The Role And Significance Of Surface And Subsurface Hydrology On Gully Head Growth In South East Spain

Carolyn Faith Francis

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CONTENTS

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ABSTRACT

This thesis seeks to extend the work of Thornes and Scoging on hillslope processes in south east Spain by paying particular attention to the relative roles of surface and sub-surface water movement in gully head change on two contrasting soil types. Gully growth is a major agent of erosion in semi-arid environments, yet studies to date have assumed the dominance of surface wash, and only speculated on the role of subsurface water movement.

The sampling design was based on the hillslopé hydrological cycle and incorporates measures of precipitation, infiltration, and runoff and subsurface flow on catchment areas above gully heads. Additional data on vegetation and some soil properties were collected. The field work was undertaken on three occasions in the summer, autumn and spring of 1982/3 to examine seasonal variations.

The analysis of surface flow was hindered by the drought which meant there were only 10 rain days in 1982, and only one storm occurred during the field sessions on 26 November 1982. Despite this several observations can be made. Wash volumes were twice as high on the marl on 26 November. Both lithologies are susceptible to relatively high erosion rates by surface wash although rates tend to be higher on the marl, and there is considerable variation on both lithologies. However for neither lithology is the amount of sediment transported sufficient to fill in the gullies in the medium term. The analysis of the subsurface hydrology shows that saturated conditions were not monitored anywhere, and maximum soil moisture values reached between 50-60% saturation on the marl and conglomerate soils. There are marked seasonal variations in soil moisture and most of the variation occurs in the upper horizons. Flux rates are negligble on the marl and dominantly in the vertical plane. On the conglomerate rates are much faster, and throughflow may well occur on occasions, and at rates exceeding evapotranspiration. This will contribute to wetter conditions around and in gullies on the conglomerate.

LIST OF FIGURES — Page

LIST OF TABLES

LIST OF PHOTOGRAPHS

CHAPTER ONE INTRODUCTION

1.1 Erosion Processes In Semi-Arid Environments

This thesis is concerned with the processes of wash and gullying in semi-arid Spain, focusing on the relative roles of surface and subsurface flow to the gully head. Such research is necessary for both academic and economic reasons as gully development is itself an interesting geomorphological issue, and leads to considerable problems in land degradation and water supply.

The significant factors associated with erosion vary at different scales (Arnett 1979). At the global scale patterns of erosion are attributed to climatic parameters and semi-arid areas appear to have relatively low erosion rates. Work by Fournier (1960) and Strakhov (1967), although varying in detail, indicate that the amount of erosion increases with precipitation. At the basin scale other parameters come into play. This is shown by Stoddart (1969) who examines the magnitude of the solid load for the largest world drainage basins using data from Strakhov. He shows that there are large sediment loads from the Ganges and Amazon basins and low rates for high latitude drainage basins in keeping with the global patterns, but also high sediment loads from semi-arid catchments such as the Tigris-Euphrates, Hwang Ho and Yangtze rivers. The largest suspended sediment load measured (according to Walling and Webb 1983) is 25 600 t $km^{-2}yr^{-1}$ for the Dali River, a tributary of the Yangtze river which drains gullied loess in a semi-arid area.

Langbein and Schumm (1958) examined sediment yield for a large number of drainage basins in the United States and suggest that the relationship between suspended sediment and precipitation forms a peaked curve with the maximum sediment yield occurring where the effective precipitation is 250-350 mm (the effective precipitation being the amount of rainfall needed to produce a known amount of runoff based on a reference temperature of 50°C). This implies maximum erosion in semi-arid areas: elsewhere erosion is limited by precipitation deficit or vegetation growth. This idea has been developed by various authors (summarised in Walling and Webb 1983), but as Walling and Kloo (1979) show, the relationship between suspended load and precipitation is very variable. This reflects the influence not only of precipitation, but also catchment size, geology, relief and land use, on erosion rates at the drainage basin scale. The effect of land use in particular can be enormous causing "accelerated" erosion rates. For example Young (1969) puts natural erosion rates between 0.0045 to 0.045 kg m^{-2} yr^{-1} for moderate to steep relief, but accelerated erosion rates on agricultural land between 4.5 to 45 kg m^{-2} yr^{-1}. Thus erosion rates may vary considerably in semi-arid areas. The erosion rate relative to the rate of soil development is also important. If the latter is slow, a medium rate of erosion may be sufficient to initiate widespread erosion and landform change. It is significant that the greatest extent of badlands lies in semi-arid areas.

Finally at the slope scale soil erosion is controlled by such factors as hillslope form, vegetation cover, surface roughness, and the predominance of surface versus subsurface flow.

Erosion occurs by a range of processes in semi-arid areas including rainsplash, wash, rilling, gullying, river flow, mass movement and wind action. They may be found in different combinations and locations within a drainage basin, they may occur at differing rates and they may have various effects on landform development. An assessment of these processes should isolate those which account for most of the erosion in semi-arid regions.

Rainsplash erosion occurs throughout the area affected by a storm. The effectiveness of the drops for detaching and transporting soil particles varies according to the type of storm, the dynamics of the drops, the soil type, the vegetation cover, the depth of surface water and the slope (Evans 1980). The impact of rainsplash has been examined in the field (Schumm 1956b, Moseley 1973, Morgan 1978, Van Asch 1983), laboratory (Bryan 1974, 1979, Noble and Morgan 1983), and by modelling (Meyer and Wischmeier 1969). Van Asch (1983) proposes that under certain circumstances rainsplash erosion can be as significant as wash for sediment transport on cohesive soils in Italy. Erosion rates by splash may be up to 100 cm^3/cm yr ^{1} on bare surfaces but

Kirkby (1980b p10) argues that with vegetation cover rates may be only 2-5 $cm^3/cm\ yr^{-1}$ even in semi-arid conditions. Kirkby suggests that rainsplash can be ignored as a major transporting agent save near the divide, or when overland flow does not occur during a storm. Rainsplash may still be a significant agent for detaching particles ready for entrainment by surface wash.

Observations of surface wash (Emmett 1970) show that it is variable in character. Flow occurs not as a uniform sheet but in concentrated anastomising paths separated by thin films of water due to the effect of the local topography and vegetation. Flow can be unsteady, laminar, turbulent, subcritical or supercritical as a result of spatial and temporal variability in rainfall and soil characteristics. Subsequently the hydraulic pattern (and from that erosion and deposition patterns) can vary considerably within a small area (Scoging 1982, Roels 1984). Production of wash depends on a variety of climatic, edaphic and biotic parameters (Kirkby 1978, Kirkby and Morgan 1980) and may cover less area than rainsplash. However surface wash transports far more sediment and is more significant as a transporting agent.

The concentrated nature of flow in rills, gullies and rivers means that although the area affected may be smaller than for wash, the erosive and transporting potential in that area is much greater. Whether gullies or surface wash contributes most to erosion in gullied basins depends partly on the extent of the gullies. Thus Piest and Spomer (1968) estimate that sheet and gully erosion average 80% and 20% respectively for a loessial region, although in other areas gullies may be responsible for most of the erosion in the drainage basin (Leopold, Wolman and Miller 1964, Seginer 1966). As most of the flow in gullies is surface wash from upslope, gullying is affected by wash erosion, but most studies suggest the dominant source of suspended sediment (and hence erosion) in gullies is from the gully sides, (as with rivers). Faber and Imeson (1982) estimate that the amount of runoff in gullies represents only a small proportion of the total runoff in many gullied river basins.

Mass movement occurs on a wide scale from soil creep to deep seated landslides in semi-arid areas. The impact of mass movement on denudation is poorly quantified (Thornes 1976). Leopold, Emmett and Myrick (1966) estimate that mass movement accounts for 0.7% only of the total sediment yield. The findings from Leopold, Wolman and Miller (1964), who measured creep using erosion pins in the Arroyo de los Frijoles near Santa Fe, New Mexico, suggest that the average rate of creep is 8.4mm per year. Others have also tried to assess the significance of creep (Schumm 1956b) for semi-arid environments, but although creep occurs it is very difficult to quantify partly because the methods used interfere with the creep itself.

There is plenty of field evidence of mass movement in semi-arid areas for example in the mountainous zones of south east Spain where large landslide scars and unstable slopes are common. On a smaller scale shallow failures on steep slopes or on river banks are possible mechanisms for gully initiation (Thornes 1976) and viscous debris flows are partly responsible for the development of the many fan complexes.

Aeolian processes too are poorly quantified for semi-arid environments outside the major dune complexes. In south east Spain field observation suggests that wind erosion and transportation are very significant, particularly during drought years.

From this brief description of relative rates of erosion processes sheet wash and gullying appear to be two of the most important processes in semi-arid areas. Surface wash occurs over large areas and according to Young (1974) is the major source of erosion by water in semi-arid environments, but for large single events gullying produces large amounts of sediment. Where gully drainage densities are high they probably contribute more to erosion than surface wash.

Gully Erosion

Gullies are difficult to define as can be seen from the relevant literature. Imeson and Kwaad (1980) list three important factors associated with gullies:

1 Gullies form where water has concentrated either naturally or through disturbance by man.

2 Gullies are often limited to unconsolidated slope materials, weak rocks eg shales, marls, and deeply weathered soils which are easily erodible.

3 Flow in gullies is ephemeral occurring during and just after storms.

Imeson and Kwaad note that the US Soil Conservation Service adds that a gully is also a form which cannot be obliterated by normal tillage.

Graf (1983) differentiates between gullies and arroyos where gullies may be 'V' shaped or 'U' shaped but rarely associated with a stream whereas arroyos are roughly rectangular in cross section with a major stream channel on the floor. However Ireland, Sharpe, and Eargle (1939) found many gullies followed old drainage lines. Thornes (1976) similarly differentiates between 'ramblas' and 'barrancos' (gullies) in the Spanish context. Ramblas are flat floored, wide ephemeral channels often with several metres of coarse alluvium, but the barrancos are smaller, unvegetated, steep sided channels without a well-developed alluvium fill.

In addition Brice (1966) suggests that a gully is partly definable by its size, being larger than a rill, although the dimensions used to define a gully are arbitrary (Bradford and Piest 1980). Brice also notes that a gully may be distinguished by recent extension of length and the steepness of the gully walls. However true these points may be, they reflect the recent erosional history and soil characteristics, especially cohesiveness. More precise morphological definitions are difficult due to wide variations in gully forms.

Gullies occur worldwide, in different climatic zones. Reports from high precipitation regions include Sarawak (Baillie 1975), and Hong Kong (Berry and Ruxton 1960, Lam 1977). In temperate regions gullies have been reported in England (Tuckfield 1964, Arnett 1979), New Zealand (Blong 1970), Eastern USA (Ireland, Sharpe and Eargle 1939), the more humid parts of Africa such as Zaire (Ologe 1972) and Rhodesia (Stocking 1977).

The bulk of reports come from semi-arid and arid areas such as the Western USA (Schumm and Hadley 1957, Tuan 1966, Beaty 1959, Heede 1970, 1974, Cooke and Reeves 1976, and Bradford and Piest 1977), Canada (Faulkner 1974, and Campbell 1982), Spain (Thornes 1976, Harvey 1982), Italy (Alexander 1982), Israel (Seginer 1966, Nir and Klein 1974, Yair et al 1980), and Africa (de Ploey 1974 in Tunisia, Imeson and Kwaad 1980 in Morocco, and Faber and Imeson 1982 in Lesotho).

It appears that areas most sensitive to gully formation tend to be semi-arid where continued development has formed extensive badlands, and conceivably such areas represent the optimal conditions for gully growth. The combination of low annual rainfall and marked seasonality leads to low vegetation cover for much of the year, but considerable erosion potential in the wet season. Together with poorly consolidated materials, steep slopes, and intervention by man these factors account for the degree of erosion in such environments. Erosion in badlands is perceived as rapid in geological terms, particularly for certain unconsolidated materials which have low limiting thresholds (Campbell and Honsaker 1982) which may result in an equilibrium between form and process. However should the badlands prove to be relic features this would not be the case. Badlands have been described as condensed, simple analogues for similar landforms elsewhere (Campbell and Honsaker 1982). Thus if the nature of gully growth is understood in its most basic form in badlands, the understanding may well be applied to less extreme environments.

The initation and extension of gullies pose a number of serious economic problems both in the gully catchment and downstream. The loss of land around gully heads may be considerable, affecting agricultural production and infrastructure for example roads, railways and bridges (Ireland et al 1939, Piest and Spomer 1968). In semi-arid areas such land may be economically marginal, but still play an important role in the regional economy. Downstream there are problems of water supply and quality. Increased sediment loads in rivers can degrade floodplain soils (Trimble 1977), particularly following flood deposition. The suspended sediment silts up reservoirs reducing water storage capacities in areas already suffering from water shortage. Furthermore the development of extensive gully networks may increase the flashiness of storm floods as the time from rainfall to flood peak falls. Water quality may also be affected by increased concentrations of soluble salts. The solid sediment load for the Rio Fardes at Posito, Guadix basin, Spain, is about 16 tonnes km^{-2} yr^{-1}, but dissolved loads are estimated at 137.5 tonnes km^{-2} yr^{-1} (Thornes 1984). Although the changes in water chemistry following excessive denudation have not been studied, greater attention is being paid to chemical weathering in semi-arid areas (Imeson, Kwaad and Verstraten 1982).

At first glance extensively gullied areas appear to have very high erosion rates. However various authors caution that gullies do not necessarily mean high erosion (Nir and Klein 1974, Imeson and Kwaad 1980). This theme is explored in a series of papers (Wise, Thornes and Gilman 1982, Thornes 1984, Gilman and Thornes 1985) for the case in south east Spain. In a number of badland areas evidence from archaeology, sediment load in rivers, reservoir silting rates, and modelling suggests overall erosion rates are in fact only about 25 tonnes km^{-2} yr^{-1}. Thornes (1984) explains this apparent paradox by two sets of possible explanations. The first set considers contemporary process and the second follows an historical or evolutionary approach. In the first set either:

1 Locally measured rates are too high due to instrumental or design errors.
2 Overall rates reflect the jerkiness of the sediment transport process.
3 Only a small part of the basin contributes to sediment production.

The second set involves processes which operated differently in the past so that the landscape evolved under a regime different from the present one. This can be explained by climatic change, and changes brought about by human activity such as deforestation, and grazing.

Within a semi-arid drainage basin erosion rates may be variable so that local erosion rates are high, for example around headwardly expanding gullies, but regional rates are on average low with large segments of relatively stable areas. Such 'stable' areas may include intensively dissected badlands where headward growth is saturated, and erosion is limited to the denudation of the small divides.

In summary gully growth in semi-arid areas can be a major source of erosion and landform change, affecting the agricultural economy of a region. Gully networks may develop to form spectacular badlands. So for both economic and geographical considerations gully growth is an important area for research.

1.2 Research Perspectives On Gully Erosion

Research on gully formation has been approached from different perspectives. Firstly, it has been related to historical sequences of events. Secondly, it has been examined with reference to morphological descriptions and process measurement. Thirdly, attempts have been made to model gully growth. These various avenues of research are now discussed to evaluate the work that has already been done, to highlight areas where future research may be profitable, and to ascertain the most appropriate ways to approach new research.

Historical Approach To Gully Growth

Gully formation has been examined by establishing the historical sequence of gully extension and relating it to environmental factors as identified in written or non-written evidence. Thus causes of gully growth are explained within the framework of the contemporaneous environment. Graf (1983) refers to this approach as the 'paradigm of origin' describing it as the earliest, most widely adopted approach which suited the theory of landform evolution of the time.

A considerable number of authors have set gully development wholly or partly within the historical context (Schumm and Hadley 1957, Tuan 1966, Cooke and Reeves 1976, and Bradford et al 1978). Cooke and Reeves (1976) assess this approach with reference to arroyos and summarise the main points.

A wide variety of data sources are available for environmental reconstruction including written records (both government and personal), interviewing 'Old Timers', and field work. The latter may involve mapping (especially from sequential photographs or maps), measuring erosion rates, and obtaining data on historical settlement patterns, archaeological evidence, stratigraphy, dendrochronology, pollen analysis and macrofossil analysis.

Cooke and Reeves (1976) tried to reconstruct the erosional history of two regions. They then tried to evaluate the relevance of different hypotheses of gully growth on a regional and local basis for both areas. Hypotheses for gully initation include the following:

1 Climatic Change. Both increases and decreases in precipitation have been cited as causes of gully development. Tuan (1966) notes that most active gullying occurs during intense storms even though such storms may temporarily improve the vegetation cover. Brice (1966) suggests that gullies develop not during droughts, but during one or two wet years within a generally dry era. On a longer time scale a change to a drier climate exacerbates gullying by reducing the protective role of vegetation. Alternatively a wetter climate, despite encouraging vegetation, may increase the amount of water available for erosion. Changes in rainfall intensity for example from light frequent rains to sporadic intense storms may also induce gullying and similar effects may also occur through changes in temperature and evapotranspiration.

2 Natural vegetation changes caused by mechanisms other than climate, for example fire or seral succession. Certain areas are more sensitive to change, for example riverine vegetation.

3 Anthropogenic changes, particularly through deforestation, overgrazing and poor agricultural practices, initiate gullying.

4 A variety of geomorphological factors induce gullying such as the sinking of ground (Rubey 1928, Buckham and Cockfield 1950), old drainage lines (Ireland et al 1939), and piping (Heede 1971, Bryan and Yair 1982).

5 Random frequency-magnitude variations such as extreme events (Wolman and Miller 1960, Thornes 1976, Harvey 1984), or differential flow in drainage basins leading to oversteepening (Schumm and Hadley 1957) may also trigger off incision. However these mechanisms only operate if the change triggers instability.

