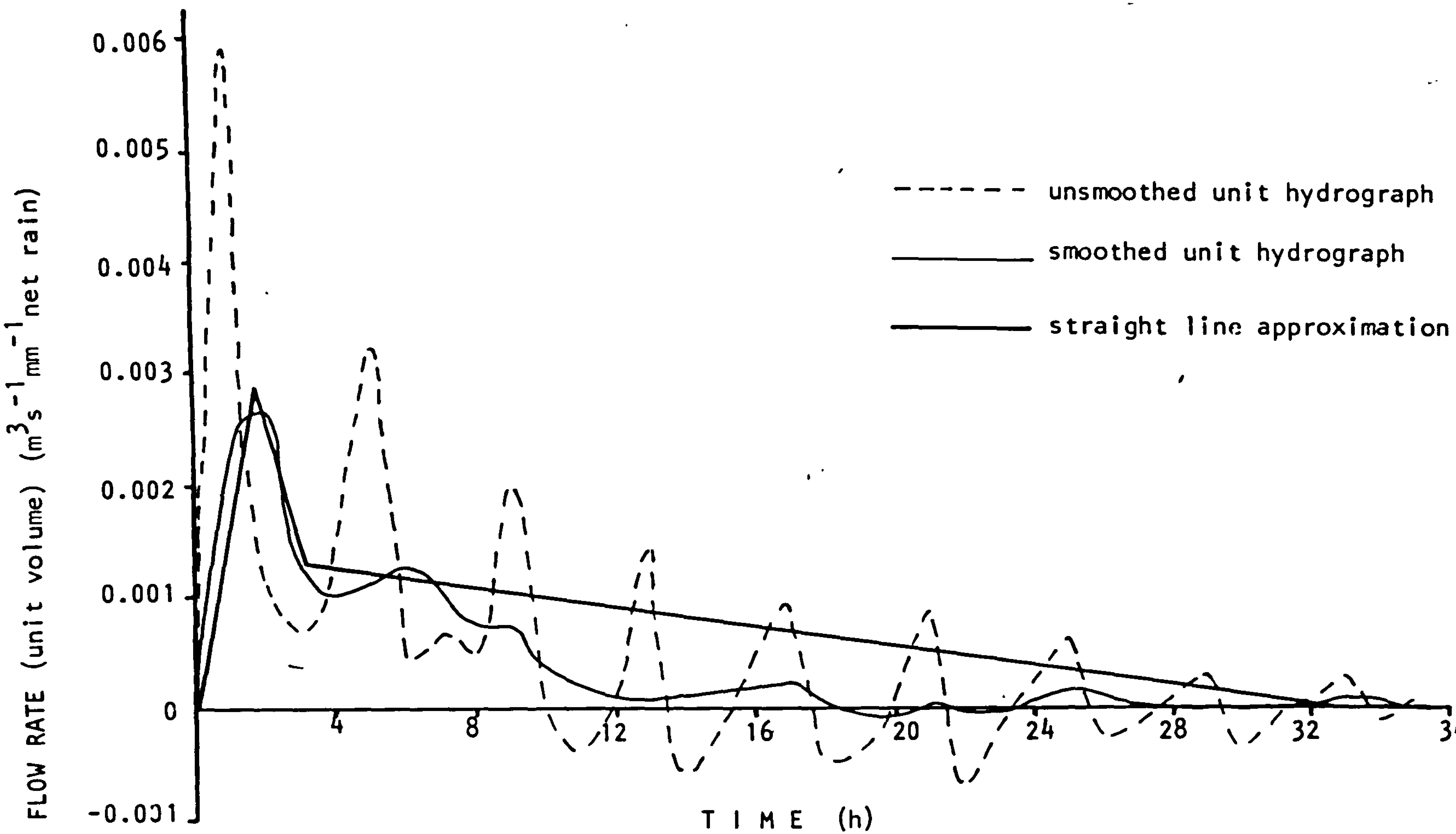


Figure 5.14 Runoff Hydrograph and attendant Rainfall for Event 1 (Pre-Burn)



DIMENSIONS: T= 1h

QP= 0.00281 $m^3s^{-1}mm^{-1}$
THALF= 0.8h
WHALF= 2.2h
BASE= 32.4h
TP= 1.7h
RECL= 30.7h

Figure 5.15 Least Squares Unit Hydrograph for Event 1, showing smoothing and Straight Line Dimensional Parameters

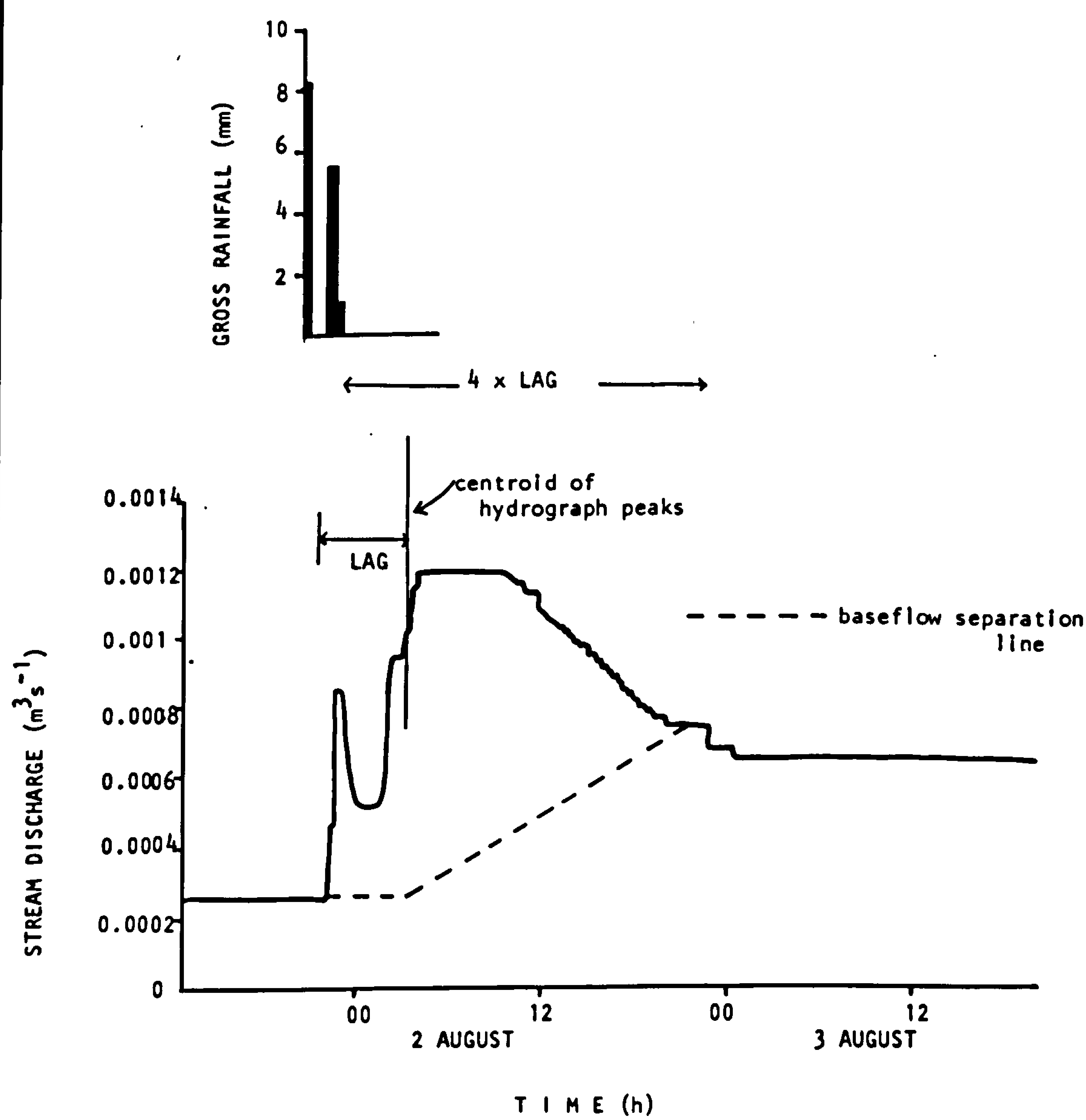


Figure 5.16 Runoff Hydrograph and attendant Rainfall for Event 4 (Pre-Burn)

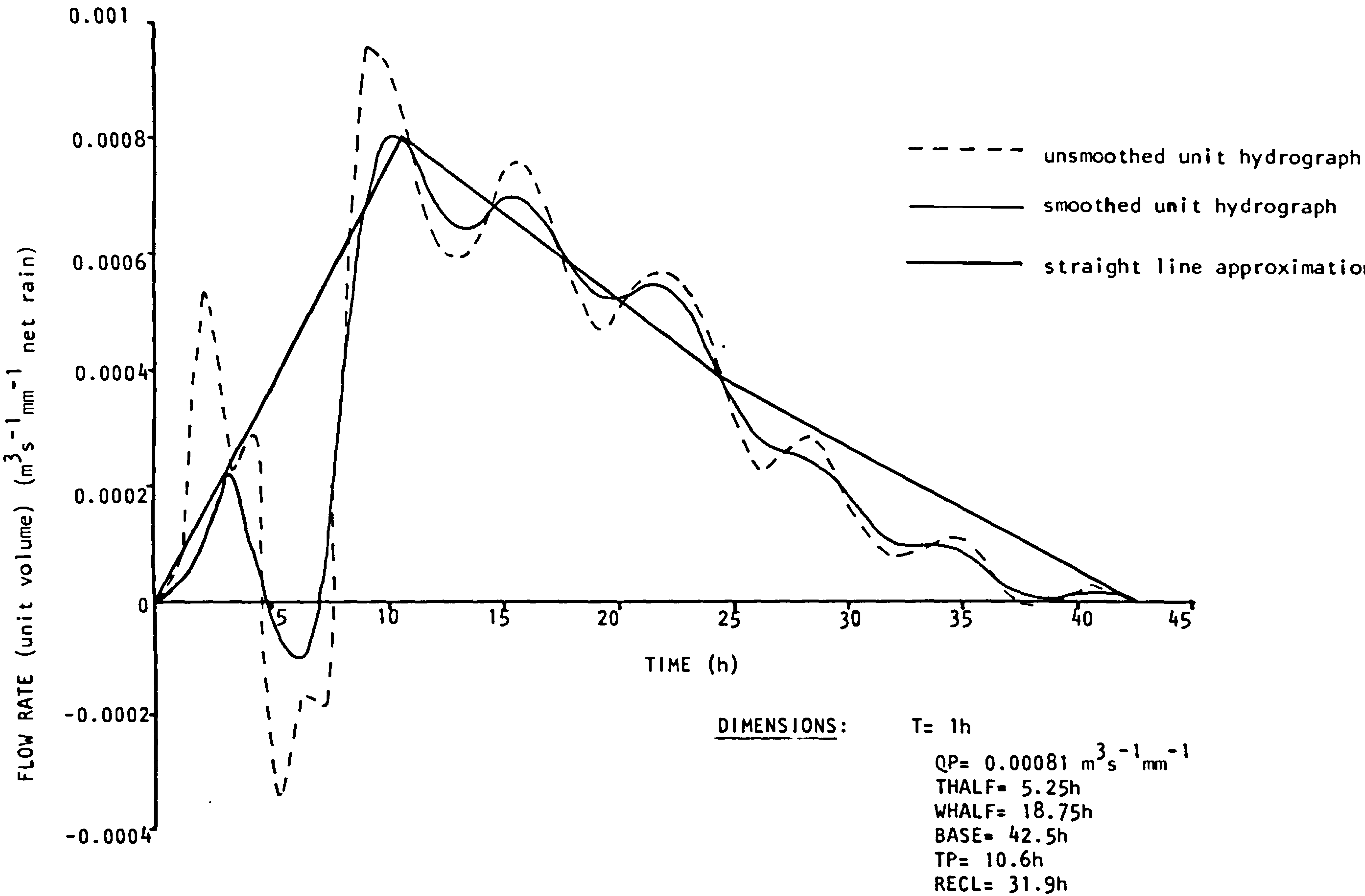


Figure 5.17 Least Squares Unit Hydrograph for Event 4

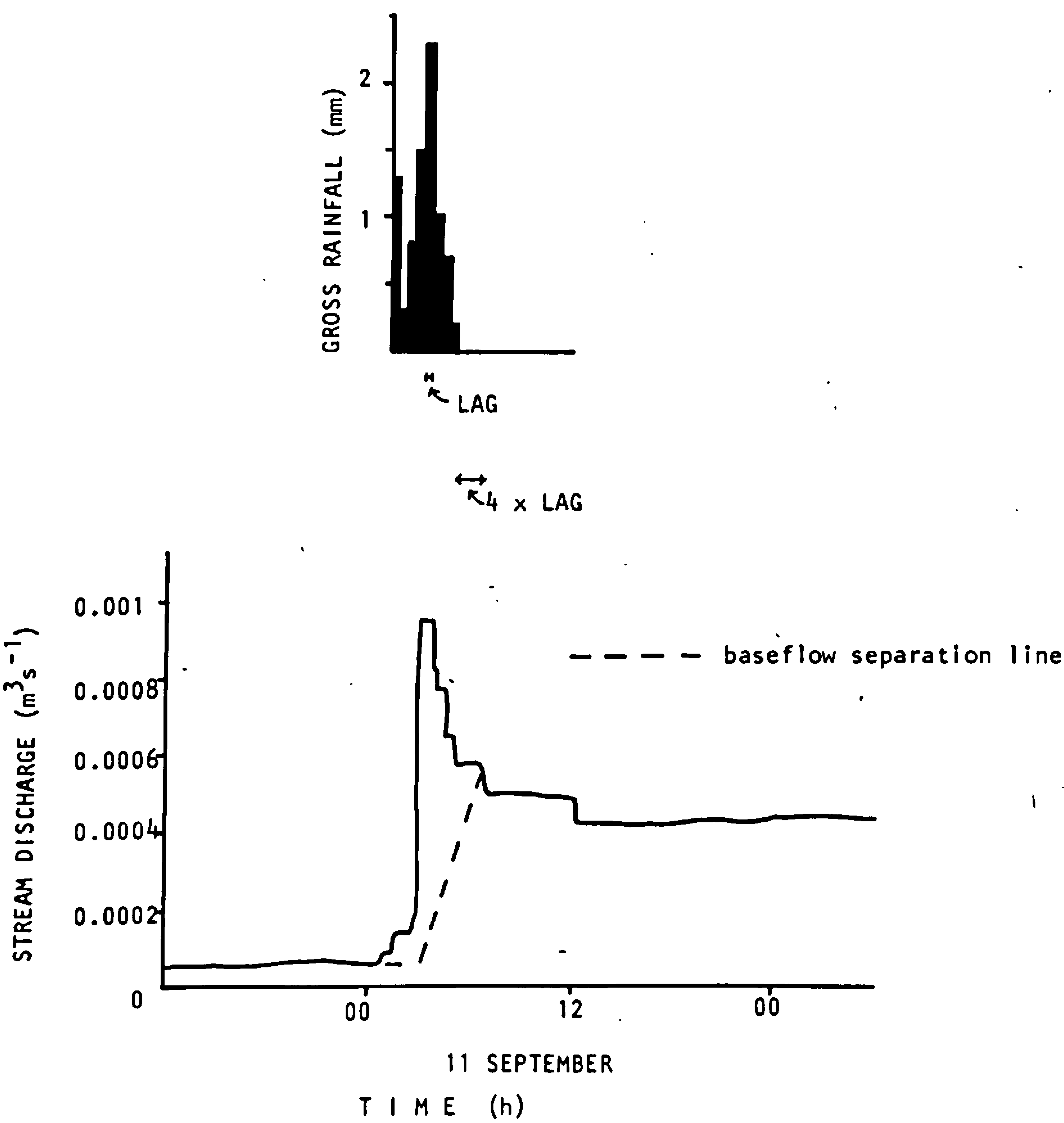
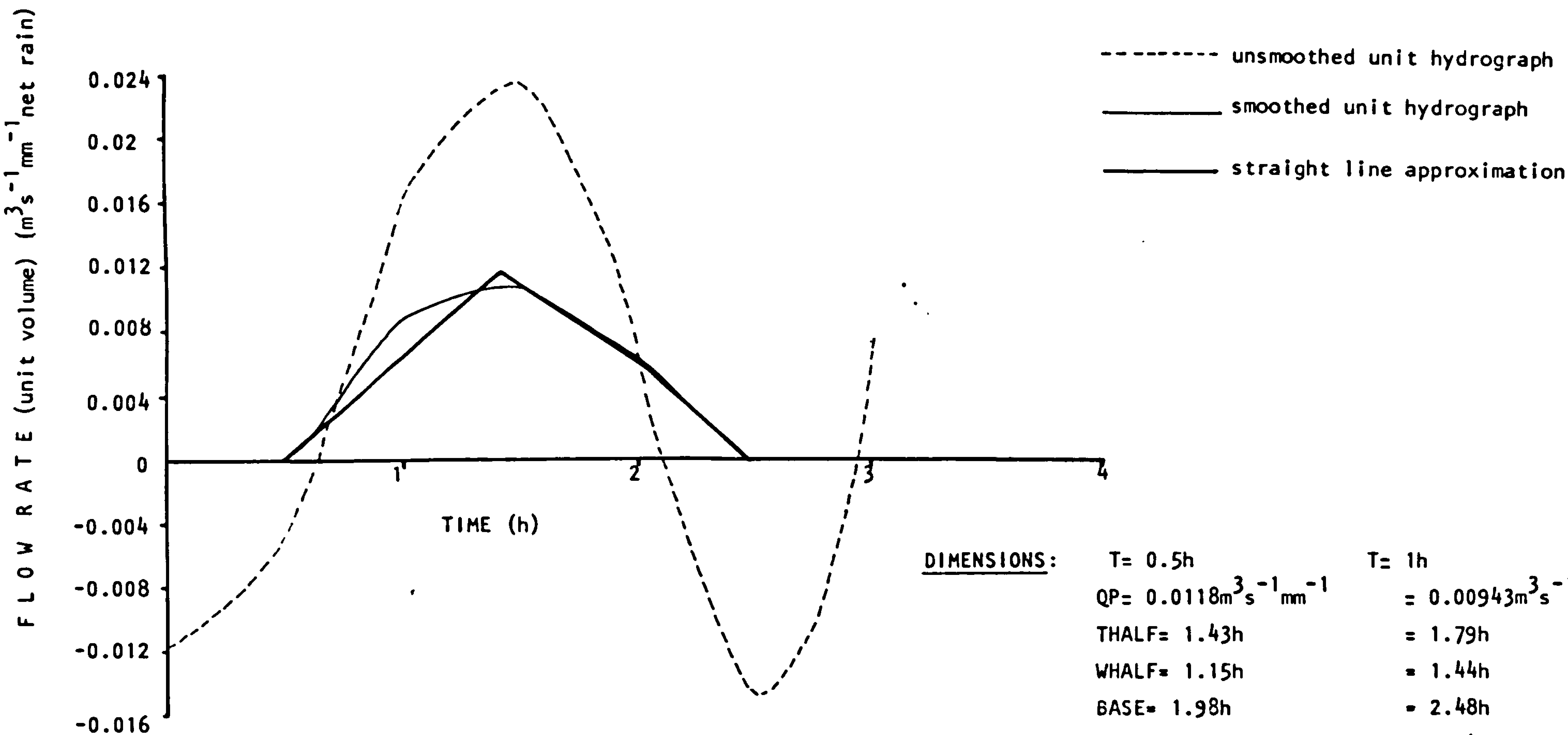


Figure 5.18 Runoff Hydrograph and attendant Rainfall for Event 1A (Post-Burn)



DIMENSIONS:	T= 0.5h	T= 1h
	QP= $0.0118m^3s^{-1}mm^{-1}$	= $0.00943m^3s^{-1}mm^{-1}$
	THALF= 1.43h	= 1.79h
	WHALF= 1.15h	= 1.44h
	BASE= 1.98h	= 2.48h
	TP= 0.91h	= 1.14h
	RECL= 1.07h	= 1.34h

Figure 5.19 Least Squares Unit Hydrograph for Event 1A

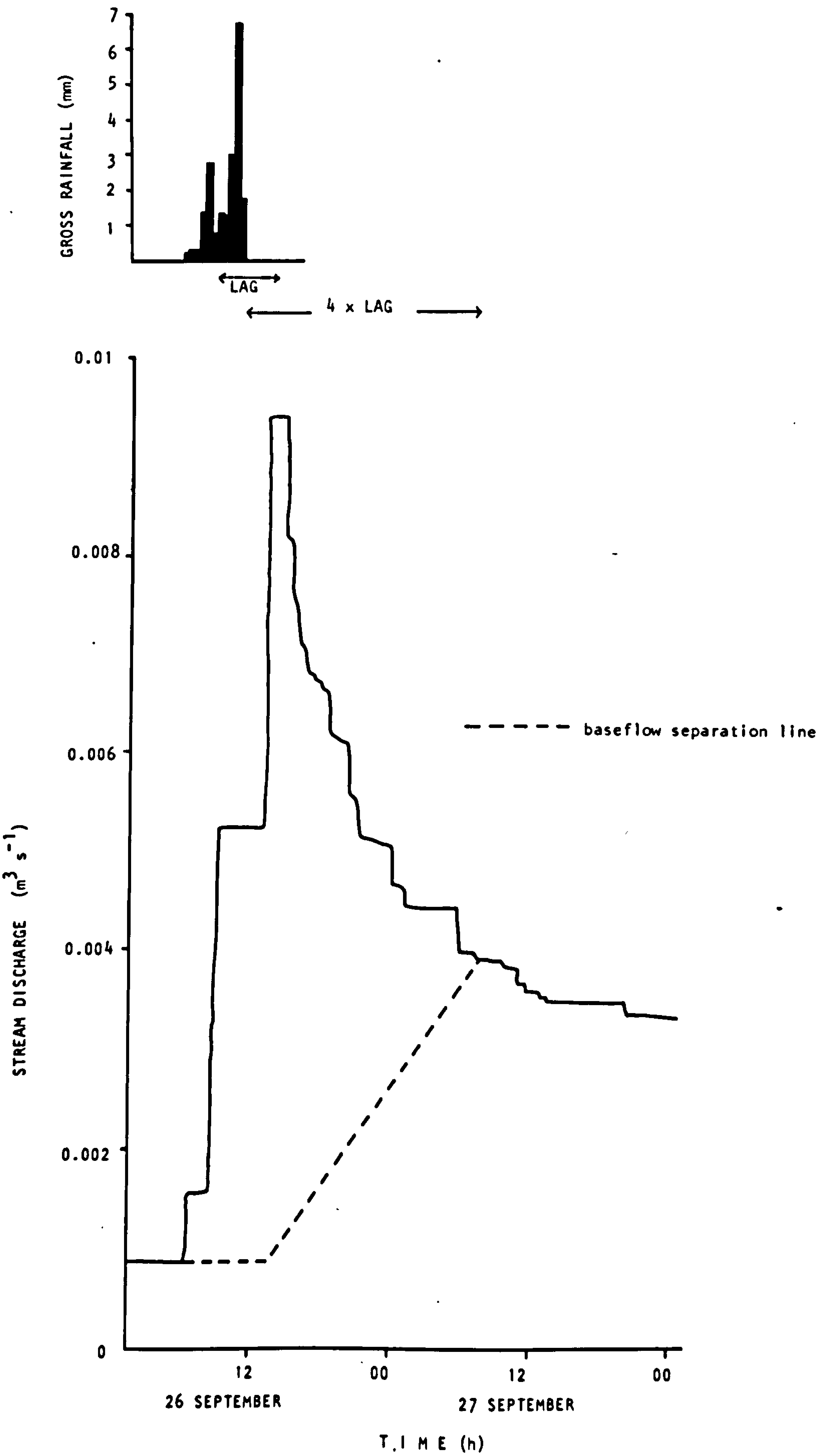


Figure 5.20 Runoff Hydrograph and attendant Rainfall for Event 2A (Post-Burn)

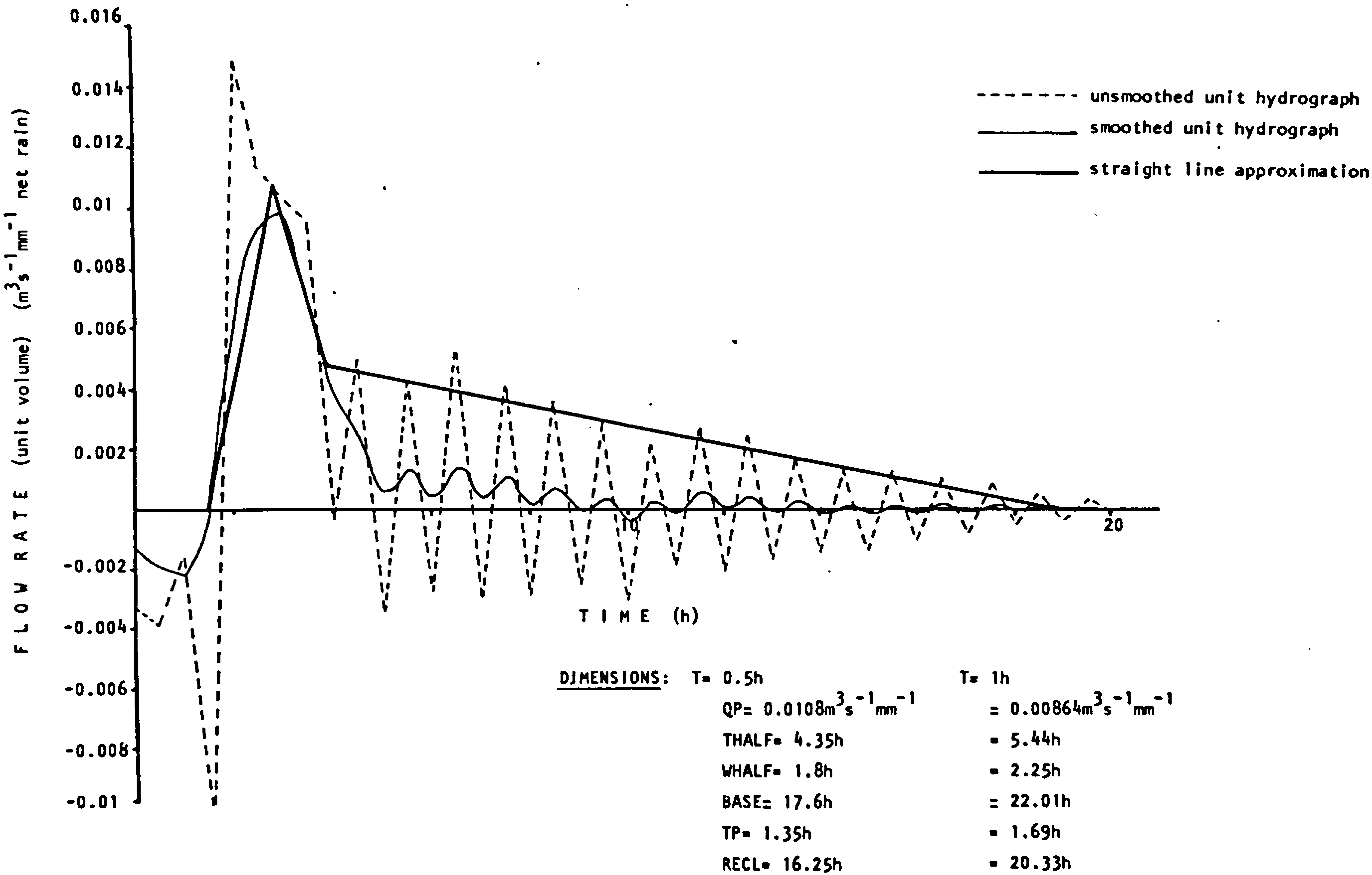


Figure 5.21 Least Squares Unit Hydrograph for Event 2A

Figures 5.14 and 5.15 being derived from the data shown in Table 5.2 (p. 251) and Appendix IV. Objective comparisons are generated by means of dimensional descriptions of the hydrographs, using the straight line approximation, defined in Figure 5.13 and exemplified by the notation in Figures 5.15, 5.17, 5.19 and 5.21.

Shape parameters, used subsequently in statistical analysis, are adjusted in cases where T is set to 0.5 h in order to make all hydrographs comparable as 1 h unit hydrographs. Parameters are re-defined by firstly increasing the value of TP (time to peak) by a factor of 0.25 (half the difference between 0.5, the original data interval, and 1 h). QP (peak discharge) is then re-calculated on the assumption that the relationship QP x TP remains the same (N.E.R.C., 1975). All remaining parameters are then re-defined assuming their ratio with TP remains constant. Only two of the thirteen unit hydrographs demonstrated a specific lag time before start of unit hydrograph and this parameter was not, therefore, included in hydrograph description and assessment and, as in the Flood Studies Report, was considered to be zero. S-curve or summation-curve methods can be used to alter unit hydrograph duration, deriving the new unit hydrograph from an S-shaped curve which represents a hydrograph of continuous effective rain falling indefinitely and at a constant rate, although success with this technique may be limited (N.E.R.C., 1975). The method is more fully described by Chow (1964), Wilson (1974) and Linsley _et al_. (1982).

5.3.4.6 Results and Physical Interpretation of Unit Hydrographs

Hydrograph shape parameters, defined for each event as described in the previous section, are used to develop both simple and multiple linear regression equations in order to determine significant relationships and predictions. Simple linear regression helps to indicate the varying importance of physical processes with vegetation change, different shape variables being related to each other for different, pre- and post-burn, catchment states. All thirteen events, however, are utilised as one complete data set for multiple regression analysis, in which hydrograph shape is regressed with storm and catchment characteristics, since the limited number of suitable storms renders separate, land-use based analyses statistically unjustifiable. Although this single, integrated approach may result in weighted equations, reflecting the greater number of events extracted for vegetated (eight) than non-vegetated (five) conditions, a combination of these two data sets effectively extends the data range of the physical parameters by inclusion of a wider set of values, and yields a set of general equations for this type of moorland catchment. Although statistical analysis is based on only small data sets, and interpretation of the results is therefore made with caution, effort is made where possible to meet the demands of the linear regression model and to recognise variation within a set of events.

a) Simple Linear Regression Analyses

All variables for proposed correlation and regression were considered firstly in terms of their hydrological interrelationships, so that dependent variables were determined in a manner consistent with physical reality. The data base was subsequently tested for the assumptions of the general linear model. Although these prerequisites are fairly restrictive, moderate departures may not necessarily result in serious errors. The more the assumptions are violated, however, the less valid are any inferences drawn from the results, and conclusions must be considered in general terms only. Goodness-of-fit of regression, and predictions made from a regression equation become more dependable as the model assumptions are satisfied.

Log transformations of the raw data were also made, both to derive more robust regression models and to overcome problems of non-normality of data. The non-parametric statistic, Spearman's rank correlation was also performed on raw data which did not comply with a normal distribution. Although the normality assumption requires both the marginal and conditional distributions of the x and y variables to be normal, the ability to test a conditional distribution relies on an ideal of several values of y for each x value: a condition which rarely applies, and thus only marginal distributions were examined. Similarly, it is seldom possible to test the assumption of zero mean of the conditional distributions of the error term in the regression equation, although small deviations from this assumption are not regarded as critical (Johnston, 1978).

Presence of one or more deviating observations may indicate departure from the assumption of homoscedascity, that is, the requirement for constant variances in the conditional distributions, although no serious violations were encountered in the present data sets. The autocorrelation assumption requires that the residuals from the regression are independent of each other, and in this respect no systematic variation was found in the sequence of positive and negative residuals, and in only a few cases were absolute values of successive residuals found to be correlated. All correlation and regression analyses were executed using the statistical computer package, 'SPSS'.

Several significant relationships transpire from correlating and regressing hydrograph shape variables with each other (Table 5.4). Regression equations are calculated only where parametric correlations are justified (Pearson product-moment correlation) and each includes the standard error of the estimate. Hydrograph shape is shown to be

I <u>RAW DATA</u>

PRE-BURN			POST-BURN		
Correlations:					
			BASE/WHALF	$r = 0.88$	(0.025)
WHALF/THALF	$r_s = 0.929$	(0.0005)	RECL/WHALF	$r = 0.903$	(0.018)
BASE/QP	$r = -0.625$	(0.049)	BASE/QP	$r = -0.955$	(0.006)
			RECL/QP	$r = -0.908$	(0.017)
WHALF/TP	$r_s = 0.946$	(0.001)			
TP/THALF	$r_s = 0.994$	(0.001)			

Regressions:

Pre-burn	Post-burn
	BASE = 18.199 + 1.942 WHALF $\pm$ 16.45 $r = 0.88$, $r^2 = 0.774$ (77.4% of variance in dependent variable accounted for by the regression) (significant at 0.025 level)
	RECL = 14.378 + 1.906 WHALF $\pm$ 14.166 $r = 0.903$, $r^2 = 0.816$ (0.018)
BASE = 42.807-5179.157QP $\pm$ 7.753 $r = -0.625$, $r^2 = 0.39$ (0.049)	BASE = 69.457-6421.31QP $\pm$ 10.252 $r = -0.955$, $r^2 = 0.912$ (0.006)
	RECL = 62.714-5831.977QP $\pm$ 13.847 $r = -0.908$, $r^2 = 0.824$ (0.017)

r_s = Spearman rank correlation coefficient

r = Pearson product-moment correlation coefficient

r^2 = coefficient of determination, the proportion of variation in the dependent variable explained by the independent variable

<u>Table 5.4 Significant Correlations and Regressions for Hydrograph Shape Parameters under Pre- and Post-Burn Conditions</u>

(Variables as defined in Figure 5.13; figures in brackets indicate level of significance)

II LOG-TRANSFORMED DATA

PRE-BURN

$WHALF = 3.357THALF^{0.858\pm0.223}$
$r = 0.785,\ r^2 = 0.615\ (0.011)$

$WHALF = 1.718TP^{0.887\pm0.217}$
$r = 0.796,\ r^2 = 0.634\ (0.009)$

$TP = 2.114THALF^{0.981\pm0.017}$
$r = 0.999,\ r^2 = 0.997\ (0.00001)$

$QP = 0.0057WHALF^{-0.664\pm0.118}$
$r = -0.897,\ r^2 = 0.805\ (0.001)$

POST-BURN

$BASE = 4.375WHALF^{0.93\pm0.34}$
$r = 0.875,\ r^2 = 0.765\ (0.026)$

$RECL = 2.825WHALF^{1.049\pm0.427}$
$r = 0.851,\ r^2 = 0.725\ (0.034)$

$BASE = 0.139QP^{-0.857\pm0.391}$
$r = -0.831,\ r^2 = 0.691\ (0.041)$

$QP = 0.016WHALF^{-1.021\pm0.091}$
$r = -0.991,\ r^2 = 0.982\ (0.0005)$

Table 5.4 (continued) Significant Regressions for Hydrograph Shape Parameters variously affected under different catchment conditions, indicating that different physical processes may assume varying degrees of importance with changing land-use. Shape variables are therefore not combined for multiple regression analysis. Using the raw data, width at half-peak (WHALF) proves to be a significant determinant of recession curve length (RECL) and base length (BASE) for the post-burn catchment only, indicating that these hydrographs may be explained by a simple triangular shape (Figs.5.22 and 5.23). The single value of a high WHALF and high BASE/RECL shown in these figures corresponds to a late October storm, and helps to extend the post-burn data sets along the axes, while the pre-burn data sets show a more constricted range of observations and fail to yield significant correlations. Predictions from these two post-burn catchment equations may prove unreliable, however, since the regression residuals show serial autocorrelation in each case.