The historical approach to gully growth may build up a picture of the sequential development of gullies and associated causes. There are however several drawbacks to this approach. Data sources available are often inadequate, of poor quality, and only indirectly related to the problem, as Cooke and Reeves point out. Written records only extend to the mid-19th century in the south west USA. Of these the government surveys are accurate, but also often inadequate as gullies are not necessarily mentioned (or recognised in the field), and surveys for roads and railways often only refer to adjoining strips of land. Even precipitation data are subject to difficult interpretation due to poor siting, and instrument or measuring errors. As with crop and cattle records, such data only infer the state of the land, and are not a measure of gullying. Interviewing is also fraught with many problems as Cooke and Reeves describe at length. The utility and coverage of field evidence depends on three groups of factors, the recording, the preservation of the data, and the interpretation.

In the first place parts of the system have to be sensitive enough to record environmental changes, and the degree of sensitivity will vary markedly for different elements (Brunsden and Thornes 1979). Thus if two types of data for two localities are chosen one may show evidence of environmental change and the other stability. The relationship between change (process) and response (morphology) can be a complicated array of time lags and phases (Allen 1974). Bradford and Piest (1980) note that gullies can develop quickly within years, but the field evidence of climatic change could take decades to develop. It is feasible that evidence from different sources may show a complicated relationship and even be incompatible. For example Tuan (1966) found that evidence from pollen and tree ring analysis supported at least two different theories of climatic change for New Mexico 1000-1400 AD.

Secondly data are preferentially preserved in certain locations for example pollen in peat bogs, and destroyed in others for example near actively eroding areas. Preservation is a patchy and precarious business subject, for example, to episodic cut and fill sequences (Schumm and Hadley 1957, Womack and Schumm 1977, and Blong 1970).

Finally these factors make interpretation difficult. On the one hand the development and preservation of evidence are severely handicapped, and on the other the initiation of degrading conditions for gully growth may not be associated with environmental change. Thresholds intrinsic to the system, if crossed, cause rapid establishment of incision cycles with no change in the external variables (Schumm 1973, and 1979). Graf (1979) alludes to the complexity of thresholds, where one may be dealing with not one but many thresholds so that within a small area there may be marked variations in morphology as different groups of thresholds are exceeded. This variation in sensitivity is a result of changes in geology, soils, vegetation, topography, microclimate, human impact etc and leads to 'complex response' (Schumm 1973) where a simple change in conditions is manifested in many different ways in different parts of the system. The last point on interpretation is that landforms may be convergent such that originally dissimilar forms evolve to similar ones, and then great care is needed to distinguish between the forms and interpret their relationships to any environmental change.

It can be difficult to determine the environmental conditions leading to gully growth by historical reconstruction. In past work hypotheses have often been too simplistic involving phenomena which are poorly understood (Thornes 1983). Furthermore this approach is inferential depending on low quality data often only indirectly related to gully extension.

Morphology/Process Approach

The second approach to studying gully growth involves the description, classification and measurement of morphology and process. Field work is reinforced by laboratory and numerical techniques, the final result being rates of processes, correlation matrices between variables and gully erosion, and multiple regression equations to estimate gully extension.

One of the earliest and best known morphological descriptions of gullies is by Ireland, Sharpe and Eargle (1939) studying gullies on the Piedmont of South Carolina. As a result of their observations they distinguished six characteristic gully forms - linear, bulbous, dendritic, trellis, parallel and compound. They also classified active gully heads either by their longitudinal section (inclined, vertical, cave or vegetated) and/or plan form (pointed, notched, rounded, and digitate). Ireland et al also describe a four stage sequence of gully initiation, growth, maturity and stabilisation. Imeson and Kwaad (1980) built on this and developed a four fold classification of gully type, each class reflecting different dominating processes. Jan de Ploey (1974) classified the gullies he studied in Tunisia into axial, digitate and frontal systems.

Detailed accounts of gully morphology are available from a number of authors. Ologe (1972) describes gully headscarps in Zaria. Brice (1966) and Bradford, Piest and Spomer (1978), for example, describe longitudinal profiles, gully walls and floors, and plunge pools in gullies. Heede, in a series of papers (1967, 1970, 1971, and 1974) describes various gully systems in the western USA with particular reference to the difference between continuous and discontinuous forms, and the morphological impact of pipes. The morphological role of pipes is also described by Baillie (1975), and Bryan and Yair (1982).

Many processes acting on the head cuts have been described, though few measured. These include abrasion, mud drips, mud trickle, washing, spalling, sloughing, puddling and transport of debris out of the plunge pool (Ologe 1972). The cycle of gully head cutting is a response to these processes plus various types of mass movement associated with flowing water (Ireland et al 1939). The latter includes undercutting of the gully floor and collapse of the walls above, collapse following tension cracks when the walls are dry, collapse in wet conditions due to pore water pressure, caving (involving a rotary rather than slumping motion), spalling and soil creep on the gully sides. Collapse due to seepage lines is also mentioned. The gully sides may also degenerate through chemical weathering, wetting and drying cycles, temperature changes and frost heaving. Surface wash which cascades over gully sides forms miniature mudslides, or coats the walls with a fine clay matrix.

The processes involved with gully wall stability are described in a number of papers (Bradford, Farrell and Larson 1973, Piest, Bradford and Wyatt 1975, Bradford and Piest 1977, Bradford Piest and Spomer 1978, and Bradford and Piest 1980). These deal especially with gully wall collapse as a result of soil moisture conditions along gully banks. Bradford et al 1973 show that the height of the watertable, cohesion of the soil, and infiltration rate are controlling factors affecting stability. Tension cracks are less important but do decrease the cohesion or strength of the soil. In the loessial gullies of the United States described by Bradford et al the ground water level is usually above the toe of the gully wall, so that the observed gully bank failures are likely to be a result of high pore water pressures.

In a field experiment by Bradford and Piest (1977) to examine gully wall failure as a function of pore water pressure the water table was controlled. The conclusion was that piezometers, carefully positioned, can be used to predict the time of mass slumping of gully walls, although the process of slumping is complicated by the role of seepage and vertical cleavage planes. In Bradford et al (1978) a sequential model of gully headwall failure is proposed where a failure or 'popout' occurs followed by column failure and gully cleanout. The growth of gullies with unstable walls depends directly on the characteristics of the soil moisture in the gully walls. Bradford et al (1978) list the controlling variables in gully wall stability as soil type, permeability, soil strength, thickness and structure. Piest and Bowie (1974) list Atterberg limits, plasticity index, soil pH, $CaCO_3$ content, dispersive minerals and dispersion ratios, and measurements of comparative and shear strength as important erodibility factors. Faber and Imeson (1982), and Imeson, Kwaad and Verstraten (1982) measured a number of physical and chemical soil properties and hydrological factors including rainfall, infiltration, conductivities, sorptivity values, soil moisture characteristic curves, volume shrinkage ratios, aggregate stability and runoff hydrographs in connection with gully form and development. Harvey (1982) measured infiltration rates and morphometric parameters in relation to piping and gully growth in south east Spain.

Gully growth itself is monitored by accurate resurveying (Heede 1967, Ologe 1972) and sequential aerial photography (Seginer 1966, and Nir and Klein 1974). From this estimates of land affected by gullying, and the volume of soil removed, can be made. Several multivariate regression equations have been developed to match the calculated erosion rates. The simplest form is suggested by Seginer (1966) where the average annual advancement is a function of the area of the catchment so that

$$E = xA^{0.5}$$

where x ranges from 2.1 to 6.0 for the examples given, E is measured in metres and A in km^2.

Piest and Bowie (1974) discuss three more predictive equations. The first is from the Soil Conservation Service in the United States (1966) where:

$$R = 1.50A^{0.46}P^{0.20}$$

where R is the average annual gully head advance (feet), A is the drainage area (acres) and P is the total annual rainfall of 0.5 inches or more over a 24 hour period that occurred during the time period, converted to an average annual basis (in inches).

Secondly Beer and Johnson (1965) used:

$$X_1 = 0.1X_4^{0.0982}\, X_6^{-0.044}\, X_8^{0.7954}\, X_{14}^{-0.2473} e^{-0.036X3}$$

where

X_1 = gully surface growth (acres)
X_3 = deviation of annual precipitation from normal (inches)
X_4 = index of surface runoff (inches)
X_6 = terraced watershed area (acres)
X_8 = gully length at beginning of period (feet)
X_{14} = length (feet) from gully head to divide.

The final example Piest and Bowie (1974) give is from Thompson (1964) where:

$$R = 0.15A^{0.49}\, S^{0.14}\, P^{0.74}\, E^{1.00}$$

R = average annual gully head advance (feet)
A = drainage area (acres)
S = slope of approach channel (%)
P = total annual rainfall of 0.5" or more during a 24 period
E = clay content of eroding soil profile (% by weight).

In all these equations the rainfall and drainage area factors are the most important components. Only Thompson's equation involves a soil parameter, and none utilises soil moisture parameters. These equations can be used to estimate future soil losses or the positions of past headcuts (Seginer 1966).

The process approach to gully development has not only led to good morphological and process descriptions, but also provided some important generalisations about the nature of gully growth which are discussed here.

Generally the amount of sediment removed by gullies is extremely variable (Piest and Spomer 1968, Ologe 1972). For example Piest Bradford and Wyatt (1975) quote from Dvorak and Heinemann that 98% of sediment eroded from a gully reach in Dry Creek, Frontier County, Nebraska, occurred during the first year of the study period April 1951 - April 1956.

The majority of sediment transported in gullies comes from within the gully itself rather than from the watershed. Thus the sediment is provided from the gully head (Piest and Spomer 1968, Bradford and Piest 1980), or channel banks (Heede 1967). Piest, Bradford and Wyatt (1975) suggest that the sediment transported is already weathered or detached as flows with similar tractive forces transport different loads.

Although gully erosion is dependent upon precipitation there is no clear relationship between rainfall parameters and gully extension (Heede 1967). Sediment transport is often weathering limited as depletion occurs after the commencement of a wetter season (Piest et al 1975, and Piest and Bowie 1974) and even within storms (Piest et al 1975). Despite this, more erosion occurs during the wetter season. Piest and Bowie (1974) showed that in ten years 80% of the sediment transport occurred in May and June when there was 55% of the annual surface runoff, and 33% annual average precipitation. Piest and Spomer (1968) noted lower sediment yield during drier years.

Ologe (1972) noted the significance of a rapid succession of storms. The first storm may do little "work", but may cause unstable conditions and may prepare the way for extensive erosion during the second storm, which may be less intense. Heede (1967) reported that a storm occurring on 7 July during which a gully head receded 7.5 feet, but a storm on 12 August (one third the size of the previous storm) caused a recession of 19.3 feet. Thus the antecedent conditions affect the stability of gully slopes as well as raising soil moisture levels to increase runoff. Superimposed on this are the effects of storm frequency and magnitude. Piest and Bowie (1974) suggested that high frequency low magnitude storms cause inwashing of sediments but low frequency high magnitude events cause gully clean out and extension. Gully flow probably requires a precipitation threshold and Ologe suggests 37mm for Zaria. Extreme events cause considerable gully wall collapse and extension (Ologe 1972).

Observing and measuring processes involve several problems. Little attention has been paid to the temporal (as opposed to spatial) sampling problem (Thornes and Brunsden 1977) and field studies (often hindered by lack of time, labour, and instruments) should be designed to examine these temporal variations. In terms of gully erosion gully extension may occur only for large, relatively rare, storms, and the onset of drought or lack of large land forming events will be detrimental. For example Heede (1967) had to wait 3½ years before flow occurred in some instrumented gullies and in a seven year period only five storms produced gully flow.

The variability of gully growth rates (due largely to antecedent conditions) means that statistically significant erosion rates depend on long term records. This is especially true if growth is dependent upon extreme events, and any attempt at establishing regression equations for growth will have to allow for large error bands. Erosion rates tend to give the impression of smooth (or continuous) gully growth rather than a disjunct and sporadic development which also makes them misleading. Even so gully growth rates are based on empirical relationships with short data records (Seginer 1966). They tend to be limited to the catchment where they were developed and do not seek to explain how or why gullies develop. Correlation matrices of variables suggest possible links between variables and extension, but do not provide causal explanations.

Process studies are still very important, particularly as so little quantified data of gully growth are available. Such studies can be used, not only to develop statistical predictions of growth and causality, but also explain the nature of gully development by careful research design.

The Modelling Approach To Gully Growth

Models form a useful research tool as they provide a framework for data collection, they form a stepping stone between theory building and law forming, they help to explain cause and affect, and they facilitate the understanding of phenomena even if at a simplified level (Chorley and Haggett 1967). A model may be defined loosely as an idea, a theory or a set of ordered thoughts. Models are simplified representations of reality which give the most significant features or relationships of that reality. In discussing geographical models Chorley and Haggett (1967) characterise a variety of models. The 'natural historical analogue model', for example, examines a group of geomorphological phenomena with regard to their assumed position in time. The 'empirical model' involves the fitting of data to simple or multiple regressions. In this "black box" approach there are inputs and outputs but little understanding of the internal operation of the system. Both these models are incorporated in discussions on the historical and the process approaches to gully growth. This section on modelling as an approach to gully growth deals specifically with what Chorley and Haggett (1967) call 'physical systems models', and in particular mathematical and hardware models.

Hardware Models

A large number of hardware models have been developed to examine a variety of hillslope, channel and drainage basin properties. In all cases such models are used as analogies for landform development as the problems of scaling, boundaries, and setting initial conditions make it impossible to recreate natural conditions.

Flume studies have been used to assess the hydrological characteristics of shallow water flow and examine the effect of that flow (and sometimes rainfall) on erosion, deposition and rill development (Savat 1977, 1980, Moss, Green and Hutka 1982, and De Ploey 1984). Larger flumes have been used to study knickpoint migration (Brush and Wolman 1960, Holland and Pickup 1976), and drainage basin evolution (Parker and Schumm 1982). The site at Perth Amboy studied by Schumm (1956a) was a manmade tip of waste and overburden from pits backfilled into an abandoned clay pit. It was created about 1929 and finally levelled in 1953. In the interim Schumm studied a range of morphological parameters and discussed the evolution of the site which (it was suggested) may have parallels with badland sites. This site could be seen as a hardware model on a large scale, or an analogy for landform evolution in other areas.

The paper by Parker and Schumm (1982) forms a model for possible experimental studies in gully migration, although their research was directed to drainage basin development rather than the growth of gully networks. They describe two modes of drainage development, firstly where channels develop instantaneously over the drainage basin and in time form an integrated network, and secondly a headward growth model as channels bifurcate and develop towards the divide. Parker and Schumm undertook two experiments to examine these modes of drainage development. For both experiments the surface was graded to two intersecting planes. In the first experiment the baselevel was lowered before the application of simulated rainfall. In the second experiment the slope gradient was steeper and the baselevel was not lowered prior to rainfall. The simulated rainfall was continued until the drainage network had ceased to grow. The effect of the different initial conditions was to cause differences in the pattern of drainage evolution. In the first case the lowered baselevel offers a site for headward migration of a knickpoint which bifurcates into a channel system, which may be similar to gully head migration. In the second experiment long tributaries develop and the area between them is later filled by additional tributary growth. Thus depending on the initial conditions, the experimental set up offers two plausible models for drainage basin evolution.

Mathematical Models And Gully Growth

Chorley and Haggett define mathematical models as those in which objects, forces and events are replaced by expressions containing mathematical variables, parameters and constants. These can be deterministic or stochastic. In the former case the relationships between variables are predictable and may be based on an understanding of the physics of the system, or on logically reasoned argument. If it is impossible to consider all variables 'noise' occurs. This may be so great that stochastic models are used instead where one or more variables has a distribution of values, thus involving a statistical element.

A large number of mathematical models have been developed for hillslope hydrological problems (Kirkby 1978), although few are specifically on gully growth (Seginer 1966). Some models related to gully growth (either directly or indirectly) are discussed briefly to show what has been and can be done by modelling.

The origin of gullies, or indeed of any erosional feature, has received relatively little attention in modelling. The basis of many models comes from Horton (1945) who examines the initiation and development of surface wash and rill erosion. He suggests a critical length from the watershed in which the erosive power of surface flow is less than the resistance of the soil surface. This assumes an increase in flow downslope by the cumulative action of increased drainage area. The critical length is assumed to be a function of:

1 The intensity of rainfall and infiltration (providing precipitation excess).
2 The length of slope (affecting the cumulative increase in flow downslope and erosion rates).
3 Slope angle.
4 Surface roughness.

Horton (1945) describes the origin of rill channels by an increase in flow depth downslope leading to greater erosive power from concentrated water and the diversion of flow into rills. If gullies are likewise caused by surface flow this description of rill development is an analogue for gully initiation.

The theoretical analysis of the conditions causing the initiation and development of a perturbation on a smooth surface was taken up by Smith and Bretherton (1972). They make three assumptions on which to build the model, firstly the drainage basin may be represented by a mathematical surface, secondly the principle of the conservation of mass is applicable, and thirdly the sediment transport at any one point is a function of the local slope and local discharge of water, that is

$$Qs = F(S,q).$$

Smith and Bretherton also introduce the concept of stability. In the unstable case a small perturbation grows with time and changes the form of the basin. This is represented mathematically as the condition where the amount of discharge of sediment is less than the rate of production of sediment. The stable case is when a perturbation is removed from the system and this occurs when the sediment discharge is equal to or greater than the rate of sediment production. On a concave segment converging flow lines increase the transporting capacity of the water and perturbations can be unstable. On straight and convex slopes fluxes move downslope or diverge. In the latter case divergence leads to deposition and a filling in of the perturbations. On a hillslope with an upper convex and lower concave form, theory suggests there will be a critical distance to the inflection point which is stable, which is similar to Horton's belt of no erosion. There will be a convex unchannelled section, a straight portion, and a channelled concave portion. Such hillslopes result from different transport processes occurring at different slope positions. Gilbert (1906) was the first to note this. Empirical support comes from Schumm (1956b) who showed that slopes on the Brule formation are concave and rilled largely as a result of water flow processes, whereas slopes of the Chadron lithology are convex as a result of creep processes. More recently Kirkby (1971) has modelled hillslope profiles as a result of different values for the 'n' and 'm' exponents representing different processes for the transport equation

$$q_s = F\ (q^n S^m).$$

Some values are m = 1 and n = 1 to 2 for slopes with soil wash without gullying and m = 2 and n = 2 for soil wash with gullying.