Base length is negatively correlated with peak discharge, in compliance with unit hydrograph theory, although the equation for the vegetated surface explains only 39% of the variance in the dependent variable. This may be due to errors in estimating BASE for pre-burn storms, although 'noise' following smoothing was common to both pre- and post-burn unit hydrographs. Alternatively, timing errors in measurement equipment may have been more common during the earlier, pre-burn stage of the experiment. A wider spread of peak discharge values is found after burning than before (Figs. 5.24 and 5.25), the highest peaks occurring in September, a month not represented in the pre-burn storm events. The restricted coverage prior to burning may therefore be a function of season, within the limited number of storms selected, or may relate to the burnt catchment being more sensitive than a vegetated surface, the vegetative layer acting as an interception 'buffer'. The idea that such relationships may lend support to the concept of variable source areas

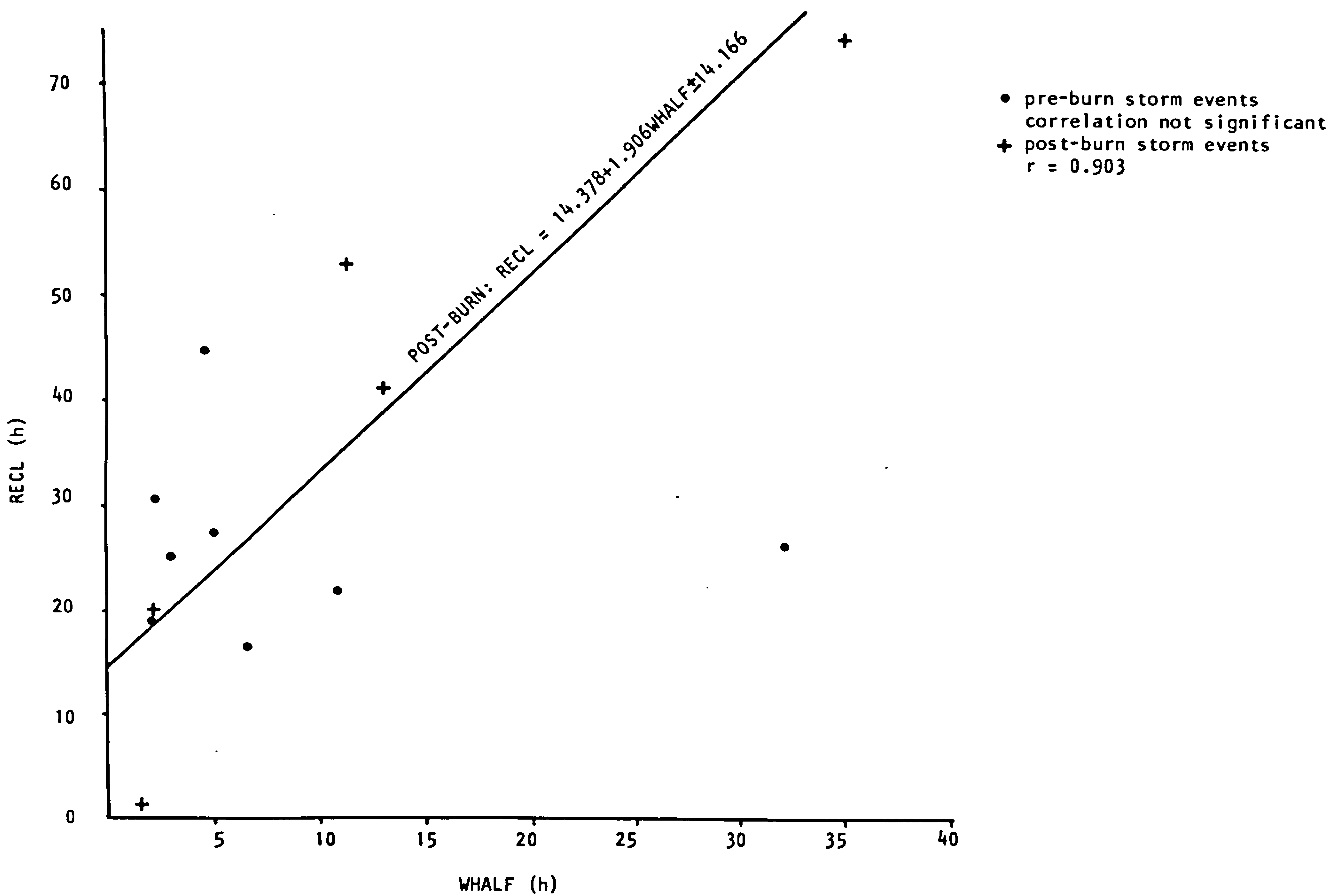


Figure 5.22 Relationship between Hydrograph Recession Length (RECL) and Width at Half-Peak (WHALF)

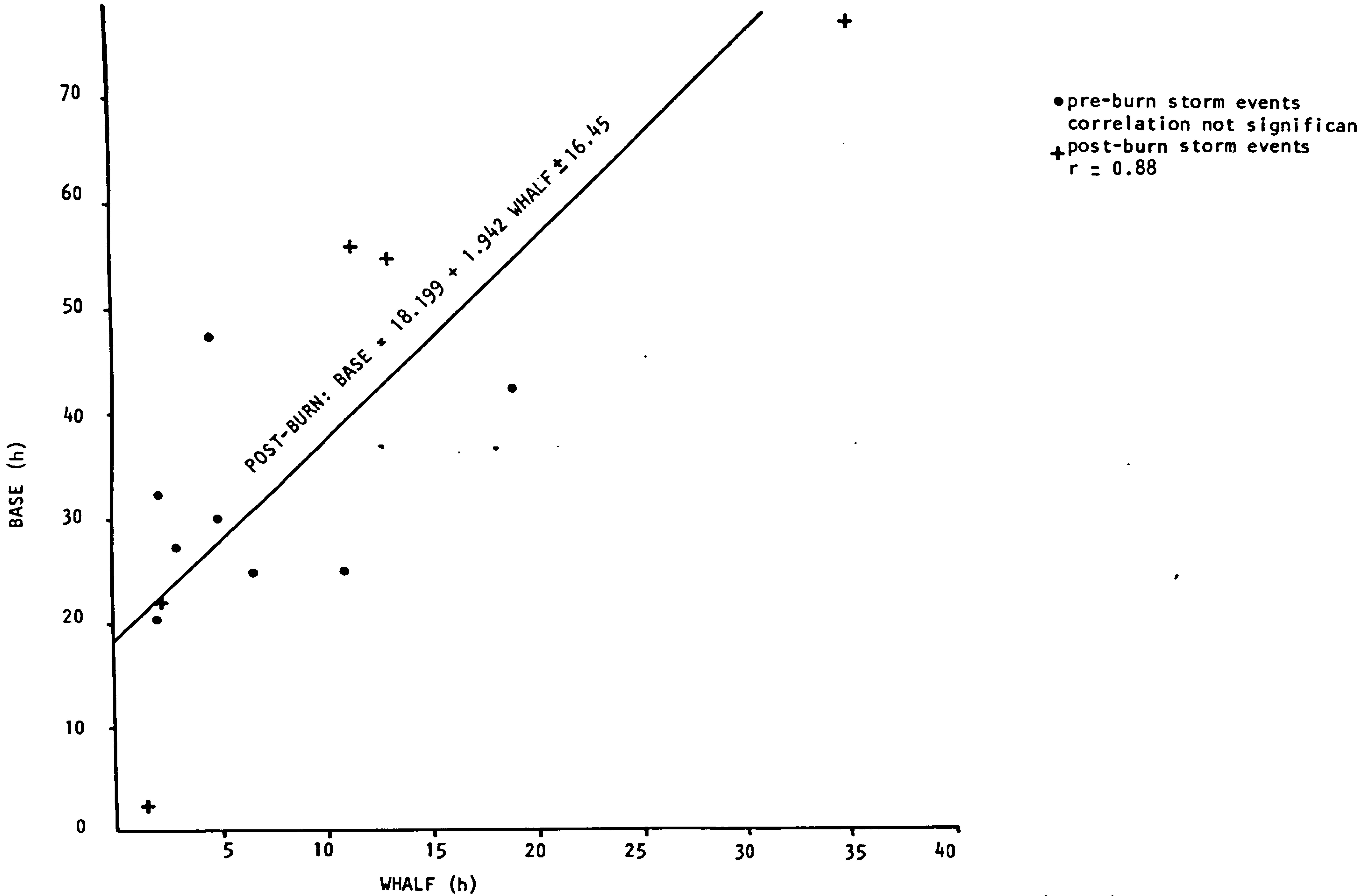


Figure 5.23 Relationship between Hydrograph Base Length (BASE) and Width at Half-Peak (WHALF)

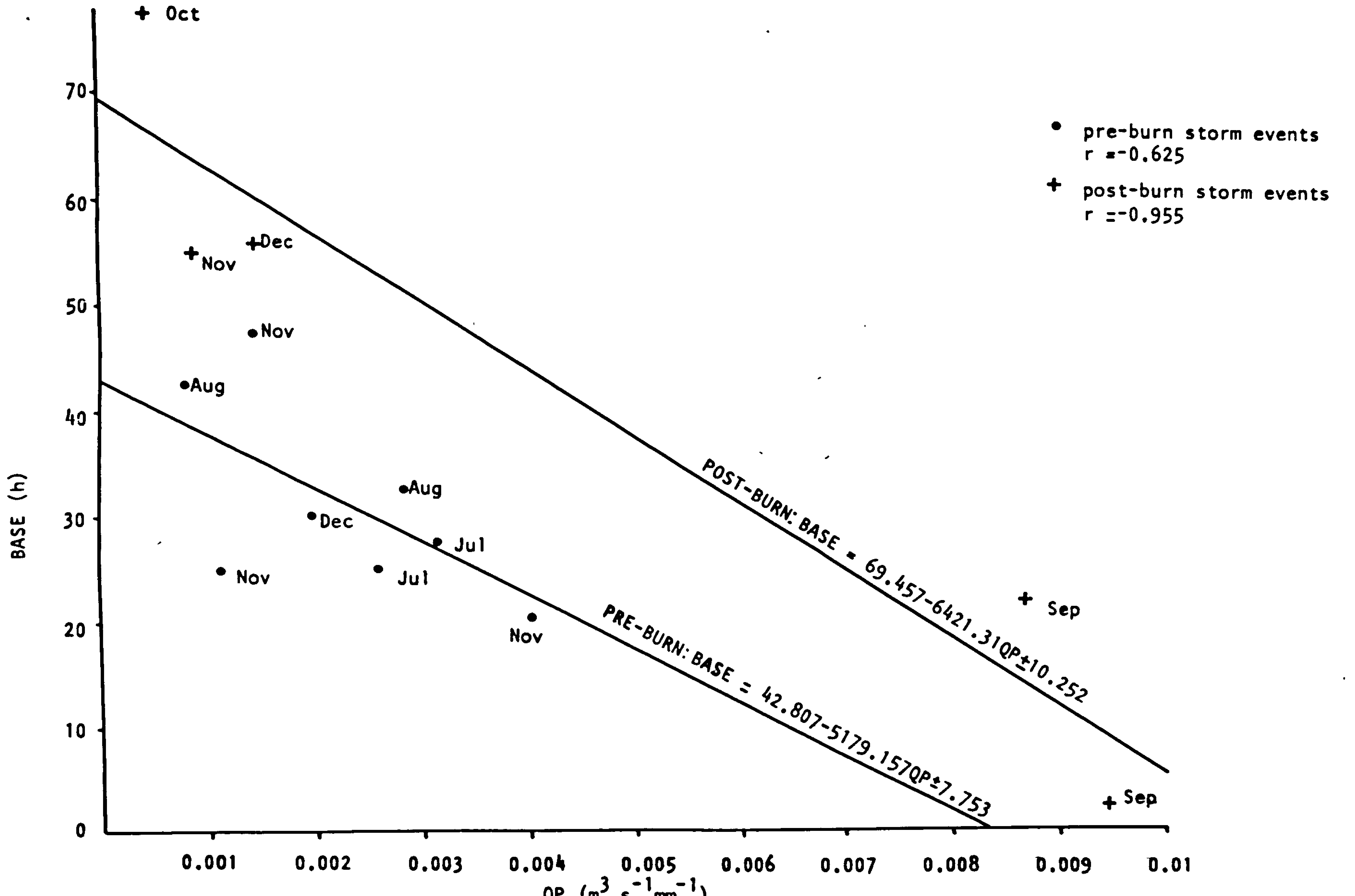


Figure 5.24 Relationship between Hydrograph Base Length (BASE) and Peak Discharge (QP) (Timing of events as indicated)

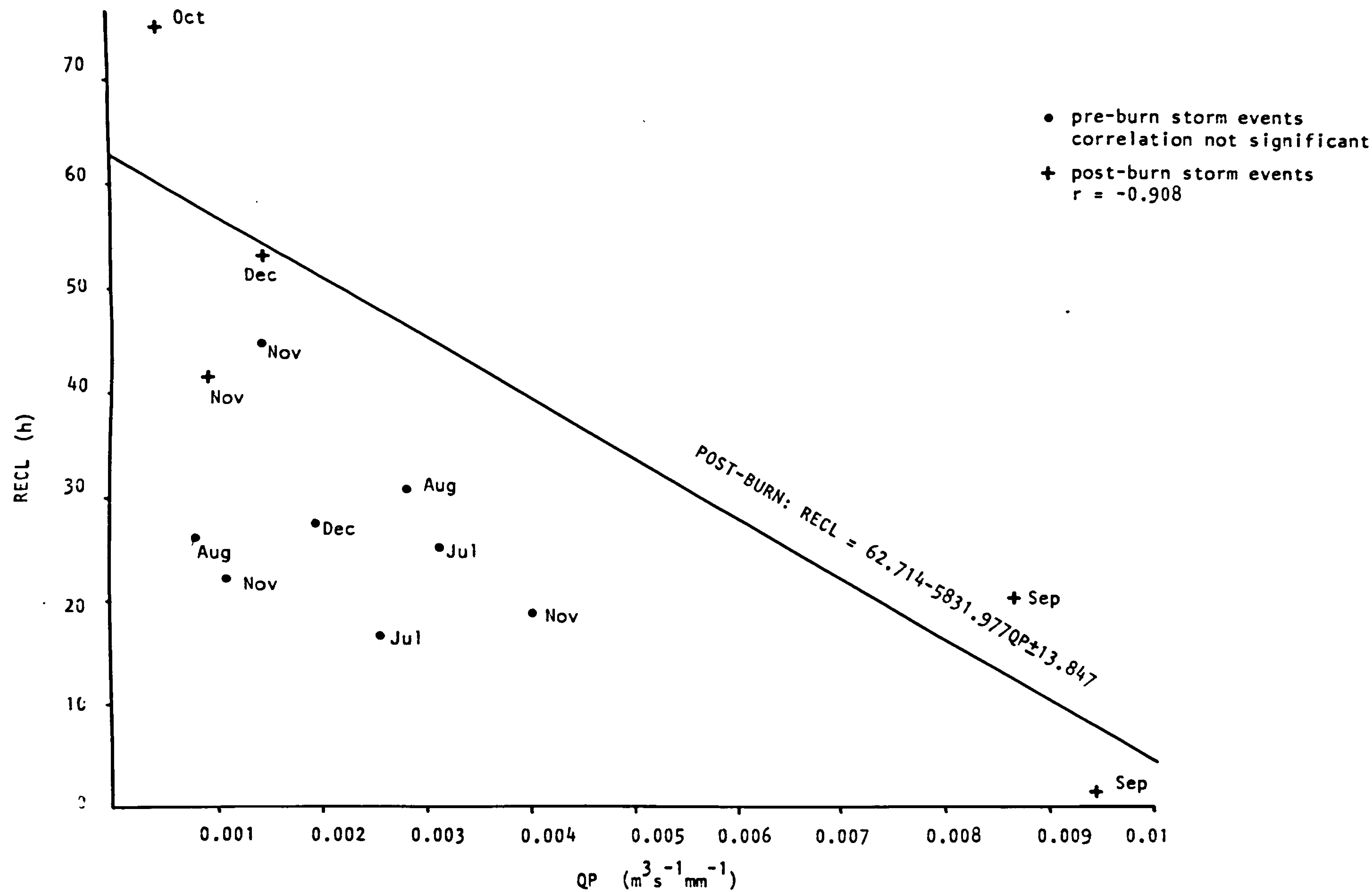


Figure 5.25 Relationship between Hydrograph Recession Length (RECL) and Peak Discharge (QP)

(Wheater et al., 1978) is considered later.

THALF, WHALF and TP show highly significant interrelationships prior to the muirburn (Table 5.4), but fail to do so for the burnt surface. This may be due to the October storm, in this case, biasing the remainder of the data set (Figs.5.26 and 5.27) since its unusually large width at half-peak, 35 h, fails to be matched by corresponding values of either THALF or TP, this storm being characterised by a steep rising limb but long recession. In certain cases, the requirement for a linear trend in the scatter of data points fails to be satisfied unequivocally, as in the case of QP/WHALF (Fig.5.28) for which a curvilinear relationship appears more appropriate. A larger data set would help to clarify such occurrences.

Log transformation of the raw data yields further significant relationships, again only one of which is common to both pre- and post-burn catchments. The highly significant relationship between $\log_{10}$ THALF and $\log_{10}$ TP prior to heather burning, $r = 0.999$ (Fig.5.29(a)) results from their physical linkage (Fig.5.13) and also from the constraints of the unit hydrograph which specify a lengthening time base with reduced peak and angle of rising limb. This correlation gives a slightly better fit to the data than the equivalent non-parametric correlation on untransformed data ($r_s = 0.994$). Similarly, Figure 5.29(b) shows strongly significant correlations between LWHALF and LQP. Data transformation fails to compensate sufficiently for the outlying observation illustrated in Figures 5.26 and 5.27, however, and width at half-peak remains correlated with both TP and THALF for the pre-burn surface only (Figs.5.30 and 5.31).

LQP is used to predict LBASE, as for the untransformed data, although the relationship for the pre-burn catchment fails to reach significance at the .05 level ($r = -0.601$, significant at 0.058 level).

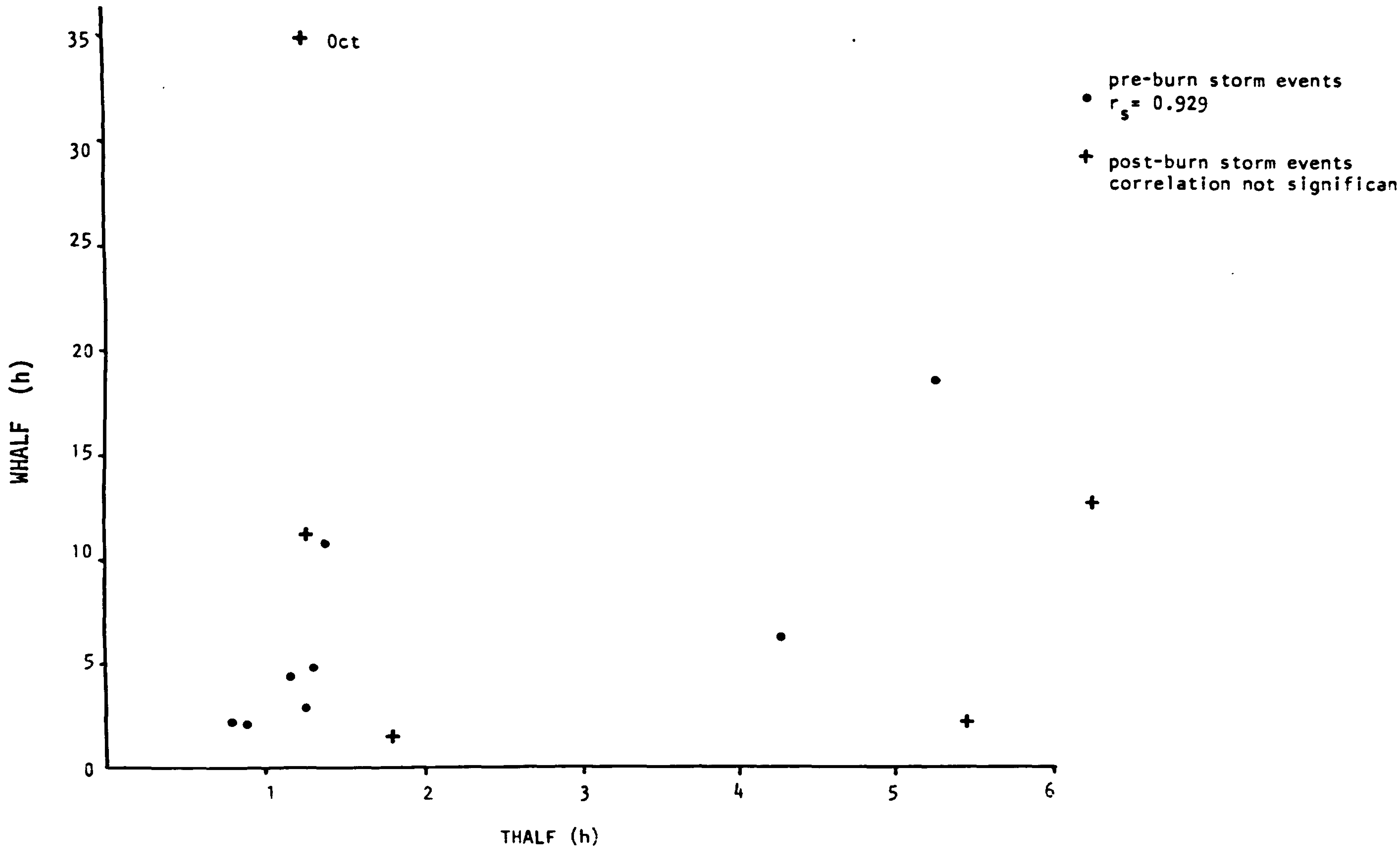


Figure 5.26 Relationship between Hydrograph Width at Half-Peak (WHALF) and Time to Half-Peak (THALF)

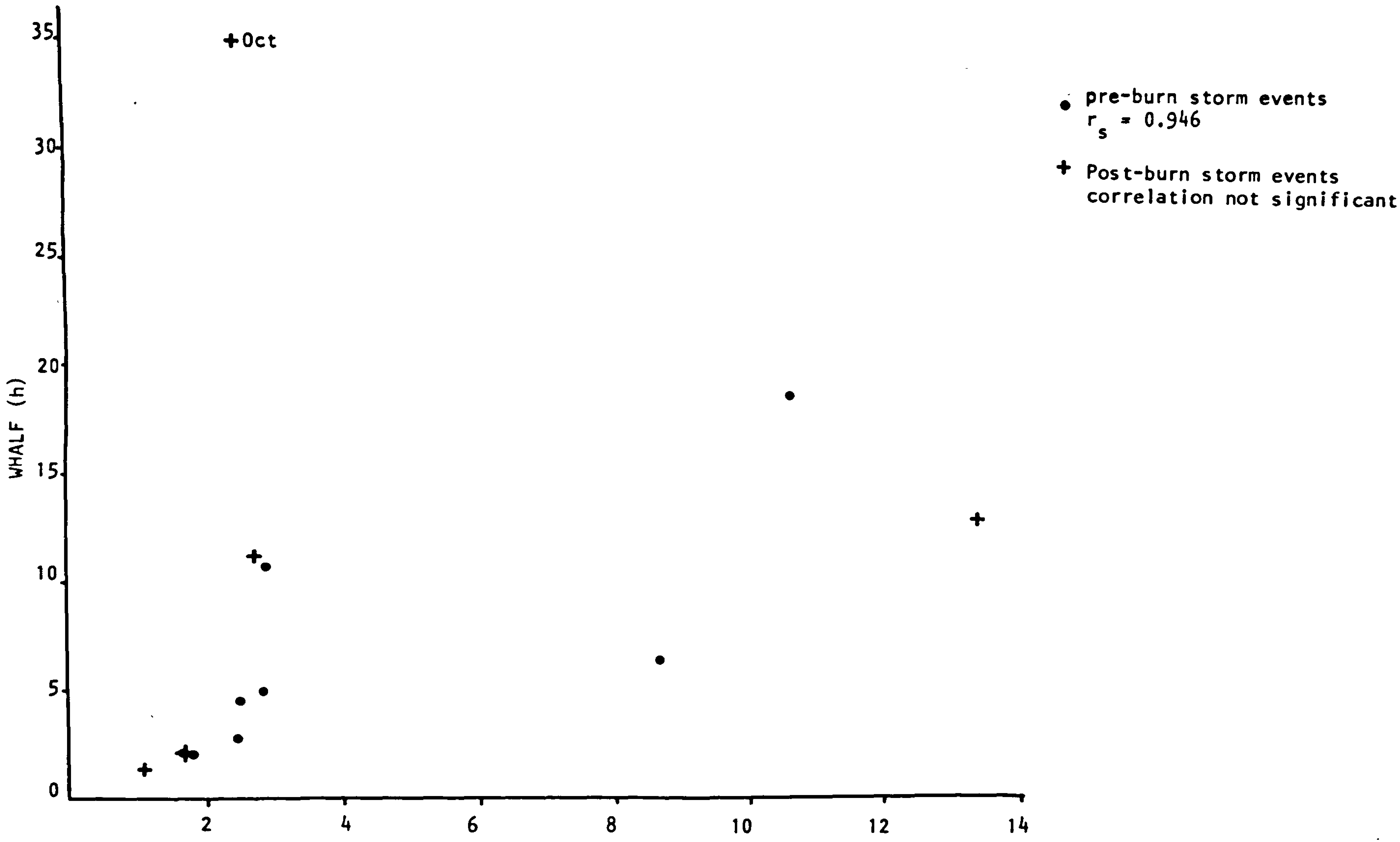


Figure 5.27 Relationship between Hydrograph Width at Half-Peak (WHALF) and Time to Peak (TP)

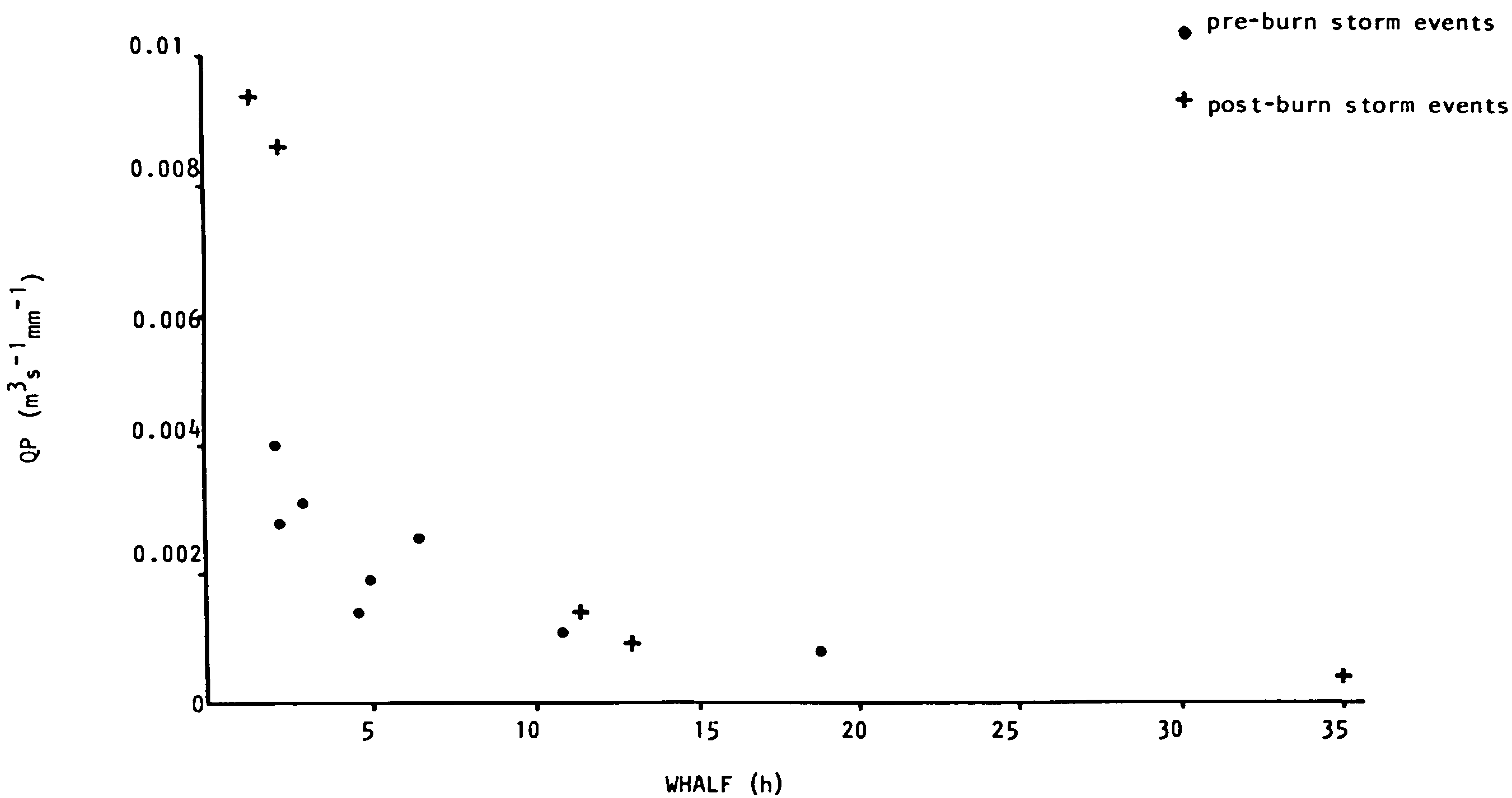


Figure 5.28 Curvilinear Relationship between Hydrograph Peak Discharge (QP) and Width at Half-Peak (WHALF)

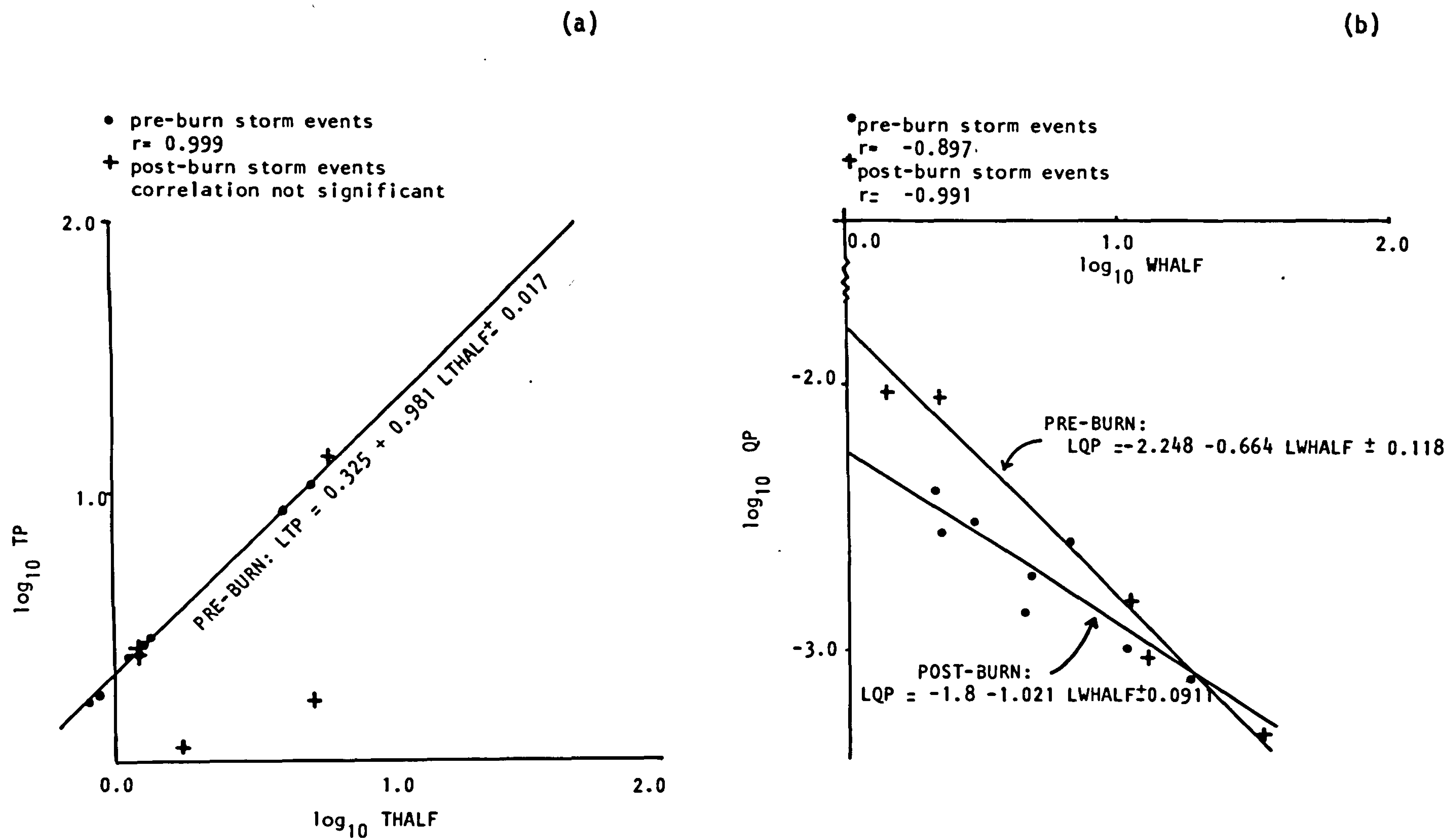


Figure 5.29 Correlations between (a) Hydrograph Time to Peak (TP) and Time to Half-Peak (THALF) and (b) Peak Discharge (QP) and Width at Half-Peak (WHALF) following Data Transformation