The conditions surrounding the stability of a perturbation are investigated further by Kirkby (1980a) with reference to the stream head, which is similar to a gully head. A small development of the hollow will locally increase its contributing area which may trigger a negative feedback causing instability. Slope profiles near divides tend to be more stable because the overland flow does not have sufficient energy to erode, however other parts of the slope may have a tendency towards instability. Kirkby (1980a) describes this change from stable to unstable conditions by a process function which allows positive and negative feedbacks between the process (mean rate of lowering) and form (area drained per unit contour length). The rate of lowering may decrease and then increase with the area drained per unit contour length (or distance downslope) for a given sediment transport law. Upslope of the minimum point of the curve a slight increase in area drained per unit contour length caused, for example, by local hollow development will lead to a decrease of the mean rate of lowering so that the perturbation will be stable. However downslope from the minimum point in the curve an increase in area drained per unit contour length will lead to an increase in the mean rate of lowering so that hollow growth becomes unstable.

Field evidence for gully development as a threshold problem comes from Schumm and Hadley (1957), and Patton and Schumm (1975) where gully growth is associated with oversteepened reaches in alluvial systems. For drainage basins over a limiting size the relationship between critical valley slope and drainage basin area seems to control the threshold. Thus local oversteepening in a valley reach may trigger incision. In areas near the threshold zone phases of incision and aggradation may be cyclical (Schumm and Hadley 1957, and Womack and Schumm 1977).

Graf (1979) suggests that another possible threshold for incision is the interation between the tractive force in channels and the

41. C.Srivanavit, J.I.Zink and J.J.Dechter, J. Amer. Chem. Soc, 99, 5876 (1977).

42 D.Gudlin and H.Schneider, Inorganica Chimica Acta, 33, 205 (1979).

43. S.O.Chan and L.W.Reeves, J. Amer. Chem. Soc. 96, 404 (1974).

44. R.W.Briggs, F.A.Etzkorn and J.F.Hinton, J. Magn. Reson., 37, 523 (1980).

45. J.F.Hinton, G.L.Turner and F.S.Millet, J. Magn. Reson., 45, 42 (1981).

46. J.E.Tanner, J.Chem. Phys., 69, 1748 (1978).

resistance of the valley floor as represented by the biomass. The measure of tractive force uses a slope variable, thereby including the inference to slope as a significant variable. He found a discriminating function for montane arroyos separating entrenched and uncut valleys depending upon the tractive force/biomass relationship.

Thornes (1984) develops a mathematical model for gully growth based on the flow to the hollow, the catchment size, and morphology of the gully head. Gully growth can result from the action of water flowing from the divide downslope or by headward migration. Thornes briefly examines both models, but elaborates on the latter. In the top-down model the length of overland flow (or catchment area) has to exceed a critical value before erosion can occur. In the headward growth model the migrating gully is seen as a shock wave moving through the hillslope. If the rate of forward migration is greater than lateral migration the gully is analogously 'supersonic'. The shape of the gully head becomes important when the apex of the gully reaches the effective length or belt of no erosion, as the distance of flow exceeds the critical length on the gully head on either side of the apex, but not at the apex itself. If the maximum erosion rate is not at the apex, but to the side (ie off centre) then the gully may bifurcate.

The implications of the shock wave model are discussed by Thornes (1984) in relation to gully development in south east Spain to explain different gully systems in different lithologies. For example gullies in sandy loams (with lower strength values) have a greater propensity to widening and bifurcation than gullies in marl. Thornes uses the analogy of subsonic shocks for easily eroded lithologies and supersonic migration in more resistant materials where the ratios of forward to lateral migration are large.

Gully system development is modelled by Faulkner (1974) in the context of allometric networks where the growth of a part of the organism (or whole) is related to the size of the organism. Assuming constant conditions of relief, climate, lithology, etc, similar growth rates between systems will cause different systems to have similar characteristics. However if growth rates vary between systems due to competition (for example), there will not be geometric relations between systems, only within them. Faulkner tested this theory on a group of gullies developing on the edge of a pediment from the river in south east Alberta. She found that the laws of drainage fitted larger gully systems (which were increasingly non-competitive) but did not fit smaller systems which were entirely competitive. This suggests any study of gully systems incorporating drainage laws should take account of competing systems.

The use of models for examining gully growth has so far been limited, and there is undoubtedly room for the development of a physically based model of headward growth taking into account surface and subsurface flows. However there are drawbacks to modelling which should be noted. As models are a simplification of reality they do not incorporate all possible variables, and several problems arise from this. The modeller has to decide on the most significant variables, that is those which account for most of the variability. Depending on the selection of variables the model may operate with varying success as reflected by the statistical tests of its output such as comparison between measured and expected results. Also the model may be particularly sensitive to some variables, with small changes in one or more causing an overreaction elsewhere in the model, alternatively some variables may contribute little to the workings of the model. Even when the model provides accurate estimates of values this does not necessarily prove that the model accurately portrays the workings of the system, so models have to be carefully designed and tested. The modelling approach has several advantages for developing theory, and establishing a framework for data collection. Provided due care is taken concerning the problems of modelling this approach offers a great deal to studying gully development.

1.3 Research Possibilities

The Research Topic

The preceding summary of gully literature highlights areas for research and the advantages and disadvantages of the different approaches. Areas already well documented include detailed descriptions of gullies and processes, analysis of the mechanical, chemical and hydrologic properties of some gullied soils, and gully advancement rates. However many of the processes identified are not quantified, and are described from observation or inferred from soil properties (the papers by Bradford and Piest are notable exceptions).

Few authors have attempted to rank processes causing gully growth in terms of importance. Water is undoubtedly the most important source of erosion but opinion differs on the significance of surface and subsurface hydrology. Implicit in definitions of gullies is the occurrence of ephemeral flow (during storms) and erosion as a result of such flow. The initiation of gullies is also generally assigned to surface flow as Morgan (1979) writes:

> "In the first stage small depressions or knicks form on a hillside...water concentrates in these depressions and enlarges them until several depressions coalesce and an incipient channel is formed."

Surface flow also dominates discussion on processes in well developed gully systems, for example Ologe (1972), in a discussion on headscarp recession, only refers to processes resulting from surface flow. The concentration on surface flow is paralleled in the theoretical field. One of the earliest hillslope hydrological models is Horton's (1945) model of surface flow production by rainfall intensity minus the infiltration rate which focuses attention on the surface rather than the subsurface component of overland flow production. Other methods of surface flow production involving subsurface flow have been developed over the past thirty years. However Hortonian flow is still considered the dominant means of overlandflow production in semi-arid areas where rainfall intensities can be high, infiltration rates low, vegetation sparse, and with thin (sometimes crusted) soils.

Simultaneously subsurface soil moisture is considered low because of the high temperatures and evaporation rates, the low annual rainfall, the shortness of storms, the assumed low infiltration rates, low storage capacity for moisture in the thin degraded soils, and rapid runoff times. Models of gully growth tend to emphasise the overriding importance of surface flow with surface discharge, distance from divide or precipitation variables figuring highly in both empirical and mathematical models.

Research during the sixties and seventies led to the increasing understanding and significance of subsurface hydrology (Chorley 1978). Overland flow is not frequently observed in many areas, particularly those characterised by appreciable quantities of soil, vegetation, litter or humus (Kirkby and Chorley 1967). Also it was noted that overland flow occurred from limited areas when rainfall intensities were less than infiltration rates, and for example near stream channels, at the base of slopes, in thin soils and hollows. This led to alternative concepts of overland flow production by return flow or saturated overland flow, and the ideas of partial and variable source areas (Betson 1964, Kirkby and Chorley 1967, Betson and Marius 1969, Dunne and Black 1970 a and b, Kirkby 1978).

Most research on subsurface flow has been conducted in temperate climates (eg Dunne and Black (1970) in Vermont, USA, and Weyman (1973) in Somerset GB) where the climatic and edaphic characteristics are most likely to favour its development. Some authors have shown that subsurface flow can contribute to the storm hydrograph (Chorley 1978 p13) but generally throughflow is too slow for this. However throughflow can maintain baseflow for some time after the storm (Dunne and Black 1970a, 1970b, Weyman 1973, Harr 1977, and Dunne 1978). Flume experiments have substantiated this idea. Hewlett and Hibbert (1963) emphasise the importance of unsaturated flow to the maintenance of baseflow although Anderson and Burt (1978) argue baseflow is sustained by a shrinking saturated wedge at the foot of the slope.

Flow through the soil not only affects the balance of the hydrological cycle, but also alters the soil itself, leading to for example the development of catenas at different slope positions. These have been described for temperate areas (Furley 1970, and Huggett 1976), but also for semi-arid areas. Muhs (1982), for example, ascribes variations in clay content to soil moisture movement on some Californian sites, and also attributes catena development in other Mediterranean climatic areas (such as Israel, Lebanon and Spain) to soil hydrology. Other evidence for soil moisture movement includes development of calcrete layers within soils, soluble silicates, salt (Conacher 1982), and concentrations of gypsum. Soil development is subject to the same soil forming processes as soils in more humid areas, although development is limited by rainfall (Buol 1965).

Preferential movement or location of water in hillslopes would be expected to produce differential mechanical and chemical weathering and various forms of evidence have been collected for temperate areas. Crabtree and Burt (1983) indicate that solution rates are preferentially greater in hollows. Water movement in permanent seepage lines could preferentially weather linear features in bedrock which can be exploited by advancing streams (Bunting 1961) or gullies. Such seepage can transport fines (Carson and Kirkby 1972, and Pilgrim and Huff 1983), and under certain circumstances develop pipes (Jones 1971).

In the semi-arid context most evidence of water movement has been largely confined to research on pipes which in some cases are quite spectacular. Drew (1982) describes pipes in southern Saskatchewan from 5 to 30 cms diameter to those large enough to enter. The larger pipes transport water at rates equivalent to surface flow velocities and in some areas account for 50% of the drainage. In the non-piped soils permeabilities are low with wetting depths of 25-30 cms after heavy rain. However in Drew's explanation of pipe development the initial stage involves water entry through macropores and cracks towards an impermeable layer, and then throughflow downdip to the point of exit. The pipes are enlarged by solutional and corrasive action. In south east Spain Harvey (1982) describes a variety of pipes occurring in gullied soils. The pipes are associated with tension cracks, but although cracks account for the entry of water, throughflow must account for the lateral movement before the pipes develop substantially.

Thus throughflow to gully heads could encourage headward growth by weathering the upslope area and weaken the gully face by increasing the pore water pressures and causing slope instability. A downslope increase in soil moisture may lead to a saturated wedge at the gully head providing base flow after storms and saturated conditions in the gully floor. Also if saturated conditions developed near the surface of gully heads the production of saturated overland flow could lead to more headward erosion by surface flow.

A number of observations on gullies have stressed the importance of subsurface flow to gully erosion. Leopold, Wolman and Miller (1964) write:

> "Sapping at the base is undoubtedly one of the most effective processes in headward extension".

Leopold et al stress the role of soil moisture in raising pore water pressures which cause collapse. A zone of throughflow would aid not only bank failure, but also the undercutting action of falling water in the plunge pool. They write that they have:

> "seldom experienced collapse of dry gully walls, nor is gully wall collapse characteristic of these short periods of storm discharge when water is actively flowing in the gully trench"(p446)

Leopold et al state that most slumping occurs several hours or the day following the storm, when the soil moisture has increased sufficiently to make the walls unstable.

The work by Bradford and Piest (various papers) quantifies and verifies the importance of collapse due to soil moisture conditions. Their study area is largely in the loess areas of the Missouri river basin where saturated hydraulic conductivities vary from low to high values (5 x 10^{-5} cm/min to 5.2 cm/min) and the annual average precipitation is 806 mm/year (Bradford and Piest 1977). These factors will favour wet soil moisture conditions, indeed the water table is often above the base of the gully wall. Perhaps it is not surprising that subsurface hydrology plays a significant role in gully advancement. In truly semi-arid areas, with precipitation between 200-500 mm/year, would subsurface hydrology still be an important factor? Leopold et al (1964) write:

> "In our experience in semi-arid regions, the erosive action of water flowing over the vertical face of a gully head generally is not among the most active agents of headward progression."(p447).

The research problem is to attempt to evaluate the relative roles of surface and subsurface flows and significant variables. This is done by quantifying the role of surface flow and by assessing the viability of subsurface flow as a significant factor contributing to gully erosion.

The choice of a suitable approach to the problem reflects the two main aims of the research project:

1 To test a hypothesis by field monitoring, using standard equations of surface and subsurface flow, and simple models to simulate the subsurface hydrological component.

2 To provide data to parameterise (at a future date) a more specific physically based model of gully growth under a semi-arid regime.

The main thrust of this thesis is directed to the first point. In order to do this successfully, and indeed to incorporate the second point, the choice of variables, techniques, and site location is set within the framework of a hydrological model. Thus two research approaches are involved, process monitoring in the field, and model building and parameterising

The Experimental Design

The experimental design involves two groups of factors, the choice of variables and techniques, and secondly the sampling scheme. These are in turn dependent upon the aims of the project which are the monitoring of surface and subsurface flow to ascertain the relative significant of both and at the same time providing data for a model. One possible type of model envisaged is a mathematical model of hillslope hydrology incorporating surface and subsurface flow to gully heads in two dimensions in the same vein as Freeze (1978). Although Freeze examined the case for a 'normal' hillslope profile rather than a gullied profile, such an approach could be used to model flow rates and routes to a gully head, with implications for the nature and simulation of gully head advancement.

The choice of variables is based upon the components of the hillslope hydrological cycle: precipitation, infiltration, surface store, and subsurface store. This cycle is itself a well thought out, monitored and documented theory (Kirkby 1978) and forms a ready framework for mathematical modelling (Freeze 1978). Within this setting the choice is further delimited by the logistics of implementation, suitability of variables for short-term monitoring, constraints of the model, and any unusual requirements imposed by the environment.

The logistics of implementing the sampling scheme are very important. The variables chosen have to be measurable by one person with relative ease and safety. This consideration is also taken into account when choosing techniques along with cost and time constraints. The variables must also occur or change within a suitable timescale so that they are not too infrequent or rapid to be measured. For example gully growth rates are known to be variable so that long records are required to obtain useful estimates. Thus measuring gully migration in one season would be unlikely to produce useful results. The model will impose constraints of simplicity and economy on the choice of variables, as the model should have as few complicating variables as possible avoiding for example pipes and vegetation.

Account should be taken of any natural characteristics of the proposed site which may influence the modelling decisions. There are three main areas of concern:

1 The impact of xerophytic vegetation on the hydrological budget and gully growth.
2 The influence of crusted and/or swelling soils on subsurface hydrology, and
3 Problems of monitoring water in a dry environment.

These topics are considered in more detail in the relevant chapters. The role of vegetation is discussed in Chapter 2, the influence of crusted soils on infiltration in Chapter 4, and the importance of swelling/shrinking soils and the problems of monitoring in dry environments are discussed in Chapter 6.

The sampling scheme involves both temporal and spatial considerations. One component of the temporal sampling scheme has already been mentioned where the frequency of monitoring must be sufficient to measure the rate of change of values. Where change is rapid and or small, a more frequent and precise measuring system must be employed. The frequency of observation should also be sufficient to separate signal and noise. On a longer time scale seasonality would be expected to alter not only inputs and outputs to the system (ie rainfall, and evapotranspiration), but affect storage component and response to rainfall so the timing of field work should take into account the range of response associated with seasonality.

The spatial sampling pattern should be sufficient to examine the background variability or distribution of values for example with depth or topographical position as well as determine trends. One way is to instrument as large an area as possible however this is uneconomic when resources are limited. At the other extreme key locations only can be monitored where the environment is most sensitive and change most likely. The experimental design is outlined here bearing in mind the four key areas for study, and the reasons for examining them.

The precipitation data are examined for three reasons, firstly to assess the major factors influencing runoff, infiltration, soil moisture content, and plant growth for the temporal record, secondly to examine the precipitation conditions during the field period, and thirdly to look at the combined effect of precipitation and evaporation on soil moisture content .

Infiltration rates have been used as a way of estimating the division of precipitation into the surface and subsurface component (Horton 1945), so the spatial and temporal variations in infiltration rates are measured using standard infiltrometers as an indirect guide to the variability of the hydrological response between and within the soils.

The quantity of overland flow and sediment transport is monitored using Gerlach troughs in order to examine four sets of factors:

1 To assess overall variation in discharge over the gully catchment and in particular estimate quantities of discharge immediately above the gully head.
2 To estimate the susceptibility of the soil to surface wash erosion and determine the overall slope denudation rates.

3 To look at the impact of both water and sediment moving into the gully.
4 To examine the long term changes of the slope which in future may lead to different responses through particularly slope armouring.

Finally the subsurface conditions are monitored to examine the development of saturated areas either with depth, hillslope position or time, using standard techniques, and calculate flux rates and angles to determine the likely magnitude of throughflow to the gully heads.

To incorporate the variability imposed by seasonality the field sessions were divided into three periods. The first session was 12 July to 30 July 1982, which was mainly used for choosing the site, mapping, setting up the instruments, and starting some sampling and observation. The second session occurred between 13 November to 9 December 1982 and was designed to monitor wetter conditions during the early winter season with the passage of winter storms. The final season was from 4 April to 3 May 1983 which was to continue monitoring the response of the last of the winter storms, and examine the effect of the drawdown of soil moisture with the onset of summer.