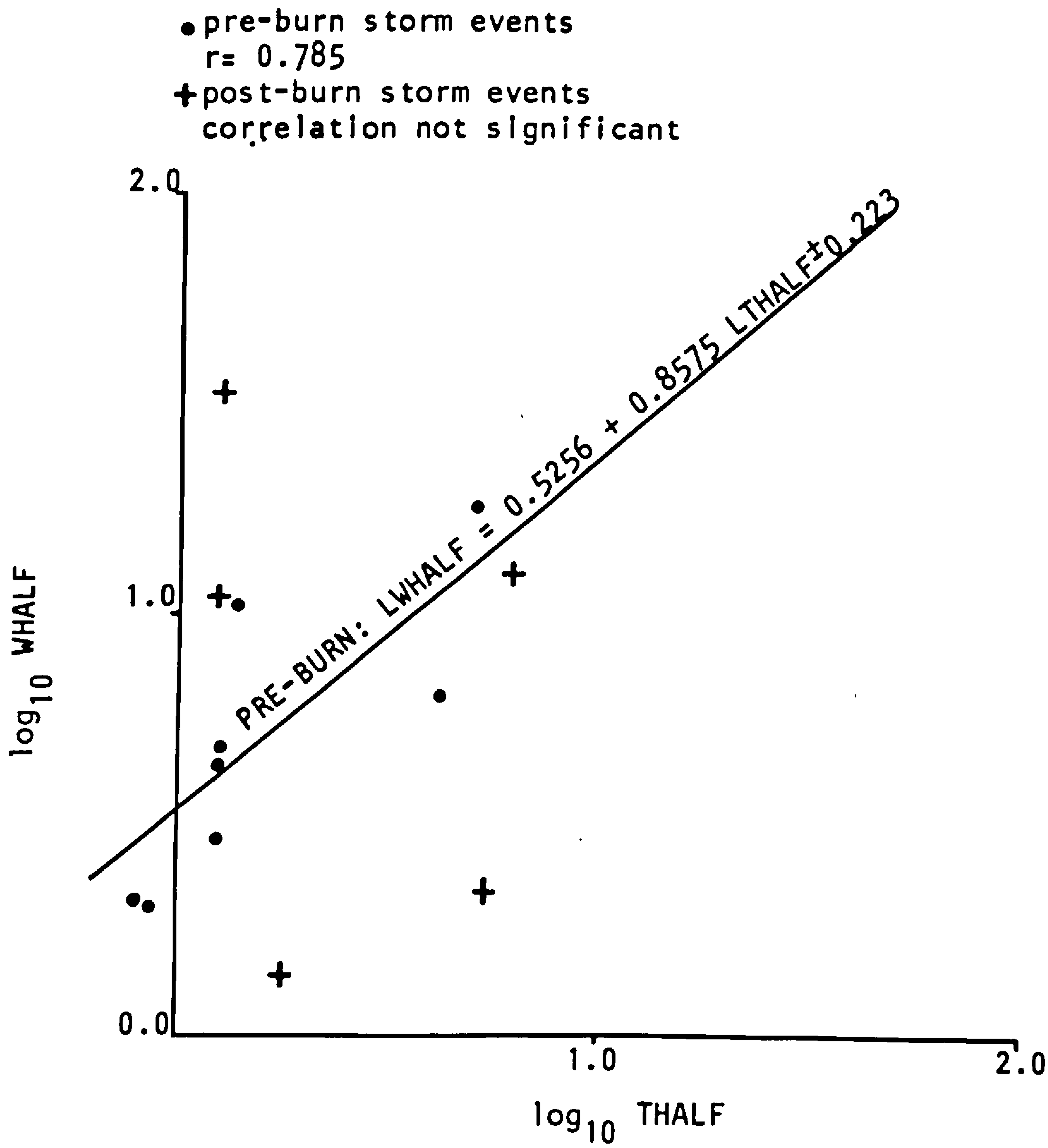


Figure 5.30 Relationship between Hydrograph Width at Half-Peak (WHALF) and Time to Half-Peak (THALF) for Transformed Data

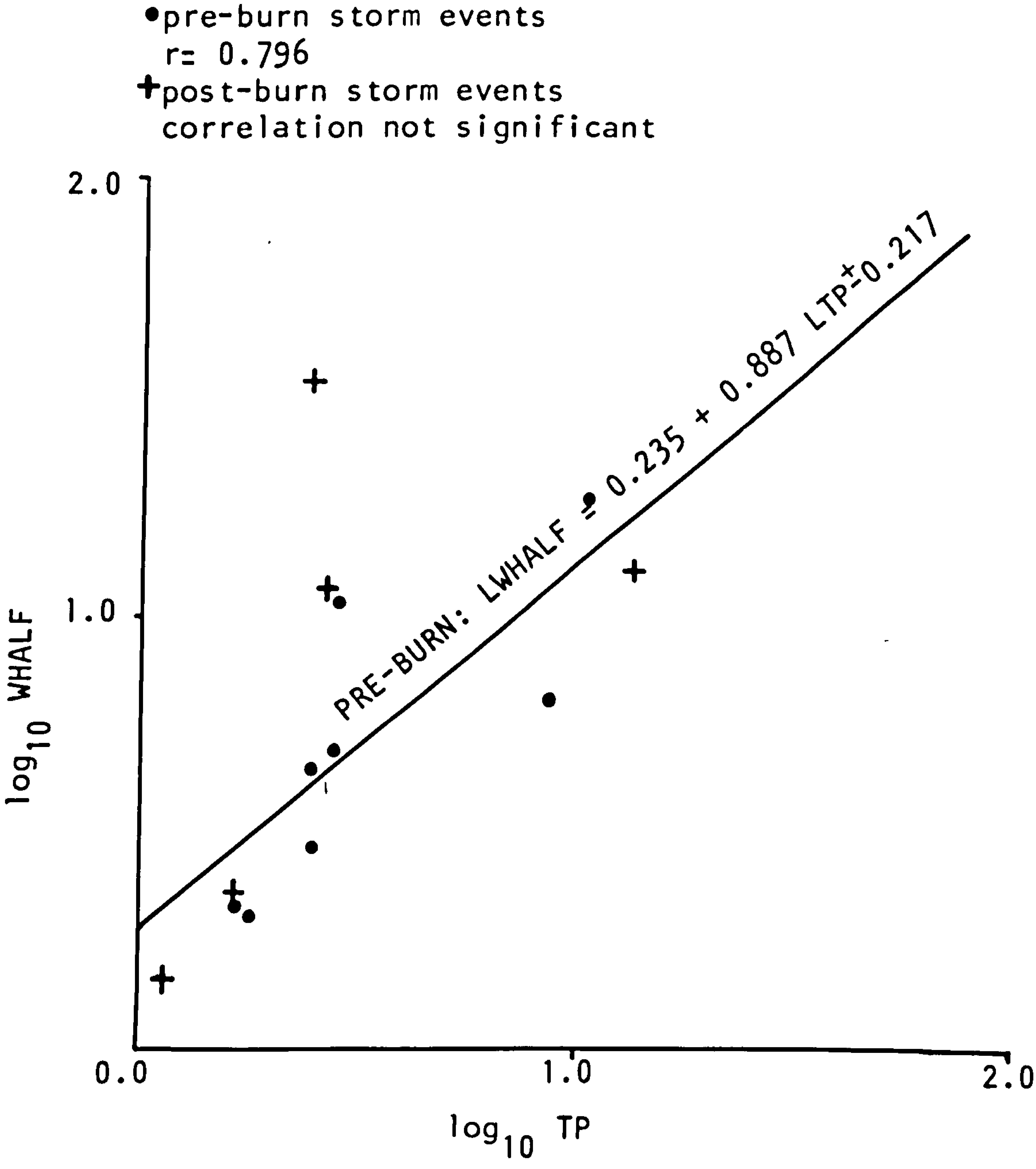


Figure 5.31 Relationship between Hydrograph Width at Half-Peak (WHALF) and Time to Peak (TP) for Transformed Data

The positive relationship between LWHALF and both LBASE and LRECL proves significant only after burning, values being confined to a restricted range for the pre-burn storms. BASE and RECL appear independent of THALF for both raw and transformed data and, indeed, for the pre-burn raw data the relationship between THALF and BASE yields a Spearman correlation coefficient of 0.0 (Fig.5.32). This is probably a consequence of the storm events chosen, since these variables should be positively related in accordance with the definition of the unit hydrograph.

b) Multiple Regression Analyses

Of the six hydrograph dimensions examined, four main variables are selected to relate hydrograph shape to storm and catchment features. QP, TP, THALF and RECL are chosen as dependent variables since these are weakly correlated with each other and with most other shape variables. The initial number of independent variables is restricted in order to facilitate process interpretation. Those selected for use are total rainfall in mm (RAIN), rainfall duration in h (DUR) and millimetres of soil moisture deficit (SMD) (Table 5.1, p. 242).Rainfall intensity is rejected as an independent variable as it is strongly correlated with both rainfall duration (Pearson's r_S= -0.679, significant at 0.005 level) and SMD (Spearman's r_S = 0.687, significant at 0.005 level) and fails to influence the dependent variables in a significant way (Figs. 5.33 and 5.34). The single outlying observation illustrated in Figure 5.33 represents a post-burn September storm event and reflects the 'flashy' response of the stream to one of the first major rainfall events following vegetation removal, at a time when the catchment was at its driest.

A 'dummy' variable is also included as a binary, independent variable. The use of dummy variables is widely documented (Mather and Openshaw, 1974; Ferguson, 1977; Johnston, 1978; Draper and Smith, 1981;

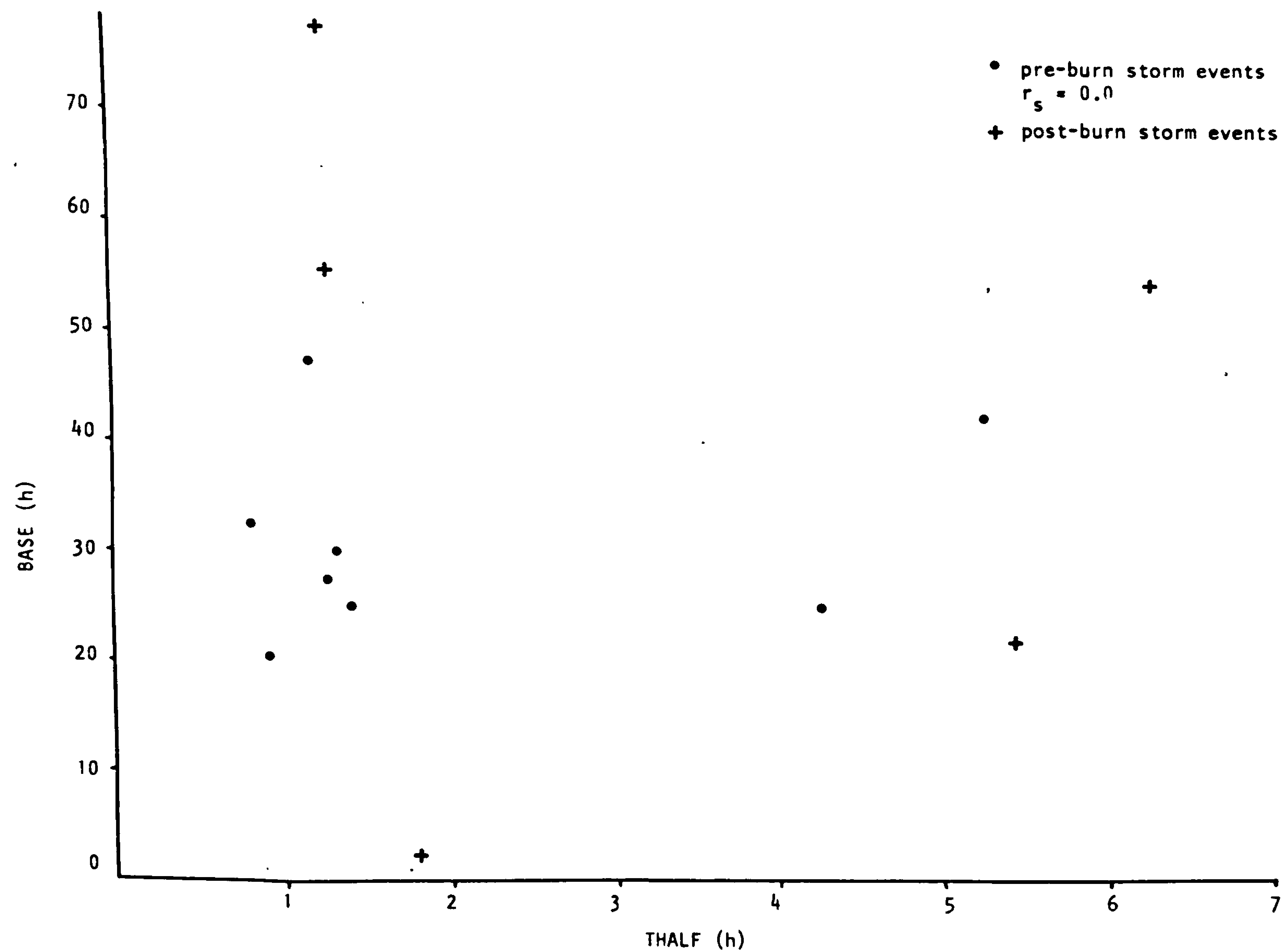


Figure 5.32 Relationship between Hydrograph Base Length (BASE) and Time to Half-Peak (THALF)

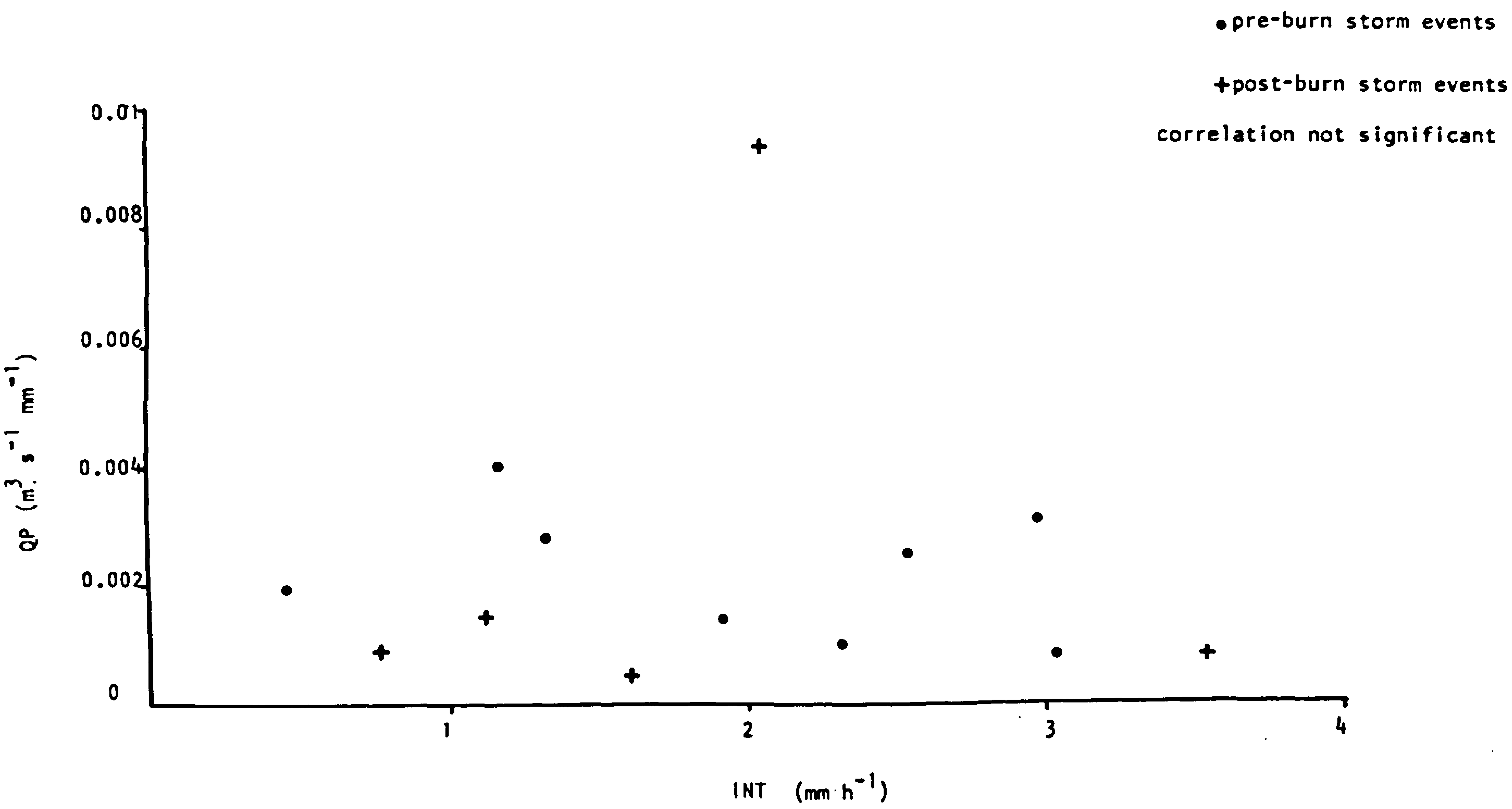


Figure 5.33 Effect of Rainfall Intensity (INT) on Hydrograph Peak Discharge (QP) (Not significant)

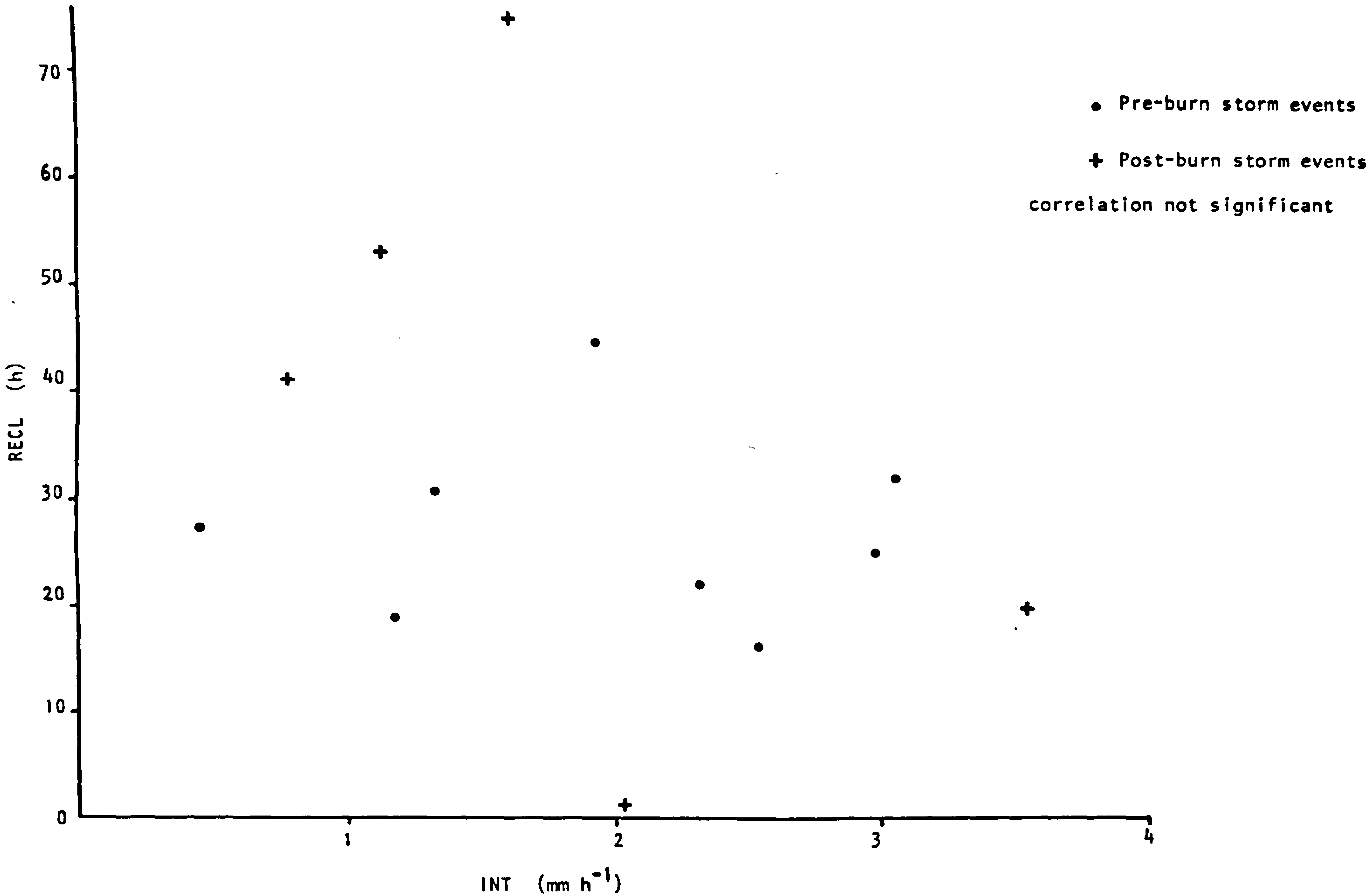


Figure 5.34 Effect of Rainfall Intensity (INT) on Hydrograph Recession Length (RECL) (Not significant)

Hewlett and Bosch, 1985) and is justified in the present instance in view of the small number of data values involved and as a means of exposing hydrograph/land-use relationships which may otherwise have been masked by variability in storm characteristics within each vegetation category. The dummy variable in this case incorporates both changes due to the removal of vegetation and variability in other ambient factors, such as climate which would have occurred irrespective of land-use patterns. By representing physical parameters which were not included in the monitoring programme, the dummy variable enables relationships to be more clearly defined since a higher degree of explanation is allowed. Inclusion of the variable also allows those independent variables in the equation which would have been omitted from separate analyses of pre- and post-burn catchments to become significant. The method therefore helps to explain differences which, otherwise, would have been left undefined (Simonett, 1967), although a larger data set or incorporation of further independent variables, such as infiltration capacity, would have produced a clearer distinction of differences resulting from the burn alone. A dummy variable has at least two discrete levels, and values of 0 and 1 are assigned here to represent burnt and vegetated states, respectively.

All the assumptions of the linear regression model are satisfied as closely as possible, the multivariate case encompassing a further requirement, that the residuals from each partial regression equation should be uncorrelated. This is usually interpreted to mean that the independent variables should be uncorrelated; otherwise collinearity exists and the partial regression coefficients become biased (Johnston, 1978). Hewlett and Bosch (1985) noted that hydrological variables characteristically exhibit collinearity, but that the 'F' Test for entering a variable into the equation is quite robust. The only transgression from the assumption in the present case lies in the weak negative correlation between soil moisture deficit (SMD) and rainfall duration, DUR (Spearman's r_s = -0.519, significant at 0.035 level). Marginal distributions of all variables proved to be normal although in the case of SMD for both basic and log transformed data, a strong normal distribution was less certain (D = 0.396 for SMD, 0.39 for LSMD, Kolmogorov-Smirnov Goodness-of-Fit test, both significant at only the 0.01 level). Scatter plots drawn prior to regression analysis show all variables to meet the linearity assumption, and in only a few cases are deviant observations apparent. These diagrams also help to illustrate the dependent variables in terms of simple functions with storm and catchment characteristics, prior to investigation of more complex relationships. Thus, soil moisture deficit is shown to have a positive effect on peak discharge (Spearman's r_s = 0.585, significant at the 0.025 level) and a negative relationship with recession length (r_s = -0.603, significant at 0.025 level) (Figs.5.35 and 5.36).

The complete set of significant multiple regression models is shown in Table 5.5. Discussion is necessarily limited to these relationships, although other, interacting and usually non-linear effects on hydrograph shape may be important (Francis, 1973). No significant linear relationships transpired to describe the variables TP or THALF. Equation 5.14, which explains 77.2% of the variation in QP, does not incorporate rainfall duration (DUR) since this adds only 0.4% to the variance explained and the 'F' ratio of the partial regression coefficient is not significant at the 0.05 level. Similarly, although inclusion of DUMMY and DUR variables in Equation 5.15 would increase the variance explained to 72.8% and 75.3%, respectively, these variables are omitted since neither regression coefficient is statistically significant. Log transformation of the raw data (Table 5.1, p.242) yields a parallel set of equations which should prove more robust than the

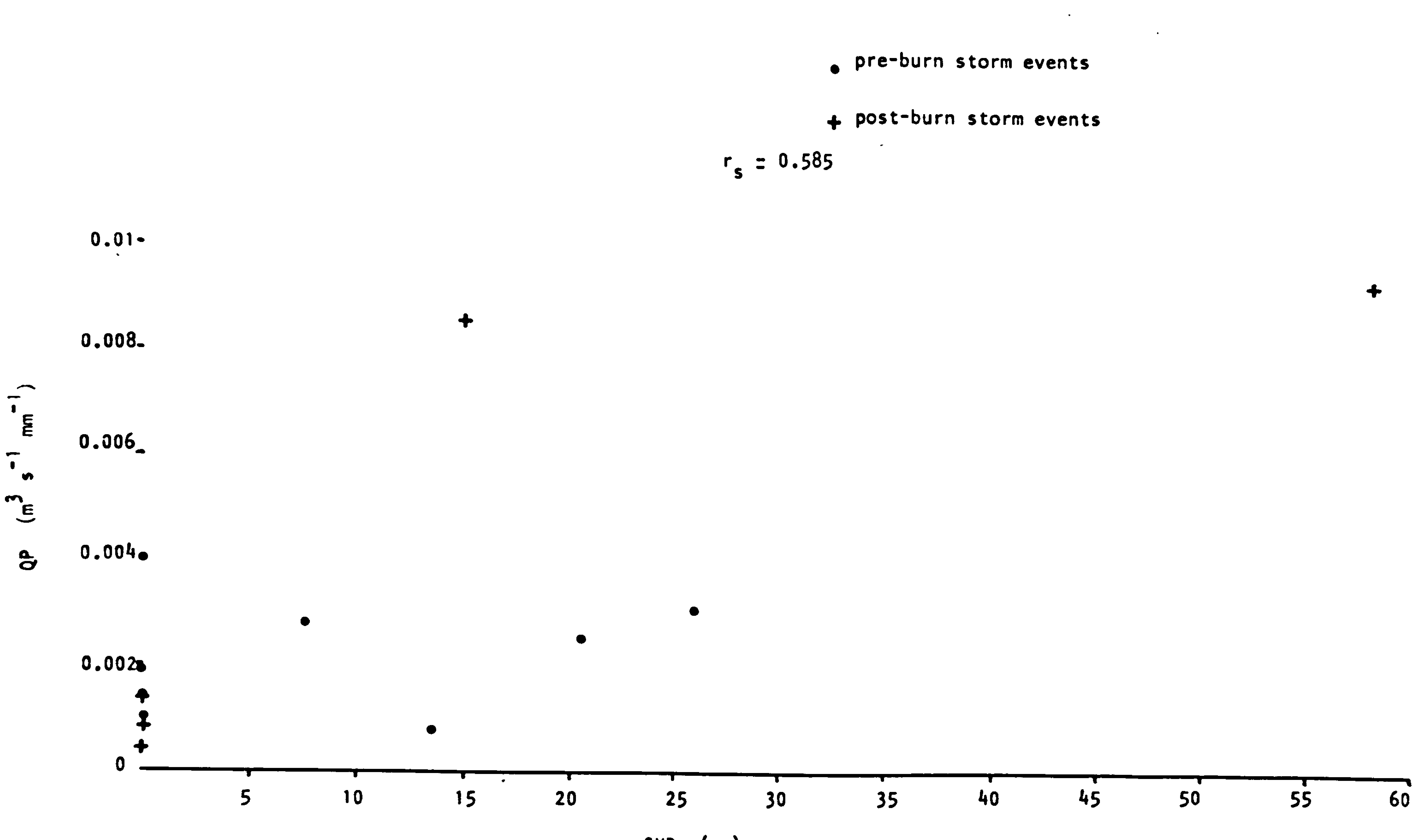
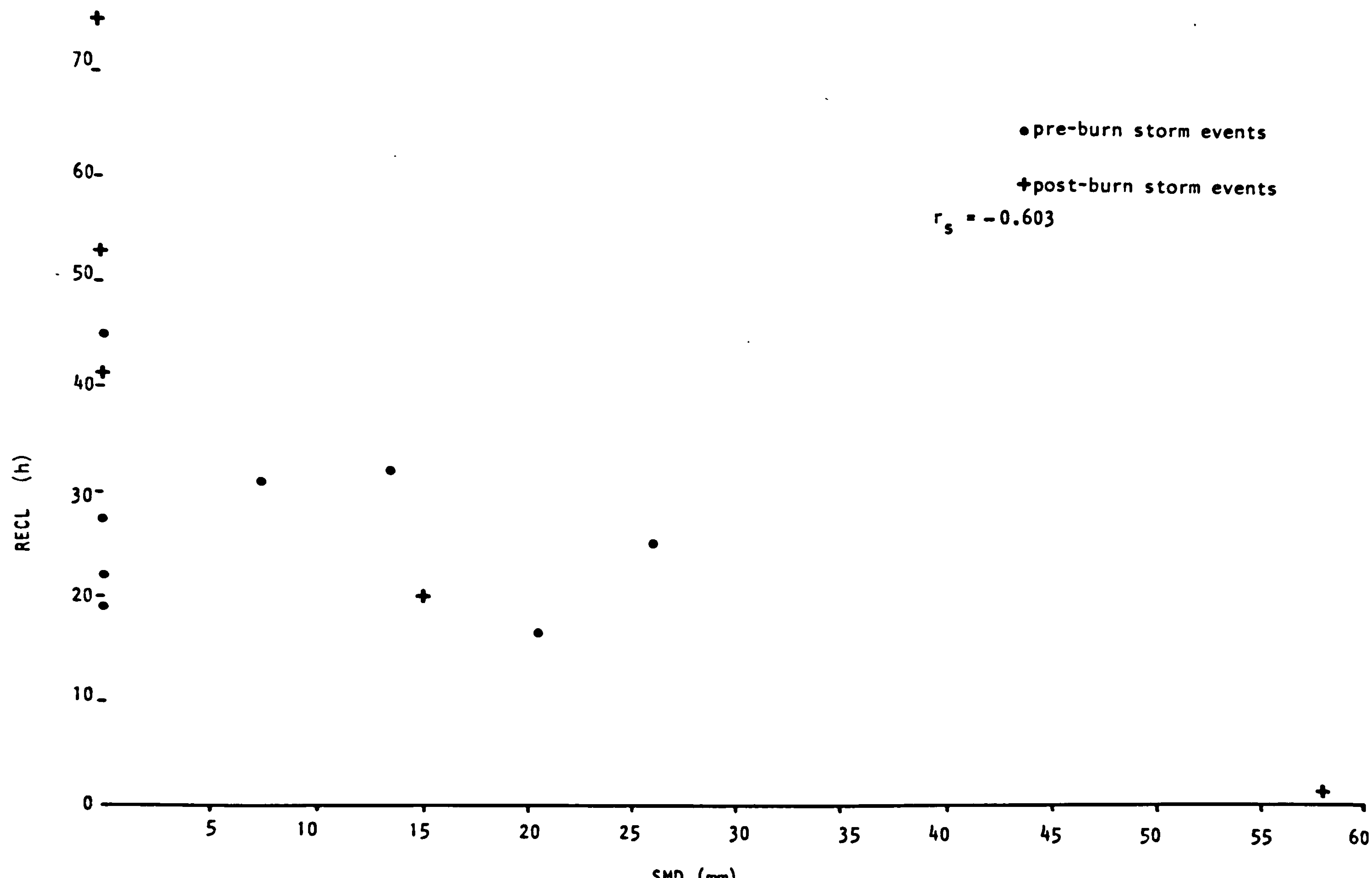


Figure 5.35 Effect of Soil Moisture Deficit (SMD) on Hydrograph Peak Discharge (QP) (Significant at 0.025 level)



<u>Figure 5.36 Effect of Soil Moisture Deficit (SMD) on Hydrograph Recession Length (RECL)</u> (Significant at 0.025 level)

Non-standardised partial regression coefficients:-

I <u>RAW DATA</u>

$QP = -0.0013 + 0.00012\ SMD + 0.00036\ RAIN - 0.00191\ DUMMY \pm 0.00159$ Eq. 5.14

$r = 0.879,\ r^2 = 0.772$

$F = 10.169$ (significant at 0.005 level)

$RECL = 68.034 - 0.745SMD - 2.482\ RAIN \pm 12.491$ Eq. 5.15

$r = 0.794,\ r^2 = 0.63$

$F = 8.508$ (significant at 0.01 level)

II <u>LOG-TRANSFORMED DATA</u>

$QP = 0.000079\ (1 + SMD/100)^{4.35}\ .RAIN^{1.189} \pm 0.269$ Eq. 5.16

$r = 0.777,\ r^2 = 0.604$

$F = 7.624$(signficant at 0.01 level)

$RECL = 42.462\ (1 + SMD/100)^{-6.023} \pm 0.233$ Eq. 5.17

$r = 0.849,\ r^2 = 0.721$

$F = 28.384$ (signficant at 0.001 level)

Standardised partial regression coefficients:-

I <u>RAW DATA</u>

$QP = 0.705\ SMD + 0.458\ RAIN - 0.336\ DUMMY \pm 0.00159$ Eq. 5.18

$RECL = -0.67\ SMD - 0.485\ RAIN \pm 12.491$ Eq. 5.19

II <u>LOG-TRANSFORMED DATA</u>

$QP = (1 + SMD/100)^{0.664}\ .RAIN^{0.433} \pm 0.269$ Eq. 5.20

$RECL = (1 + SMD/100)^{-0.849} \pm 0.233$ Eq. 5.21

<u>Table 5.5 Significant Multiple Regression Models</u>

equivalent untransformed models. An improved equation results for RECL, explaining 72.1% of the variance in the dependent variable, and giving a simple regression as the best fit (Eq. 5.17). A slightly worse fit pertains in the case of peak discharge (Eq. 5.16) with, again, one less variable than in the basic equation.

Standardised partial regression coefficients, or beta weights, are also determined, in order to overcome problems of mixed units of measurement in the independent variables, and therefore to illustrate the relative effects of each independent variable on the dependent, on a standard scale. The overriding importance of soil moisture deficit is thus indicated by Equations 5.18, 5.19 and 5.20 since in relative terms peak discharge increases, and RECL decreases, at a greater rate with a given increase in SMD than at the same rate of increase in total rainfall. Relationships between runoff characteristics and antecedent soil moisture are often established as significant (Dickinson and Whiteley, 1973; Lynch, 1977). Wheater and Weaver (1980) concluded that for lowland, clay catchments around Gloucester, soil moisture deficit derived from the Grindley model is the most significant parameter in determining hydrograph shape. For the present catchment, SMD is highly correlated with QP in Equation 5.14 (partial regression coefficient significant at 0.005 level) and is strongly negatively correlated with RECL (Eq. 5.15), suggesting an overall quicker response from a drier catchment. Log SMD is particularly highly correlated with log RECL in Equation 5.17 (significant at 0.001 level). These results minister to the proposition of variable contributing areas, since under high moisture deficit conditions only a small area of the catchment contributes to streamflow, whilst on a wetter catchment, subsurface flow becomes more important and source areas expand, increasing times of travel. The hydrograph is therefore lengthened and peak discharges are reduced

(Wheater, et al., 1978, 1982). Possible source areas on the present study site include regions close to the main channels and along the slope base, and localised saturation patches towards the catchment boundary. The fast response of a dry catchment may be counteracted by greater rainfall losses to soil moisture replenishment under high deficit conditions, however, this concept being examined later in terms of its importance relative to prevailing land-use, when separate, 'average' unit hydrographs are derived for each land-use type. Other hillslope parameters, not specified in this study, may also be relevant to the runoff generation process. Freeze (1980), for example, in a stochastic-conceptual study, concluded that the spatial distribution of hydraulic conductivity is significant in determining the statistics of runoff events and that it should be included in representations of the unit hydrograph.

As multiple regression equations are not found to be significant in describing hydrograph time to peak, TP, time to half-peak, THALF, or their log equivalents, other storm or geomorphological factors may be important in determining these variables. Time taken to peak was related to quickflow-generating mechanism and catchment area by Dunne (1978), while McCaig (1983, Fig.4) summarised these relationships diagrammatically, from which the principal mode of storm runoff generation for the present study site is concluded to be subsurface stormflow. Difficulties, in some cases, in locating the precise time of start of the rising limb, due to 'noise' in the hydrograph, may also contribute to the absence of significant correlations with storm and catchment features. A larger data set may help to elucidate valid relationships for TP and THALF, since existing data show clustering (Fig.5.37) and scatter (Fig.5.38). In particular, evidence to corroborate the idea of speed of runoff response as a function of

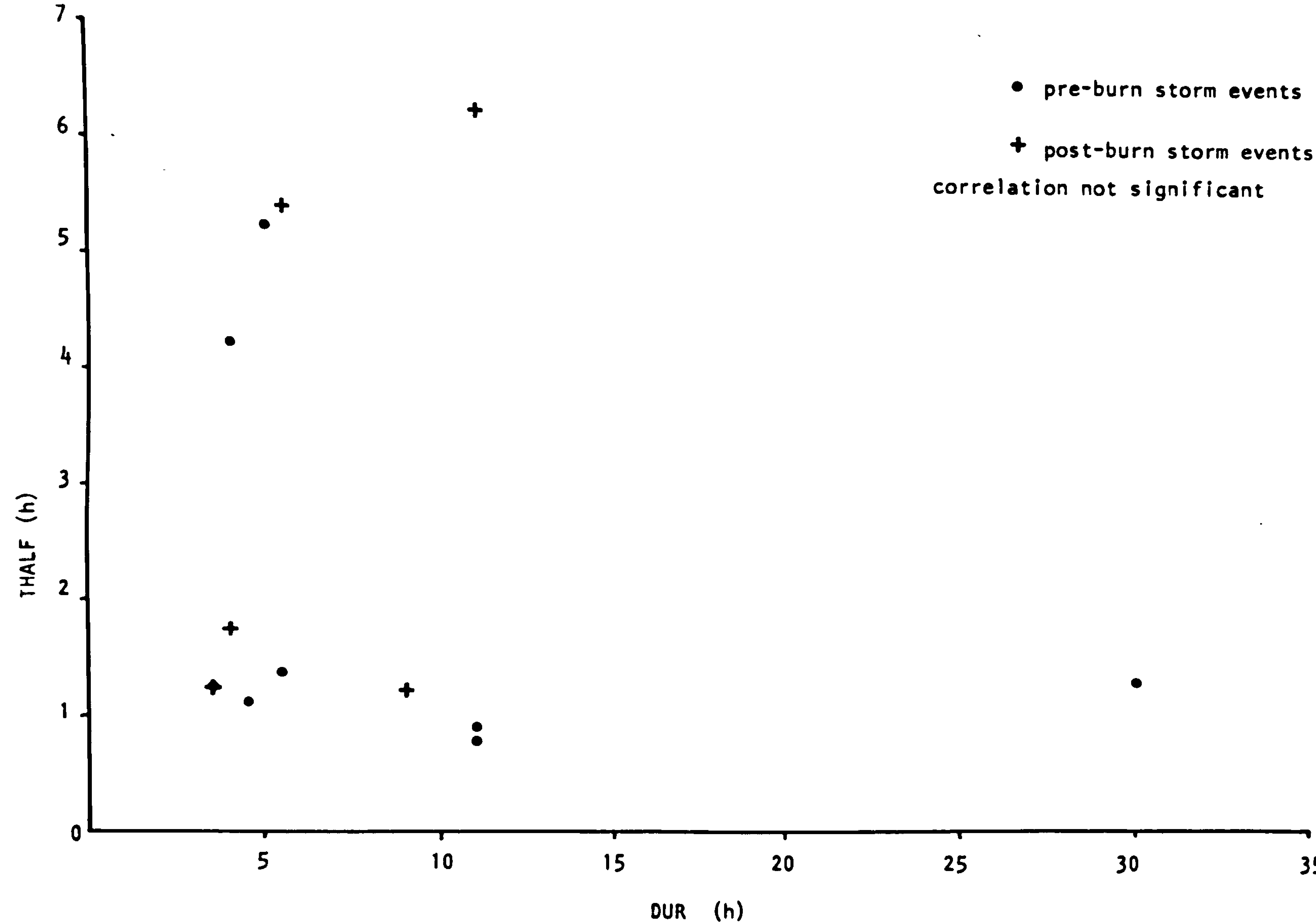
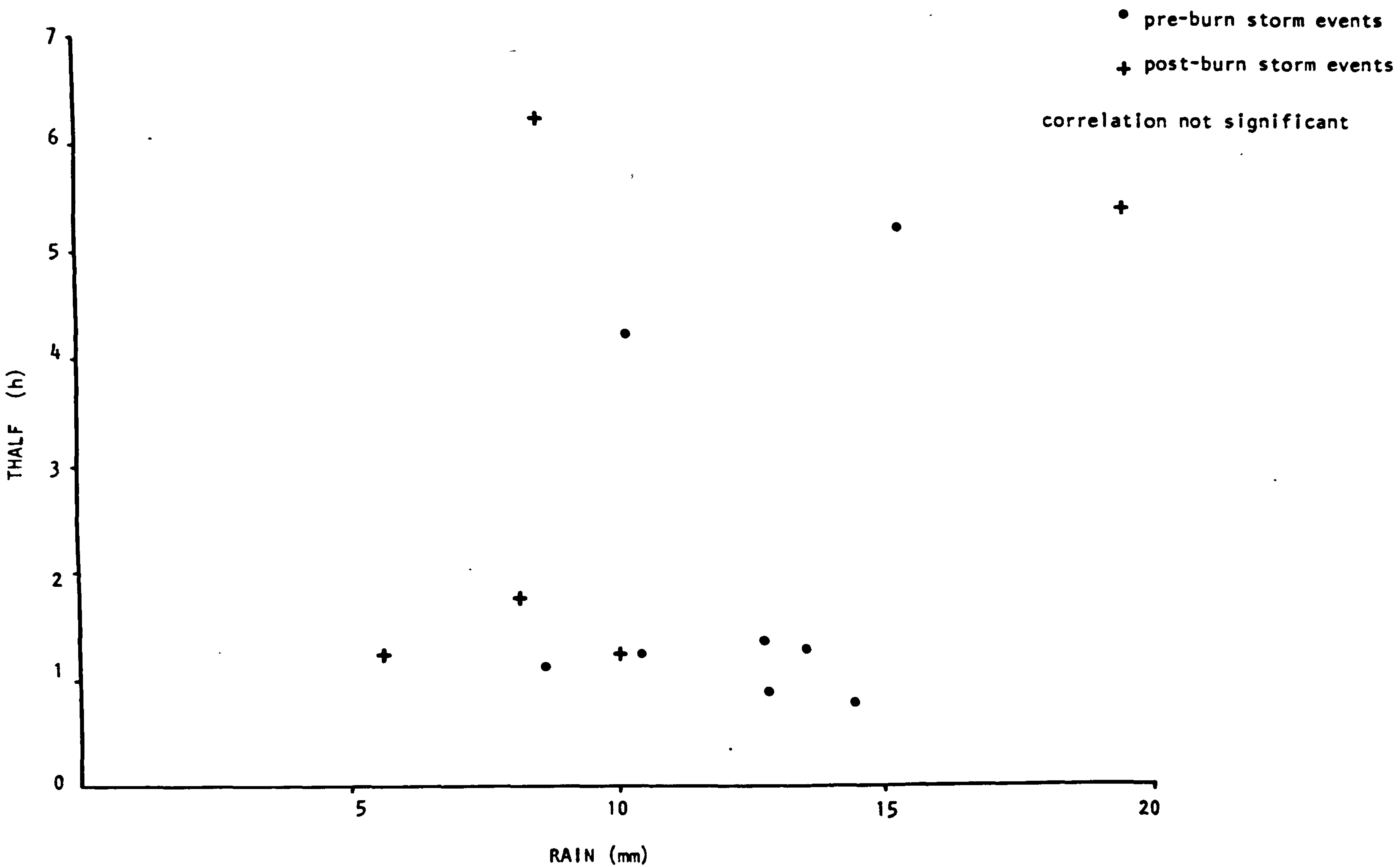


Figure 5.37 Relationship between Hydrograph Time to Half-Peak (THALF) and Rainfall Duration (DUR) Showing Clustering



<u>Figure 5.38 Relationship between Hydrograph Time to Half-Peak (THALF) and Gross Rainfall (RAIN) Showing Scatter</u>

antecedent catchment wetness, may be forthcoming with further observations pertaining to soil moisture deficits exceeding zero (Fig.5.39).

Multiple regression models are used below in conjunction with plots of 'average' unit hydrographs to yield final conclusions regarding predicted hydrograph shape for vegetated and burnt moorland areas. Although the precision of regression predictions may be limited, when used in combination with average unit hydrographs which give consistent results, they do permit valid general conclusions to be drawn. As there is no exclusive definition of an 'average' unit hydrograph, no universally acceptable method of obtaining an average plot is available (Boorman and Reed, 1981). Average changes in hydrograph dimensions with land-use are determined here by plotting average times to peak (TP) and peak discharges (QP) for pre- and post-burn events separately, and then aligning each set of hydrographs with their peaks coincident at the average point. Average values for the ordinates on either side of the peak then generate the typical unit hydrograph (N.E.R.C., 1975) (Fig.5.40). This 'peaks aligned' technique counteracts the tendency to underestimate individual unit hydrograph peaks which is characteristic of the Flood Studies method of unit hydrograph derivation as used here. Use of median rather than mean values to determine averages minimises the effect of outlying points and maintains the characteristic unit hydrograph shape. Although the technique may also yield an average hydrograph with a volume of less than unity (Boorman and Reed, 1981) this was not found to be the case in the present study, where the post-burn hydrograph overestimates unit volume by 1% and the pre-burn plot underestimates by only 9%. The number of unit hydrographs used to derive

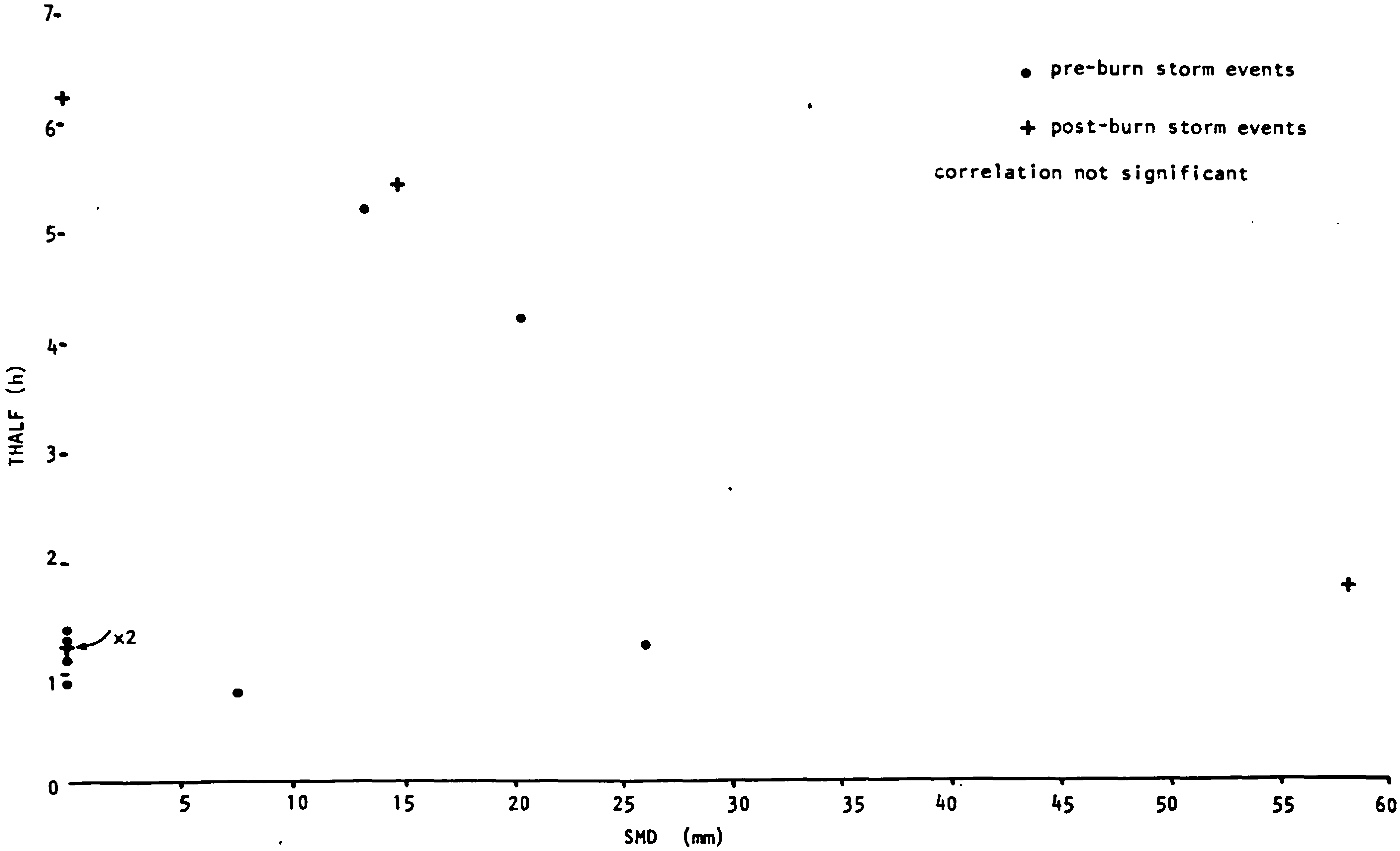


Figure 5.39 Effect of Soil Moisture Deficit (SMD) on Hydrograph Time to Half-Peak (THALF) (Not significant)

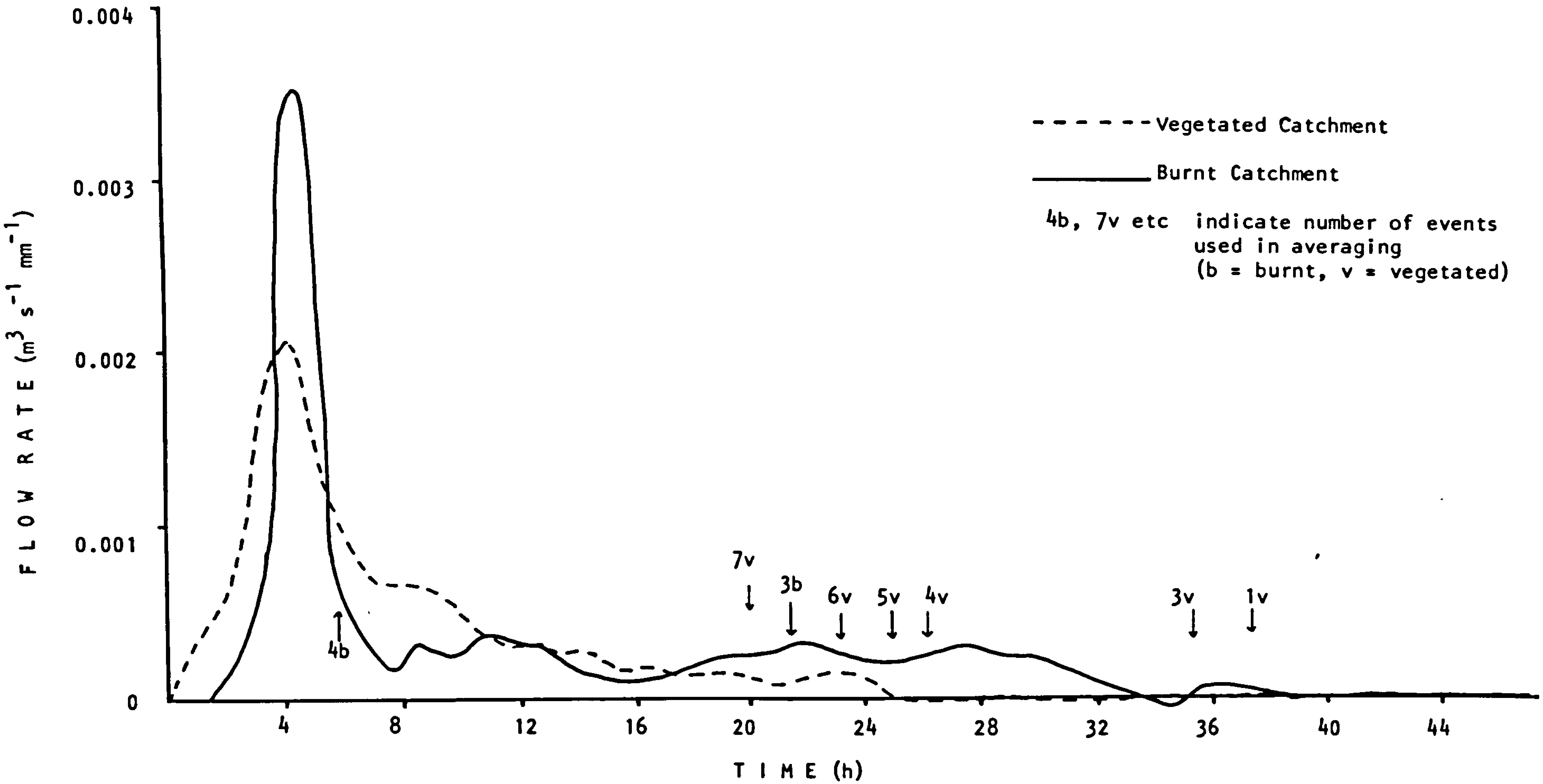


Figure 5.40 Median Peaks Aligned Average Unit Hydrographs (All events used in averaging for initial time periods except where indicated)

the median values is indicated on Figure 5.40.

By substituting values for both mean SMD and mean total rainfall for the storm events studied, and for the dummy variable, Equation 5.14 may be used to predict peak discharge ($m^3\ s^{-1}\ mm^{-1}$) for a comparable storm under different land-use conditions. To predict the dependent variable from independent values not present in the original sample, Johnston (1978) recommended inclusion of the 'standard error of the forecast'. This comprises the standard error of the estimate, given in the equations in Table 5.5, the standard error of the mean of the dependent variable and the standard error of the regression coefficient. The last gives a zero result for mean values of independent variables and is therefore eliminated in the present case, while the error of the mean in the *y* variable is so small for QP that the final standard error of the forecast is equivalent to the standard error of the estimate. Thus, for mean SMD = 10.82 mm, mean RAIN = 11.49 mm,

Vegetated catchment (DUMMY = 1):

$$\widehat{QP} = -0.0013 + 0.00012(10.82) + 0.00036(11.49) - 0.00191 \pm 0.00159$$

$$= \underline{0.00223 \pm 0.00159}$$

$$\therefore \underline{\widehat{QP} = 0.000638 \text{ to } 0.00381}\ m^3\ s^{-1}\ mm^{-1}$$

Burnt catchment (DUMMY = 0):

$$\underline{\widehat{QP} = 0.00414 \pm 0.00159}$$

$$\therefore \underline{\widehat{QP} = 0.00255 \text{ to } 0.00573}\ m^3\ s^{-1}\ mm^{-1}$$

Although the relatively large standard error promotes a region of overlap between the two predicted peak discharge values, average QP for the post-burn catchment remains almost twice that for a totally vegetated area.

Since the regression model for recession length (Eq. 5.15) excludes the land-use variable, DUMMY, perhaps because of the small number of cases, no definitive predictions are valid for RECL. The effects of vegetation removal on hydrograph recession length may have been masked by variability within each of the two storm groups and more observations may enable distinction of differences due to vegetation burning alone, since DUMMY is the next variable to enter the model. General explanations for differences between pre- and post-burn hydrographs are offered below, however, in the light of conclusions drawn thus far.

c) Land-Use Effects on the Storm Hydrograph

Unit hydrograph analysis has disclosed a number of important differences between runoff from vegetated, and from burnt moorland surfaces on the Egton catchment. Post-burn hydrographs are generally characterised by higher peak flows and narrower width prior to main recession than is found before burning. In contrast, the vegetated catchment typically yields a wider hydrograph with reduced peak and shallower rising limb. Average times to peak are comparable for the two surfaces (4.12 h for pre-burn, 4.44 h for post-burn) while lag time was deemed indistinguishable from zero for almost every event. Although a reduced recession limb is indicated for the vegetated catchment (Fig.5.40) noise during hydrograph derivation and the reduced number of events included in averaging renders conclusions regarding the precise length of recessions somewhat speculative, as supported by the neutral results of Equation 5.15. In the light of results from soil moisture analyses, it is suggested that differences in rainfall losses to the soil outweigh any direct change in soil surface permeability following the muirburn, particularly in view of the generally small to moderate size of most storms, and because of a degree of protection which is afforded by remaining litter and vegetation on burnt ground. The possibility of surface compaction due to the impact of falling rain breaking down soil aggregates and closing soil voids, and thus reducing porosity and infiltration rates on the burnt surface, is therefore excluded as an explanation of rapid flow response and higher peak discharges, in line with earlier findings relating to subsurface flow. Rather it is suggested that the narrower width and higher peak of the typical post-burn unit hydrograph arise from rapid replenishment of soil profile deficits, resulting in a faster response as the moisture holding capacity of the soil is reduced and saturation is attained quickly. These findings conform with those of Lockwood and Venkatasawmy (1975) for a Pennine catchment under rough grazing, from which runoff volume was controlled by soil moisture state rather than by infiltration capacity.

Variable contributing areas and subsurface runoff are both sources of surface stormflow for the site, with subsurface flow forming perhaps the dominant mechanism. Reduced interception accentuates subsurface flow under the burnt catchment and this runoff component may make a more significant contribution to the hydrograph than under vegetated conditions. This may be interpreted to suggest that the slightly enhanced recession limb of the post-burn average hydrograph may indicate a slow release of water from the catchment, from a subsurface source. It is proposed, however, that evidence for this effect is inconclusive not only in view of difficulties in determining specific base lengths, but also in the light of unit hydrograph shape theory, by which base length is reduced with increasing peak discharge. Furthermore, although the potential for temporary storage and slow release of water is also indicated by hydrograph recessions remaining above the pre-storm discharge level, this is equally applicable to both pre- and post-burn events.

Greater rainfall losses on the characteristically higher deficit, pre-burn surface reduce peak flow and since vegetation generally restricts flow and hinders channel production and erosion (Penman, 1963), extended flow paths and travel times result in a shallower and wider hydrograph than that found after vegetation removal, regardless of catchment wetness. In contrast, exposure of rill channels through burning, and the concomitant increase in flow velocity which these channel systems generate, aid in accelerating the speed of response of the devegetated moor at Egton. It is therefore suggested that the significant effects of soil moisture deficit on both peak discharge and recession length referred to earlier (p.292), apply independently of land-use type, and underlie the main control of catchment response by rainfall replenishment of moisture deficits, as related to surface vegetation covering.

5.4 CONCLUSIONS

The overriding importance of specific components of the hydrological cycle and of physical catchment conditions, in their effects on runoff systems for the Egton headwater area, is demonstrated by the foregoing account. Variations in measured subsurface flow under three vegetation covers are explicable through differences in losses to interception and transpiration. At the scale of experiment considered here, this effect outweighs the expected trend of enhanced subsurface runoff in vegetated areas.

Effective use has been made here of the unit hydrograph approach to flood assessment. This technique has been used successfully for several decades, although modern computational proficiency has now surpassed the basic demands of the method and more realistic approaches, such as sophisticated simulation modelling are succeeding it. The method remains a useful tool when more sophisticated techniques are impractical however, yielding favourable and dependable results through a combination of automatic data analysis and human judgement (N.E.R.C., 1975). Change in soil moisture storage capacity with land-use is found to be the most significant factor in determining surface runoff hydrograph form at Egton, with storm rainfall amount as a secondary contributing agent. The most evident effect of controlled heather burning is that on hydrograph peak discharge, with a burnt surface displaying an overall quicker runoff response, a greater degree of sensitivity and thus a higher potential for resource loss than comparable heather-covered ground.

The present analysis has attempted to provide the most complete evaluation of the limited number of storm events available, although a larger data base would help to eliminate the effects of anomalous observations and, specifically, may clarify the significance of heather burning for the hydrograph recession. The regression models derived by the current investigation may be applicable to areas of similar physical and storm characteristics, as well as for the Egton site under simulated rainfall and moisture deficit conditions. In the next chapter those hydrological variables considered relevant here, are examined further in the context of a catchment water balance. In particular, the potential significance of the interception and transpiration components is pursued, in terms of the implications of land management patterns for magnitudes of catchment water use.

CHAPTER 6

TOWARDS A WATER BALANCE

6.1 INTRODUCTION AND OBJECTIVES

Water balance studies represent essential analytical techniques in elucidating the importance of different components in the hydrological regime of an area. Such an approach enables quantitative determination of both the water resources of an area and the potential effects of land-use change (Sokolov and Chapman, 1974) by generating information on the timing and magnitude of potential surpluses and deficits (Thornthwaite and Mather, 1955).

The assessment of hydrological 'losses' under different land-use conditions forms an underlying theme throughout this dissertation. Using a 'water-balance-accounting' procedure for the soil profile, soil moisture models were applied in Chapter 4 to estimate daily soil moisture deficits, providing some indication of actual evaporative loss for woodland and both vegetated and burnt moorland, and forming a basis for the explanation of vegetation-induced differences in runoff hydrographs in Chapter 5. The present chapter attempts to consolidate and evaluate further some of these findings. Soil moisture budget models are reassessed in terms of their ability to predict accurately 'observed' evapotranspiration levels in the catchment, and the results are reviewed in the light of conclusions already drawn for model accuracy in soil moisture deficit prediction. The consequences of changing land-use for the evapotranspiration component are examined, especially with reference to the relative importance of its two constituent aspects, evaporation of intercepted water and transpiration.

Water balance components are determined on a monthly basis for the moorland area as a whole, yielding a single annual budget. Values of precipitation, runoff and changes in soil moisture storage are determined from measurements, while actual evapotranspiration is calculated as the remaining 'unknown' term in the equation. Ratios of actual to potential evapotranspiration are also discussed in the context of prevailing land-use, particularly with respect to crop wetness and resistances.

6.2 THE WATER BALANCE EQUATION

6.2.1 DEVELOPMENT OF AN EQUATION

Synthesis of a water balance involves employment of the Continuity Equation to assess the equilibrium between water input and output (Sokolov and Chapman, 1974). The equation represents an integration of relationships between the major elements of the hydrological cycle over a finite time period. Since in the present analysis inflow and outflow are balanced on a small catchment (headwater) scale most references in the current chapter are to catchment-based water balances. Ideally, every component of the water budget should be measured or calculated, using independent checks wherever possible in order to verify the accuracy of the balance, although where this is not practical one constituent can be assessed by an accounting process of elimination.

The type of water balance equation adopted reflects both the characteristics of the area under study and the time period involved. For records consisting of several years' data the water balance may be simplified as a summation of mean annual runoff and mean annual evapotranspiration equating with mean annual precipitation. An estimate of change in water storage should additionally be included for shorter period balances (annual, seasonal or monthly), the importance of this component increasing with reduced time scales. Although the storage term is restricted for the present site to changes in soil moisture, it may in other cases include surface retention change (Pegg, 1970), changes in groundwater storage and river channel storage in large basins.

The water balance of the Egton catchment is evaluated in terms of measured components and a single unknown term, evapotranspiration, which is calculated as follows:

$$AE = P - Q \pm \Delta S \qquad \text{Eq 6.1}$$

where:

AE = actual evapotranspiration

P = precipitation

Q = stream runoff

ΔS = changes in soil moisture storage

Variables are represented as an equivalent depth of water over the catchment (mm) although volume (m^3) or flow rate (m^3s^{-1}) are suitable alternatives. Estimation of evapotranspiration as the residual term in the water balance equation is often adopted in this way for small catchment experiments, the underlying philosophy being that this would be the least accurately measured component of the equation and, following establishment of a well-instrumented and non-leaking catchment, satisfactory results may be obtained by this means (Tang and Ward, 1982). Due to inherent difficulties in estimating the soil moisture storage term, annual balances often commence at a time of minimum storage in order to reduce measurement errors. December 1980 is chosen for the beginning of the budget in this study. Balances are derived for monthly periods, defined by availability of data, to the end of 1981, while summation provides an annual budget. Short-term balances of between two and five weeks are also incorporated, as dictated by data availability and reliability. Lack of runoff data for the woodland area confines calculations to the moorland zone, and conclusions made with respect to an anticipated situation under woodland are therefore drawn from the results of other related work. As the study period includes a change in vegetation cover, the relative magnitudes of annual evapotranspiration and runoff may differ from those expected had land-use remained constant. All other analyses however, such as monthly balances and actual/potential evapotranspiration relationships, which comprise the major part of the chapter, do account for the change in vegetation.

Undetected leakage of water either into or from a study area represents a potentially significant error in water balance determination and it is therefore important to ascertain that the catchment is watertight. Paired catchments can be used to overcome the leakage problem, since partial solutions of the water balance equation may be computed by combination. Alternatively, a leakage term may be specified in the equation itself (McGowan et al., 1980). Changes in hydraulic head which induce flow at the zone of leakage are, however, usually temporary and minor, and flow is generally directed to the stream channel prior to measurement (Hewlett et al., 1969) although cumulative errors may be significant. Soil Survey analysis of the present study area reveals an impermeable clay layer (Carroll and Bendelow, 1981) whose mineral composition, largely non-smectitic, precludes marked volumetric changes and hence the possibility of leakage cracks. The clay overlies shales, calcareous shales and fine silts, completing an impermeable seal to the catchment.

In view of errors involved in estimating each component of the balance, Sokolov and Chapman (1974) advocated additional inclusion of a residual 'discrepancy' or error term, in the catchment equation. Although the magnitude of this term should ideally remain low, this may merely indicate coincidental balancing of other components. In the present case, included errors, discussed below are absorbed by default into the single unknown term, 'AE'.

6.2.2 DATA CONSOLIDATION AND SOURCES OF ERROR

The individual terms of the water budget and their inherent errors are outlined briefly in the following sections in the specific context of water balance construction. Inaccuracies in determining these variables accrue from two general sources; the spatial and temporal variability of each component, and systematic and random errors in the measurement and calculation of components.

6.2.2.1 Precipitation (P)

Daily values monitored by the Sneaton automatic weather station are used to determine total precipitation, both for reasons discussed in the previous chapter (p.228) and because model-predicted values of evapotranspiration are determined on the basis of Sneaton High Moor precipitation data. A break in the record occurs at the end of 1981, lasting for a period of three weeks, and this effectively terminates the full water balance calculations. Shorter gaps of one or two days are filled on the assumption that a constant ratio is maintained with the Egton High Moor catch for the few days either side of the missing value, and where localised storms prevailed, Egton values were directly substituted for Sneaton data. In the light of other sources of error involved in precipitation measurement, it is not expected that discrepancies incurred during these alterations will have a significant effect on final values. In particular, accuracy of rainfall estimation depends largely on gauge design and exposure, and errors from this source should be minimal as ground-level gauges, used as part of the automatic monitoring scheme, give the best estimates (Rodda, 1967). Extrapolation from point measurements to areal assessment also requires the gauge site to be representative of the wider area, although, as discussed in Section 4.3.4.3, no significant errors are expected to arise from this source. Generally, an attempt has been made to provide a realistic estimate of precipitation without invalidating predicted values of actual evapotranspiration.

6.2.2.2 Runoff (Q)

The procedure for obtaining stream stage values and converting to discharge and equivalent runoff depth follows that described in the previous chapter for records used in hydrograph analysis. For the purposes of water balance calculations, stage values are defined for hourly intervals. Exact specification of this component poses the least difficulties, since it is usually the most accurately measured. Care must be taken however, to avoid leaks past the gauging structure and sediment accumulation upstream (Edwards, 1970) although leaks are reduced by effective use of well-built structures (Hewlett *et al*., 1969). Degrees of expected accuracy have been discussed in Chapter 3 and depend on factors such as reliablity of flow measurement, flow variability and the length of period under consideration (Sokolov and Chapman, 1974).

6.2.2.3 Moisture Storage (ΔS)

Confinement of the study area by a clay seal limits the most significant changes in water storage to those occurring within the soil layer. Although, ideally, changes should be monitored for the complete soil profile, to the water table, or to the deepest wetting front, measurements made in the upper 1 m, the upper rooting zone, give an approximate assessment of soil moisture content (Sokolov and Chapman, 1974). Measurements of soil moisture in the top 80 cm (Chapter 4) are therefore assumed here to account for the essential changes in storage and these readings are confined to the zone above the base of the clay seal. Such changes are calculated from differences in moisture content between the beginning and end of each accounting period, as determined from neutron probe measurements. Many of the problems previously inherent in assessing changes in soil moisture content have been reduced by the advent of the neutron probe technique, which facilitates measurement replication over space and time. Representative measuring sites are still required however, although care in the installation of access tubes minimises bias and enhances validity of results. Particular attention was given both to tube installation and depth relocation of the probe during measurement, since for water balance analyses, errors incurred from negligence during these procedures may prove more serious than those originating from probe calibration (McGowan and Williams, 1980). Changes in soil moisture storage under examination here, however, are generally small in relation both to other facets of the water balance and to total profile moisture content. The maximum monthly change of 46 mm during September 1981, for example, represents only about 10% of total moisture content.