The layout of instrumentation was along a single flow line to the gully head to minimise the equipment needed and maximise the data collection. The experimental design was employed for two gullies each on a different lithology in order to introduce some lithological comparison (Thornes 1984). The layout was designed to be as economical as possible to obtain the most useful amount of information in a one-person operator situation.

CHAPTER TWO FIELD SITE DESCRIPTION

2.1 The Study Area - Choice and Description

South east Spain is one of the semi-arid regions of the Mediterranean which has been intensively gullied, for example in the Guadix, Tabernas, Vera, and Murcia basins, and in several smaller montane valleys. One of the latter is the Ugijar basin, where hillslope and channel processes have already been studied in detail (Thornes 1976, Scoging 1976, and 1982, Butcher and Thornes 1978). The extensive gully erosion around Ugijar and the already existing background information makes this a suitable location for the research.

The Ugijar basin is in the Province of Granada (figure 2.1) lying among the eastern end of the Baetic Cordillera. To the north lies the Sierra Nevada-Filabride complex rising to 3478m on Mulhacen. To the south are a series of folded Triassic nappes called collectively the Alpujarride complex. These are, from west to east, the Sierra de Lujar (1844m), the Sierra la Contraviesa (1508m), the Murtas nappe (1511m), and the Sierra de Gâdor (2145m). Between the two sets of mountainous relief lies a series of montane valleys in an east-west line which form three main river basins. The Rio Guadalfeo drains the western end and the Rio Andarax the eastern part of the Sierra Nevada/Alpujarride trough. The central section is drained by the Rio Adra which incorporates the Rios Ugijar, Yator, and Chico (figure 2.2).

Photograph 2.1 is an aerial view of the central section of the valley around Ugijar. This shows the heavy dissection of parts of the Ugijar basin, particularly on the slopes and headwaters of the ramblas where a series of parallel gullies flows from the watershed to the rambla floor. Elsewhere there are also single linear gullies which bifurcate occasionally, and run into relatively undissected areas.

The Ugijar basin seems to be conducive to gully erosion for a number of reasons. Firstly the gullied morphology partly reflects the geological control with certain lithologies more prone to

Figure 2·1 Location Of Study Site

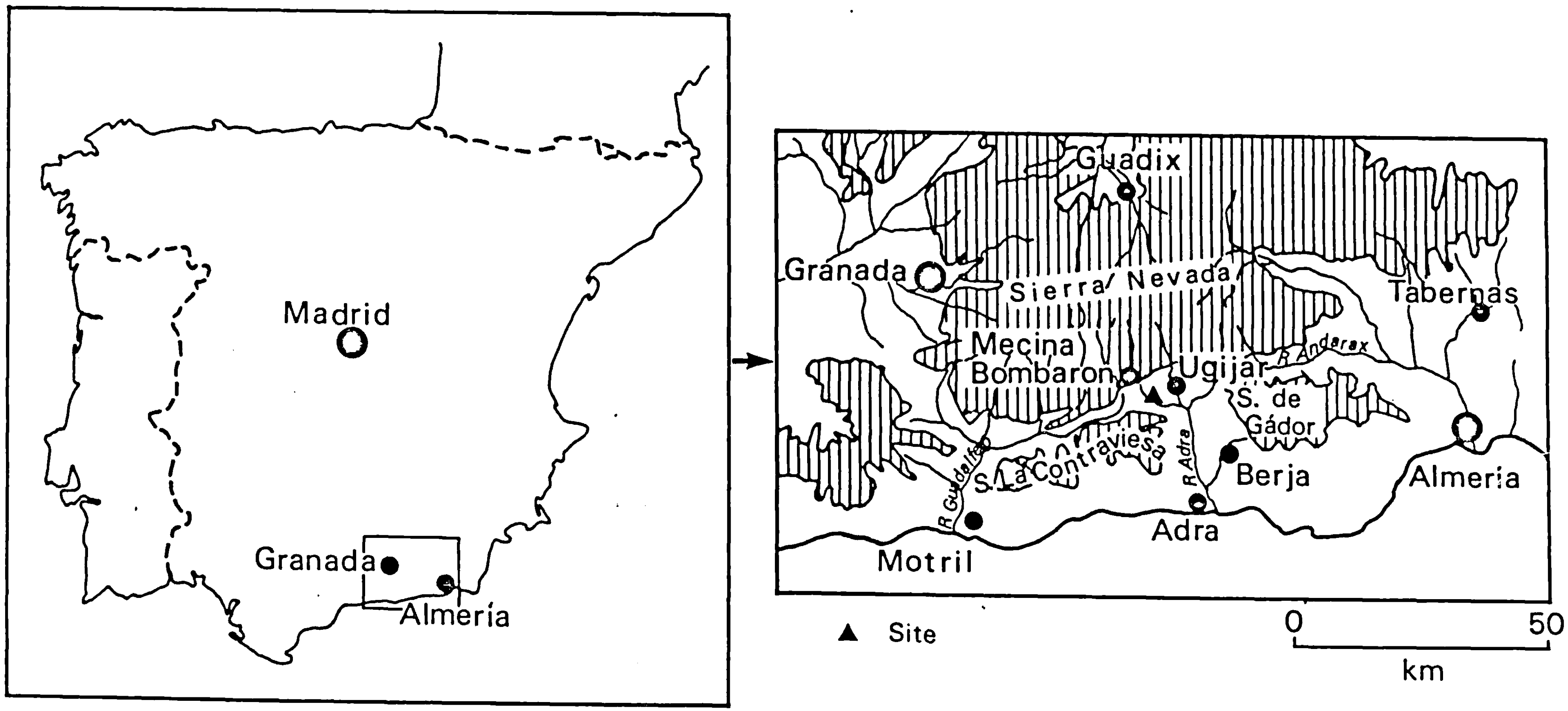
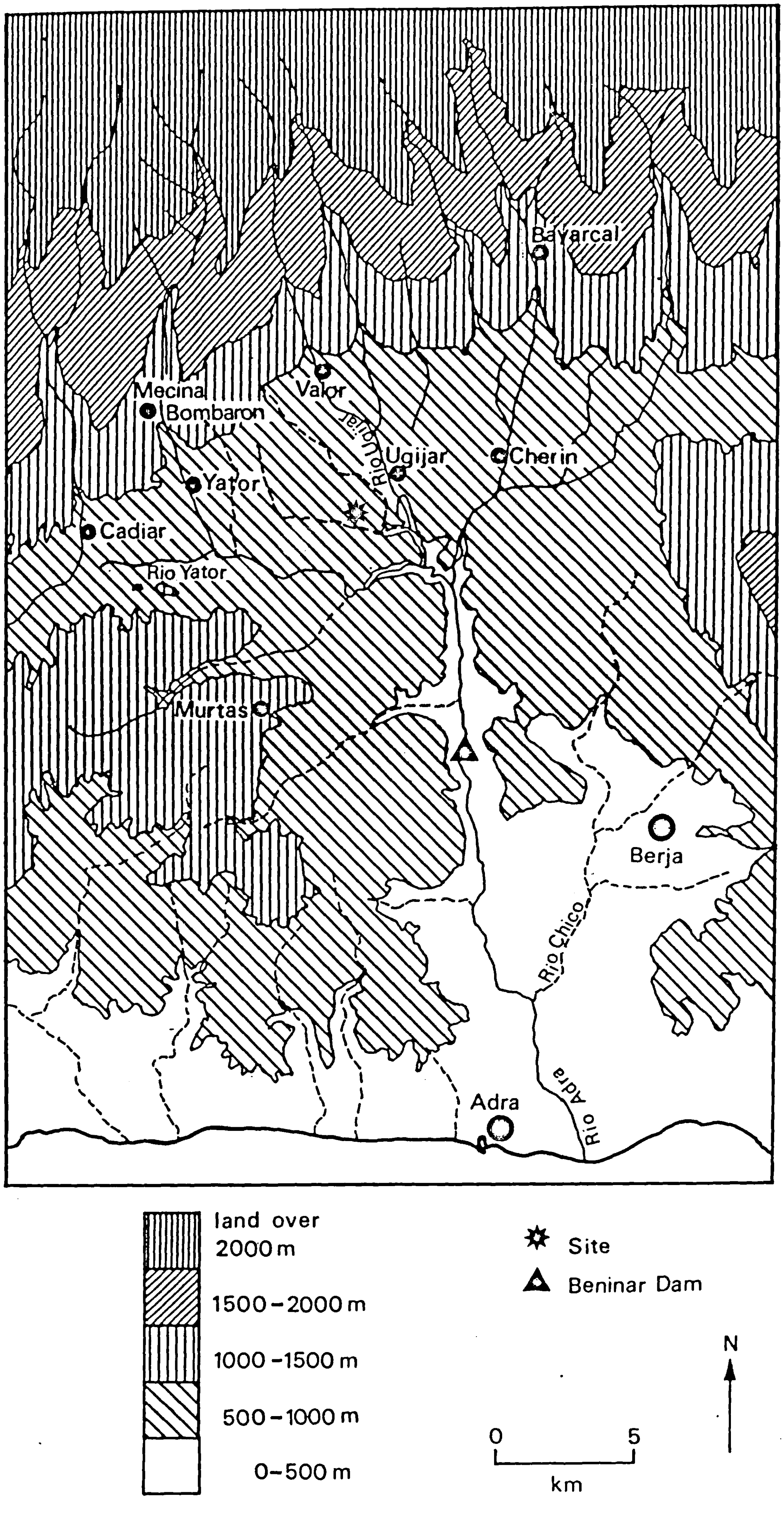
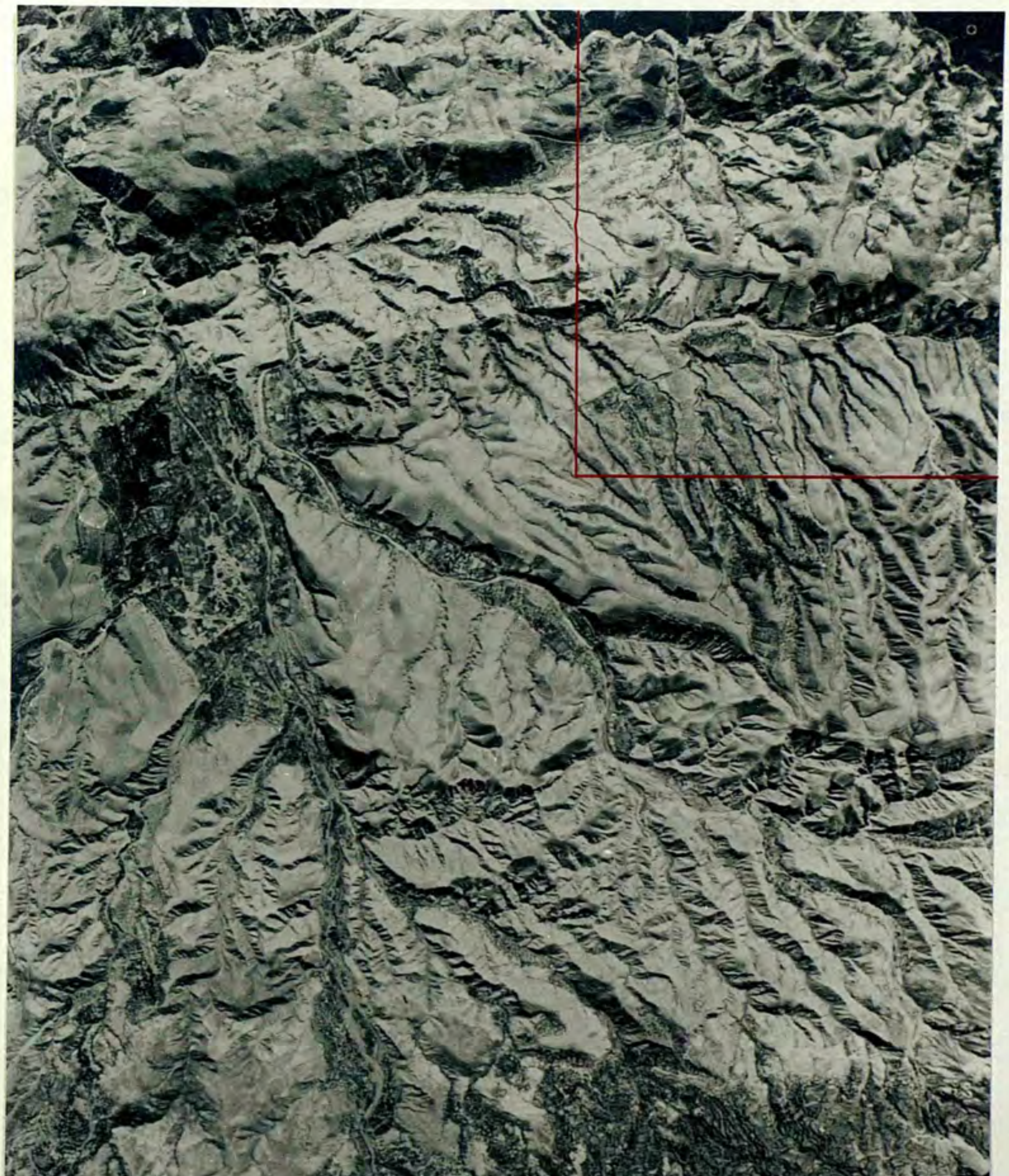
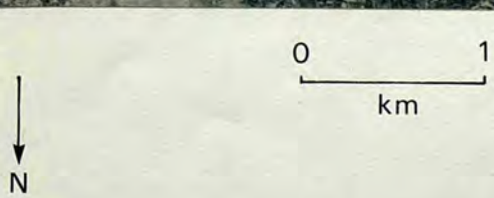


Figure 2·2 Topographic map of the Ugijar Basin



Photograph 2·1 Aerial photograph of the Ugijar Basin





gullying than others. Secondly the climatic regime is markedly seasonal with low annual rainfall falling mainly in winter but varying greatly between and within years (the climate of the region is discussed further in Chapter 3). Thirdly the type and density of vegetation cover help to make the soil prone to erosion, particularly in the cultivated areas. Finally the agricultural practices undertaken aid erosion processes.

The geology of the area is very complex with a variety of lithologies reflecting a complicated orogenic history. Figure 2.3 shows a generalised map of the geology of the region at 1:200 000. The geology of the basin broadly follows the structural units and can be divided into three groups on this basis.

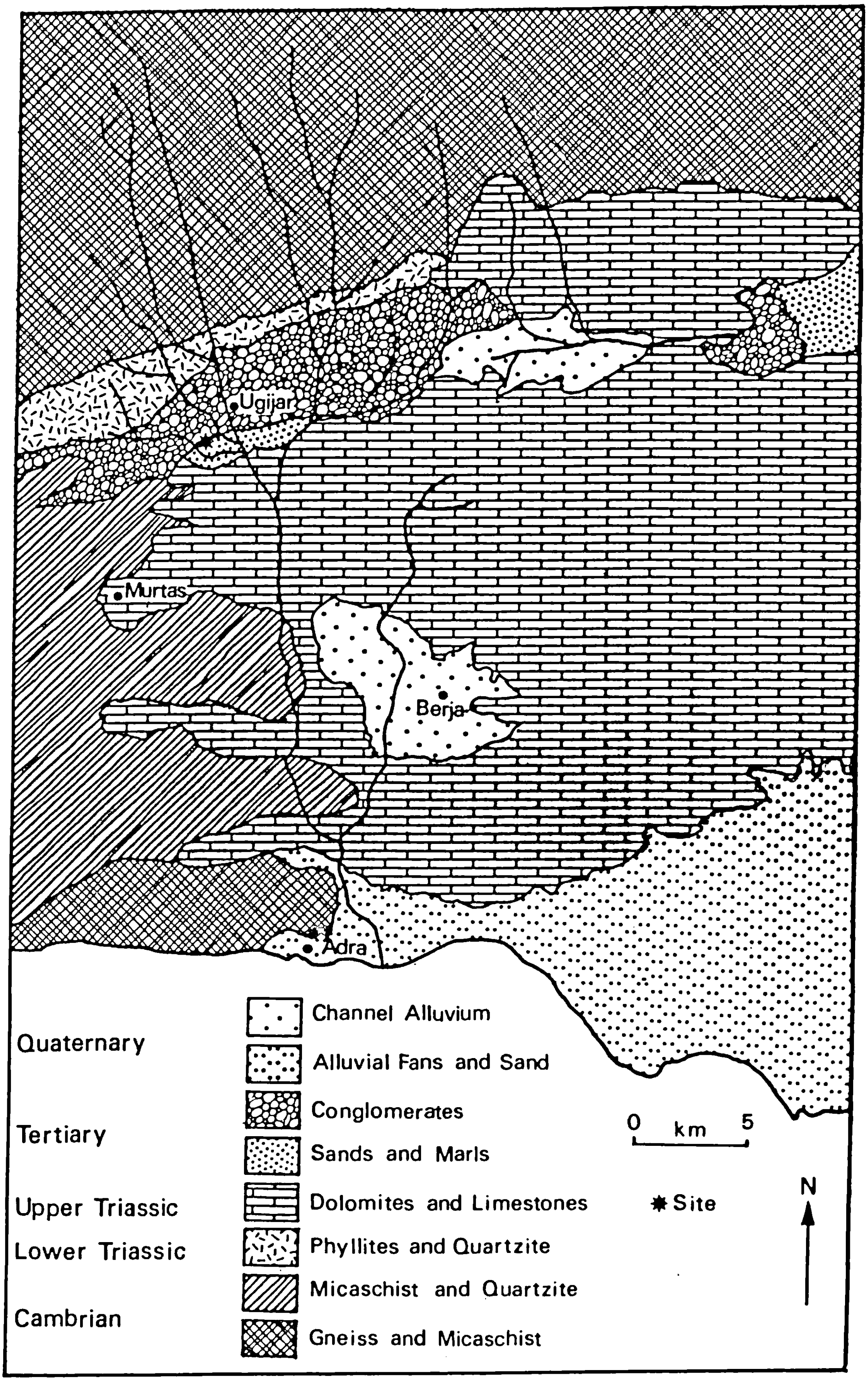
Firstly the Sierra Nevada system is an intensely folded and metamorphosed structure with Cambrian to early Tertiary rocks such as micaschists, quarzites, amphibolites, slates, and gneisses. Some major faults lie east-west along the southern slopes marking the junction between the mountains and valleys.