6.2.2.4 Actual Evapotranspiration (AE)

Six types of actual evapotranspiration estimate are calculated by summation of the daily values predicted by soil moisture models and each is compared with the water balance-derived estimate. Due to model constraints, actual evapotranspiration is predicted for one complete year only (1981). Each estimate is defined by model type, potential evaporation function and assumed maximum extraction depth (total or layer moisture deficits) and is explained in a later section (6.3.2). Daily values of both Penman and Penman-Monteith potential evaporation estimates are determined as described in Chapter 4. As with changes in soil moisture storage, values of model-predicted evapotranspiration and Penman-Monteith potential evapotranspiration are computed for the post-burn period on the basis of weighted areal means for the two moorland zones, burnt and vegetated.

In conjunction with errors encountered in the measurement of hydrological variables, as considered above, misjudgements in determining catchment characteristics may also affect water balance results. Accurate assessment of catchment area, for example, is an important prerequisite to reliable runoff estimation and depends ultimately on rigorous surveying of the catchment boundary. In order to minimise errors in water balance construction, it is necessary firstly, to establish a well-instrumented catchment, with representative measurement sites, both in number and location (Chapter 3). Secondly, the specific difficulties involved with short time periods should be recognised, since the shorter the time interval, the more precisely should each of the hydrological components be evaluated (Sokolov and Chapman, 1974). Thirdly, reliable determination of one component as the residual term of the water budget depends on the accuracy of the remaining elements. Thus, in the present instance, water balance-derived values of actual evapotranspiration incorporate an error constituent, possibly including an amount of water not explained by the balance equation employed. The calculated component therefore represents more of a relative than an absolute definition.

In order to avoid such errors exceeding the magnitude of the residual component itself, care should be exercised to ensure accurate measurement of the other components. Finally, it may be possible to verify calculated water budgets where all components are measured. Thus, Ward (1972) used calculations over different time lengths within the same run of data in an attempt to eliminate coincidental balancing of measurement discrepancies, and thereby to determine calculation accuracy and check for catchment leaks.

6.3 RESULTS AND INTERPRETATION

Annual and monthly water budgets for Egton are shown in Table 6.1 and Figure 6.1. Both total and mean daily values of each component are calculated and values of potential evapotranspiraton are included for comparison. Predicted estimates of actual evapotranspiration for the woodland are given in Appendix VI.

6.3.1 ANNUAL AND SEASONAL WATER BALANCES

The annual water budget covers virtually a complete calendar year, from 17 December 1980 to 8 December 1981, the latter date being selected in view of missing automatic weather station data for the last three weeks of December 1981. As soil moisture model-predicted estimates of actual evapotranspiration are unavailable for the latter part of December 1980, and soil moisture data are missing for the first fortnight of the following year, however, a second 'annual' balance was also calculated for the period between 14 January 1981 and 8 December 1981. The results allow comparison between total predicted values of actual evapotranspiration (monthly and annual) and those derived by difference from the water balance, and form the basis of discussion in Section 6.3.2. The five-day period between 23 July and

PERIOD AND DAY NUMBER (approx. 28 days unless indicated)	P	Q	ΔS	AE	PE (PENMAN)	PE (PENMAN-MONTEITH)
26.11.80 - 16.12.80 331-351 [21]	85.0 (4.1)	74.2 (3.5)	-1.0 (-0.1)	11.8 (0.6)	5.6 (0.3)	19.6 (0.9)
17.12.80 - 13.1.81 352-13	44.0 (1.6)	30.7 (1.1)	+4.0 (+0.1)	9.3 (0.3)	6.4 (0.2)	15.7 (0.9)
14.1.81 - 10.2.81 14-41	28.5 (1.0)	23.6 (0.8)	-6.0 (-0.2)	10.9 (0.4)	7.5 (0.3)	10.8 (0.4)
11.2.81 - 9.3.81 42-68	69.5 (2.6)	48.1 (1.8)	+9.0 (+0.3)	12.4 (0.5)	12.3 (0.5)	13.3 (0.5)
10.3.81 - 7.4.81 69-97	134.5 (4.6)	114.0 (3.9)	-4.0 (-0.1)	24.5 (0.9)	28.0 (1.0)	22.0 (0.8)
8.4.81 - 5.5.81 98-125	54.0 (1.9)	35.0 (1.3)	+0.4 (+0.01)	18.6 (0.7)	56.7 (2.0)	35.1 (1.3)
6.5.81 - 2.6.81 126-153	53.5 (1.9)	36.3 (1.3)	-9.7 (-0.4)	26.9 (1.0)	89.1 (3.2)	55.3 (2.0)
3.6.81 - 1.7.81 154-182	34.5 (1.2)	3.1 (0.1)	-29.2 (-1.0)	60.6 (2.1)	92.1 (3.3)	56.2 (1.9)
2.7.81 - 22.7.81 183-203 [21]	42.0 (2.0)	3.8 (0.2)	-2.0 (-0.1)	40.2 (1.9)	60.5 (2.9)	40.0 (1.9)
<Q Data missing>						
28.7.81 - 2.9.81 209-245 [37]	51.0 (1.4)	14.8 (0.4)	-30.2 (-0.8)	66.4 (1.8)	111.8 (3.0)	72.7 (2.0)
3.9.81 - 29.9.81 246-272	112.0 (4.2)	31.1 (1.1)	+46.0 (+1.7)	34.9 (1.3)	54.9 (2.0)	41.7 (1.5)
30.9.81 - 27.10.81 273-300	84.5 (3.0)	50.5 (1.8)	+6.2 (+0.2)	27.8 (1.0)	25.8 (0.9)	22.2 (0.8)
28.10.81 - 24.11.81 301-328	51.5 (1.8)	32.5 (1.2)	-1.3 (-0.1)	20.3 (0.7)	12.0 (0.4)	10.5 (0.4)
25.11.81 - 8.12.81 329-342 [14]	14.5 (1.0)	16.4 (1.2)	-5.1 (-0.4)	3.2 (0.2)	4.0 (0.3)	1.5 (0.1)
<AWS data missing>						
20.1.82 - 2.2.82 20-33 [14]	18.0 (1.3)	32.0 (2.3)	-	-	8.0 (0.6)	-
3.2.82 - 2.3.82 34-61	49.5 (1.8)	8.9 (0.3)	-	-	18.0 (0.6)	-
ANNUAL 14.1.81 - 8.12.81	730.0	409.2	-17.9	346.7	554.7	381.3
17.12.80 - 8.12.81	774.0	439.9	-21.9	356.0	561.1	397.0

AWS automatic weather station
- data unavailable

Table 6.1 Water Balance for Egton Catchment
All figures are in mm except those in round brackets, which indicate daily mean equivalents (mm day $^{-1}$)

(approx. 28 days unless indicated)		Q	ΔS	AE	PE (PENMAN)	PE (PENMAN-MONTEITH)
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11.2.81 - 9.3.81 42-68	69.5 (2.6)	48.1 (1.8)	+9.0 (+0.3)	12.4 (0.5)	12.3 (0.5)	13.3 (0.5)
10.3.81 - 7.4.81 69-97	134.5 (4.6)	114.0 (3.9)	-4.0 (-0.1)	24.5 (0.9)	28.0 (1.0)	22.0 (0.8)
8.4.81 - 5.5.81 98-125	54.0 (1.9)	35.0 (1.3)	+0.4 (+0.01)	18.6 (0.7)	56.7 (2.0)	35.1 (1.3)
6.5.81 - 2.6.81 126-153	53.5 (1.9)	36.3 (1.3)	-9.7 (-0.4)	26.9 (1.0)	89.1 (3.2)	55.3 (2.0)
3.6.81 - 1.7.81 154-182	34.5 (1.2)	3.1 (0.1)	-29.2 (-1.0)	60.6 (2.1)	92.1 (3.3)	56.2 (1.9)
2.7.81 - 22.7.81 183-203 [21]	42.0 (2.0)	3.8 (0.2)	-2.0 (-0.1)	40.2 (1.9)	60.5 (2.9)	40.0 (1.9)
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14.1.81 - 10.2.81 14-41	28.5 (1.0)	23.6 (0.8)	-6.0 (-0.2)	10.9 (0.4)	7.5 (0.3)	10.8 (0.4)
11.2.81 - 9.3.81 42-68	69.5 (2.6)	48.1 (1.8)	+9.0 (+0.3)	12.4 (0.5)	12.3 (0.5)	13.3 (0.5)
10.3.81 - 7.4.81 69-97	134.5 (4.6)	114.0 (3.9)	-4.0 (-0.1)	24.5 (0.9)	28.0 (1.0)	22.0 (0.8)
8.4.81 - 5.5.81 98-125	54.0 (1.9)	35.0 (1.3)	+0.4 (+0.01)	18.6 (0.7)	56.7 (2.0)	35.1 (1.3)
6.5.81 - 2.6.81 126-153	53.5 (1.9)	36.3 (1.3)	-9.7 (-0.4)	26.9 (1.0)	89.1 (3.2)	55.3 (2.0)
3.6.81 - 1.7.81 154-182	34.5 (1.2)	3.1 (0.1)	-29.2 (-1.0)	60.6 (2.1)	92.1 (3.3)	56.2 (1.9)
2.7.81 - 22.7.81 183-203 [21]	42.0 (2.0)	3.8 (0.2)	-2.0 (-0.1)	40.2 (1.9)	60.5 (2.9)	40.0 (1.9)
<Q Data missing>						
28.7.81 - 2.9.81 209-245 [37]	51.0 (1.4)	14.8 (0.4)	-30.2 (-0.8)	66.4 (1.8)	111.8 (3.0)	72.7 (2.0)
3.9.81 - 29.9.81 246-272	112.0 (4.2)	31.1 (1.1)	+46.0 (+1.7)	34.9 (1.3)	54.9 (2.0)	41.7 (1.5)
30.9.81 - 27.10.81 273-300	84.5 (3.0)	50.5 (1.8)	+6.2 (+0.2)	27.8 (1.0)	25.8 (0.9)	22.2 (0.8)
28.10.81 - 24.11.81 301-328	51.5 (1.8)	32.5 (1.2)	-1.3 (-0.1)	20.3 (0.7)	12.0 (0.4)	10.5 (0.4)
25.11.81 - 8.12.81 329-342 [14]	14.5 (1.0)	16.4 (1.2)	-5.1 (-0.4)	3.2 (0.2)	4.0 (0.3)	1.5 (0.1)
<AWS data missing>						
20.1.82 - 2.2.82 20-33 [14]	18.0 (1.3)	32.0 (2.3)	-	-	8.0 (0.6)	-
3.2.82 - 2.3.82 34-61	49.5 (1.8)	8.9 (0.3)	-	-	18.0 (0.6)	-
ANNUAL						
14.1.81 - 8.12.81	730.0	409.2	-17.9	346.7	554.7	381.3
17.12.80 - 8.12.81	774.0	439.9	-21.9	356.0	561.1	397.0

AWS automatic weather station
- data unavailable

Table 6.1 Water Balance for Egton Catchment

All figures are in mm except those in round brackets, which indicate daily mean equivalents (mm day $^{-1}$)

(approx. 28 days unless indicated)					(PENMAN)	(PENMAN-MONTEITH)
26.11.80 - 16.12.80 331-351 [21]	85.0 (4.1)	74.2 (3.5)	-1.0 (-0.1)	11.8 (0.6)	5.6 (0.3)	19.6 (0.9)
17.12.80 - 13.1.81 352-13	44.0 (1.6)	30.7 (1.1)	+4.0 (+0.1)	9.3 (0.3)	6.4 (0.2)	15.7 (0.9)
14.1.81 - 10.2.81 14-41	28.5 (1.0)	23.6 (0.8)	-6.0 (-0.2)	10.9 (0.4)	7.5 (0.3)	10.8 (0.4)
11.2.81 - 9.3.81 42-68	69.5 (2.6)	48.1 (1.8)	+9.0 (+0.3)	12.4 (0.5)	12.3 (0.5)	13.3 (0.5)
10.3.81 - 7.4.81 69-97	134.5 (4.6)	114.0 (3.9)	-4.0 (-0.1)	24.5 (0.9)	28.0 (1.0)	22.0 (0.8)
8.4.81 - 5.5.81 98-125	54.0 (1.9)	35.0 (1.3)	+0.4 (+0.01)	18.6 (0.7)	56.7 (2.0)	35.1 (1.3)
6.5.81 - 2.6.81 126-153	53.5 (1.9)	36.3 (1.3)	-9.7 (-0.4)	26.9 (1.0)	89.1 (3.2)	55.3 (2.0)
3.6.81 - 1.7.81 154-182	34.5 (1.2)	3.1 (0.1)	-29.2 (-1.0)	60.6 (2.1)	92.1 (3.3)	56.2 (1.9)
2.7.81 - 22.7.81 183-203 [21]	42.0 (2.0)	3.8 (0.2)	-2.0 (-0.1)	40.2 (1.9)	60.5 (2.9)	40.0 (1.9)
<Q Data missing>						
28.7.81 - 2.9.81 209-245 [37]	51.0 (1.4)	14.8 (0.4)	-30.2 (-0.8)	66.4 (1.8)	111.8 (3.0)	72.7 (2.0)
3.9.81 - 29.9.81 246-272	112.0 (4.2)	31.1 (1.1)	+46.0 (+1.7)	34.9 (1.3)	54.9 (2.0)	41.7 (1.5)
30.9.81 - 27.10.81 273-300	84.5 (3.0)	50.5 (1.8)	+6.2 (+0.2)	27.8 (1.0)	25.8 (0.9)	22.2 (0.8)
28.10.81 - 24.11.81 301-328	51.5 (1.8)	32.5 (1.2)	-1.3 (-0.1)	20.3 (0.7)	12.0 (0.4)	10.5 (0.4)
25.11.81 - 8.12.81 329-342 [14]	14.5 (1.0)	16.4 (1.2)	-5.1 (-0.4)	3.2 (0.2)	4.0 (0.3)	1.5 (0.1)
<AWS data missing>						
20.1.82 - 2.2.82 20-33 [14]	18.0 (1.3)	32.0 (2.3)	-	-	8.0 (0.6)	-
3.2.82 - 2.3.82 34-61	49.5 (1.8)	8.9 (0.3)	-	-	18.0 (0.6)	-
ANNUAL 14.1.81 - 8.12.81 17.12.80 - 8.12.81	 730.0 774.0	 409.2 439.9	 -17.9 -21.9	 346.7 356.0	 554.7 561.1	 381.3 397.0

AWS automatic weather station
\- data unavailable

Table 6.1 Water Balance for Egton Catchment
All figures are in mm except those in round brackets, which indicate daily mean equivalents (mm day $^{-1}$)

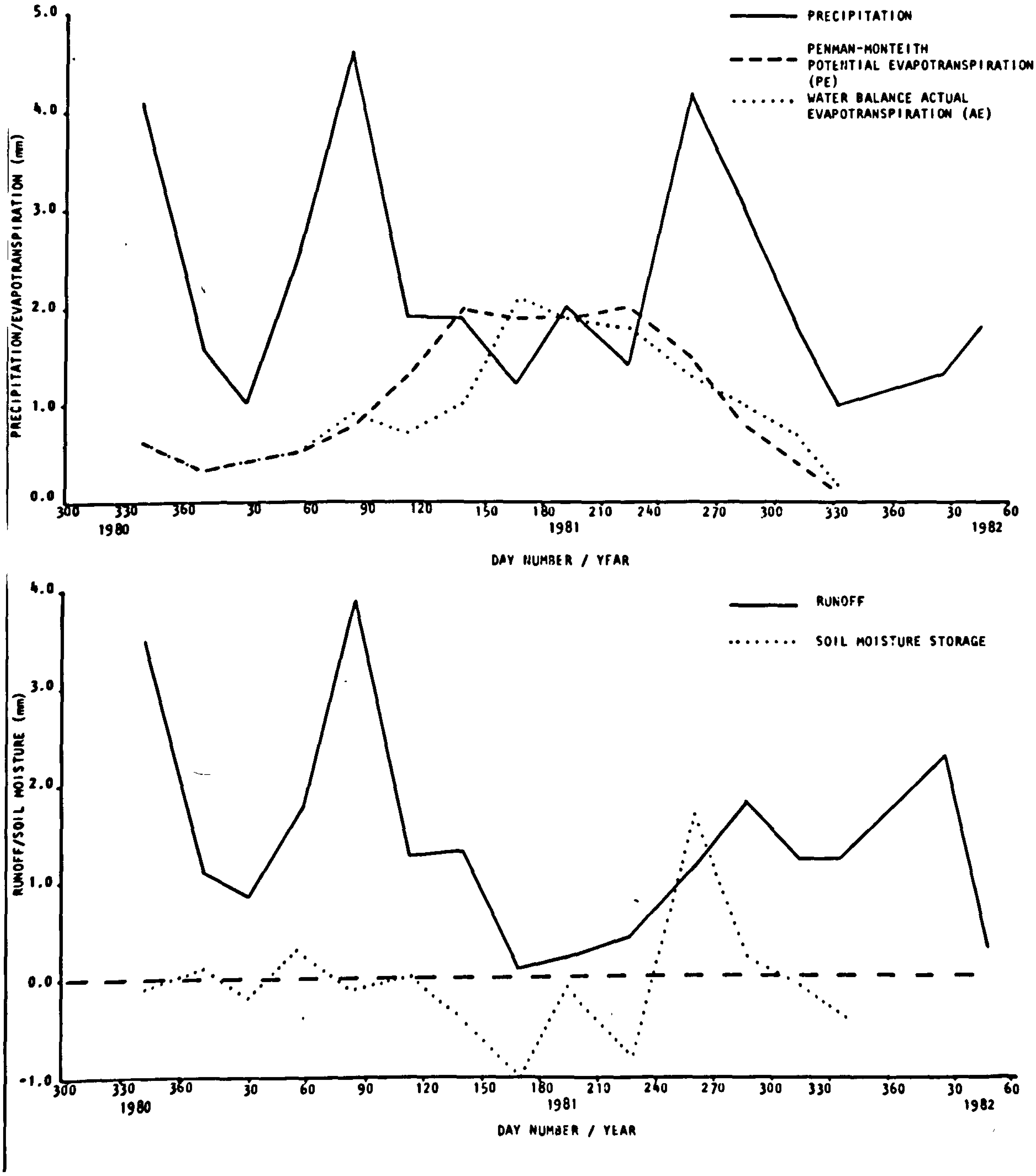


Figure 6.1 Egton Catchment Water Balance (Mean Daily Values)

27 July 1981 (days 204 to 208) is omitted from all calculations, since although records of rainfall for this time are complete for the Sneaton weather station (76 mm measured), comparative measurements from Egton are unavailable for confirmation, and a gap is also present in the runoff record.

Over the period 17 December 1980 to 8 December 1981 annual rainfall is fairly evenly distributed in its contributions to runoff and actual evapotranspiration, with 57% being accounted for by stream runoff and 46% by evapotranspiration as derived from the water balance. There is an overall reduction in soil moisture storage of approximately 3% (22 mm) over the year, while the difference between maximum and minimum storage (February/March to August) amounts to about 75 mm (Fig. 6.2).

Moisture storage starts to diminish during May, and shows more rapid reduction during the low rainfall months of June and August. Evapotranspiration exceeds rainfall during June, July and August, reaching its maximum for the year in June, with a mean of 2.1 mm day^{-1} . This period correspondingly marks both the time of minimum streamflow, mean daily runoff being less than 0.5 mm, and the main soil moisture deficit period. Typically, one-third of rainfall is lost to runoff during summer. Enhanced rainfall during September rapidly increases moisture storage, and greater proportions of rainfall are lost to runoff as soil moisture deficits are reduced throughout autumn and winter, at the expense of evapotranspiration rates (Fig.5.8). Evaporative losses reach minimum levels during December for both 1980 and 1981, and comprise only about 20% of rainfall.

__Figure 6.2 Soil Moisture Storage Variations (1981)__ (28-day periods unless indicated by bracketed figures)

6.3.2 ASSESSMENT OF ACTUAL EVAPOTRANSPIRATION

'Observed' catchment water balance data are used in this section to validate independent assessments of evapotranspiration simulated by soil moisture models. Discussion is confined to patterns of water use through evapotranspiration loss, since the constraints imposed on model estimation of runoff are considered too restricting to allow quantitative validation of this component. Thus, the Grindley model allows runoff to occur only when soil moisture deficit reaches zero. Under the assumptions of MORECS, drainage, which may comprise runoff, drainage to lower layers or direct groundwater recharge, may occur during deficit, but under the proviso that the top layer of the model is full and more than 10 mm of overflow occurs from it on any one day. Existing documented predictions of catchment runoff therefore vary in reliability. Davies (1981) showed that river flow response to rainfall in a Shropshire catchment failed to be mirrored in MORECS' predictions of soil moisture deficit, which remained moderately high despite attainment of field capacity. Effective rainfall, which largely equates with catchment runoff, was significantly underestimated by the model. An additional, catchment storage, model was engaged by Greenfield (1981) to allow comparison of generated and observed river hydrographs and, with the qualifications that the soil moisture model allowed for summer percolation and that a reliable drying curve was used, reasonable results were obtained.

In the present analysis predictions of annual evapotranspiration comply with the results of other studies in terms of direction of model prediction (Headworth, 1970; Kitching _et al_., 1977; Davies, 1981). Thus in all but one case, that determined by the MORECS model using 'layer' soil moisture deficits, actual evapotranspiration is overestimated by the models (Table 6.2). Calculation of soil moisture deficits from a specified field capacity value ('total' profile deficits) has already been shown to promote overestimation of true deficit, as a consequence of included drainage (Chapter 4). Moisture deficit determined in this way, therefore, is not a true indication of moisture loss from the profile by evaporation. In addition, inadequacies and misrepresentations by Penman estimates of potential evaporation contribute to exaggerated predictions of actual evapotranspiration in cases where this formula is applied.

Unlike studies such as those of Headworth and Kitching et al., however, model estimates in the present investigation are based on optimised parameters. Thus, having removed the potential deficiency of inaccurate parameter specification, comparisons can be synthesized to assess model performance and the sensitivity of input variables. Most annual predictions therefore constitute satisfactory estimates of catchment evapotranspiration (Table 6.2) and four of the six totals are within 5% of that derived by the water budget. The most successful type of annual estimate is that produced by the Grindley model using Penman-Monteith potential evapotranspiration data and layer soil moisture deficits, yielding a residual of only 4 mm, a 1% error (AE_4). In comparison, Kitching et al. (1977) discovered that, using the recommended root constant for grass, actual evaporation was overestimated by the Grindley calculation by 13% over a three-year period, although the authors maintained that this degree of error is not unreasonable for an evaporation estimate. Headworth (1970) concluded that with a recommended root constant the Grindley model overestimated mean annual actual evaporation from short-rooted vegetation by a margin of 7% over that using Headworth's advocated

PERIOD	AE	AE_1	AE_2	AE_3	AE_4	AE_5	AE_6
14.1.81 - 10.2.81	10.9 (0.4)	7.5 (0.3)	7.5 (0.3)	10.8 (0.4)	10.8 (0.4)	10.8 (0.4)	10.8 (0.4)
11.2.81 - 9.3.81	12.4 (0.5)	12.3 (0.5)	11.5 (0.4)	13.3 (0.5)	13.3 (0.5)	13.3 (0.5)	13.3 (0.5)
10.3.81 - 7.4.81	24.5 (0.9)	28.0 (1.0)	27.0 (0.9)	22.0 (0.8)	22.0 (0.8)	22.0 (0.8)	22.0 (0.8)
8.4.81 - 5.5.81	18.6 (0.7)	53.3 (1.9)	44.2 (1.6)	35.1 (1.3)	33.1 (1.2)	35.1 (1.3)	30.4 (1.1)
6.5.81 - 2.6.81	26.9 (1.0)	71.6 (2.6)	59.3 (2.1)	55.3 (2.0)	51.0 (1.8)	56.1 (2.0)	47.3 (1.7)
3.6.81 - 1.7.81	60.6 (2.1)	42.8 (1.5)	39.7 (1.4)	52.5 (1.8)	50.0 (1.7)	52.5 (1.8)	46.1 (1.6)
2.7.81 - 22.7.81	40.2 (1.9)	45.9 (2.2)	42.3 (2.0)	38.7 (1.8)	36.4 (1.7)	39.5 (1.9)	35.0 (1.7)
28.7.81 - 2.9.81	66.4 (1.8)	71.1 (1.9)	47.1 (1.3)	69.0 (1.9)	61.1 (1.7)	67.3 (1.8)	53.4 (1.4)
3.9.81 - 29.9.81	34.9 (1.3)	39.5 (1.5)	37.2 (1.4)	38.3 (1.4)	38.6 (1.4)	32.5 (1.2)	38.1 (1.4)
30.9.81 - 27.10.81	27.8 (1.0)	25.6 (0.9)	25.2 (0.9)	22.2 (0.8)	22.3 (0.8)	22.2 (0.8)	21.9(0.8)
28.10.81 - 24.11.81	20.3 (0.7)	11.9 (0.4)	12.1 (0.4)	10.5 (0.4)	10.7 (0.4)	10.5 (0.4)	10.5 (0.4)
25.11.81 - 8.12.81	3.2 (0.2)	4.0 (0.3)	4.0 (0.3)	1.5 (0.1)	1.5 (0.1)	1.5 (0.1)	1.5 (0.1)
ANNUAL 14.1.81 - 8.12.81	346.7	413.5	357.1	369.2	350.8	363.3	330.3

Potential Evapotranspiration/ Model Combination:-

AE Water balance actual evapotranspiration

AE_1 Penman/Grindley Total profile soil moisture deficits

AE_2 Penman/Grindley Layer soil moisture deficits

AE_3 Penman-Monteith/Grindley Total profile soil moisture deficits

AE_4 Penman-Monteith/Grindley Layer soil moisture deficits

AE_5 Penman-Monteith/MORECS Total profile soil moisture deficits

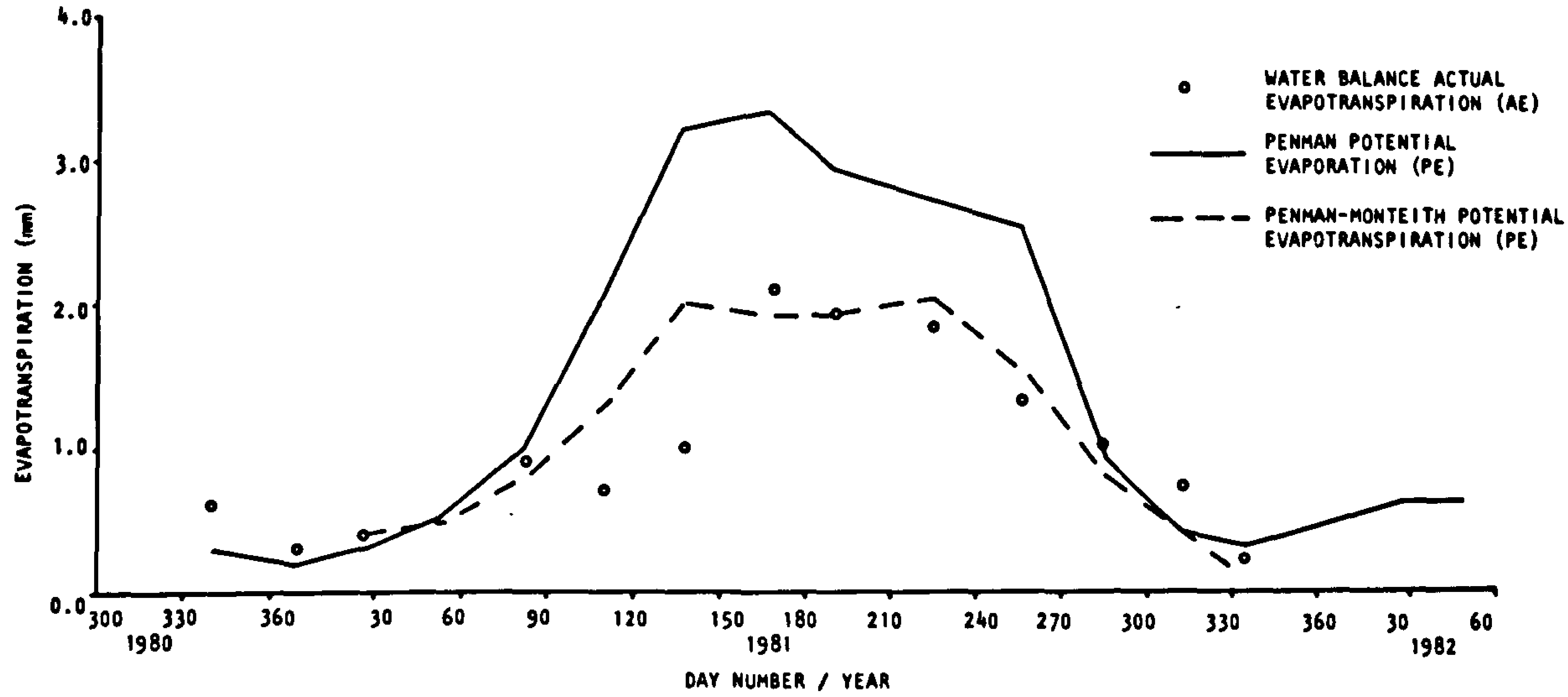
AE_6 Penman-Monteith/MORECS Layer soil moisture deficits

All figures in mm except those in brackets which indicate daily mean equivalents (mm day^{-1})

Table 6.2 Soil Moisture Model-Predicted and Water Balance Actual Evapotranspiration root constant. Nevertheless, even in the present study, accuracy of prediction varies with potential evaporation and moisture deficit functions. As may be expected from conclusions drawn earlier, the Grindley model prediction based on Penman evaporation and total profile deficits deviates the most from the water balance-derived value (the model overestimates the annual total by a margin of 19%). The nature of these influences is discussed below in the context of seasonal predictions of evapotranspiration, on the premise that single, annual estimates may be misleading.

General comparisons of seasonal variations in actual evapotranspiration estimates are illustrated by Figure 6.3. 'Observed' values match potential rates until March, and resume again in late September or early October. Catchment evapotranspiration is also close to Penman-Monteith potential during June, July and August. Instances of actual values exceeding potential demand may result from overestimated resistances in potential evapotranspiration calculations (Chapter 4), or from other sources such as measurement errors and inherent defects of evaporation formulae, discussed below. Actual:potential evapotranspiration ratios are evaluated in further detail, firstly, in the present context in explanations of discrepancies between observed and predicted values, and secondly, in the following section (6.3.3) in terms of the assumptions of the evaporation formulae in relation to the processes of transpiration and evaporation of intercepted water. The accuracy of predicted estimates of catchment evapotranspiration is shown qualitatively for monthly totals and daily mean values by Figures 6.4 to 6.6. Deviation from observed levels is quantified in terms of the root mean square error (RMSE), calculated as for soil moisture deficit predictions (Chapter 4, p.111) and illustrated in Table 6.3.

(a)



(b)

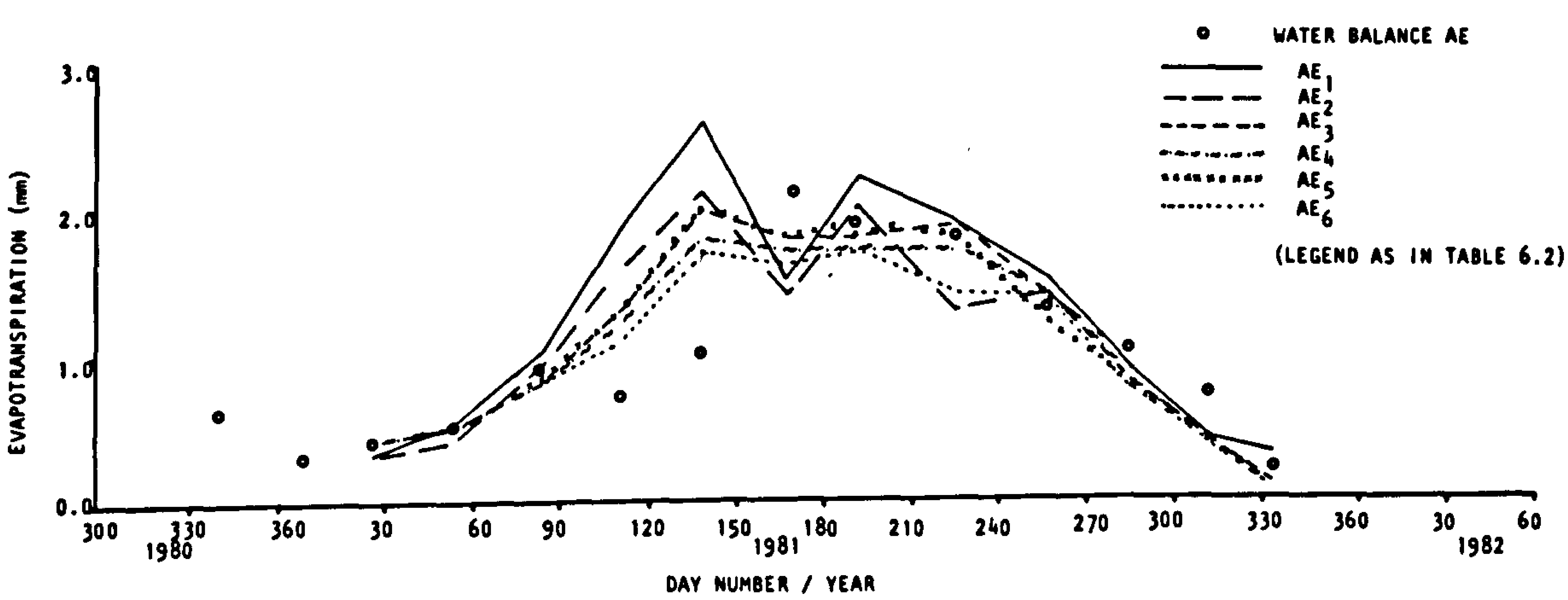


Figure 6.3 Comparisons of 'Observed' Actual Evapotranspiration with (a) Potential Demand and (b) Soil Moisture Model-Predicted Values (AE_1 to AE_6) (mean daily values)

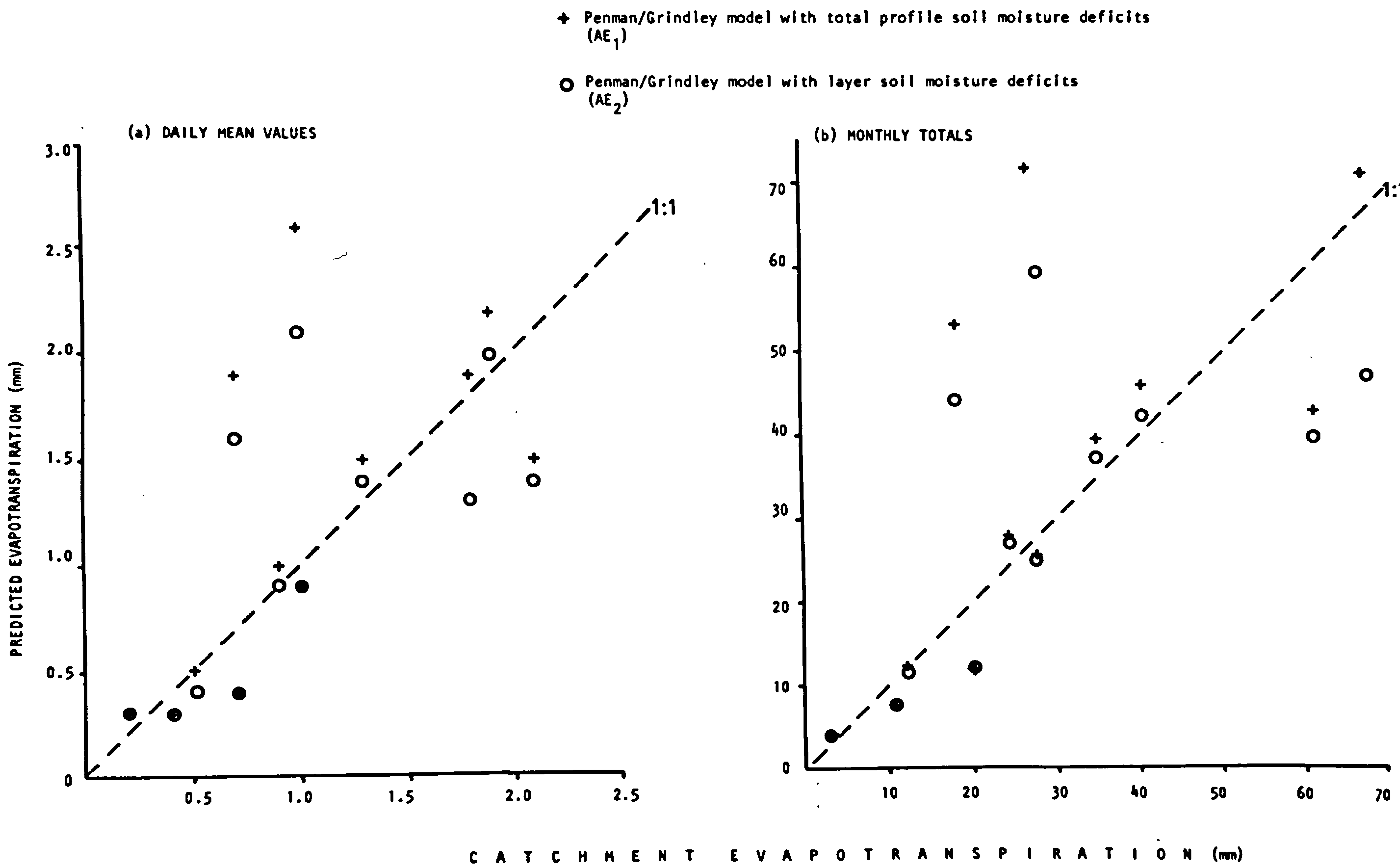


Figure 6.4 Soil Moisture Model Accuracy in Evapotranspiration Prediction (AE_1 and AE_2)

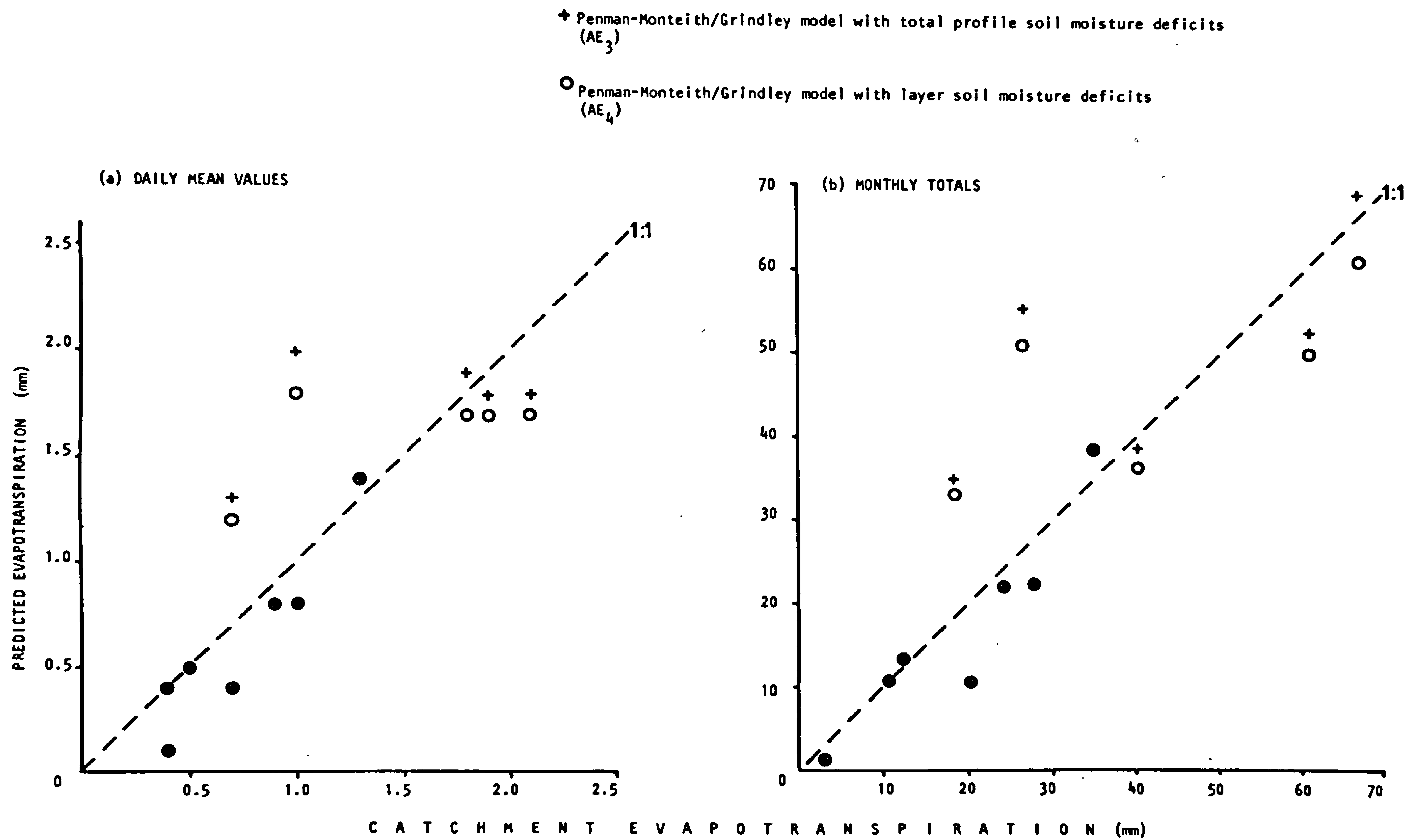


Figure 6.5 Soil Moisture Model Accuracy in Evapotranspiration Prediction (AE_3 and AE_4)

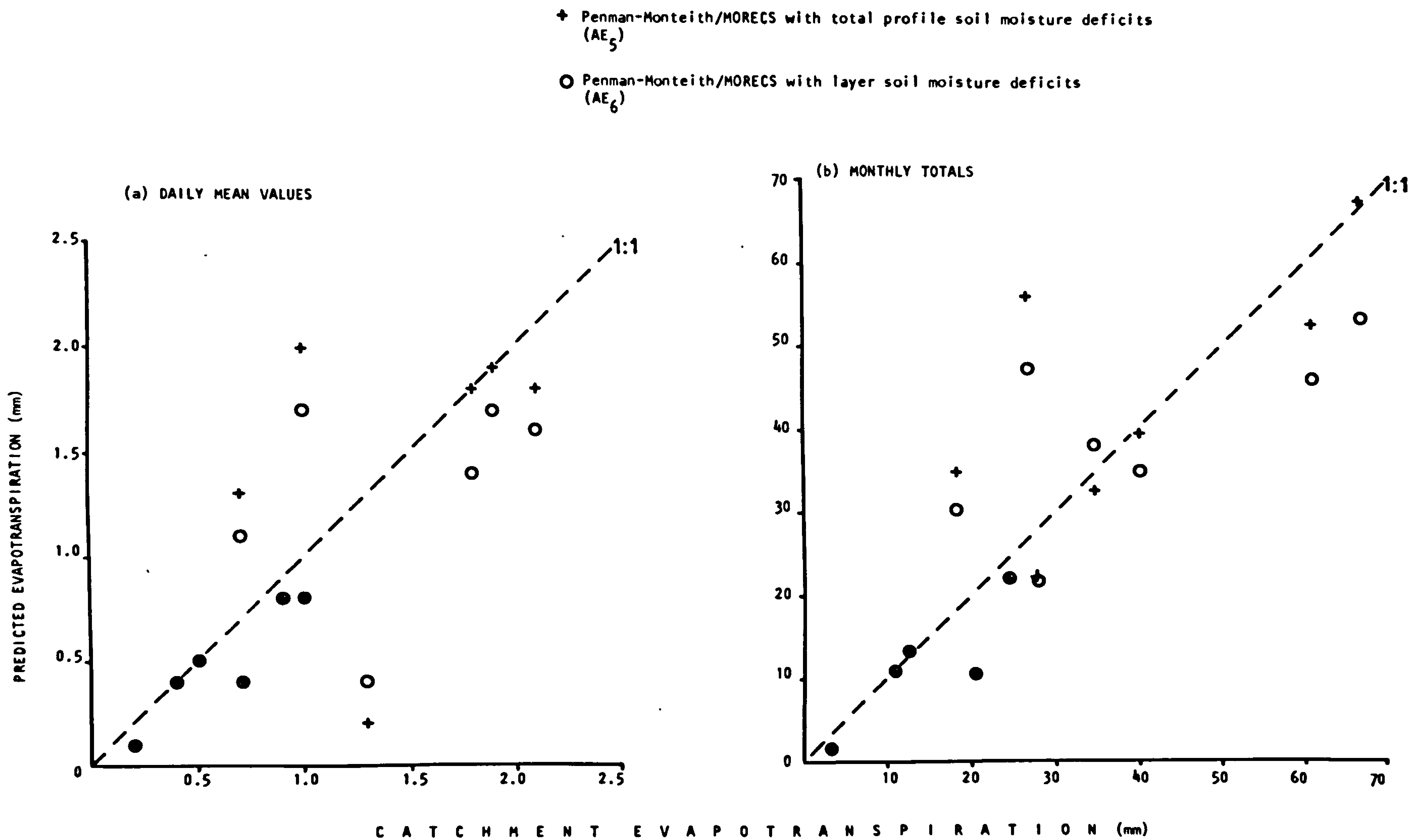


Figure 6.6 Soil Moisture Model Accuracy in Evapotranspiration Prediction (AE_5 and AE_6)

PREDICTED ACTUAL EVAPOTRANSPIRATION	ROOT MEAN SQUARE ERROR DAILY MEAN VALUES	MONTHLY TOTALS
AE_1		
(Penman PE/Grindley Model Total profile moisture deficits)	0.621	17.547
AE_2		
(Penman PE/Grindley Model Layer moisture deficits)	0.492	14.769
AE_3		
(Penman-Monteith PE/Grindley Model Total profile moisture deficits)	0.369	10.418
AE_4		
(Penman-Monteith PE/Grindley Model Layer moisture deficits)	0.324	9.539
AE_5		
(Penman-Monteith PE/MORECS Total profile moisture deficits)	0.366	10.547
AE_6		
(Penman-Monteith PE/MORECS Layer moisture deficits)	0.324	9.63

Table 6.3 Error Terms for Soil Moisture Model Estimates of Actual Evapotranspiration

The results signify the importance of several fundamental influences on catchment evapotranspiration prediction. Firstly, all estimates are subject to inaccuracies in the measured components rainfall, runoff and soil moisture characteristics. This factor pertains in a number of instances where discrepancies are found between 'water balance-derived' components and 'independent' estimates, as well as in cases where disagreements occur within a water balance and large residual errors accumulate.

'Model-predicted' evapotranspiration is to a large extent a function of potential demand, and the importance of selecting a suitable potential evapotranspiration (PE) formula is demonstrated by comparing predictions of the Penman PE/Grindley model combination (AE_1 and AE_2) with the remaining results. The former estimates prove the poorest overall (RMSE = 17.547, 14.769 for monthly estimates of AE_1 and AE_2, respectively). The need to interpret model predictions on a monthly or seasonal basis is exposed since, from annual totals alone, AE_2 is shown to overestimate catchment evapotranspiration by only 10 mm (an error of 3%). Annual potential evapotranspiration derived from the Penman formula (561.1 mm) is almost one and a half times that estimated by Penman-Monteith calculations (397 mm), while enhanced estimates of potential rates during the early season relative to those later in the summer induce exaggerated predictions of actual evapotranspiration by both AE_1 and AE_2 during late spring and early summer (Fig.6.3). For the immediate post-burn period (April, May) AE_1 is almost three times that determined from the water balance equation (Table 6.2). Application of these estimates of actual evapotranspiration would therefore result in underestimation of runoff for this time of year and overestimation by AE_1 leads, on two occasions, to the proposal of negative stream runoff. Reduction of

Penman potential values during late summer may explain the late season underestimation by AE_2. This underrating is not apparent for AE_1 estimates which are larger overall than their layer deficit counterparts.

These discrepancies between estimates of actual evapotranspiration as based on Penman or Penman-Monteith potential demand arise from inherent differences in evaporation specification. Penman potential evapotranspiration is calculated on the assumption of a complete cover of vegetation throughout the year; enhanced rates of evapotranspiration are therefore expected to apply during periods when, in reality, plant cover is reduced due to seasonal growth patterns. In contrast, the Penman-Monteith formula accounts for reduced cover in spring through varied surface resistances. No allowance is made, however, for seasonal variations in root development and growth, or for periods of root dormancy in any of the six estimates of evapotranspiration. Only a single root constant value is used in the Grindley model and, similarly, the TOPLYR:TOPLYR+BOTLYR ratio used in MORECS is not permitted to vary with season. The Penman-Monteith equation enables definition of the effects of a crop change; a vital rationale in the present experiment. The surface and aerodynamic resistance terms are therefore altered to reflect vegetation removal, and improved simulation of evapotranspiration by the Grindley model is promoted (Fig. 6.5), with a reduction in the early season overestimation and late season underestimation characteristic of Penman-based estimates.

Parameter optimisation minimises the effects of model structure on evapotranspiration assessment, as shown by comparing error terms from the MORECS model for AE_5 and AE_6 with those from equivalent Grindley predictions, AE_3 and AE_4, respectively (Table

6.3). The overall best estimates prove to be those based on Penman-Monteith evapotranspiration data combined with layer soil moisture deficits, model type being largely unimportant.

Errors in soil moisture deficit calculation form a further important influence on evapotranspiration prediction. The inadequacy of representing evaporative loss from the soil profile by total profile deficits, based on a single field capacity value, is demonstrated here in a number of ways. As indicated in Table 6.3, errors associated with evapotranspiration estimates are consistently reduced, by 10% to 20%, following replacement of total profile by layer moisture deficits. Soil moisture models based on field capacity-derived deficits generally make no allowance for summer drainage and may thus culminate in predicted negative runoff. In order to ensure exclusion of all profile drainage, an exacting method of drainage separation needs to be applied to overcome the difficulty of defining the seasonally changing evaporation/drainage demarcation: the zero flux plane (Wellings and Bell, 1980; Bell, 1981). The extent of drainage inclusion in field capacity-based deficits for the study catchment is demonstrated by Figures 6.7 and 6.8 in comparing total profile and layer moisture deficits, and by Figures 4.5 and 4.6 (p. 106,107) which illustrate the difference between a draining profile and an evaporating one. True deficit is overestimated both for heather by inclusion of up to 50% as drainage, but principally under de-vegetated conditions, for which only a very small extraction depth pertains, resulting in errors of 65% to 90% in summer. Total profile deficits may occasionally be in error by 100% for either land-use type in winter when the whole profile is draining.

Further support is thus provided for the general conclusions reached in Chapter 4, both models being sensitive to potential

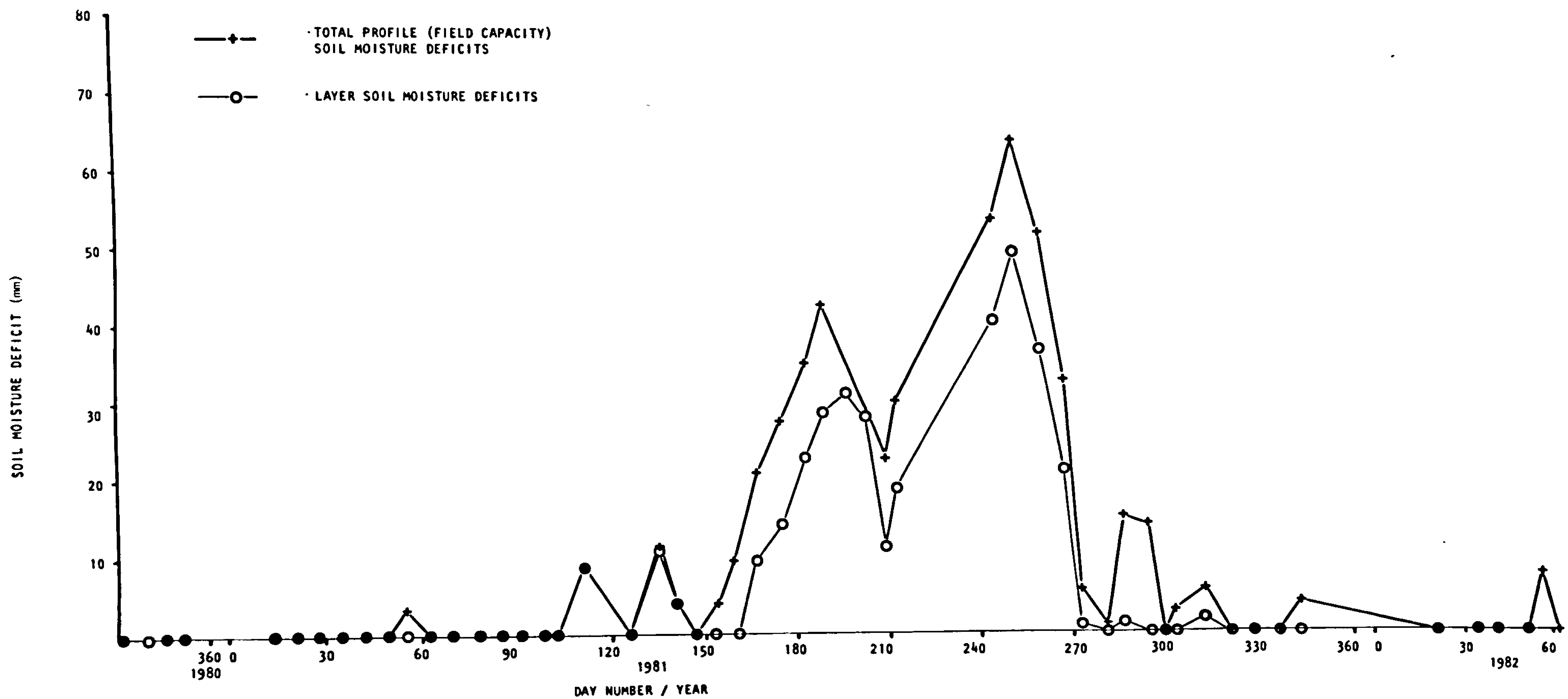


Figure 6.7 Comparison of Total Profile and Layer Soil Moisture Deficits under Heather Moorland

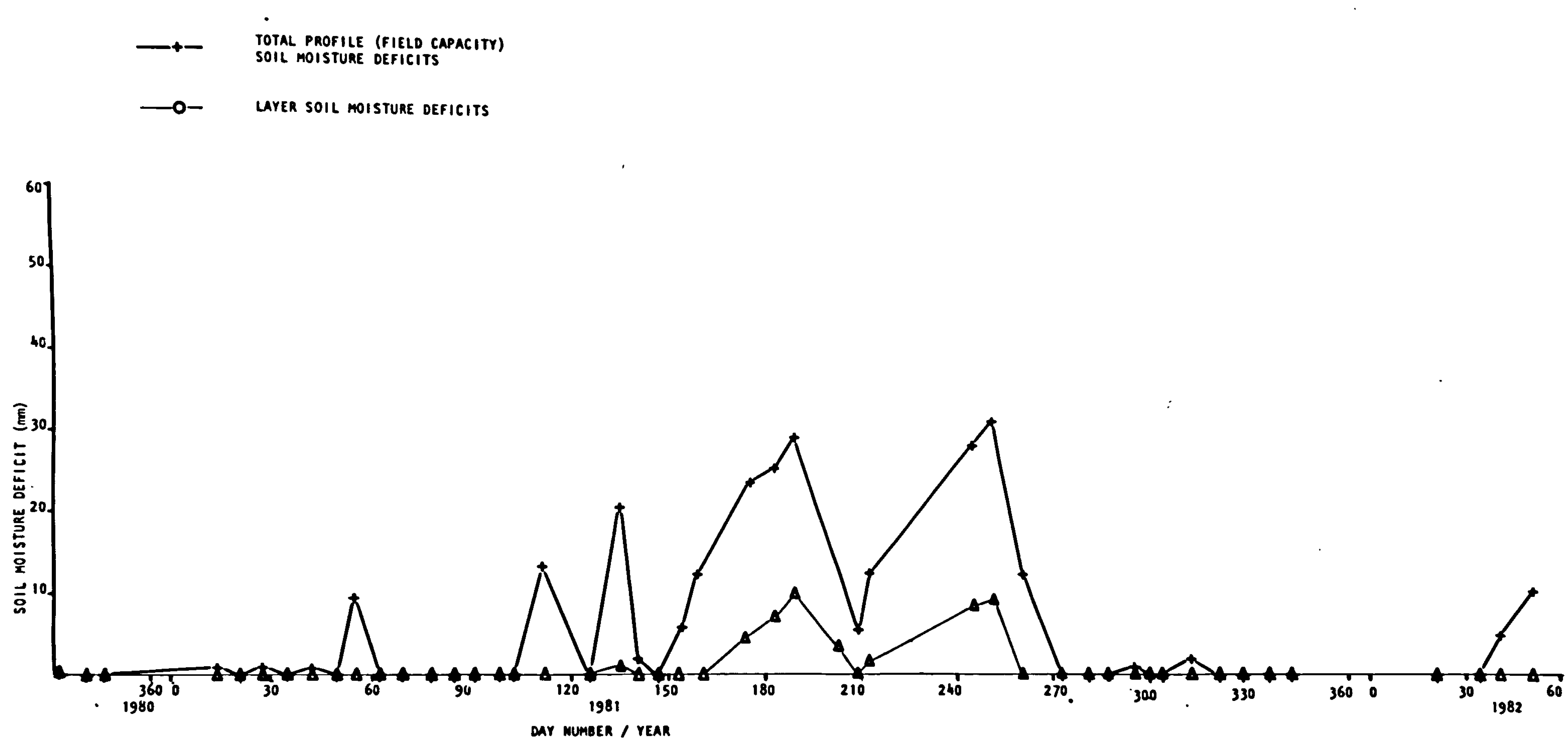


Figure 6.8 Comparison of Total Profile and Layer Soil Moisture Deficits under Burnt Moorland

evapotranspiration and soil moisture deficit data. The best simulations of actual evapotranspiration, and therefore the most accurate catchment water balances, are derived from soil moisture models using Penman-Monteith potential evapotranspiration data in conjunction with 'layer' soil moisture deficits. General overestimation of evapotranspiration in the early season and the more limited underestimation later in the year again suggest application of a seasonally variable root constant for the Grindley model, or TOPLYR:TOPLYR + BOTLYR ratio for MORECS, to improve further the simulations of catchment evapotranspiration and runoff. The importance of the potential evapotranspiration function and its effect on actual:potential evapotranspiration ratios is examined in the following section. The overall significance of evapotranspiration in the water balance, and subsequent implications of land-use change are also discussed.

6.3.3 ACTUAL/POTENTIAL EVAPOTRANSPIRATION RELATIONSHIPS

Annual evapotranspiration from the moorland is estimated from the water balance equation as 356 mm (17 December 1980 to 8 December 1981). This represents 90% of Penman-Monteith potential demand, but only 64% of the equivalent Penman value. Seasonal variations in actual/potential evapotranspiration ratios are illustrated in Table 6.4 for water balance evapotranspiration against both Penman and Penman-Monteith potentials. Relative evapotranspiration reaches maximum levels during the winter months (January, February, March, October, November and December) usually exceeding both Penman and Penman-Monteith potential rates. Ratios with Penman-Monteith evapotranspiration also surpass unity during June and July, although this may be explicable in terms of overestimated resistances. The

PERIOD	DAY NUMBER (approx. 28 days unless indicated)	AE : PE (Penman)	AE : PE (Penman-Monteith)	DAILY MEAN RAINFALL (mm day^{-1})
26.11.80-16.12.80	331-351 (21 days)	2.0	0.67	4.1 (85.0)
17.12.80-13.1.81	352-13	1.5	0.33	1.6 (44.0)
14.1.81-10.2.81	14-41	1.33	1.0	1.0 (28.5)
11.2.81-9.3.81	42-68	1.0	1.0	2.6 (69.5)
10.3.81-7.4.81	69-97	0.9	1.13	4.6 (134.5)
8.4.81-5.5.81	98-125	0.35	0.54	1.9 (54.0)
6.5.81-2.6.81	126-153	0.31	0.5	1.9 (53.5)
3.6.81-1.7.81	154-182	0.64	1.11	1.2 (34.5)
2.7.81-22.7.81	183-203 (21 days)	0.66	1.0	2.0 (42.0)
28.7.81-2.9.81	209-245 (37 days)	0.6	0.9	1.4 (51.0)
3.9.81-29.9.81	246-272	0.65	0.87	4.2 (112.0)
30.9.81-27.10.81	273-300	1.11	1.25	3.0 (84.5)
28.10.81-24.11.81	301-328	1.75	1.75	1.8 (51.5)
25.11.81-8.12.81	329-342 (14 days)	0.67	2.0	1.0 (14.5)

Rainfall figures in brackets indicate totals (mm)

Table 6.4 Actual:Potential Evapotranspiration Ratios Calculated using Daily Mean Values anomalous ratio of 2.0 during the November/December period may also originate from this inaccuracy. Minimum ratios, at about one-third of Penman and one-half of Penman-Monteith potential rate, are reached during spring and early summer. Soil moisture stress is dismissed as a causal factor for these lower ratios, since layer deficits reach a maximum of only 12 mm under vegetated moorland and 1 mm under burnt during this period.

The pattern of actual:potential evapotranspiration (Penman) relationships found here agrees in part with that found for a fifteen-year old stand of heather on Sneaton High Moor (Wallace et al., 1982). Variation from unity in these ratios was related to amount of rainfall, with low ratios corresponding to dry months, and, to a certain extent the Egton data fit this hypothesis, the main exception being for January/February (days 14 to 41) with a low daily mean rainfall (1.0 mm) but one of the higher actual:potential evapotranspiration ratios (1.33). Actual evapotranspiration, estimated in the Sneaton study from lysimeter measurements, was closer than the Egton result to annual Penman potential rate : about 87% for both 1980 and 1981, although differences both in estimated rainfall and drainage, and in mode of evaluation of evapotranspiration (rainfall minus drainage from a small lysimeter) explain the margin between the two sites.

Process control by the relative magnitudes of aerodynamic and surface resistances ('r_a' and 'r_s') explains the importance of rainfall amount and canopy wetness. For a wet heather canopy actual evapotranspiration rate, if calculated by the Penman-Monteith formula, may be expected to be high, since surface resistance is close to zero and, therefore, the only resistance to evaporation is a small aerodynamic resistance (Wallace et al., 1982). As transpiration ceases under wet foliage, evaporation is in the form of interception loss. Evaporation of intercepted water proceeds at a rate often to in excess of Penman potential evaporation and is therefore undervalued by the Penman equation. High actual:Penman potential evaporation ratios therefore pertain during winter months, Penman PE underrating water loss. Underestimation of woodland evaporation using the Penman formula, has also been explained by McGowan et al. (1980) in terms of comparatively high rates of loss by interception. Since aerodynamic resistance affects the rate of evaporation but not the absolute quantity of water lost, long-term evaporation values are less sensitive to aerodynamic resistance (Wallace et al., 1984). Penman-Monteith actual evapotranspiration remains lower on dry days when surface resistance is no longer zero, and transpiration is curtailed. Transpiration is overestimated by Penman potential calculations which assume a grass crop and, thus, implicitly too low a surface resistance value (typically 50 sm^{-1}, as opposed to 100 sm^{-1} to 150 sm^{-1} for heather) and actual transpiration remains below potential demand in summer. Heather therefore behaves in a way more comparable to coniferous forest than to grass in this respect, r_s values for heather and conifers being similar. Transpiration from grass should be greater than that from heather and coniferous species because of its lower r_s.

6.3.4 IMPLICATIONS FOR LAND-USE CHANGE

Much of the evidence presented to substantiate the idea that reduction in vegetation cover enhances water yields and that afforestation decreases yields (Law, 1957a,b; Clarke and Newson, 1978; Calder, 1979; Calder and Newson, 1979) relies on the varying magnitudes of interception loss. The importance of interception in moorland hydrology has been demonstrated in the present dissertation through its influence on runoff and in terms of its contribution to actual evapotranspiration rates. For a given vegetation type, the proportion of rainfall intercepted varies with regional climate and from season to season (Rutter et al., 1971; Gash and Morton, 1978). Interception loss depends both on canopy capacity, the amount of water on the canopy after cessation of rainfall and throughfall, and on the duration of evaporation during saturated canopy conditions (Gash et al., 1980). A number of studies have shown, provisionally for medium-height vegetation such as heather, and in particular for forest species, that interception holds quantitative significance in the water balance. Interception losses of up to 50% of gross annual rainfall have been recorded for pine and spruce species and in winter intercepted water may evaporate to in excess of Penman's open water estimate (Leyton et al., 1967; Rutter, 1963; Rutter and Fourt, 1965; Roberts et al., 1982). Corresponding magnitudes have been reported for moorland communities in some instances. Aranda and Coutts (1963), for example, found that 40% to 50% of precipitation was intercepted by heather, results comparable to those of nearby pine and spruce trees.

Much of the discussion on the potential importance of the evaporation of intercepted water is centred around its significance in relation to transpiration. Opinions vary, however, as to whether interception contributes an additional evaporative loss or is merely in balance with transpiration in the water budget. Penman (1963, p.38), for example, contended that evaporation of intercepted water represents a 'slightly luxurious' alternative, and not a supplement to transpiration loss. It has also been found however, that on wet crops

intercepted water may evaporate up to five to ten times faster than transpired water from dry vegetation, while both processes may occur simultaneously from one plant, should parts of the canopy be wet and parts dry (Hewlett, 1982).

Variations in relative rates of transpiration and interception under the same environmental conditions relate to the comparative magnitudes of vegetative resistances, referred to earlier, and thus to vegetation type (Fig. 6.9). 'Relative transpiration rate' (Monteith, 1965) E_t/E_i is close to unity for short vegetation such as grass, since r_s and r_a are of similar magnitudes. Evaporation of intercepted water should therefore be counterbalanced in the water budget by a reduction in transpiration (Burgy and Pomeroy, 1958; McMillan and Burgy, 1960; Stewart and Thom, 1973) and the interception component remains relatively less important. As r_a is reduced relative to r_s, however, interception rate begins to exceed loss by transpiration. For shrubs and particularly trees, therefore, there is strong evidence to suggest both the importance of interception in its own right and the plausibility of interception loss rates exceeding those of transpiration. Separate estimation of these two components may therefore be expedient in certain water balance and water resources studies (Leyton et al., 1967; Rutter, 1975; Stewart, 1977; Gash et al., 1980) with specific modelling of the interception component.