Secondly are the Alpujarride nappes with three major suites of rocks which affect the patterns of erosion (Thornes 1976). These are (a) the micaschists and quartzites, (b) the phyllites and quartzites, and (c) the limestones-dolomites.

The micaschists and quartzites occur on the nappes to the south west and south of Ugijar. These consist of quartz, oligoclase, and white mica and biotite. In the upper part of the series there is also sodic amphibolites, chlorites, and chloritoids. In the upper most part of the series there is an increase of quartzite and reworked micashists. The latter series is particularly susceptible to erosion because they are weakly consolidated and composed of minerals formed in the early part of the chemical reaction series (Thornes 1976).

The phyllites and quartzites lie to the north along the Manto de Castaras (an Alpujarride nappe lying along the footslopes of the Sierra Nevadas) and to a lesser extent to the south of Ugijar. The phyllites are metamorphosed argillaceous sediments of quartz, chlorite and muscovite, and are found in alternate bands of fine

Figure 2·3 Geology map of the Ugijar Basin



quartzite and calcic layers. Finally the limestone-dolomite suite lies mainly to the south and south east of Ugijar and is heavily metamorphosed.

The third main geologic component is the Neogene and Quaternary deposits in the Ugijar valley. The sequence starts with Tortonian sands, conglomerates and limestones with marls deposited on top. These are of marine origin and have been subjected to some tectonic activity. The marls are a whitish calcareous shale or clay formed by the mixing of fines and calcium carbonate. This is an important lithology in south east Spain, and elsewhere contains high levels of gypsum and sodium salts. The chemistry of marls is described for the Murcia area (Perez et al 1982) but the chemical compositon of the marls near Ugijar is not known.

The marls are sometimes covered with more conglomerate material resulting from Quaternary erosion. In places this appears more like a soil or slope deposit than bedrock. However in several areas road cutting sections show that the deposit can be deep with bedding planes or layers varying in slope. These consist of a range of particle sizes up to large boulders. Lastly in the larger river beds Quaternary alluvium has been deposited by erosion from the surrounding hills.

The soils of the area have not been studied fully and few details are available. Thornes (1976) quotes the means for dispersion ratios and colloid ratio as 76.6 and 0.56 respectively from ICONA for a variety of soils in the area. Although the sample sizes are small and there is no significant difference between soil types, this does suggest that the soils are susceptible to erosion.
The Ministerio de Agricultura (1982) have estimated sediment loss using the Universal Soil Loss Equation for south east Spain. In the Ugijar Basin the estimated sediment loss varies from 0 to 200 tonnes ha^{-1} yr^{-1} reflecting the variation in geology, slopes, vegetation cover and agricultural practice.

The Alpujarran villages of the region are old Moorish settlements that have been in existence for over 1000 years (Bosque Maurel, 1973). The surrounding land has also been cultivated and grazed for a long time and there are probably no areas of 'natural' vegetation left. A better distinction is between cultivated and fallow-land vegetation, the development of the latter often reflecting the time lapse since previous cultivation, pyric succession, and the intensity of grazing and erosion.

The structure of vegetation communities in noncultivated areas of the Ugijar Basin are affected by two other points, a rapid increase in altitude with horizontal distance, and the proximity to the coast. Polunin and Smythies (1973) divide the vegetation communities into five types ranging from low level Mediterranean plants to the alpine herbaceous perennials. The two communities of interest are the Mediterranean zone (0-1200m) and the Mediterranean-montane zone (1200-1700m). The former is characterised by shrubs such as Lygos sphaerocarpa, Cytisus sp., and Genista umbellata. Oranges grow to c1100m and olives to c1300m and towards the upper limits there are spiny thickets of deciduous shrub. The Mediterranean-montane zone was once probably natural oak, which still occurs in pockets for example near Bayarcal, and elsewhere woodland associated plants persist. Although each community has characteristic plants there is considerable variation in communities within and between zones. The type of vegetation in the Ugijar basin is matorral with two subdivisions 'high' and 'low' distinguishing between the state of the communities rather than differences in community composition. High matorral, in its most developed state forms dense thickets of evergreen shrubs 2 to 4m high. The dominant species for high matorral include Arbutus unedo, Erica arborea, Cistus monspeliensis, Quercus sp., Juniperus sp., Rosmarinus officinalis, Olea europaea, Phillyrea sp., and Asparagus sp. In low matorral the bushes may be only 0.5 to 1.5m high with scattered patches of bare ground. The main species are Rosmarinus officinalis, Stipa lagascae, Thymus vulgaris, and Cistus monspeliensis.

The main control on natural plant growth in Mediterranean ecosystems is water stress, and to a lesser extent nutrient and heat stress (Kruger, Mitchell, and Jarvis 1983). Miller (1983) describes the cyclic relation between precipitation and soil moisture for Mediterranean areas with respect to plant growth. In late autumn and winter precipitation and soil moisture are high. During the spring precipitation is low, but soil moisture remains relatively high for a while. By the summer both precipitation and soil moisture are low. However plant growth is limited in winter by low temperatures especially at elevation. Miller (1983) found that the predominant species of chaparral require mean daily temperatures greater than 10°C for growth, and Mooney (1983) puts the required temperatures for optimum photosynthesis rate for Californian flora at 16-29°C.

Adaptions to these stresses occur in two ways, the evolution of xerophytic characteristics, and the development of certain life cycle patterns. Xerophytic adaptations reduce water losses, increase heat tolerance, and improve water intake against high soil tensions. For example Shachori, Rosenzweig and Polkjakoff-Mayber (1967) found that maquis shrub (French equivalent to high matorral) and pine (in Israel) have the greatest capacity for moisture removal extending its effect to 8m in the soil. However they find little variation in terms of moisture depletion in the top metre of soil between various cover types.

Certain life cycle patterns have evolved to make most use of the harsh conditions. Some plants, for example Thymelaea hirsuta continue to photosynthesise throughout the year, while at the other extreme herbaceous annuals complete their life cycle within a relatively short growing season in spring (Orshan 1983). In between there is a variety of possible responses, for example Genista umbellata and Lygos sphaerocephala shed their leaves under water stress although their green stems can continue to photosynthesise, whilst other shrubs and herbaceous plants are effectively deciduous in habit and 'shut down' for the summer.

Despite a growing literature on Mediterranean ecosystems, the consumptive use of water for many species and the effect of plants on the hydrological cycle are still poorly understood. Furthermore much of the literature centres on analogous communities in California, Australia, South Africa, South America, and other Mediterranean countries like France, Italy and Israel (Harrison, Small and Mooney 1971, Gray 1982 and 1983, and Kruger et al 1983). References to Spanish ecosystem (with some exceptions like Pineda 1981) are largely confined to classification and description. Nevertheless it is evident that water and temperature stresses cause seasonal variations in the quantity and quality of plant cover, which is most in the spring (Miller 1983, Specht et al 1983) and least in late summer. This truism holds for the plant communities of the Ugijar basin and has significant implications for erosional processes. The occasional summer storm can cause a considerable amount of damage as the interception, storage and protective roles of plants are less than in winter and spring.

In more specific terms vegetation growth inhibits gully development. Graf (1979) formalises the role of vegetation in terms of changing the relationship between the threshold value of resistance on valley floors and the tractive force of flowing water. A decrease in vegetation cover on valley floors may sufficiently lower resistance to erosion to cross a threshold and induce entrenchment of the channel which in turn lowers the watertable and can accelerate vegetation decline in a positive feedback. Alternatively dense vegetation increases the resistance of valley floors by extensive root development and increasing surface litter creating greater roughness to flow, higher infiltration rates, and more sedimentation.

Bull (1979) also describes aggrading and degrading conditions as responses to thresholds of critical power. He suggests that semi-arid stream systems in fine-grained, easily eroded material are particularly sensitive to changes as a result of crossing threshold boundaries. The initiation of a minor channel decreases the residence time of ephemeral sheet flow and consequently infiltration falls and vegetation dies. Badlands can develop rapidly and the vegetation is unable to re-establish itself.

The effect of vegetation on gully growth has been observed by other authors. Ireland et al (1939) noted that erosion was accelerated along stream banks cleared of vegetation. Piest and Bowie (1974) found lower erosion rates from catchments with vegetated gullies than from those with nonvegetated gullies. Heede (1974) found that gullies did not usually occur in densely vegetated areas unless they invaded the area, and also noted that root systems tended to inhibit gully growth, although they could support piping. Revegetation of gully floors helps to stabilise them and is a standard technique for gully control (Ireland et al 1939, Hudson 1981). The extensive gullying of the Ugijar basin suggests that in many areas the vegetation cover biomass lies below or very close to the conditions for stability.

Finally the land use and agricultural practices (including the types of crops) are conducive to extensive erosion. There are two types of agricultural systems (Thornes 1976):

1 Secano or dry farming of tree crops (almonds and olives) and cereals.
2 Regadío or irrigation-dependent agriculture which maintains vines and relatively high value 'salad' crops.

The regadío is confined to the valley bottoms and the proximity of villages as it depends on water and the maintenance of irrigation systems. Several regadío harvests a year are possible, and although they are important economically regadío agriculture occupies a small area. Nunez Noguerol (1969) estimates that in most Alpujarran municipalities regadío, secano and uncultivated land account for 6.14%, 14.83% and 79.00% respectively of the total area.

More important are land use practices on the secano land as this covers a greater area and is more prone to erosion. Tree crops are usually planted in grid patterns over hillslopes or on terraces down steeper slopes (these patterns come out clearly on the aerial photographs). Soil between the trees is often kept deliberately clear of vegetation and virtually unprotected from the agents of soil erosion. Occasionally cover is provided by weeds or plants such as Capparis. Fields are ploughed (though not always along the contour) to break up rills which form readily on the loose soil. Tree and cereal crops are frequently cultivated to the very edge of gully systems to maximise the available space (photograph 2.2) however such practices probably encourage the development of the gullies. Cereals are harvested in the summer, leaving the ground unprotected during the autumn and winter rains. Furthermore a system of año-y-vez is employed whereby land may be left fallow for up to seven years (Thornes 1976). The Ministerío de Agricultura (1982) has published values for the 'C' component in the Universal Soil Loss Equation reflecting the

Photograph 2.2 Tree crop agriculture and gully erosion near Ugijar

affect of the crop on erosion. These are shown in table 2.1 and bring out the susceptibility of cultivated trees and vines to erosion, with very small values for forest, matorral and regadío.

Other practices encouraging erosion include irrigation methods, for example flooding fields, failing to maintain terraces, severe fragmentation of holdings, and rural depopulation. Much of the uncultivated land is used for grazing animals (sheep and goats), and overgrazing is a frequently cited cause of accelerated erosion. The aggregate of cultural, socio-economic and natural factors explains the considerable erosion of the Ugijar valley.

2.2 The Field Site - Choice and Description

The actual study site has to fit three basic requirements: a propensity towards gullying, simplicity of variables, and accessibility.

Firstly the site should have not only incipient or small gullies growing into it but also have a propensity towards extensive gully development. Gullies may, given the right conditions, appear on a wide range of lithologies, but certain lithologies are more prone to gullying than others. For example following an extreme storm, a gully could form on a certain lithology, but not develop or last very long. On such lithologies the gully does not become (or stay) unstable. Some rock types eg loess (USA, China), shales and sandstones (Canada), and marls (Spain) are very susceptible to gullying and badland development.

Badlands as such are not suitable for the project. The very high drainage densities mean gully catchment areas are small and the relative magnitudes of surface and subsurface hydrology could be difficult to monitor. Excessive competition between systems may mean erosion rates are partly determined by the degree of success of the competition so that rates are inhibited (Faulkner 1974). It is preferred to examine non-competing gully systems with well defined catchment areas. Also evidence suggests that high erosion rates in badlands are localised, confined to areas where headward growth is still possible as opposed to areas where downwasting of

Table 2.1 Values for the 'C' component in the Universal Soil Loss Equation for different types of vegetation (Ministerio de Agricultura 1982)

Vegetation Type	'C'
Dense Forest	0.01
Matorral - good cover	0.08
Matorral - poor cover	0.20
Regadío	0.04
Cultivated annuals and herbaceous plants	0.25
Cultivated trees and vines	0.40

divides is the main result of erosion (Seginer 1966). The gully under consideration should be actively eroding and not be a fossilised form "touched up" occasionally by erosion processes to look fresh.

Secondly the site should have as few complicating variables as possible in order to suit the proposed model and sampling scheme. Therefore the site should be based on a relatively straightforward geological structure with a "uniform" lithology. Slope profiles and plans should also be simple so that flow lines can be drawn relatively easily from gully head to divide.

The field site chosen for the research had to be one where gullies develop readily into 'typical' forms unhindered by constraints. What constitutes a typical gully is very difficult to define considering the variety of morphologies, modes of origin and dominant processes. Many gully systems are a result of human activity where runoff is often concentrated by the construction of roads, railways, ditches, terraces and terrace outlets, culverts, contour ploughing and the subsequent breakdown of furrows, and ground compacted by stock or heavy machinery (Ireland et al 1939). Such concentrations of water can be seen as a peculiar hydrological case for gully growth, not necessarily replicated naturally to the same extent. One approach would be to choose gullies which develop into the most simple forms by 'natural' means, then only the most fundamental processes are in operation. This is the most straightforward case to study and the research could be extended later to include more 'atypical' forms.

The gullies should not be formed by unusual natural mechanisms either such as the sinking of ground (Rubey 1928, Buckham and Cockfield 1950) or pipes (Heede 1971, Imeson and Kwaad 1980, Harvey 1982) both of which would compromise the sampling design. The presence of rills would complicate the area upslope of the gully and should also be absent.

Ideally a non-vegetated site should be used because vegetation is a complicating factor, and is still poorly understood. The site should also be uncultivated as the disturbance of the soil surface and structure by crops, heavy machinery or construction will tend to exacerbate the problem of accelerated soil erosion.

Thirdly the site should be accessible by vehicle so that it can be reached daily, easily, sometimes with heavy equipment or water barrels. Yet it should still be "out of the way" to reduce sight-seeing and vandalism.

These considerations represent the ideal case, and not necessarily the actual site. The next step was to find the most suitable site which met as many of the stated criteria as possible.

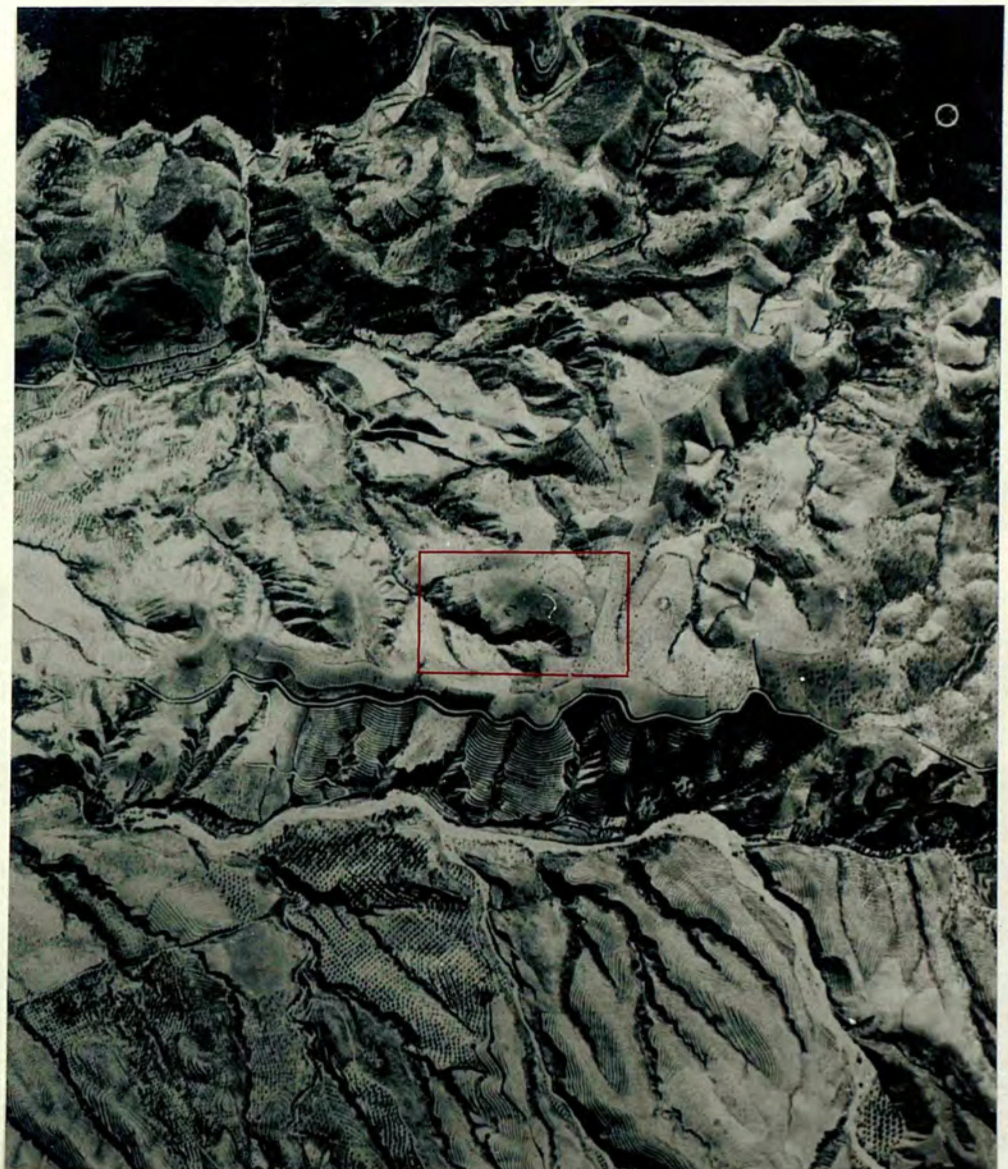
The Location, Topography and Geomorphology of the Study Site

The study site is in the southern part of the Ugijar basin approximately 4km from Ugijar, 7.5km from Mecina Bombaron, and 8.25km from Cadiar (figure 2.2) at 3° 5' W and 36° 57' N. The hillslope is easily accessible by road and dirt track, although it is hidden from the main road which goes between Ugijar and Yator. During the entire period of field monitoring there were no signs of vandalism or disturbance to the installed equipment or slope.