That interception loss rates can physically exceed those of transpiration may be explained in terms of differences in energy availability. Rutter (1967, 1975) proposed that an excess input of sensible heat supplies the extra energy required for intercepted water to be evaporated faster. This idea was supported by his finding wet leaves to be cooler than the surrounding air and thus the establishment of a temperature gradient enables generation of an extra downward heat flux from the air. Other workers, for example, Pereira (1967) have questioned the availability of this additional energy, while

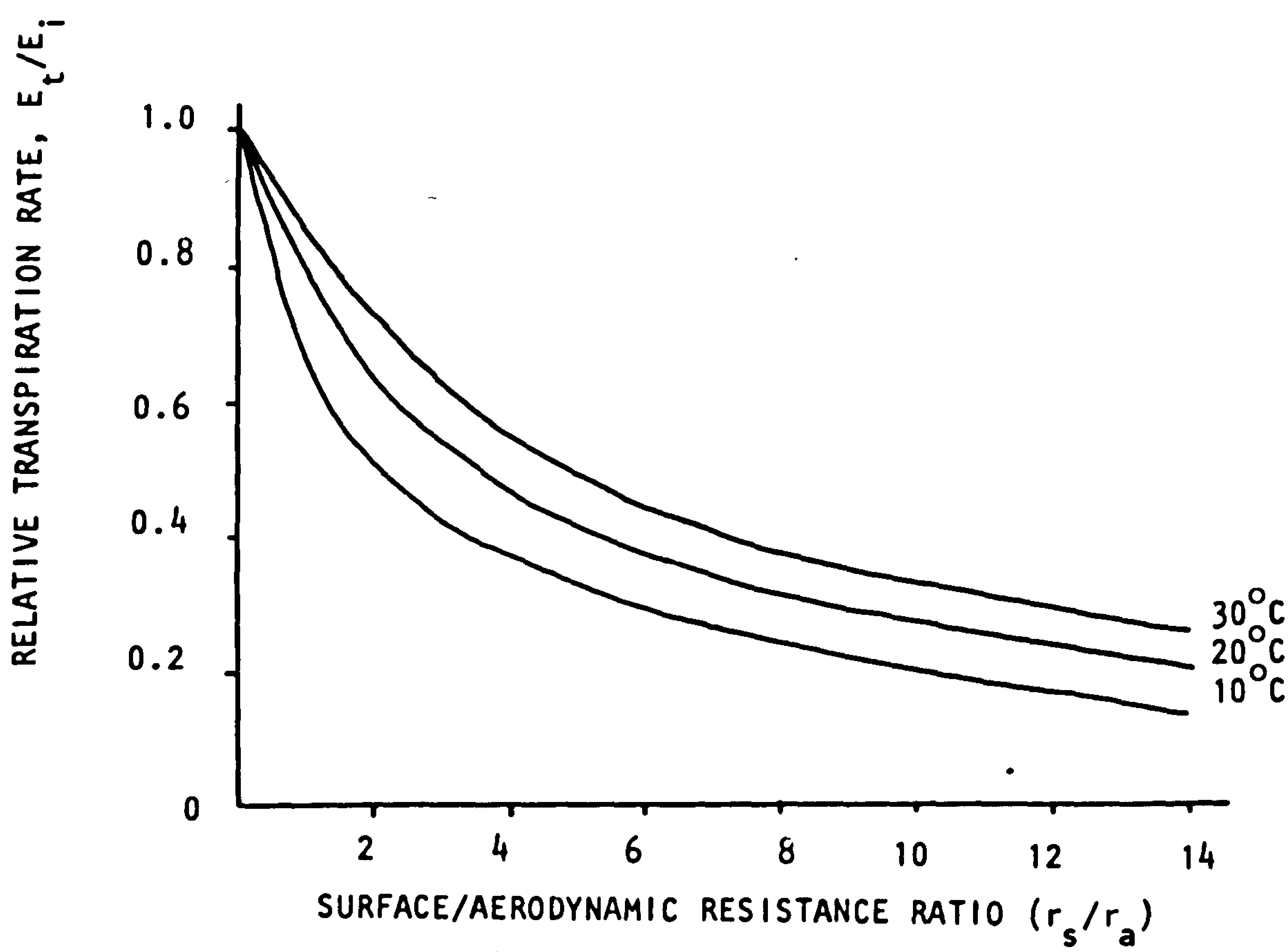


Figure 6.9 Change of Relative Transpiration Rate (E_t/ E_i) with Surface/Aerodynamic Resistance Ratio (r_s/r_a) and with Temperature [after Monteith (1965)]

suggestions that it may arise from leading edge advection were refuted by Stewart (1977) using an experimental site near the centre of a large (70 km^2) forest block at Thetford. Evaporation of intercepted water in Thetford used, on average, 127% of available energy, compared to only 41% of available energy used in transpiration (Gash and Stewart, 1975).

Specific differences in magnitudes of transpiration and interception determine comprehensive changes in total evaporation, and thus in water use with changing land-use. Because heather and coniferous species have generally similar surface resistances, broadly comparable rates of transpiration should be expected (Wallace et al., 1982). Slightly higher transpiration losses may in fact be occurring from Wintergill Plantation than from heather, however, since an annual reduction of 50 mm of soil moisture is recorded for the woodland, as opposed to 26 mm for the heather plot over the same period (14 January 1981 to 8 December 1981). Values of combined transpiration and interception for Egton Moor (Table 6.1) remain comparable with, or even lower than, reported transpiration rates for conifers, the latter being of the order of 3 mm day^{-1} to 3.5 mm day^{-1} or about 340 mm yr^{-1} to 350 mm yr^{-1} (Milne, 1979; Whitehead and Jarvis, 1981; Roberts, 1983). Winter transpiration rates under conifers may be as low as 0.03 mm day^{-1}, however (Calder, 1978), while the state of the present site in the post-spring period as only partially vegetated also leads to moderated levels of evapotranspiration during this time.

Using the most reliable model predictions of total evapotranspiration, the Grindley model with Penman-Monteith potential evapotranspiration estimates and layer moisture deficits, it is possible to compare losses for different, simulated land-uses at Egton. Thus, had the muirburn not taken place, an evapotranspiration loss of 390 mm is predicted for the period between 14 January 1981 and 8 December 1981, in comparison to that of 347 mm actually recorded. The volume of water available for runoff would have been accordingly reduced, assuming a negligible change in soil moisture storage over the year. An annual increase in total evapotranspiration of 70 mm or 18% is tentatively predicted for a land-use change from a complete covering of heather to coniferous woodland (Appendix VI). This magnitude of change is in agreement with that of 10% to 20% predicted by Wallace et al. (1982) for afforestation of Sneaton High Moor.

Although general magnitudes of interception under moorland vegetation have in some instances been found to be comparable to those of coniferous forests (Aranda and Coutts, 1963; Leyton et al., 1967), interception losses under forest may more typically be expected to exceed those from heather because of the lower aerodynamic resistance of the former (Wallace et al., 1982). Much of the predicted loss for Wintergill Plantation may therefore comprise evaporation of intercepted water, depending on the relative transpiration rates between woodland and moorland, referred to earlier. An increase in burnt area from that measured here, approximately one-third of the moorland study area, to 100% would lead to an increase in runoff loss, and a reduction in evapotranspiration of about 98 mm, being 28% of the recorded loss for the partly burnt moorland. The difference in loss between a heather catchment and a completely burnt moor is expected to be in the region of 141 mm, a reduction following burning of about one-third of the total value expected under heather.

All such results must be viewed in the context of local climate since both the relative contributions of interception and transpiration, and the absolute amount of interception loss in particular, vary with rainfall regime. Thus, Gash and Stewart (1977) recorded a total annual evapotranspiration loss from a Scots pine stand in Thetford Forest which was only 88% of that expected from grass, while in central Wales, an interception loss of four times that for Thetford has been recorded for spruce trees (Calder, 1976). More definitive research with medium-height species is needed before unequivocal conclusions can be drawn regarding their relative transpiration rates as compared to those of other crops.

6.4 CONCLUSIONS

The hydrology of the Egton Moor study area is examined in the foregoing discussion by means of a general water balance equation. Budgets are determined for monthly intervals as well as on an annual basis for 1981. Although a satisfactory water balance may be developed for a particular catchment, however, difficulties may arise in applying results and conclusions to other areas (Ward, 1967b). Earlier demonstration of the representativeness of the study area would probably allow limited extrapolation to adjacent areas of the North York Moors showing similar climatic and soil/vegetation regimes. Although annual rainfall was below the long-term (1941 to 1970) average (Table 2.3, p. 26), monthly totals for both Egton and Sneaton sites closely follow records for comparable sits in the region for 1981 (Fig.2.4). Values of Penman potential evaporation (Sneaton) match long-term records in all but the mid-summer months (Fig.2.5).

The residual term of the equation, actual evapotranspiration, is used to evaluate the ability of soil moisture models to predict this component of the water balance. The results generally confirm conclusions drawn both in Chapter 4 and by other workers. Model predictions of annual evapotranspiration form satisfactory estimates of 'observed' values, with all predicted results except those from the

Penman PE-Grindley model/total moisture deficits combination giving estimates to within $\pm 7\%$ of that determined from the water balance. Evapotranspiration is overestimated and runoff underestimated in the early season, while the best estimates are derived from models using Penman-Monteith potential evapotranspiration data and layer soil moisture deficits. Combined with a variable drying specification parameter, this model profile may be expected to provide realistic assessments of catchment evapotranspiration.

The significance of evapotranspiration in the water balance has been emphasised here and the important role of interception, potentially in moorland and particularly in forest environments has been acknowledged in terms of its relationship with transpiration. The idea that interception may constitute an additional evaporative loss to transpiration in the moorland water budget and that it may require separate evaluation on this catchment forms a topic for further investigation. Some published work has indicated broadly comparable rates of interception loss under moorland vegetation and woodland species, although, following afforestation in the type of environment under study here, a small, 10% to 20%, increase in water use is probable.

The results presented in this chapter represent a provisional attempt at describing and interpreting the general hydrology of the study area. More exhaustive assessment of interrelationships between water balance components requires further, detailed data for catchment characteristics, while comprehensive mapping of source area locations in relation to the distribution of land-use would aid analysis of runoff characteristics and would facilitate prediction of hydrological changes brought about by further alterations in vegetation.

CHAPTER 7

SUMMARY AND CONCLUSIONS

A single, manageable headwater experiment in the North York Moors has been used to observe and evaluate the immediate hydrological changes initiated by the current moorland management practices of afforestation and, particularly, controlled heather burning. Changes in hydrological relationships with land-use have been the subject of numerous previous studies dating from the last century. The present dissertation deviated from the mainstream of investigation through its concentration on quantification of the responses of the natural, medium-height vegetation typical of dry British moorlands (largely Calluna) to which comparatively little attention has been directed, most previous research being centred on forest or grassland hydrology. The salient features of the hydrological cycle in this environment were examined utilising, where possible, objective modelling techniques to elucidate important relationships, and culminating in the construction of an annual water budget. Special attention has been paid to evaporative losses and to changes in soil moisture regimes, aspects of the moorland system for which information is particularly deficient. The findings of the dissertation suggest certain implications for the future development and management of this type of moorland. In view of the catchment's typifying much of the North York Moors area, in terms of its geology, soil type, vegetation community and land management, the general nature of the results may be regarded as typical for much of the region.

Quantification of the effects of land-use on soil moisture regimes was accomplished via empirical modelling using a moisture extraction specification and a potential evaporation function. This type of technique was selected in preference to a theoretical approach, since the latter both requires acquisition of specialist information relating to factors such as root and soil resistances, moisture potentials, or root geometries, and is less suited to a practical, field-based treatment. Models were calibrated in the present study for the year 1981 by comparing predicted soil moisture deficits with 'measured' values. Deficits may then be predicted for succeeding years on the basis of rainfall, potential evaporation and land-use data only; a type of application which is beneficial to water authorities, engineers and agriculturalists in the assessment of flood levels, reservoir replenishments and irrigation needs. 'Actual' soil moisture deficits were expressed relative to the soil moisture 'constant' field capacity, abstracted from neutron probe measurements of moisture content made to a depth of 0.8 m. The neutron scattering technique facilitates frequent and precise measurements on intensive temporal, spatial and depth scales without site destruction, and its use is invaluable in applications of this nature. The shortcomings of the field capacity concept were acknowledged in the dissertation, although its value and convenience in practical situations prompted trial of its applicability for this case. Values of mean moisture content over two winters represented field capacity for the total measured profile, for separate land-use categories (heather moorland, burnt moorland and coniferous woodland).

The single-layer Penman-Grindley soil moisture model, used until recently by the Meteorological Office as a 'standard' model, and its two-layer successor, MORECS were implemented to simulate deficits and to predict actual evapotranspiration from each land-use plot. Using the single parameter (root constant) drying curve postulated by Penman (1949), root constants as recommended by Grindley (1969) and 'field capacity' moisture deficits, the Penman-Grindley model generated poor overall simulations, with high error terms for all land-uses. 'Observed' deficits, which were comparable under heather and woodland over the year, with both of these exceeding those found after heather burning, were, in general, overestimated by the model for the late spring/early summer period and, in counteraction, were undervalued later in the year. Misrepresentation of plant water use as well as inappropriate drying curve shape may partly explain these discrepancies. Timing of autumn re-wetting was found to be adequate, and that of summer runoff was generally accurate.

Elimination of spurious drainage by re-running the model for 'layer' (maximum extraction) deficits worsened fits since predicted deficits remained unaltered. Layer deficits were determined after the method of McGowan (1974), through the separation of drainage from evapotranspiration by identifying deviations in the rate of change of moisture content over time. It was provisionally concluded from drainage separation that heather burning reduces moisture deficits such that on the study site maximum root extraction depth was reduced from 50 cm under heather to 20 cm under the burnt plot. A value of 50 cm also applied to the woodland. Thus, the maximum 'layer' soil moisture deficit for 1981 was 53 mm under heather, but reduced to only 11 mm on the burnt moorland. Enhanced transpiration and evaporation of intercepted water from the vegetated plots explains much of this difference. The main summer deficit period lasted longest under woodland, at approximately 150 days, and least under the burnt area (about 100 days).

In an attempt to improve goodness of model fit, a further set of simulations was implemented, based on the optimum root constant value for each land-use type. Using total profile (field capacity) moisture deficits, marked improvements resulted for the woodland and burnt moorland plots, although late season deficit underestimation was still apparent. No real changes followed for heather. Overall model performance was refined even further (error of fit was almost invariably reduced) for both Grindley-recommended and optimised root constants following repeated model runs with potential evapotranspiration calculated from the Penman-Monteith, as opposed to the Penman formula. The most notable feature of the simulations for both moorland conditions was the improved reproduction of early season deficits. This was a direct result of the reduced estimates of evapotranspiration incurred by implementing the Penman-Monteith equation. The resistance terms used to represent heather and woodland were probably overestimates and thus predicted values of soil moisture deficit remained low for the later part of the year for both sites, and also during the early season under woodland. For further enhancement of fit, therefore, subsequent trials should be conducted with reduced 'r_s' and 'r_a' values.

The best fits of all Grindley model runs resulted from inputs of Penman-Monteith potential evapotranspiration and actual 'layer' moisture deficits, in combination with an optimised root constant. The Penman-Monteith formula accounts for plant-atmosphere resistances, within the limitations discussed previously, while soil moisture deficits calculated on the basis of 'layer' moisture content avoid the inaccuracies of 'total' profile deficits. The latter included a proportion of drainage of up to 100% for winter months when the whole soil profile is undergoing drainage, whilst in summer, true deficit was overestimated particularly for the burnt plot, for which maximum extraction depth was only 20 cm. More realistic methods of deficit determination, therefore, such as those based on McGowan (1974), as used here, or Wellings and Bell (1980) are advocated. Finally, the root constant values recommended by Grindley were found to be particularly inappropriate for the type of high rainfall and low moisture deficit environment under study here. The amount of evapotranspiration occurring at potential demand was shown to be limited and thus optimisation was to much lower values than those 'recommended'. In several cases, notably for the burnt moorland area, less than 25 mm of moisture was evaporated at the potential rate, resulting in a negative root constant. The results are in accordance with a number of previous findings that recommended root constants, and thus actual evapotranspiration, are overestimated for specified crops.

With optimised drying specifications, employment of the more physically realistic, two-layer soil moisture model, MORECS, yielded only slightly improved reproduction of actual moisture deficits than did the Grindley model. MORECS therefore provided little of the anticipated degree of improvement over its single-layer counterpart and for this type of environment at least, for 'regional' prediction of deficits, it is suggested that the soil moisture estimation model used by the Meteorological Office for almost twenty years can still be relied upon without recourse to its double-layered replacement. Although only limited additional information was forthcoming from the MORECS simulations, conclusions from the first model were corroborated. Thus, evapotranspiration at the potential rate was shown to be limited, with recommended drying parameters again proving to be too high for this environment, while inclusion of

Penman-Monteith evapotranspiration estimates almost always improved model performance, and, with simulations based on 'layer' moisture deficits, reduced error terms were achieved.

Results revealed that evaporation continued at higher deficits under vegetated than under burnt moorland, indicating a lower available water capacity in the latter case. Following a change from heather moorland to burnt ground, deficits and evaporative losses at the potential rate were reduced throughout the year, and were generally maintained or increased as a result of coniferous afforestation. It is proposed that the typical seasonal pattern, apparent for several simulations by both soil moisture models, and comprising overestimation of early season deficits followed by underestimation later in the year, may be avoided or at least partially eradicated by allowing the drying specification parameter to vary throughout the year. Although the best overall fits of all model runs proved to be those based on the MORECS model using an optimised drying specification, 'layer' moisture deficits and Penman-Monteith evapotranspiration, this model provided only moderate improvements against each set of comparable specifications for the Grindley model. Type of input data and specified drying parameters were therefore found to be more important than model structure in deficit simulation.

Most model runs provided accurate predictions of the timing of summer runoff. Enhanced interception, transpiration and moisture abstraction by roots explain the limited number of returns to field capacity in summer under woodland, in comparison to the two moorland plots, while observations of subsurface flow confirmed that greater volumes were generated both from the burnt area in comparison to woodland, and on the moorland after heather burning. Subsurface flow was promoted under all land-use types by the presence of an impermeable clay layer underlying a permeable organic horizon. Root channels may assist the process under woodland, although, since the definition of subsurface linear features depends partly on size, the essential significance of definitive pipeflow in the area was questioned. Rather, it is tentatively suggested that the headwater is characterised largely, but probably not exclusively, by dispersed matrix flow. Timing of throughflow response corresponded to periods of measured surface storm runoff most notably for the burnt area. It is suggested that summer storm rainfall is used preferentially in soil moisture deficit reduction under woodland and the vegetated moorland, whereas diminished evaporation from the burnt plot renders more water available for runoff generation. The low extraction depth characteristic of the burnt moorland encourages early attainment of profile saturation and thus a comparatively rapid runoff response. These effects, in conjunction with typically low rainfall intensities, minimise any potential influence which a bare, compacted surface may have in modifying runoff through reduced infiltration.

Spatial variation in soil moisture content within and between land-use categories was interpreted through contour interpolation. Data for the immediate surface layers of the soil profile (0 cm to 7.5 cm) were accumulated by thermogravimetric means and spatial variation in these layers was explained not only with respect to differences in vegetation cover, but also through small-scale variations in other factors, such as storm characteristics, antecedent moisture conditions and topography (position on slope). Differences in total profile moisture (0 cm to 80 cm) over the catchment were shown to be fairly uniform throughout the total measuring period of almost two years and were interpreted in terms of potential variable runoff-generating areas, using topographical evidence (position on slope and contour shape). The slope base was identified as an area of flow convergence and, therefore, a runoff-contributing zone, and the area of younger heather as a spur and one of generally good drainage. Catchment boundary zones were specified as poorly drained areas, characterised by a levelling of slope angle and reduced moisture fluxes.

Having considered the repercussions of vegetation cover change for both soil moisture and subsurace runoff regimes, the third major analytical section of the dissertation was presented to evaluate surface runoff responses to rainfall inputs. Calluna moorland hydrology was assessed in terms of the relationship between rainfall and runoff volumes on both a monthly and an annual basis. Total measured runoff for 1981 comprised 57% of annual rainfall. This is a blanket figure, however, giving only a general indication of a typical rainfall:runoff ratio for the area, since the year includes the moorland land-use change of early April. Seasonal variations for a burnt catchment are therefore more meaningful and in this respect approximately one-third of gross rainfall was shown to runoff during the complete summer period, being reduced to less than 10% during the mid-summer months of June and July. Runoff response was rapid under the saturated catchment conditions of winter with roughly 70% of rainfall contributing to total runoff. One constraint to expressing rainfall:runoff ratios on a monthly or annual basis was demonstrated by measured volumes of runoff apparently exceeding measured preciptation input for a particular period: this was explicable in terms of snowmelt.

Partial system synthesis, using the deterministic 'unit hydrograph' model, was utilised to quantify and analyse changes in flood hydrograph features generated as a result of the muirburn.

Despite the limitations of the unit hydrograph technique, it remains a useful tool in many practical circumstances and was justifiably applied in the present case, especially since some of its assumptions prove less constraining for small catchments, particularly when storm selection, and hydrograph and rainfall separation are executed with care. Thirteen storm events were suitable for analysis and were defined by one-hour unit hydrographs, derived by matrix inversion with smoothing. Each unit hydrograph was approximated by one straight line for the rising limb and two for the recession. These segments, so derived, permitted quantitative comparisons of pre- and post-burn hydrographs through depiction by dimensional shape parameters. The latter were used to formulate simple linear regression models, which generally proved to be land-use specific.

The roles of storm and non-vegetational catchment characteristics in hydrograph formation were examined by multiple regression equations. This type of model may be used in preliminary predictions of the hydrograph form for ungauged catchments similar to those used in this study, whilst specification of factors such as the area of catchment modified by man may promote greater degrees of accuracy. The complete data set (pre- and post-burn values combined) was used for this analysis, effectively extending the range of physical variables involved in each data set. Further, a 'dummy' variable was included as an independent variable, both as an index of catchment state to represent responses arising from land-use change, and to reflect variations in other environmental factors. Although four dependent variables were initially included to describe hydrograph shape, only two dimensions, recession curve length (RECL) and peak flow (QP), proved to be significant in these models.

In agreement with previous work, it was demonstrated that peak discharge increased markedly after devegetation and that antecedent moisture regime was significant in determining hydrograph shape. Thus, using the multiple regression models as predictors of average hydrograph shape, peak flow was almost doubled as a result of heather burning. This increased response was related both to rapid replenishment of soil moisture deficit resulting in early attainment of profile saturation, and to enhanced subsurface flow. Under a completely vegetated catchment, lower, wider hydrographs with shallower rising limbs arose from greater rainfall losses both to the soil profile (due to higher moisture deficits) and to the evapotranspiration component. An additional, underlying control of hydrograph shape by catchment wetness acts independently of land-use type so that for a given reduction in soil moisture deficit, peak discharge is lowered and recession length is increased, resulting in a wider general unit hydrograph for a wet catchment, while under drier conditions the catchment responds more quickly to rainfall input. This may reflect the operation of dynamic contributing areas, travel times increasing and the hydrograph widening as these zones extend. Whilst recession curve length is predictable from soil moisture deficit by a simple regression model, the best fit equation for predicted peak flow showed the latter to be determined additionally by total rainfall amount (RAIN) as well as land-use. As rainfall input increases, absolute magnitude of the peak flow variable is amplified.

The results can be interpreted in terms of their implications for the physical and ecological stability of the catchment, and thus for moorland management. Vegetation burning is found to promote exaggeration of the observed hydrograph peak, while increased rates of infiltration and throughflow enhance laterally draining water, which can result in the development of gullies and seepage faces, through undermining and collapse (Imeson, 1971). Thus, through channel formation and expansion, drainage density ultimately increases, channel erosion and movement of surface material downslope are enhanced (Arnett, 1980) and speed of response is increased further. The current results therefore indicate the initial stages of catchment degradation and it is imperative, therefore, that controlled burning is maintained at frequencies sufficient to allow vegetational recolonisation and system stabilisation.

As a final analysis, the physical controls of water use with changing vegetation cover were discussed in Chapter 6. Monthly and annual catchment water balances were presented, although data restrictions confined the calculations to represent the whole moorland slope area, including the change of surface cover. Further calibration would be required for a more specific and quantitative comparison of complete pre- and post-burn balances. All terms of the water balance equation were measured except for the actual evapotranspiration component (AE), which was computed by elimination. A degree of error is incorporated in a term derived by difference, due to accumulated inaccuracies in the remaining measured components, and therefore the water balance-derived ('observed') estimates of evapotranspiration were presented initially as relative rather than absolute values. Potential rates of evapotranspiration (PE) pertained until March and applied again from late September or early October. Fifty-seven per cent (440 mm) of measured rainfall for 1981 contributed to total runoff and 46% (356 mm) was lost to evapotranspiration. Showing a reduction of only 3%, annual change in soil moisture storage was insignificant both in comparison to other water budget components and in relation to the total measured moisture content of the soil profile.

The specific ability of empirical soil moisture models to predict accurately catchment evapotranspiration was tested by comparing model-predicted estimates with those 'observed' for the two phases of moorland land-use. Model evapotranspiration was determined by adjusting crop specific potential evapotranspiration in accordance with changes in soil moisture content (the drying curve). Six types of estimate were reviewed, derived using Grindley and MORECS models, Penman and Penman-Monteith PE, and layer and total profile soil moisture deficits, and, in order to generate the most accurate estimates of catchment evapotranspiration, only optimised model parameters were used. Although annual evapotranspiration was generally overestimated by model predictions and, by implication, stream runoff was undervalued, most types of estimate predicted the water balance-derived value to within 5%. The smallest errors arose from model simulations using Penman-Monteith potential evapotranspiration combined with 'layer' soil moisture deficits, providing acceptable values for inclusion in a catchment water budget. A seasonally varying drying specification parameter is expected to improve predictions further. The poorest assessment, produced by the Grindley model using Penman evapotranspiration and total profile moisture deficits, overestimated actual evapotranspiration by 19%. As discussed previously, total 'deficits' were found to include spurious drainage, while the Penman formula fails to represent changing plant water use with season. Indeed, of all six model estimates of catchment evapotranspiration over the year, the highest errors appertained to predictions based on Penman PE. In particular, evapotranspiration was overestimated during the early season, when potential rates remained high in comparison to the later summer values; this feature led to the postulation of negative streamflow for certain periods. Penman/Grindley predictions based on layer soil moisture deficits were too low during the later part of the year, although replacement of Penman with Penman-Monteith potentials improved simulations since the latter calculations incorporate plant resistance terms which were altered in accordance with changing moorland land-use.

Predictions from the two-layered soil moisture model, run only with Penman-Monteith evapotranspiration, generally failed to show improvement over the analogous Grindley model simulations, since, as noted earlier for soil moisture deficit prediction, evapotranspiration and actual moisture deficit specification proved more critical than model structure. Thus, inclusion of layer moisture deficits reduced error of prediction by about 10% over that found for total profile deficits for this model.

Tentative losses by evapotranspiration may be postulated for simulated land-uses on the catchment. Thus, assuming the moorland area had remained completely vegetated, an increase in annual evapotranspiration of approximately 12% would have resulted. A concomitant reduction in stream discharge may also be proposed, since higher losses by transpiration and evaporation of intercepted water are expected to reduce water available for runoff. Increasing the existing burnt area to cover the whole moorland slope would reduce actual evapotranspiration by about 98 mm or 28% of the measured, water balance, value. An attendant increase in stream runoff may be provisionally offered. More significantly, a land-use change from an all vegetated catchment to a completely burnt area would be expected to produce a reduction in annual actual evapotranspiration of approximately one-third (141 mm). In terms of land management, therefore, the area of vegetation actually removed from the moorland has resulted in a notable reduction in water lost by evapotranspiration, while burning a larger proportion of the catchment would promote more significant changes. In agreement with general predictions for Sneaton High Moor (Wallace et al., 1982) an increase in evapotranspiration loss of 18% (70 mm)is expected as a result of afforestation of the heather moor at Egton.

Relative evapotranspiration rate, actual evapotranspiration expressed as a proportion of potential demand, was at a maximum during January, February, March, October, November and December for both Penman and Penman-Monteith potentials, when the actual:potential ratio often exceeded unity. Although this ratio was minimised during spring and early summer, at one-third of Penman potential demand and one-half of the corresponding Penman-Monteith value, soil moisture stress was not found to be a prevailing influence. Rather, seasonal variation in the ratio of actual:potential values was explained in terms of a relationship with rainfall amount, and thus canopy wetness, with higher ratios pertaining during wet conditions. These variations were qualified in terms of the relative magnitudes of surface and aerodynamic resistances. Hence, on wet days, when surface resistance is negligible and aerodynamic resistance is low, interception loss, the only form of evaporation, is relatively high. Actual evaporation values should therefore be close to Penman-Monteith potential estimates. That some exceed potential may be due to overestimated resistance values in the Penman-Monteith calculations used here. The Penman equation underestimates evaporation of intercepted water and thus potential estimates from this formula are often lower than actual values in winter. Under dry conditions, when surface resistance can reach up to 100 sm^{-1} to 150 sm^{-1} under heather, the Penman formula, assuming a grass crop and therefore by implication underestimating the magnitudes of surface resistance applicable to both burnt and vegetated moorland, yields overestimated values of transpiration, and potential values remain well above those actually found. Since more realistic resistance values are incorporated into the Penman-Monteith formula, actual evapotranspiration is closer to calculated potential. Heather and coniferous species are expected to produce similar results in this respect, since both types of vegetation have broadly similar r_s values. For a burnt moorland surface, a reduction in evapotranspiration is expected over that found for heather and woodland, as discussed above, particularly in view of the relatively high aerodynamic resistance of bare ground and higher soil moisture storage of the latter (Chapters 4 and 5).

Explanations of changing water use with vegetation cover rely heavily upon the effects of evaporation of intercepted water and the relationship between interception and transpiration. It is therefore important to determine the signficance of interception for different species. The potential importance of this water balance component for a moorland catchment has been discussed in qualitative terms both with regard to its degree of importance in total evapotranspiration and with respect to variations in runoff responses under changing land-use. Further work needs to be conducted on natural medium-height vegetation, however, before definitive conclusions are drawn on the specific quantitative significance of interception in regulating water use. Most previous work has been confined to examination of taller crops, demonstrating the overriding importance of interception over transpiration for these species. Evidence for the relative significance of interception in moorland communities, however, is conflicting. Aranda and Coutts (1963) and Leyton et al. (1967), for example found amounts of intercepted water under heather to be comparable to those under coniferous species, while Wallace et al. (1982) predicted a small increase in plant water use as a result of afforesting the Calluna moor at Sneaton, higher interception losses being promoted under coniferous woodland by a lower aerodynamic resistance. The climatological and geomorphological characteristics of an area should be considered before examining the wider applicability of such results, however, interception and transpiration contributions varying, for example, with rainfall regime. Thus, evaporation losses from forest and grassland are similar in low rainfall environments, whilst under high rainfall conditions, the margin is more distinct because of greater losses to interception under forests (Gash and Stewart, 1977).

7.1 SUGGESTIONS FOR FURTHER WORK

Proposals for further study are made both with reference to the headwater area investigated here and in a wider sense for application to other moorland regions. Potential topics of research involve development of the themes already examined in this report, as well as consideration of hydrogeomorphological issues not covered here.

Simple empirical soil moisture budget models were calibrated in the current study using observed soil moisture deficit data for three types of land-use. Their ability to predict soil moisture regime and actual evapotranspiration was assessed, and, for the purposes of a general hydrological study, reasonable simulations of soil moisture deficit were derived. The significance of soil moisture regime and changing soil moisture availability and evaporative losses under man-modified, medium-height vegetation environments deserves further investigation, however, since plant water use studies have been directed strongly towards tree species. Improved soil moisture balance predictions could result from inclusion of seasonally varying drying parameters or different drying relationships which could be developed and tested to extend the analysis. Following calibration for a particular catchment area, models may then be used to predict a number of hydrological characteristics for several years. It would be profitable, for example, to review the accuracy of predicted drainage, using models which represent this component realistically.

Further research is required into the importance of the interception component for the water balance of natural, medium-height species such as Calluna. The quantitative relationship between rates of transpiration and evaporation of intercepted water, and the relative significance of each for total evapotranspiration have implications both for the importance of soil moisture availability in its determination of actual transpiration and in the definition of water use under low-growing vegetation in relation to that of taller species.

Analysis of runoff generation processes in moorland environments should be further elucidated, assessment of rainfall-runoff relationships being particularly facilitated for headwater areas by the lack of complicating routing effects and tributary flows found in larger river systems (Ward, 1984). The results of the present study would be augmented by more specific identification of spatially and temporally changing saturated source areas (using soil moisture potential measurements) in relation to moisture regimes and moisture flux patterns. The latter may be modelled using hydraulic gradients and can be analysed in terms of relationships with soil characteristics.

A more sophisticated, intensive and continuous subsurface flow monitoring system would enable changes in this component to be defined quantitatively and its potential significance for stream discharge to be defined. With an extended data base and calibration period, total storm runoff response may be examined in relation to season, vegetation recolonisation, or source area location within specific land-use categories. Relating hydrograph parameters to physical catchment characteristics enables subsequent flood forecasting for ungauged basins using catchment features alone. 'Area of burnt ground' supplies a possible independent variable to extend the applicability of the present study, in developing these so-called 'synthetic hydrographs', although the need for strict experimental control is emphasised here since the importance of small-scale variations in topography, infiltration characteristics and soil factors has to be defined.

To analyse the hydrograph at points other than the gauging station, hydrograph behaviour as the flood wave moves downstream can be determined by flood routing techniques, which indicate flood wave attenuation using storage in stream channel sections (Wilson, 1974). An analogous procedure, runoff routing, 'the process of routing rainfall-excess (or surface runoff) through catchment storage to produce an outflow that is an estimate of the surface runoff hydrograph of a catchment' (Laurenson, 1964, p. 142), can account for non-linearity of response and spatial rainfall variations (Mein et al., 1974; Laurenson, 1964) and is especially useful for application to unusually shaped basins (Linsley et al., 1949). Routing based on kinematic wave theory, which has made a particularly important contribution to understanding of the rainfall-runoff conversion, uses kinematic wave equations to generate hydrographs from rainfall and catchment characteristics (Wooding, 1965a,b, 1966; Eagleson, 1972; Woolhiser and Liggett, 1967).

Geomorphological adjustments to land-use change also warrant qualification and quantification. In aiming to produce sediment and solute budgets, rates of transport and removal of material should be determined in relation to channel erosion and changes in slope surface material. Sheet erosion, rill development and stream bank changes require evaluation over short- and long-term periods, while trends in factors such as soil temperature have important connotations for frost-heave processes and may need monitoring. On a different scale, changes in particle size distributions with land-use also merit consideration in terms of repercussions for resource depletion. Changing rates of removal of organic matter from the system and the consequences of such changes for the moorland ecosystem are worthy of particular observation.

Small experimental catchments provide a valuable means of interpreting and predicting hydrological changes brought about by land-use alteration. Sophisticated methods of analysis are diversifying the objectives of early catchment studies, however, and in this respect the data and conclusions of the present investigation represent a basis from which a complete conceptual, and ultimately a mathematical hydrological system model may be developed to portray prospective consequences of current or proposed moorland management practices. Specific implications may be modelled by monitoring a series of catchments or plots, which, for example, exhibit various stages of afforestation or have been subjected to different intensities of muirburn. Although an important objective is to establish an accurate and predictive tool for land management application, equally, attention must be directed towards the identification and analysis of those processes which generate the effects evident after land-use change.

APPENDIX I

MOORLAND SOIL PROFILE PROPERTY ANALYSIS

	Horizon		
	Oh	Eg	Bg
% coarse fraction (by weight) (> 2 mm)	29.99	6.66	0
% fine fraction (< 2 mm)	70.01	93.34	100.00
% sand (0.063 - 2.0 mm)	78.81	33.82	22.69
% silt (0.002 - 0.063 mm)	7.84	51.23	32.14
% clay (<0.002 mm)	13.35	14.95	45.17
pH	3.3	3.4	3.4
loss-on-ignition (%)	77.65	9.23	7.44

(Fullen, 1981)

APPENDIX II

SOIL MOISTURE VOLUME FRACTIONS

The data shown are moisture volume fractions (M.V.F.) for representative neutron probe access tube sites from each land-use plot, selected from the total sample of 27 measuring sites.