The slope forms part of the headwaters of an ephemeral channel or rambla draining into the Rio Yator (photograph 2.3). Figure 2.4 gives a good indication of the topography of the study site whereas figure 2.5 shows some detailed geomorphological features and the layout of the sampling scheme. The shape of the hillslope varies along its length. The eastern end of the slope forms a spur which is steepest in the middle part of the profile and flattens towards the watershed. The base of the slope forms a small concave depression in both plan and profile above the gullied footslopes. This is site 1 for the sampling scheme. In the middle section of the slope, between the landslide and the second instrumented part, the hillslope is convex in profile, but rectilinear in plan form with a north easterly aspect. Towards the northern end the hillslope rounds into a concave basin forming one half of the headward catchment for the rambla.

Slope lengths from the watershed to rambla floor vary between 80-250m and altogether the hillslope covers an area approximately $0.08km^2$. The Mapa Militar de España at 1:50 000 scale shows the

Photograph 2·3 **Detailed aerial photograph of the site**



0 0·5
km

N

Figure 2·4 Topographic map of the study site

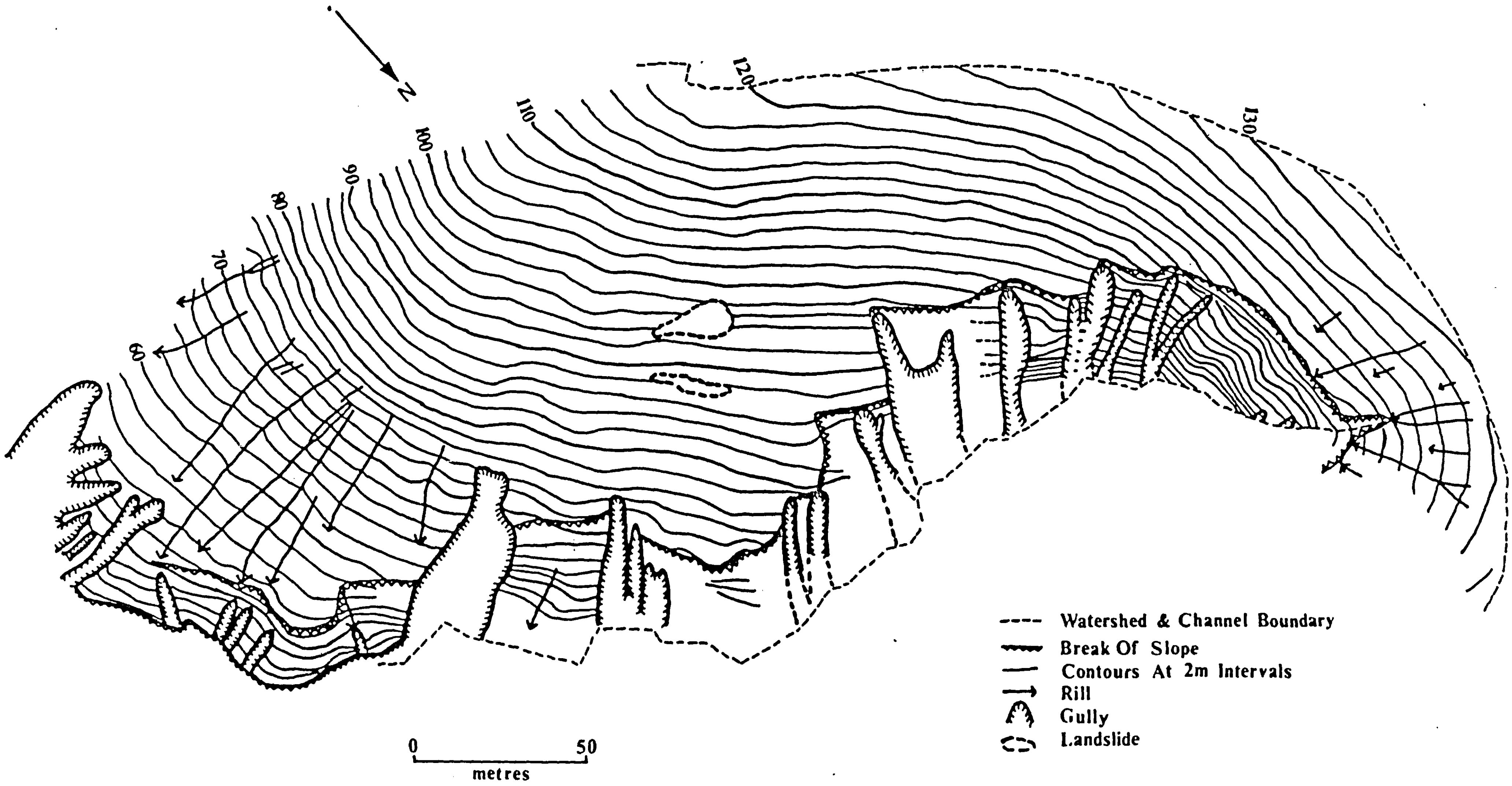
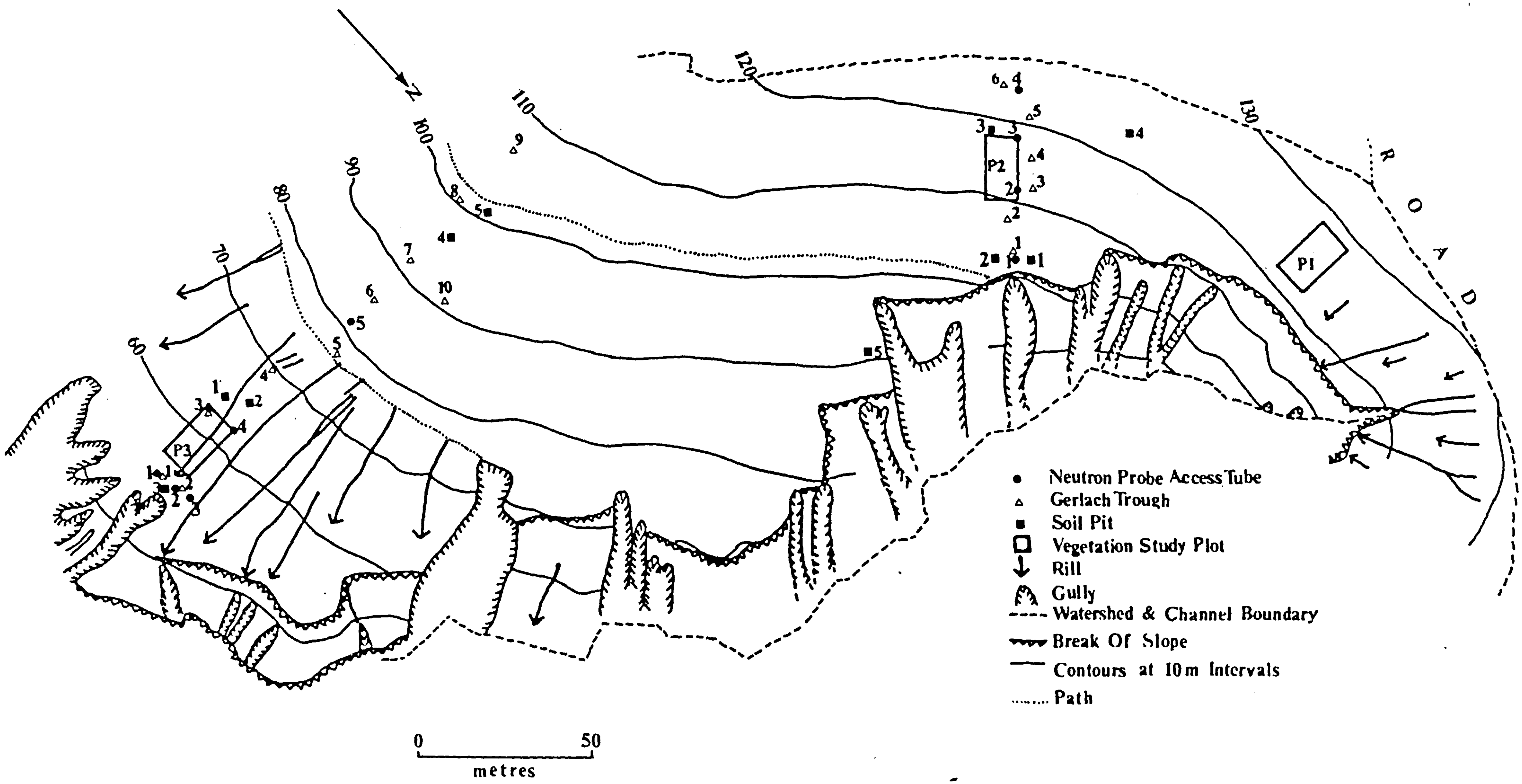


Figure 2·5 Map of the study site showing the sampling layout



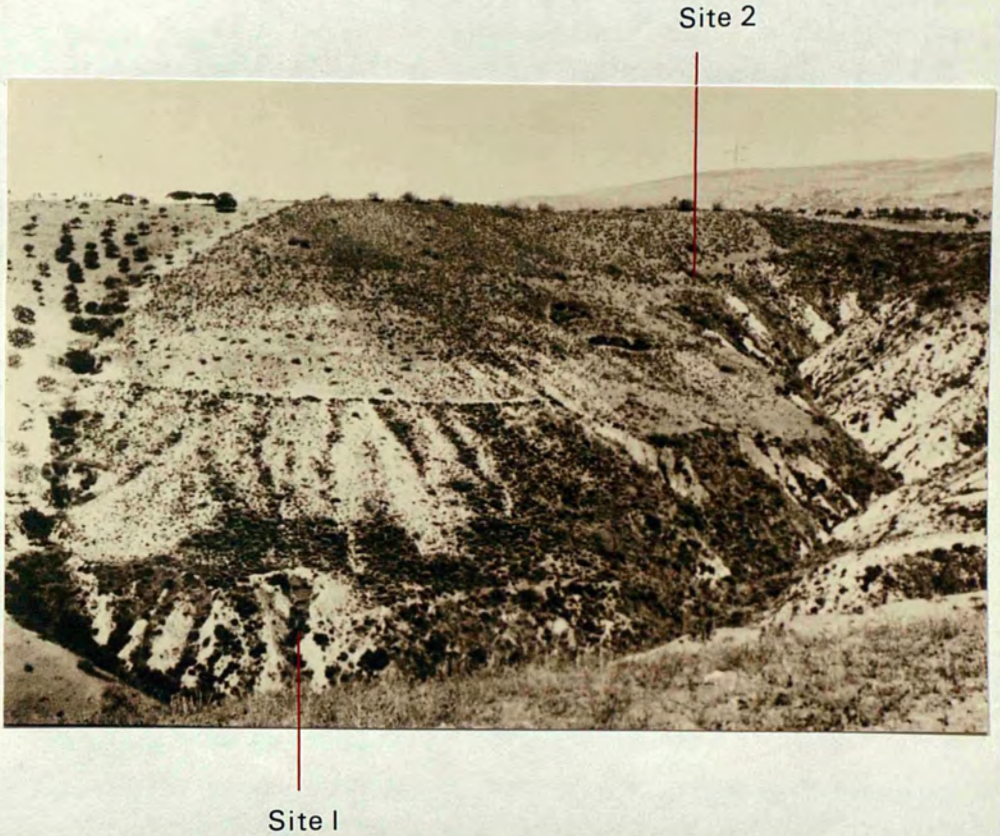
700m contour dissecting the hillslope, however the field survey was not linked to the national network and contours are relative. The maximum difference in altitude on the slope is approximately 90m.

The site is bounded by some distinctive features. The watershed is sufficiently delineated although towards the northern end the watershed is cut by a dirt track which may be responsible for the rills in this area. The lower boundary was taken as the rambla channel bed, and the two lateral bounds by rills.

There is a notable break of slope on the site which runs for its entire length and approximately divides the gullied footslopes from the relatively unscarred hillslopes above. Slope angles above the break are 20° to 28°, and below the break 35°, as is clearly shown in the survey. Several sections of the footslopes could not be mapped as they were too steep to work on safely, were thickly vegetated in parts, and unstable near the gullies. In photograph 2.4 the footslopes are brought out not only by the gullies, but also by the marked increase in vegetation. This lower section may have been steepened by incision from ephemeral flow in the rambla and gully growth, and the break of slope may have then been intensified by repeated ploughing.

There are two paths which cut across the slope. One, about a third of the way down from the watershed, although marked by the vegetation, has not interfered appreciably with the processes and forms on the slope. The second path dissects the lower third of the marl spur at the head of a series of rills. The path has apparently affected the surface hydrology of the slope by channelling some of the surface flow into rills. The rills have a stepped profile with short reaches and waterfalls with plunge pools occuring downslope. Towards the flatter portion of the slope the morphological form of the rills tends to disappear. At the maximum they are up to c40cms deep, and although not seen in action, are probably capable of transporting significant quantities of water and sediment. One of these rills enters the instrumented area on site 1, and peters out about 10m to the eastern side of the main gully. Although this undoubtedly increases water and sediment load above the gully, the extent of

Photograph 2.4 Photograph of the study slope



its influence is unknown. The second groups of rills is found in the concave depression at the northern end of the slope. These do not interfere with the monitored sections.

Gullies only occur along the footslopes with the deepest ones in the middle section. Two gullies were chosen for analysis. One at the eastern end is eroded solely in marl, it has bifurcated and has a small tributary on the southern side. Photograph 2.5 shows site 1 on the marl which brings out the gullies at the foot of the slope as well as the vegetation patterns and geomorphological features. The main gully head is a regular arcuate shape (photograph 2.6) with vertical walls. Tension cracks have developed around the gully head. The gully slopes around the bifurcation are shallower. The gully bottom is covered in loose marl debris and is vegetated. The second gully is largely cut in a reddish conglomeratic soil towards the northern end of the slope. It has more gently sloping sides, and is less sharply defined. It is also less deep but more densely vegetated (phogograph 2.7).

There is no clear topographical relationship between the rills and gullies. The largest gullies are not fed by rills, indeed few rills flow into gullies, and the two types of features are developing independently. A small landslide feature shows that parts of the hill are potentially unstable. Few soil slopes are stable in the long term for such steep angles. The only evidence of deposition is the unconsolidated deposits in the larger gullies and rambla channel bed. The hillslope is dominated by erosional processes, of these sheetwash is probably the most widespread although gully erosion looks impressive.

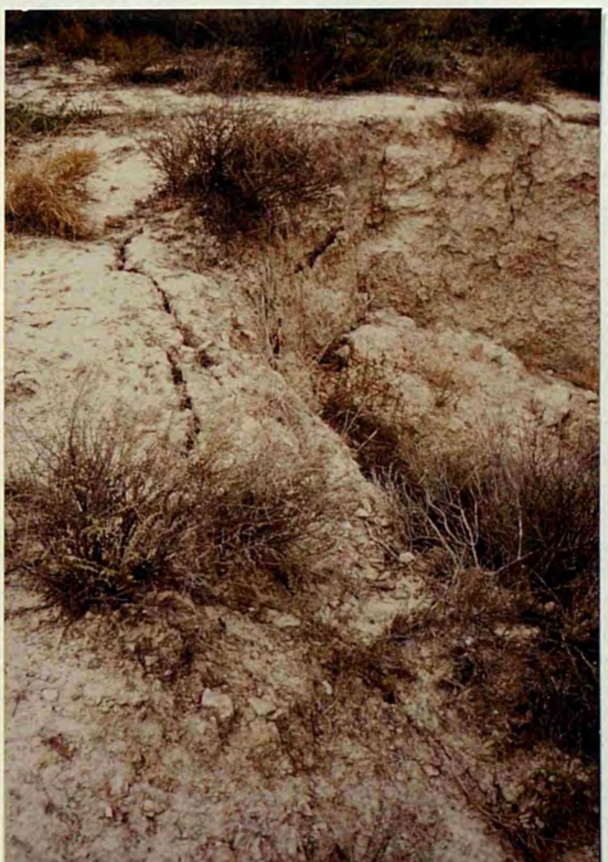
Soils

The underlying bedrock is the late Tertiary whitish-grey marl. This is overlain on the upper sections of the hillslope and towards the northern end by a reddish conglomerate-derived material which may be the remnants of one of the sandy lithologies or some slope deposit (figure 2.6). Near surface profiles are examined for both sites, one on marl and the other on the conglomerate soil, and descriptions are based on the Soil Survey

Photograph 2·5 Site 1 - Marl

Gully Head

Photograph 2·6 Gully head on site 1



Photograph 2·7 Site 2 - Conglomerate

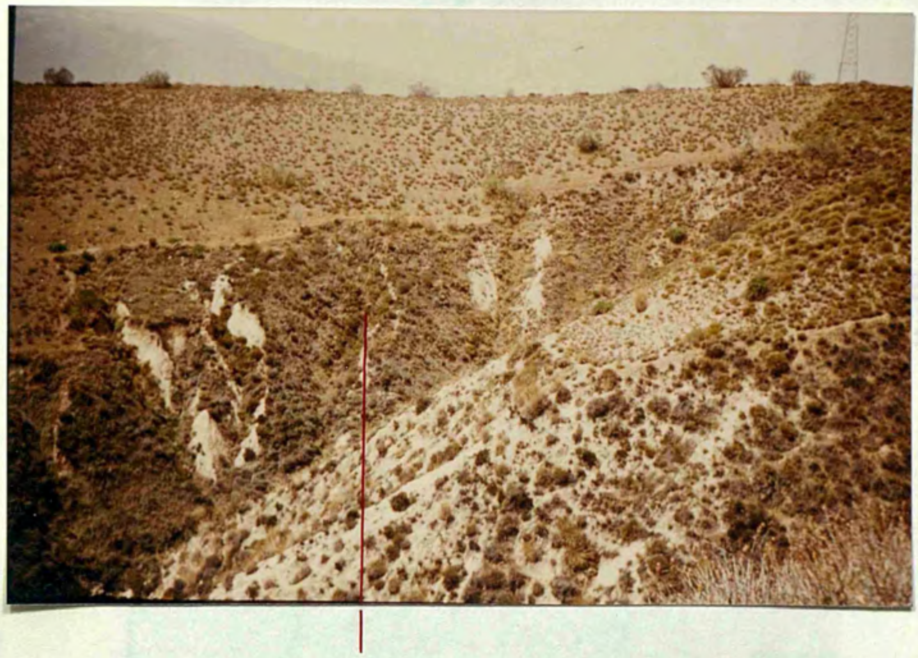
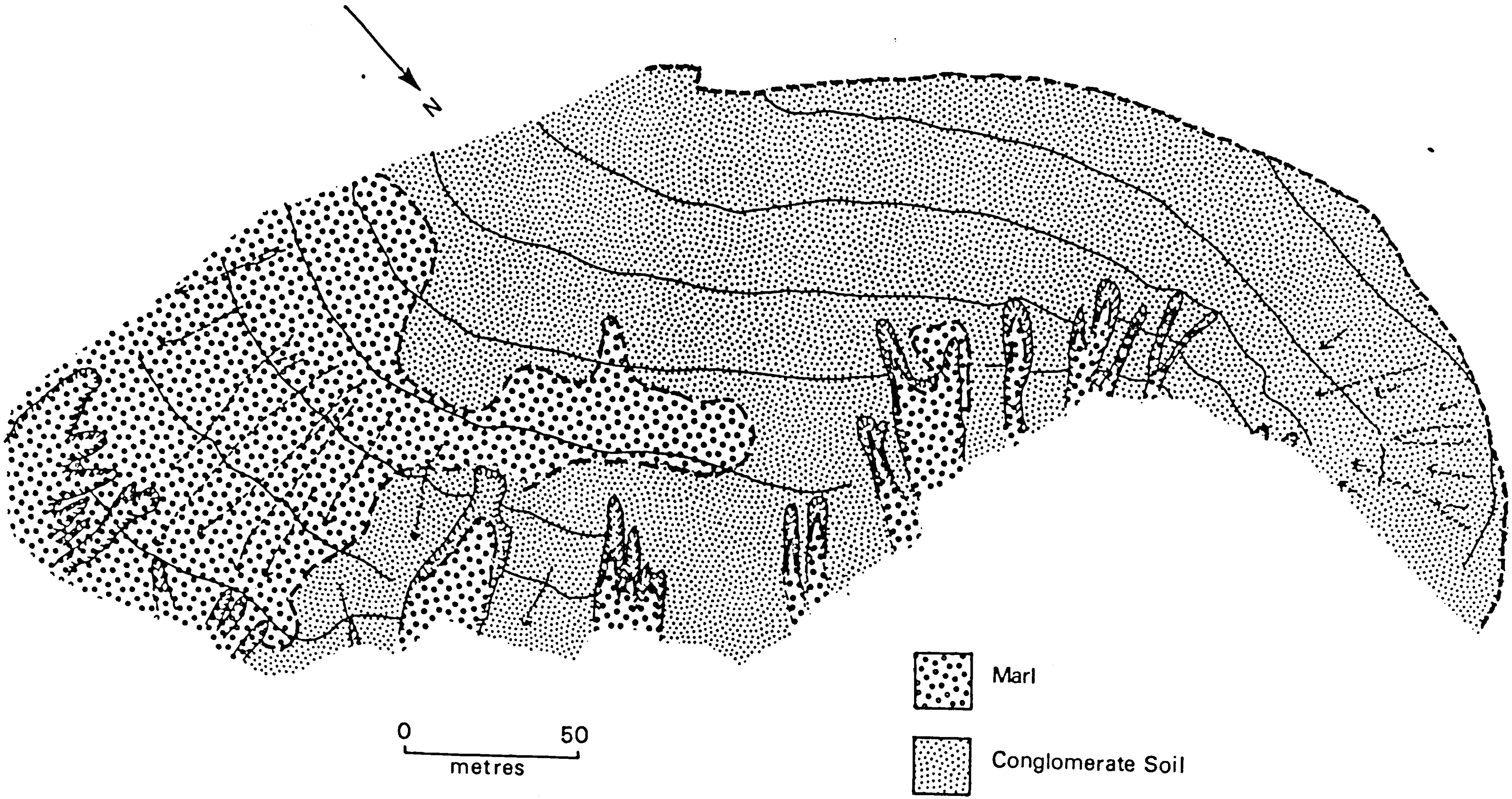


Figure 2·6 Soils on the study site



Field Handbook (Hodgson 1976). Further analysis includes bulk density, calculations of porosity, % stoniness, and % $CaCO_3$ content.

Marl

Profile descriptions were taken from soil pits 1, 2, and 6 (which is actually at tube 1 on site 1) to depths 131 cms, 64 cms, and 65.5 cms respectively on the lower section of the slope (profiles for pits 1 and 6 are shown in figure 2.5). The term 'soil' is used loosely to describe the material although it is probable that the profile consists of weathered marl and bedrock. There are two horizons which are distinguished in the field by a marked change in structure. The upper horizon extends to 10-15cms, and the lower continues beyond 131cms (photograph 2.8, and table 2.2). The boundary between the two horizons is abrupt (taking place within 0.5-2.5cms), wavy and dips parallel to the slope.

The upper horizon is capped by a thin crust, less than 5mm thick and composed of fine silt or clay material. A similar description is given by Farres (1978). Below this the soil consists of moderately well developed aggregates up to 5cms long, which are loosely packed and separate easily when disturbed. They are also composed of fine material.

In the lower horizon the marl is cemented together to form a solid, massive bedrock. This shatters into angular shards with conchoidal fractures along horizontal planes in the formation. Mechanical disturbance also causes plenty of dust.

The visual differences in horizons are the greater porosity of the upper horizon supported by estimates of porosity calculated from the bulk density of samples and measurement of the specific density of particles (Black 1965). The values for specific density were calculated using a pyncometer (Black 1965) and as table 2.3 shows the values are consistent within and between soil types. Using the estimates for bulk density and specific density, porosity values are calculated for the profile where

$$P = 1 - (\text{bulk density}/\text{specific density}) \times 100$$

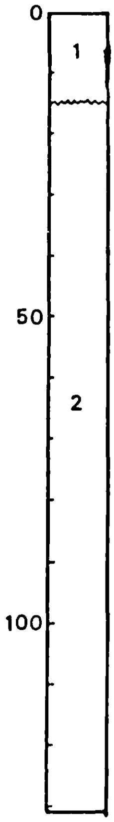
The porosity of the marls decreases with depth for example mean

Photograph 2·8 Soil section on marl

Photograph 2·9 Close up of marl bedrock

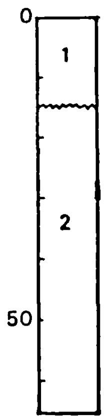
Table 2·2 Soil section on marl

Pit 1



Soil Profile	Sample Depth cms	Bulk Density	Porosity %
	18·6	1·47	45·1
		1·48	44·8
		1·43	46·6
		1·47	45·1
	38·6	1·49	44·4
		1·70	36·6
		1·63	39·2
		1·65	38·4
	58·6	1·66	38·1
		1·68	37·3
		1·71	36·2
		1·61	39·9
	78·6	1·81	32·5
		1·73	35·4
		1·73	35·4
	98·6	1·75	34·7
		1·82	32·1
		1·97	26·5
	118·6	1·73	35·4
		1·77	34·0
		1·74	35·1

Pit 6



Soil Profile	Sample Depth cms	Bulk Density	Porosity %
	25·5	1·43	46·6
		1·41	47·4
	45·5	1·30	51·5
		1·31	51·1
	65·6	1·60	40·3
		1·60	40·3

Table 2.3 Values for the specific density of soils

Conglomerate	Clay	Marl
2.70	2.75	2.67
2.77	2.72	2.69
2.68	2.69	
2.73	2.74	
2.71		
2.79		

values at pit 1 are 45.4% at 18.6 cms, 39.7% at 38.6 cms, 34.4% at 78.6 cms, and 34.8% at 118.6 cms. There is also a marked difference in the strength of aggregates between horizons. In horizon 1 individual soil aggregates easily crumble between the fingers whereas shards from horizon 2 can only be crushed underfoot.

The marl is a whitish grey throughout, but in pit 2 some yellow streaking 5-10mm wide with grey edging 1-2mm wide was discovered (photograph 2.9).

There are few stones in the profile and there is a marked decrease of stoniness with depth from 7.54% (between 0-15cms), 3.60% (between 15-30cms) and 0.12% (between 30-45cms) in the stone content greater than 2mm by percent weight (from pit 3). These stones are small, ranging from 6-20mm and are subrounded to subangular quartz fragments.

Some very fine roots and one or two woody roots occur in the profile. In pit 2 the roots were found lying between horizontal planes or cracks in the marl, however as the pits were deliberately sited away from bushes, the placing of the pits would have an effect on the density of root networks observed.

Calcium carbonate values were determined following standard preparation and using the volumetric calcimeter method (Black 1965). Samples were analysed from the surface and with depth. Table 2.4 shows that on the marl values for percent calcium carbonate range between 40.9 and 67.4%. The mean value for surface samples is 47.4% (with a standard deviation of 4.64), and 55.1% (with 7.83 standard deviation) for values with depth, including the marl values on site 2 (table 2.7). A difference of means test between all marl samples measured at depth and those measured on the surface shows that the percent of $CaCO_3$ is significantly greater at the 0.001 significance level with samples taken from depth. Within the depth samples, those from pit 3 have lower values than elsewhere. This suggests that whilst the marl is highly calcareous, the amount of $CaCO_3$ falls slightly for weathered parts like the surface. Values for $CaCO_3$ are variable. The analysis shows that within any one horizon the

Table 2.4 $CaCO_3$ Determinations on the marl

(a) Percent $CaCO_3$ from surface samples

43.30	56.63	51.08	50.83	53.97	52.89	49.81
51.86	52.24	62.28	45.83	49.65	50.71	54.28
47.00	46.75	46.71	44.80	43.29	44.50	42.01
46.13	47.19	41.73	48.15	47.55	41.75	47.75
46.46	50.54	41.77	45.63	43.18	49.60	46.02
42.99	43.09	41.04	43.46	42.89		

Number of observations	40
Mean	47.43
Standard deviation	4.69

(b) Percent $CaCO_3$ from depth samples

Pit 1		Pit 6	
Depth cms	$CaCO_3$ %	Depth cms	$CaCO_3$ %
18.6	50.97	20.5	45.04
	47.13		42.26
	55.10		45.30
	56.91		40.93
	52.64		45.39
78.6	67.43	34.5	47.89
	66.81		46.15
	65.78		48.52
	60.44		48.98
	62.43		49.45
118.6	55.98	76.5	45.36
	53.09		45.24
	58.35		
	58.32		
	57.05		

measured value for $CaCO_3$ may range by up to 9% within five samples. This may reflect either operator error or actual variations in the soil. Perez, Garcia, and Tapia (1982) present $CaCO_3$ data varying from 42.4% to 70.7% for 12 sites in Murcia but do not give standard deviations at each site. Such variation may almost occur at one site and so does not necessarily represent variations in the composition of the marl between sites.

Conglomerate Slope Deposit

The profile descriptions for the conglomerate soil are derived from soil pits 1, 2, and 3 (figure 2.5) which are 147cms, 156cms and 113cms deep respectively. The pits when taken together show up to four horizons (photograph 2.10 and table 2.5). The first horizon, for all locations, extends to 16-20cms. The depth of the second horizon is more variable. In pit 3 (furthest upslope) it continues beyond 113cms, but in pits 1 and 2 it finishes between 60-65cms and 100-105cms respectively. The third horizon in pit 1 continues from 60-65cms to 85-95cms, but goes beyond 156cms for pit 2. Finally a fourth horizon, only visible in pit 1, extends from 85-95cms to over 145cms. Generally the boundaries between horizons vary from 2.0-6.0cms in a wavy line parallel to the surface. However in pit 1 the boundary between horizons 3 and 4 is very abrupt occurring within 0.5cms, although the form of the interface is irregular.

Horizon 1 is a brownish-reddish colour with no mottling. In pit 4, 40.6% of the stones (by weight) are greater than 2mm, and tend to range between 2-60cms along the longest axis. These stones are mainly fragments of micaschist with garnets or a white and red quartz stone, transported from the surrounding highland areas. The shape of the stones varies with lithology and degree of recent disturbance. Many micaschist fragments are rotten and can be broken easily into platy pieces. Recently disturbed micaschist pebbles are platy and angular, whereas less disturbed pebbles tend to be sub-rounded. The quartz grains are variable in shape but tend to be subrounded. Calculated porosity values are between 36.3% to 46.5% (table 2.5), with pores between 0.5 to 2mm in diameter. There was no evidence of piping or macropores. Aggregates have low strength values and crumble easily. Fibrous

Photograph 2.10 Soil section on conglomerate

Table 2·5 Soil section on conglomerate

Pit 1

Soil Profile	Sample Depths cms	Bulk Density	Porosity %
1		1·57	42·5
	15	1·64	39·9
2		1·53	44·0
		1·71	37·4
	35	1·63	40·3
		1·74	36·3
		1·67	38·8
	55	1·72	37·0
3		1·55	43·2
		1·55	43·2
	75	1·46	46·5
		1·58	42·1
4		1·42	47·5
	95	1·25	53·8
		1·44	46·8
		1·44	46·8
		1·40	47·8
	115	1·40	[illegible]
		1·56	41·8
	135	1·47	45·1

Pit 2

Soil Profile	Sample Depths cms	Bulk Density	Porosity %
1		1·74	36·3
	5·5	1·70	37·7
		1·46	46·5
2		1·69	38·1
	25·5	1·64	39·9
		1·69	38·1
	45·5	1·68	38·5
		1·68	38·5
	65·5	1·71	37·4
		1·70	37·7
	85·5	1·71	37·4
		1·77	35·2
3	105·5	1·56	42·9
		1·58	42·1
	125·5	1·57	42·5
	145·5	1·50	45·1

roots often less than 1mm diameter are common, although there are a few larger woody roots. There is no evidence of other recognisable organic matter such as partly decomposed leaves or twigs incorporated in the soil.

Horizon 2 is similar except for an observed increase in stoniness, decrease in voidage and a change in colour. Table 2.6 shows increases in stone content of 37.4% to 61.3% in pit 3 and 40.8% to 61.5% in pit 4, representing changes over some 60cms depth. The stones appeared to be aligned with the long axis horizontal, but there are no definable strata or lines. Towards the bottom of this horizon some large boulders appear, again horizontally orientated and exceeding 60cms in length. The number of roots decreases to an estimated 1 to 25/100cms^2, although this again reflects the distance of the pits from bushes. In pit 1 there were 2 roots greater than 0.5cms, but the rest are fine and fibrous. Ants are active in both horizons 1 and 2.

Horizon 3 consists of a yellow clayey matrix, sometimes with greyish green mottles 2 to 5 mm wide. This has a very low stone content, visual observation suggesting less than 5%. Pore sizes are very small, with calculated porosities of 42.1 to 46.5%, and there were less than 10 roots/100cms^2.

Horizon 4 is the whitish marl bedrock, very compact and hard rather than weathered. Again the rock shatters into small shards on impact. There are few stones but some fibrous roots do reach this horizon. Porosity values ranged from 41.8 to 47.8%.

Table 2.7 shows $CaCO_3$ values for the conglomerate with depth and on the surface. The values for horizons 1, 2 and 3 vary between 0.0 to 8.2% and a difference of means test shows that there is no significant difference in values for calcium carbonate for these horizons between soil pit 1 and 2 with a mean and standard deviation of 3.4% and 2.0 in pit 1, and 2.4% and 2.5 in pit 2. Surface values are also low with a mean of 0.2% (standard deviation equals 0.20). The low quantity for horizon 3 is particularly interesting suggesting that the clay horizon is not derived from the marl itself - however, neither does this show that it is developed from the conglomerate. In pit 1 the calcium

Table 2.6 Variations in stone content greater than 2mm on the conglomerate soil

Depth cms	Pit 3 %	Pit 4 %	Pit 2 %
0 - 13	37.35	40.77	40.64
14 - 27	3.77	45.61	37.73
28 - 44	55.16	60.61	
45 - 55	61.26	61.54	

Table 2.7 $CaCO_3$ Determinations on the conglomerate soil

(a) Percent $CaCO_3$ from surface samples

0.11	0.22	0.46	0.31
0.07	0.34	0.26	0.10
0.00	0.00	0.14	0.00
0.00	0.00	0.73	0.22
0.22	0.00		

Number of observations 18
Mean 0.18%
Standard deviation 0.20

(b) Percent $CaCO_3$ from depth samples

Pit 1 Depth cms	%	Pit 2 Depth cms	%
15	3.59	5.5	2.13
	3.11		6.33
	0.00		1.20
	3.59		0.27
	1.22		2.13
35	0.75	45.5	0.00
	4.94		3.06
	4.95		2.60
	5.40		8.17
	4.98		2.13
75	3.11	85.5	0.00
	1.23		3.05
	7.29		1.18
	4.01		0.00
	2.60		0.00
115	58.17	125.5	0.00
	64.14		0.00
	62.46		5.46
	61.28		4.49
	63.53		5.88
135	59.89		
	62.69		
	62.03		
	63.39		
	62.75		

carbonate values for the marl horizon 4 vary between 58.2 to 64.1%, in line with values recorded for marl on site 1.