Figures in brackets show uncorrected values for 10 cm depth, corrected M.V.F.'s being derived as shown in Chapter 3 (p.56).

Mean winter count rates (R/R_S) used in Equation 3.3 for each access tube site are given below:-

	Access Tube Number		
	9 (Heather Moorland)	12 (Burnt Moorland)	W2 (Woodland)
Mean winter count rate at 10 cm (R/R_S)	.494	.56	.352
Mean winter count rate at 20 cm (R/R_S)	.63	.655	.603

Date/Day Number : 16.7.80/198 — 23.7.80/205

Profile Depth (cm)	9 (Heather Moorland) 16.7.80/198	12 (Burnt Moorland) 16.7.80/198	W2 (Woodland) 16.7.80/198	9 23.7.80/205	12 23.7.80/205	W2 23.7.80/205
10	-	.624(.534)	-	.55(.431)	-	-
20	-	.592	-	.572	-	-
30	-	.6	-	.568	-	-
40	-	.552	-	.576	-	-
50	-	.542	-	.556	-	-
60	-	-	-	-	-	-
70	-	.438	-	.458	-	-
80	-	.42	-	-	-	-

Profile Depth (cm)	9 2.8.80/215	12 2.8.80/215	W2 2.8.80/215	9 3.9.80/247	12 3.9.80/247	W2 3.9.80/247
10	.698(.547)	.641(.548)	-	.578(.453)	.618(.529)	-
20	.634	.645	-	.572	.617	-
30	.639	.635	-	.596	.653	-
40	.599	.582	-	.607	.573	-
50	.546	.514	-	.556	.536	-
60	.485	.46	-	.492	.461	-
70	.455	.435	-	.463	.443	-
80	.452	.514	-	.434	.412	-

Profile Depth (cm)	10.9.80/254 9	12	W2	17.9.80/261 9	12	W2
10	.664(.521)	.622(.532)	-	.603(.473)	.739(.632)	-
20	.548	.597	-	.64	.63	-
30	.548	.573	-	.644	.644	-
40	.562	.548	-	.604	.598	-
50	.541	.527	-	.544	.529	-
60	.494	.46	-	.489	.453	-
70	.463	.438	-	.456	.434	-
80	.437	.426	-	.434	.422	-

Profile Depth (cm)	24.9.80/268 9	12	W2	1.10.80/275 9	12	W2
10	.569(.446)	-	-	.554(.517)	.576(.611)	-
20	.582	-	-	.554	.576	-
30	.624	-	-	.557	.567	-
40	.598	-	-	.568	.525	-
50	.541	-	-	.557	.528	-
60	.494	-	-	.495	.457	-
70	.451	-	-	.457	.439	-
80	.434	-	-	.45	.428	-

Profile Depth (cm)	8.10.80/282 9	12	W2	15.10.80/289 9	12	W2
10	.634(.545)	-	-	.56(.544)	.577(.617)	.578(.323)
20	.634	-	-	.56	.577	.578
30	.649	-	-	.571	.561	.457
40	.609	-	-	.594	.525	.412
50	.551	-	-	.554	.521	.455
60	.5	-	-	.478	.46	.457
70	.466	-	-	.459	.434	.44
80	.439	-	-	.538	.427	.41

Profile Depth (cm)	22.10.80/296 9	12	W2	29.10.80/303 9	12	W2
10	.641(.522)	.66(.591)	.617(.363)	.641(.508)	.662(.577)	.616(.358)
20	.641	.66	.617	.641	.662	.616
30	.646	.66	.484	.644	.653	.49
40	.613	.592	.432	.612	.591	.431
50	.552	.53	.472	.561	.533	.472
60	.5	.457	.477	.489	.458	.48
70	.458	.439	.455	.465	.439	.457
80	.443	.423	.425	.45	.414	.417

5.11.80/310 | 11.11.80/316

Access Tube Number

Profile Depth (cm)	9	12	W2	9	12	W2
10	.642(.464)	.665(.583)	.602(.364)	.656(.491)	.665(.583)	.606(.332)
20	.642	.665	.602	.656	.665	.606
30	.643	.662	.491	.66	.662	.491
40	.61	.594	.42	.62	.594	.433
50	.559	.539	.473	.569	.539	.485
60	.497	.455	.477	.489	.455	.48
70	.464	.431	.442	.46	.431	.451
80	.445	.43	.416	.45	.43	.419

19.11.80/324 | 26.11.80/331

Profile Depth (cm)	9	12	W2	9	12	W2
10	.645(.512)	-	-	.642(.503)	.687(.59)	.612(.366)
20	.645	-	-	.642	.687	.612
30	.65	-	-	.651	.66	.499
40	.609	-	-	.617	.601	.443
50	.549	-	-	.566	.538	.476
60	.491	-	-	.494	.458	.486
70	.46	-	-	.468	.442	.472
80	.439	-	-	.43	.429	.423

5.12.80/340 | 11.12.80/346

Profile Depth (cm)	9	12	W2	9	12	W2
10	.637(.511)	.677(.599)	.628(.38)	.632(.51)	.699(.613)	.617(.365)
20	.637	.677	.628	.632	.669	.617
30	.658	.66	.5	.671	.666	.496
40	.614	.597	.451	.624	.592	.434
50	.561	.533	.485	.565	.536	.475
60	.504	.463	.495	.492	.461	.496
70	.457	.436	.472	.466	.446	.457
80	.44	.425	.424	.444	.425	.424

17.12.80/352 | 14.1.81/14

Profile Depth (cm)	9	12	W2	9	12	W2
10	.644(.51)	-	-	.639(.529)	.662(.581)	.603(.375)
20	.644	-	-	.639	.662	.603
30	.651	-	-	.662	.663	.495
40	.609	-	-	.609	.591	.432
50	.554	-	-	.559	.543	.468
60	.5	-	-	.498	.45	.474
70	.47	-	-	.454	.439	.451
80	.442	-	-	.426	.422	.422

21.1.81/21				28.1.81/28		
	Access Tube Number					
Profile Depth (cm)	9	12	W2	9	12	W2
10	.655(.495)	.677(.584)	.606(.364)	.645(.471)	.646(.526)	.604(.345)
20	.655	.677	.606	.645	.646	.604
30	.664	.669	.482	.648	.654	.489
40	.614	.594	.434	.61	.592	.43
50	.55	.543	.47	.556	.543	.481
60	.5	.458	.471	.494	.455	.477
70	.456	.427	.447	.457	.445	.452
80	.439	.42	.415	.447	.428	.416

4.2.81/35				11.2.81/42		
Profile Depth (cm)						
10	.66(.507)	.673(.553)	-	.646(.49)	.649(.545)	.607(.354)
20	.66	.673	-	.646	.649	.607
30	.655	.659	-	.661	.66	.484
40	.612	.59	-	.603	.588	.433
50	.556	.545	-	.559	.542	.467
60	.493	.45	-	.503	.453	.476
70	.454	.45	-	.459	.444	.448
80	.437	.427	-	.437	.425	.422

18.2.81/49				24.2.81/55		
Profile Depth (cm)						
10	.647(.497)	.662(.535)	.618(.362)	.593(.464)	.608(.524)	.599(.357)
20	.647	.662	.618	.593	.608	.599
30	.656	.669	.483	.651	.601	.475
40	.605	.594	.434	.612	.573	.436
50	.567	.539	.465	.552	.524	.463
60	.494	.559	.478	.493	.461	.474
70	.455	.525	.444	.457	.44	.451
80	.437	.525	.415	.446	.417	.42

3.3.81/62				10.3.81/69		
Profile Depth (cm)						
10	.671(.535)	.667(.6)	.607(.373)	.648(.525)	.674(.638)	.616(.375)
20	.671	.667	.607	.648	.674	.616
30	.653	.676	.501	.656	.668	.488
40	.607	.601	.443	.615	.597	.442
50	.563	.544	.478	.561	.543	.476
60	.492	.464	.487	.491	.462	.481
70	.464	.436	.465	.459	.437	.448
80	.438	.431	.427	.434	.42	.418

11.3.81/78					26.3.81/85	
	Access Tube Number					
	9	12	W2	9	12	W2
Profile Depth (cm)						
10	.65(.506)	.67(.604)	.606(.354)	.665(.537)	.668(.609)	.613(.381)
20	.65	.67	.606	.665	.668	.613
30	.66	.662	.494	.658	.658	.494
40	.621	.596	.436	.611	.595	.435
50	.564	.534	.473	.557	.543	.486
60	.491	.456	.484	.495	.46	.489
70	.465	.438	.462	.456	.434	.459
80	.436	.42	.425	.432	.418	.43
1.4.81/91					8.4.81/98	
Profile Depth (cm)						
10	.685(.537)	.735(.629)	.657(.384)	.652(.511)	.696(.384)	.639(.373)
20	.645	.675	.616	.661	.616	.602
30	.671	.675	.49	.661	.49	.478
40	.617	.594	.434	.608	.434	.436
50	.555	.536	.474	.553	.474	.472
60	.49	.461	.478	.487	.478	.486
70	.463	.449	.449	.468	.449	.456
80	.437	.431	.415	.444	.415	.43

12.4.81/102					22.4.81/112	
Profile Depth (cm)						
10	.662(.519)	.697(.596)	.587(.343)	.573(.449)	.607(.519)	.565(.33)
20	.667	.652	.612	.585	.598	.594
30	.654	.657	.481	.583	.598	.481
40	.616	.595	.433	.604	.58	.437
50	.559	.531	.472	.564	.548	.478
60	.5	.46	.475	.495	.463	.487
70	.459	.436	.454	.457	.439	.454
80	.451	.418	.421	.438	.423	.426

6.5.81/126					15.5.81/135	
Profile Depth (cm)						
10	.644(.505)	.713(.61)	.599(.35)	.566(.444)	.608(.52)	.57(.333)
20	.653	.653	.601	.581	.621	.596
30	.664	.652	.484	.609	.635	.484
40	.616	.601	.448	.594	.588	.428
50	.558	.53	.474	.563	.534	.472
60	.5	.464	.479	.486	.467	.468
70	.462	.445	.455	.457	.437	.442
80	.446	.424	.421	.437	.42	.424

	Access Tube Number					
	20.5.81/140			27.5.81/147		
Profile Depth (cm)	9	12	W2	9	12	W2
10	.599(.47)	.613(.524)	.553(.323)	.631(.495)	.714(.611)	.628(.367)
20	.604	.638	.558	.664	.663	.604
30	.655	.654	.48	.673	.665	.482
40	.596	.578	.432	.607	.599	.424
50	.56	.541	.476	.549	.545	.475
60	.498	.461	.473	.491	.461	.47
70	.459	.438	.452	.454	.435	.44
80	.442	.611	.416	.44	.413	.419

	3.6.81/154			10.6.81/161		
Profile Depth (cm)						
10	.575(.451)	.615(.526)	.572(.334)	.574(.45)	.603(.516)	.544(.318)
20	.607	.626	.606	.577	.596	.568
30	.65	.659	.481	.588	.6	.458
40	.595	.591	.429	.589	.576	.419
50	.562	.541	.464	.558	.535	.457
60	.488	.459	.464	.5	.46	.464
70	.457	.436	.445	.452	.437	.445
80	.434	.424	.415	.443	.419	.41

	17.6.81/168			24.6.81/175		
Profile Depth (cm)						
10	.542(.425)	-	-	.538(.422)	.571(.488)	.5(.292)
20	.552	-	-	.532	.583	.544
30	.557	-	-	.557	.57	.436
40	.588	-	-	.577	.555	.406
50	.548	-	-	.551	.531	.454
60	.493	-	-	.492	.456	.466
70	.457	-	-	.456	.436	.455
80	.435	-	-	.449	.419	.411

	2.7.81/183			8.7.81/189		
Profile Depth (cm)						
10	.524(.411)	.573(.49)	.43(.251)	.499(.391)	.551(.471)	.396(.231)
20	.538	.568	.533	.518	.575	.494
30	.541	.57	.434	.537	.569	.42
40	.558	.547	.398	.546	.546	.386
50	.557	.533	.427	.545	.548	.429
60	.491	.457	.47	.487	.474	.457
70	.464	.441	.444	.457	.448	.434
80	.441	.43	.409	.434	.411	.405

	15.7.81/196			22.7.81/203		
	Access Tube Number					
Profile Depth (cm)	9	12	W2	9	12	W2
10	.491(.385)	-	-	.486(.381)	.56(.479)	.457(.267)
20	.522	-	-	.515	.576	.496
30	.526	-	-	.523	.578	.406
40	.555	-	-	.551	.536	.369
50	.54	-	-	.55	.535	.419
60	.49	-	-	.485	.48	.451
70	.447	-	-	.46	.437	.431
80	.439	-	-	.442	.419	.414

	28.7.81/209			31.7.81/212		
Profile Depth (cm)						
10	.534(.419)	.622(.532)	.556(.325)	.524(.411)	.597(.511)	.509(.297)
20	.564	.668	.595	.544	.622	.574
30	.576	.654	.475	.548	.663	.468
40	.589	.597	.419	.564	.604	.426
50	.56	.534	.457	.551	.546	.457
60	.499	.457	.474	.504	.464	.47
70	.461	.434	.443	.46	.444	.447
80	.442	.425	.413	.433	.417	.414

	3.9.81/246			9.9.81/252		
Profile Depth (cm)						
10	.481(.377)	.555(.475)	.442(.258)	.477(.374)	.559(.478)	.413(.241)
20	.509	.567	.527	.49	.57	.497
30	.516	.571	.432	.511	.558	.411
40	.549	.544	.397	.545	.534	.385
50	.55	.528	.458	.538	.537	.44
60	.501	.46	.459	.497	.463	.457
70	.46	.435	.436	.457	.437	.425
80	.431	.414	.411	.435	.424	.411

	17.9.81/260			24.9.81/267		
Profile Depth (cm)						
10	.51(.4)	.586(.501)	.485(.283)	.525(.412)	-	-
20	.525	.583	.551	.538	-	-
30	.519	.583	.434	.546	-	-
40	.547	.572	.391	.571	-	-
50	.532	.539	.432	.548	-	-
60	.491	.458	.455	.484	-	-
70	.467	.444	.431	.469	-	-
80	.434	.431	.411	.439	-	-

30.9.81/273 7.10.81/280

	Access Tube Number					
Profile Depth (cm)	9	12	W2	9	12	W2
10	.641(.503)	.703(.601)	.558(.326)	.646(.513)	.669(.596)	-
20	.639	.671	.575	.646	.669	-
30	.645	.655	.475	.653	.666	-
40	.6	.598	.41	.602	.594	-
50	.55	.532	.452	.549	.547	-
60	.494	.456	.459	.499	.471	-
70	.453	.441	.436	.457	.44	-
80	.445	.423	.415	.436	.416	-

14.10.81/287 20.10.81/293

Profile Depth (cm)						
10	.562(.439)	.656(.556)	.594(.33)	.549(.424)	.649(.518)	.579(.311)
20	.562	.656	.594	.549	.649	.579
30	.564	.649	.479	.553	.655	.471
40	.594	.585	.424	.59	.596	.425
50	.555	.54	.47	.544	.528	.463
60	.495	.461	.477	.484	.457	.473
70	.461	.44	.449	.457	.441	.453
80	.436	.416	.41	.432	.421	.422

28.10.81/301 4.11.81/308

Profile Depth (cm)						
10	.661(.467)	.669(.593)	.594(.338)	.609(.459)	.662(.558)	.59(.339)
20	.661	.669	.594	.609	.662	.59
30	.662	.672	.477	.658	.649	.487
40	.596	.593	.432	.611	.596	.423
50	.548	.538	.479	.546	.543	.468
60	.502	.457	.466	.492	.455	.467
70	.462	.441	.452	.46	.439	.448
80	.439	.417	.415	.439	.427	.424

11.11.81/315 18.11.81/322

Profile Depth (cm)						
10	.58(.443)	.659(.527)	.589(.335)	.649(.524)	.675(.626)	.574(.343)
20	.58	.659	.589	.649	.675	.574
30	.603	.659	.474	.659	.662	.482
40	.605	.592	.432	.602	.601	.428
50	.556	.534	.471	.565	.542	.461
60	.5	.465	.472	.497	.474	.464
70	.459	.443	.443	.461	.448	.45
80	.445	.415	.424	.448	.425	.421

	25.11.81/329			2.12.81/336		
	Access Tube Number					
Profile Depth (cm)	9	12	W2	9	12	W2
10	.653(.463)	.665(.585)	.615(.355)	.643(.5)	.68(.606)	.602(.348)
20	.653	.665	.615	.643	.68	.602
30	.653	.662	.486	.654	.655	.48
40	.611	.594	.428	.595	.598	.439
50	.555	.536	.474	.568	.546	.477
60	.49	.457	.483	.49	.452	.479
70	.463	.437	.45	.46	.434	.45
80	.447	.427	.421	.448	.43	.42

	9.12.81/343			20.1.82/20		
Profile Depth (cm)						
10	.595(.455)	.662(.573)	.593(.348)	.649(.511)	-	-
20	.595	.662	.593	.649	-	-
30	.643	.676	.489	.654	-	-
40	.614	.604	.434	.604	-	-
50	.575	.525	.468	.554	-	-
60	.493	.461	.478	.492	-	-
70	.458	.433	.453	.458	-	-
80	.441	.424	.414	.439	-	-

	3.2.82/34			10.2.82/41		
Profile Depth (cm)						
10	.633(.468)	.66(.551)	.599(.351)	.653(.472)	.652(.527)	.589(.348)
20	.633	.66	.599	.653	.652	.589
30	.651	.655	.497	.65	.672	.48
40	.602	.589	.444	.612	.606	.433
50	.546	.539	.478	.55	.538	.474
60	.493	.452	.477	.503	.457	.475
70	.471	.425	.45	.457	.434	.455
80	.434	.421	.43	.451	.426	.416

	17.2.82/48			24.2.82/55		
Profile Depth (cm)						
10	.616(.455)	.617(.54)	.606(.334)	.572(.434)	.6(.512)	.597(.326)
20	.616	.617	.606	.572	.6	.597
30	.644	.619	.478	.587	.571	.475
40	.616	.575	.428	.604	.559	.423
50	.556	.525	.469	.559	.537	.459
60	.497	.445	.471	.491	.464	.477
70	.469	.426	.449	.468	.434	.435
80	.448	.417	.417	.444	.422	.406

	3.3.82/62		
Profile Depth (cm)			
10	.649(.525)	-	-
20	.649	-	-
30	.649	-	-
40	.594	-	-
50	.553	-	-
60	.503	-	-
70	.466	-	-
80	.435	-	-

APPENDIX III

SOIL MOISTURE CONTENT (SURFACE LAYERS)

Values are expressed as a percentage of wet weight of soil

Date/Day Number :	18.11.80/323		13.1.81/13	
Sample Location No.	Depth(cm)		Depth (cm)	
(Fig 2.2)	0-2.5	2.5-7.5	0-2.5	2.5-7.5
1A	74.13	72.07	67.2	57.74
2A	73.63	70.21	71.25	77.23
3A	62.03	59.28	53.48	38.74
4A	44.26	26.65	30.5	27.46
5A	49.22	42.25	60.34	58.03
6A	77.9	63.63	78.18	74.59
7A	79.02	74.88	84.31	78.41
8A	36.26	32.73	74.89	50.79
9A	67.58	68.59	70.25	71.14
10A	64.44	63.85	70.99	64.83
W1A	40.7	40.16	43.48	39.26
W2A	57.91	53.58	47.32	32.97
W3A	-	-	-	-
W4A	26.85	33.22	70.24	70.2

Date/Day Number :	3.2.81/34		24.2.81/55	
1A	24.41	26.85	50.75	23.57
2A	68.34	71.79	76.6	60.95
3A	61.64	67.21	75.1	60.73
4A	38.6	23.01	73.37	57.7
5A	31.99	28.93	80.67	68.88
6A	74.0	53.22	77.18	71.55
7A	57.69	57.0	81.34	60.26
8A	67.64	68.03	81.67	75.04
9A	70.24	71.57	81.46	95.7
10A	67.23	69.0	81.25	79.59
W1A	65.17	65.63	77.92	36.61
W2A	37.61	35.99	79.53	47.85
W3A	61.98	47.98	78.94	59.11
W4A	51.16	58.94	78.64	59.76

Date/Day Number : Sample Location No.	16.3.81/75 Depth(cm) 0-2.5	2.5-7.5	7.4.81/97 Depth (cm) 0-2.5	2.5-7.5
1A	70.03	68.97	74.28	46.43
2A	72.34	64.27	67.87	62.32
3A	75.77	72.19	69.84	66.33
4A	85.24	58.73	78.44	63.74
5A	87.18	87.61	87.77	84.11
6A	78.23	84.1	78.58	82.13
7A	78.05	78.68	79.52	79.75
8A	77.32	71.35	84.16	83.18
9A	84.27	82.55	77.07	74.53
10A	79.83	74.72	74.59	72.98
W1A	51.86	29.92	65.78	34.67
W2A	57.56	50.79	53.09	33.08
W3A	72.08	76.88	77.76	60.98
W4A	21.72	34.8	78.2	36.93

Date/Day Number :	13.4.81/103		14.5.81/134	
1A	70.91	44.31	47.93	28.53
2A	69.16	63.61	56.49	45.36
3A	69.91	68.5	43.95	53.61
4A	60.24	23.11	55.93	45.3
5A	90.07	90.78	76.73	76.94
6A	97.81	84.95	49.32	71.75
7A	75.19	69.98	57.9	53.56
8A	78.13	71.13	51.75	75.05
9A	77.61	68.51	64.28	54.95
10A	76.99	73.49	70.4	58.99
W1A	70.61	40.19	65.83	56.17
W2A	66.91	40.59	70.92	46.28
W3A	67.0	73.83	69.58	74.28
W4A	61.78	73.88	70.75	58.75

Date/Day Number : Sample Location No.	18.6.81/169 Depth(cm) 0-2.5	18.6.81/169 Depth(cm) 2.5-7.5	16.7.81/197 Depth (cm) 0-2.5	16.7.81/197 Depth (cm) 2.5-7.5
1A	64.72	64.29	35.96	43.69
2A	53.41	54.61	42.28	59.9
3A	62.43	65.69	59.17	48.1
4A	72.27	75.76	71.07	44.77
5A	44.98	67.93	65.21	64.61
6A	61.63	71.27	55.09	69.1
7A	57.54	65.0	38.84	48.5
8A	53.82	69.16	67.73	75.97
9A	67.81	74.95	63.7	70.25
10A	76.53	69.47	67.95	64.02
W1A	72.59	25.05	66.23	29.69
W2A	62.85	34.27	73.9	34.25
W3A	54.42	47.02	68.84	41.19
W4A	54.69	33.04	51.14	36.92

Date/Day Number :	10.9.81/253		8.10.81/281	
1A	19.9	60.09	52.46	29.59
2A	57.42	65.84	63.4	65.44
3A	62.26	64.72	74.05	65.77
4A	30.28	18.94	53.38	22.14
5A	14.92	60.96	77.7	64.13
6A	19.42	74.69	86.11	84.09
7A	55.13	40.54	77.68	76.28
8A	66.42	72.66	82.79	75.48
9A	66.85	73.43	75.54	78.84
10A	52.3	73.93	83.04	72.91
W1A	30.63	28.11	77.5	47.28
W2A	27.33	41.89	77.71	69.02
W3A	25.56	48.74	75.36	58.39
W4A	30.68	44.66	78.66	49.99

Date/Day Number : Sample Location No.	3.11.81/307 Depth(cm) 0-2.5	2.5-7.5	1.12.81/335 Depth (cm) 0-2.5	2.5-7.5
1A	70.59	-	63.93	59.36
2A	70.52	70.08	73.31	71.65
3A	73.81	71.52	77.56	68.83
4A	74.2	42.25	81.99	39.11
5A	79.74	83.61	87.57	50.3
6A	82.44	81.14	80.25	84.31
7A	75.54	64.21	70.57	57.09
8A	72.22	72.72	82.14	74.77
9A	74.72	76.21	76.55	75.38
10A	74.16	80.39	76.94	76.89
W1A	68.91	59.96	74.24	40.95
W2A	49.74	39.97	73.27	74.63
W3A	78.36	65.26	72.76	47.6
W4A	77.12	51.65	75.11	68.89

Date/Day Number :	26.1.82/26		25.2.82/56	
1A	68.45	59.25	-	-
2A	75.11	72.63	-	-
3A	76.02	59.56	70.77	69.98
4A	85.72	73.35	-	-
5A	83.1	86.89	-	-
6A	78.69	83.39	-	-
7A	80.28	77.68	73.6	62.34
8A	81.06	71.7	-	-
9A	76.95	67.12	78.31	66.76
10A	80.4	77.41	-	-
W1A	74.33	25.33	-	-
W2A	74.16	51.07	66.84	56.11
W3A	69.92	51.17	-	-
W4A	69.08	59.25	-	-

APPENDIX IV

DERIVATION OF RAINFALL MATRIX FOR UNIT HYDROGRAPH ANALYSIS

Rainfall matrix, |p| for flood event number 1, shown in Table 5.2 (p.251)& Figs. 5.14 and 5.15 (p.258,259).Number of rows = number of response runoff observations. The matrix is composed by 'filling out' with zeros the one column matrix of net rainfall values.

	(0.05)	0	0	0	.	.	.
	(0.05)	0.05	0	0	.	.	.
	(0.2)	0.05	0.05	0			
	(0.4)	0.2	0.05	0.05			
	(0.5)	0.4	0.2	0.05			
column of	(0.3)	0.5	0.4	.			
net rain	(0.2)	0.3	0.5	.			
increments	(0.2)	0.2	0.3	.			
	(0.05)	0.2	0.2				
	(0.3)	0.05	0.2				
	(0.2)	0.3	0.05				
	0	0.2	0.3				
	0	0	0.2				
	0	0	0				
	0	0	0				
	0	0	0				
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	0	0	0				

APPENDIX V

MEASURED RAINFALL UNDER WOODLAND (WINTERGILL PLANTATION)

(Standard Meteorological Office Mk. II gauge)

PERIOD AND DAY NUMBER	RAINFALL (mm)
9.10.80 - 14.10.80 283-288	5.1
15.10.80 - 21.10.80 289-295	25.2
22.10.80 - 28.10.80 296-302	14.7
29.10.80 - 4.11.80 303-309	2.6
5.11.80 - 11.11.80 310-316	16.2
12.11.80 - 18.11.80 317-323	13.7
19.11.80 - 25.11.80 324-330	8.1
26.11.80 - 16.12.80 331-351	55.5
17.12.80 - 23.12.80 352-358	9.6
24.12.80 - 20. 1.81 359-20	18.7
21. 1.81 - 27. 1.81 21-27	3.1
28. 1.81 - 3. 2.81 28-34	4.6
4. 2.81 - 17. 2.81 35-48	17.7
18. 2.81 - 10. 3.81 49-69	Record missing
11. 3.81 - 1. 4.81 70-91	58.0
2. 4.81 - 7. 4.81 92-97	7.6
8. 4.81 - 14. 4.81 98-104	9.1
15. 4.81 - 22. 4.81 105-112	0.0
23. 4.81 - 28. 4.81 113-118	15.2
29. 4.81 - 5. 5.81 119-125	15.7
6. 5.81 - 12. 5.81 126-132	4.6

PERIOD AND DAY NUMBER	RAINFALL (mm)
13. 5.81 - 19. 5.81 133-139	4.1
20. 5.81 - 27. 5.81 140-147	18.2
28. 5.81 - 2. 6.81 148-153	11.6
3. 6.81 - 9. 6.81 154-160	3.1
10. 6.81 - 16. 6.81 161-167	3.6
17. 6.81 - 23. 6.81 168-174	1.6
24. 6.81 - 30. 6.81 175-181	7.1
1. 7.81 - 7. 7.81 182-188	5.6
8. 7.81 - 14. 7.81 189-195	8.1
15. 7.81 - 21. 7.81 196-202	11.1
22. 7.81 - 28. 7.81 203-209	22.7
29. 7.81 - 4. 8.81 210-216	0.0
5. 8.81 - 10. 8.81 217-222	39.4
11. 8.81 - 18. 8.81 223-230	0.0
19. 8.81 - 25. 8.81 231-237	3.6
26. 8.81 - 2. 9.81 238-245	0.0
3. 9.81 - 8. 9.81 246-251	0.0
9. 9.81 - 15. 9.81 252-258	11.1
16. 9.81 - 22. 9.81 259-265	8.6
23. 9.81 - 29. 9.81 266-272	12.1
30. 9.81 - 6.10.81 273-279	Record missing
7.10.81 - 13.10.81 280-286	10.6
14.10.81 - 20.10.81 287-293	1.6
21.10.81 - 27.10.81 294-300	22.2

PERIOD AND DAY NUMBER	RAINFALL (mm)
28.10.81 - 4.11.81 301-308	4.1
5.11.81 - 10.11.81 309-314	Trace
11.11.81 - 17.11.81 315-321	Trace
18.11.81 - 24.11.81 322-328	13.2
25.11.81 - 1.12.81 329-335	8.1
2.12.81 - 9.12.81 336-343	4.1
10.12.81 - 19. 1.82 344-19	100.0 (minimum - collecting bottle full)
20. 1.82 - 26. 1.82 20-26	Record unreliable
27. 1.82 - 3. 2.82 27-34	4.1
4. 2.82 - 9. 2.82 35-40	3.1
10. 2.82 - 16. 2.82 41-47	2.1
17. 2.82 - 23. 2.82 48-54	2.1
24. 2.82 - 2. 3.82 55-61	7.6
ANNUAL 17.12.80 - 9.12.81 (28 days missing)	399.4

PERIOD AND DAY NUMBER	PE(P)	AE_1	AE_2	PE(PM)	AE_3	AE_4	AE_5	AE_6
26.11.80 - 16.12.80 331-351	5.6 (0.3)			31.5 (1.5)				
17.12.80 - 13.1.81 352-13	6.4 (0.2)			40.4 (1.4)				
14.1.81 - 10.2.81 14-41	7.5 (0.3)	7.5 (0.3)	7.5 (0.3)	26.6 (1.0)	26.6 (1.0)	26.6. (1.0)	26.6 (1.0)	26.6.(1.0)
11.2.81 - 9.3.81 42-68	12.3 (0.5)	12.3 (0.5)	11.5 (0.4)	30.8 (1.1)	30.8 (1.1)	30.8 (1.1)	30.8 (1.1)	30.8 (1.1)
10.3.81 - 7.4.81 69-97	28.0 (1.0)	27.0 (0.9)	27.0 (0.9)	51.4 (1.8)	51.4 (1.8)	51.4 (1.8)	51.4 (1.8)	51.4 (1.8)
8.4.81 - 5.5.81 98-125	56.7 (2.0)	56.7 (2.0)	45.9 (1.6)	35.0 (1.3)	35.0 (1.3)	35.0 (1.3)	35.0 (1.3)	35.0 (1.3)
6.5.81 - 2.6.81 126-153	89.1 (3.2)	81.5 (2.9)	58.4 (2.1)	50.5 (1.8)	50.5 (1.8)	50.5 (1.8)	50.5 (1.8)	50.5 (1.8)
3.6.81 - 1.7.81 154-182	92.1 (3.3)	38.8 (1.3)	37.8 (1.3)	62.7 (2.2)	62.7 (2.2)	62.4 (2.2)	62.7 (2.2)	62.7 (2.2)
2.7.81 - 22.7.81 183-203	60.5 (2.9)	42.3 (2.0)	36.5 (1.7)	37.9 (1.8)	37.9 (1.8)	37.8 (1.8)	37.9 (1.8)	37.9 (1.8)

PE(P) Penman potential evapotranspiration

PE(PM) Penman-Monteith potential evapotranspiration

Potential Evapotranspiration/ Soil Moisture Model Combination:-

AE_1 Penman/Grindley Total profile soil moisture deficits

AE_2 Penman/Grindley Layer soil moisture deficits

AE_3 Penman-Monteith/Grindley Total profile soil moisture deficits

AE_4 Penman-Monteith/Grindley Layer soil moisture deficits

AE_5 Penman-Monteith/MORECS Total profile soil moisture deficits

AE_6 Penman-Monteith/MORECS Layer soil moisture deficits

APPENDIX VI : POTENTIAL AND ACTUAL EVAPOTRANSPIRATON DATA FOR WOODLAND
[AE predicted from soil moisture models.
All figures in mm except those in brackets which indicate daily mean equivalents (mm day $^{-1}$)]

PERIOD AND DAY NUMBER	PE(P)	AE_1	AE_2	PE(PM)	AE_3	AE_4	AE_5	AE_6
28.7.81 - 2.9.81 209-245	111.8 (3.0)	89.0 (2.4)	49.0 (1.3)	69.4 (1.9)	69.4 (1.9)	69.4 (1.9)	69.4 (1.9)	69.4 (1.9)
3.9.81 - 29.9.81 246-272	54.9 (2.0)	37.1 (1.4)	37.3 (1.4)	47.9 (1.8)	47.9 (1.8)	42.9 (1.6)	47.9 (1.8)	45.7 (1.7)
30.9.81 - 27.10.81 273-300	25.8 (0.9)	25.5 (0.9)	25.3 (0.9)	36.4 (1.3)	36.4 (1.3)	36.4 (1.3)	36.4 (1.3)	36.4 (1.3)
28.10.81 - 24.11.81 301-328	12.0 (0.4)	12.0 (0.4)	12.0 (0.4)	16.6 (0.6)	16.6 (0.6)	16.6 (0.6)	16.6 (0.6)	16.6 (0.6)
25.11.81 - 8.12.81 329-342	4.0 (0.3)	4.0 (0.3)	4.0 (0.3)	0.0 (0.0)	0.0 (0.0)	0.0 (0.0)	0.0 (0.0)	0.0 (0.0)
20.1.82 - 2.2.82 20-33	8.0 (0.6)			22.4 (1.6)				
3.2.82 - 2.3.82 34-61	18.0 (0.6)			48.3 (1.7)				
ANNUAL								
14.1.81 - 8.12.81	554.7	434.0	357.4	465.2	465.2	459.8	465.2	463.0
17.12.80 - 8.12.81	561.1			505.6				

APPENDIX VI (continued) : POTENTIAL AND ACTUAL EVAPOTRANSPIRATON DATA FOR WOODLAND

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