Vegetation

The most common shrubs and herbaceous plants growing on the site at Easter 1982 have been identified (Botany Dept, Bedford College, and Natural Science Museum) and are listed in tables 2.8 a and b. Some specimens could not be identified at all, or only at the family (rather than species) level. Shrub families found include Compositae, Labiatae, Leguminosae and Thymelaeaceae, and the commonest herbaceous families are Leguminosae, Compositae, and Cruciferae. The species composition and the size and spacing of bushes suggests a "low matorral" community.

Although a proper analysis of the effect of vegetation cover on the hydrological budget does not fall within the scope of this thesis, the spatial distributions of soil moisture and organic content of surface soil samples were analysed with reference to the bush vegetation cover to estimate the affect of different plant densities on patterns of soil moisture and nutrient pools. The exercise was carried out in March 1984 two or three days after a heavy downpour, when vegetation cover would be approaching the spring maximum biomass.

Three vegetation plots 10m x 20m were established with one on marl and two on the conglomerate soil (figure 2.5). Plots 1 and 2 on the conglomerate soil were dominated by Genista umbellata and Artemisia sp respectively, but the vegetation densities were different. Plot 3 on the marl was vegetated by a mixture of species with a relatively high density cover similar to plot 1.

On each plot 40 random coordinates were used as a soil sampling basis. For each point the distance to, and diameter of the crowns of the four nearest bushes were noted, and a surface soil sample was collected using a Kubiena tin (7.5 cms x 6.2 cms x 3.7 cms). Soil moisture contents of the samples were determined thermogravimetrically (Black 1965). The samples were then divided into two groups using a 2mm seive. In the coarser fraction the "macro-organics" were picked out by hand and weighed. The percent

Table 2.8 Vegetation on the study site.

(a) Shrubs

Family	Name	Habitat
Compositae	Artemisia sp.	Limestone places
Labiatae	Lavandula stoechas	Dry stony places, sunny hillsides, pine woods, south & central Spain.
Labiatae	Phlomis sp.	
Leguminosae	Genista umbellata	Dry stony hillsides, south & southeast Spain
Leguminosae	Lygos sphaerocephala	
Thymelaeaceae	Thymelaea hirsuta	Dry places, maritime sands, Mediterranean region.

(b) Annual and perennial herbaceous plants found on the site

Family	Name	Habitat
Bromeae	Bromus rubens	Cultivated ground, dry waste places
Capparidaceae	Capparis spinosa	Rocks, waste places, littoral
Compositae	Calendula arvensis	Cultivated and unculti-vated ground, widespread
	Chrysanthemum segetum	Cornfields, cultivated ground, widespread
	Leontodon manoccanus	
Cruciferae	Alyssum sp.	Stony places
	Biscutella auriculata	
Leguminosae	Lathyrus clymenum	Sandy and stony places, Mediterannean Europe
	Trifolium arvense	
	Trifolium campestie	
	Vicia lutea	Grassy places, widespread
Liliaceae	Muscari comosum	Rocky ground, fields, vineyards, widespread
Papaveraceae	Papaver	Waste places, circum Mediterranean
Plantaginaceae	Plantago afra	
Polygonaceae	Rumex acetosella	
Primulaceae	Anagallis foemina	Cultivated places, fields, & track sides
Resedaceae	Reseda lutea	Fallow ground & stony places, especially limestone, circum Mediterranean
Rosaceae	Sanguisorba minor	Widespread
Rubiaceae	Galium verrucosum	

of the "micro-organics" were determined for the fine fraction by weight loss on ignition, with a correction for $CaCO_3$ content on the marl samples. The total organic content was expressed as the percent of the dry weight of the soil sample.

Sketch maps of the overall vegetation for each plot were made using tapes and drawn in figures 2.7, 2.8, and 2.9. Mean bush crown diameters are 55.8cms, 62.8cms, and 31.8cms on plots 1, 2, and 3, with mean distances between bushes being 59.1cms, 155.5cms and 43.4cms respectively. Frequency diagrams of bush diameters (figure 2.10) show that although the largest bushes are on plot 1, so is the greatest variation in class size. On plot 2 bush sizes are skewed to the right with a predominance of the larger sizes, and on plot 3 bush sizes tend to be skewed to the left favouring the smaller sizes. Figure 2.11 shows that the variation in bush spacing is greatest for plot 2 without a strong modal position. On plot 3 there is a peaked distribution with common distances around 30 to 40cms but on plot 1 the variation is flatter with most bushes 40 to 70 cms apart.

An estimate of cover provided by the vegetation is obtained by dividing the mean diameter of the bushes by the mean stem distance between them. These values are 0.94, 0.40, and 0.73 for plots 1, 2, and 3, bearing out the observation that plot 1 is the most densely vegetated site, and plot 2 the least vegetated one.

The results (means and standard deviations) for the soil moisture and organic data are given in table 2.9, and the results from the difference of means tests are in table 2.10 The amount of macro-organics is highest on plots 1 and 3, and both are significantly higher than macro-organic contents of soils on plot 2. Micro-organic and total organic content (which is greatly influenced by the micro-organic content) are highest for plot 1 and least for plot 3. The soil moisture contents are higher for the marl at the 0.001 significance level. However on the two conglomerate plots there was also a significant difference in the soil moisture content of the soils by the 0.05 level. Mean values were 20.1% for marl and 11.6% and 10.8% (with standard deviations of 3.05, 1.36 and 2.20) for plots 2 and 1 respectively.

Figure 2·7 Vegetation cover on plot 1

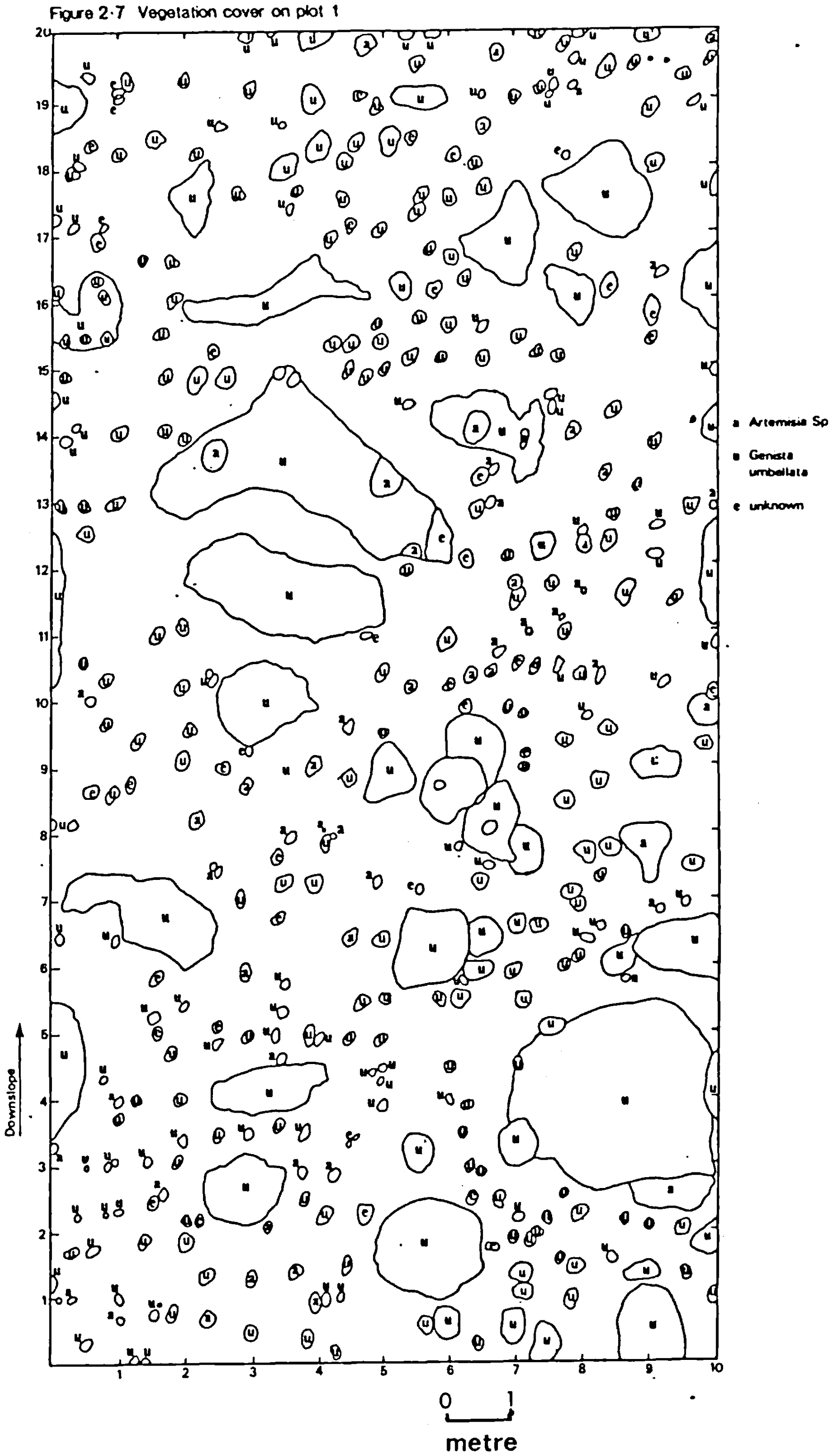


Figure 2·8 Vegetation cover on plot 2

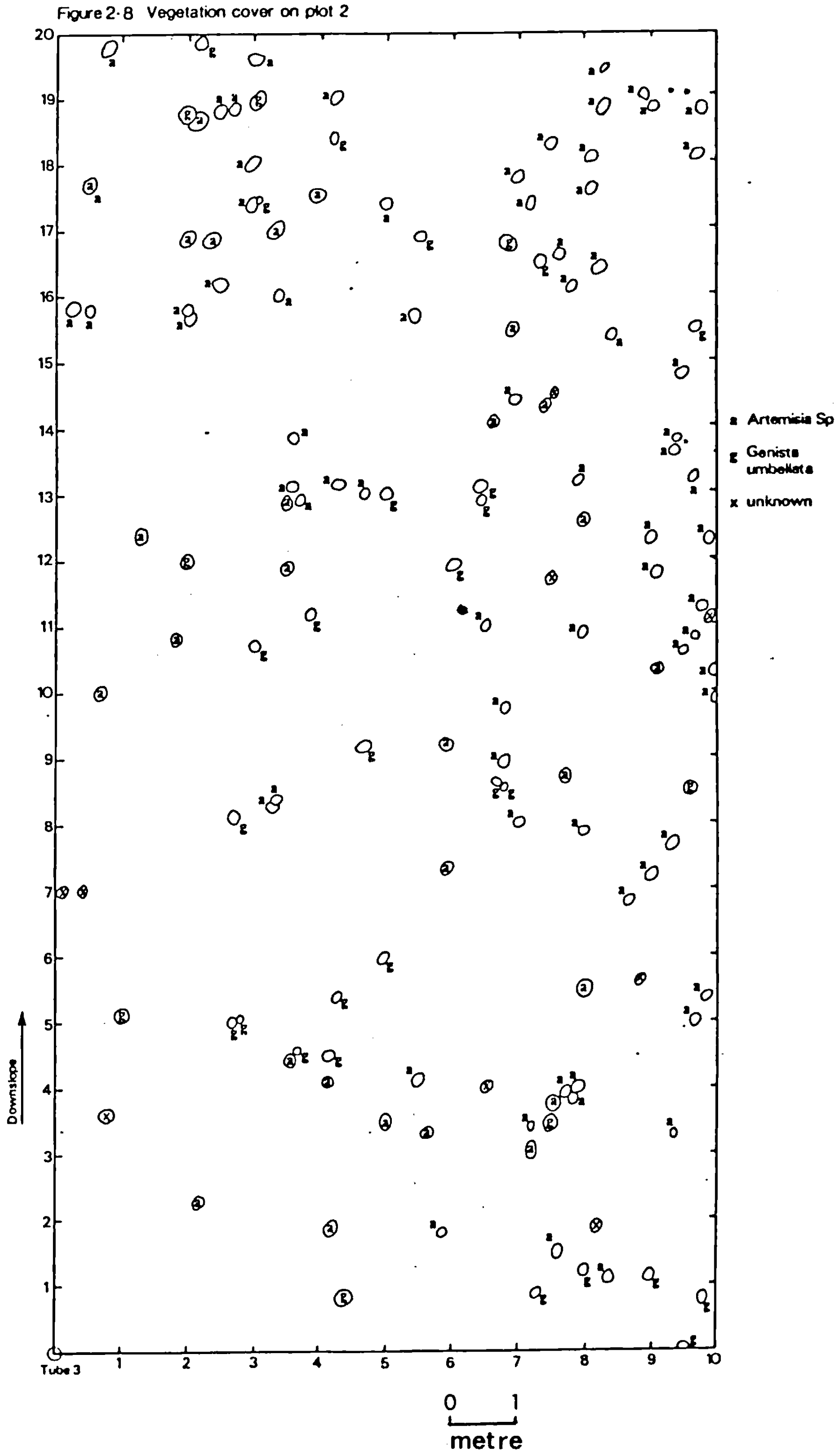


Figure 2·9 Vegetation cover on plot 3

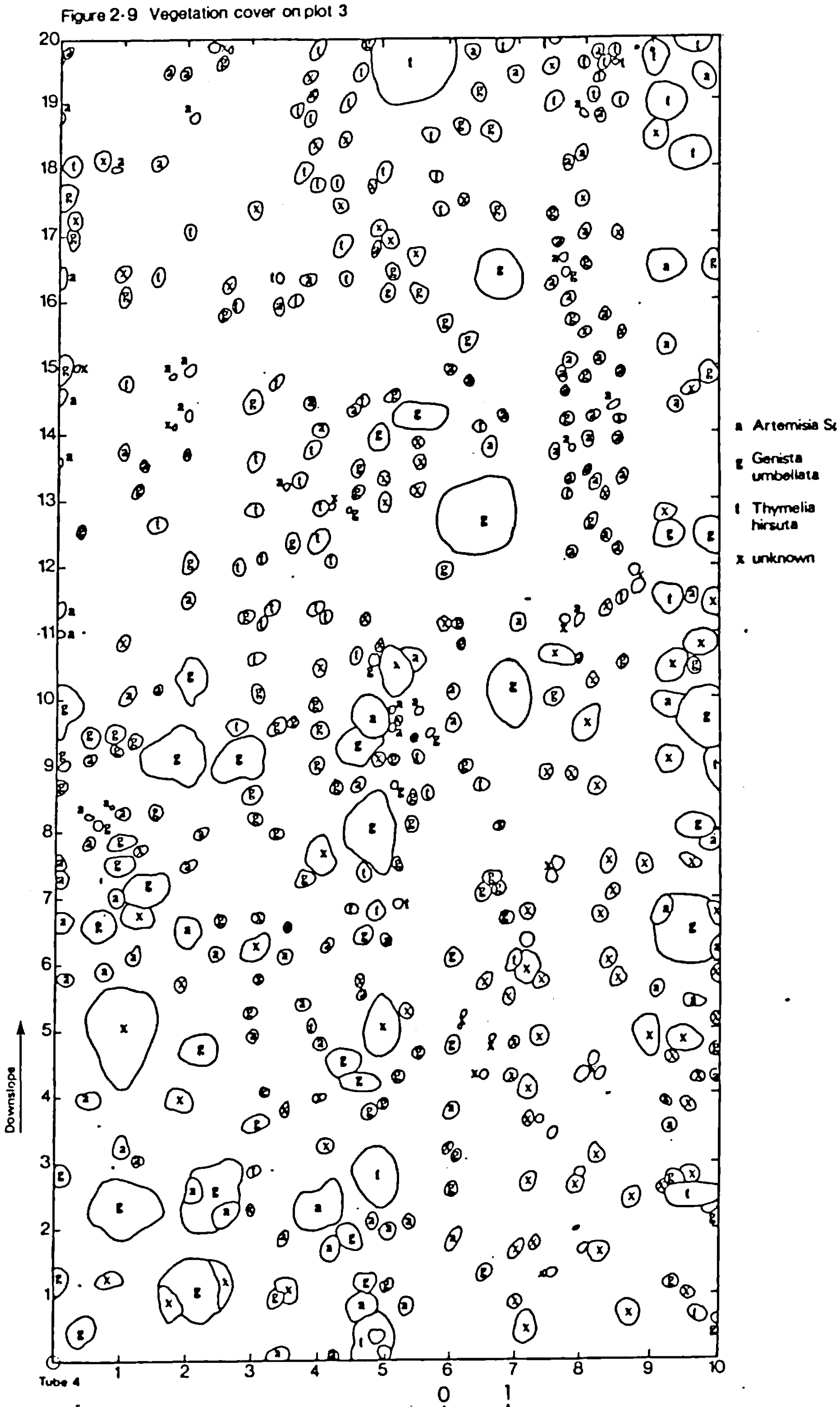


Figure 2·10 Histograms of bush size on plots 1, 2, and 3

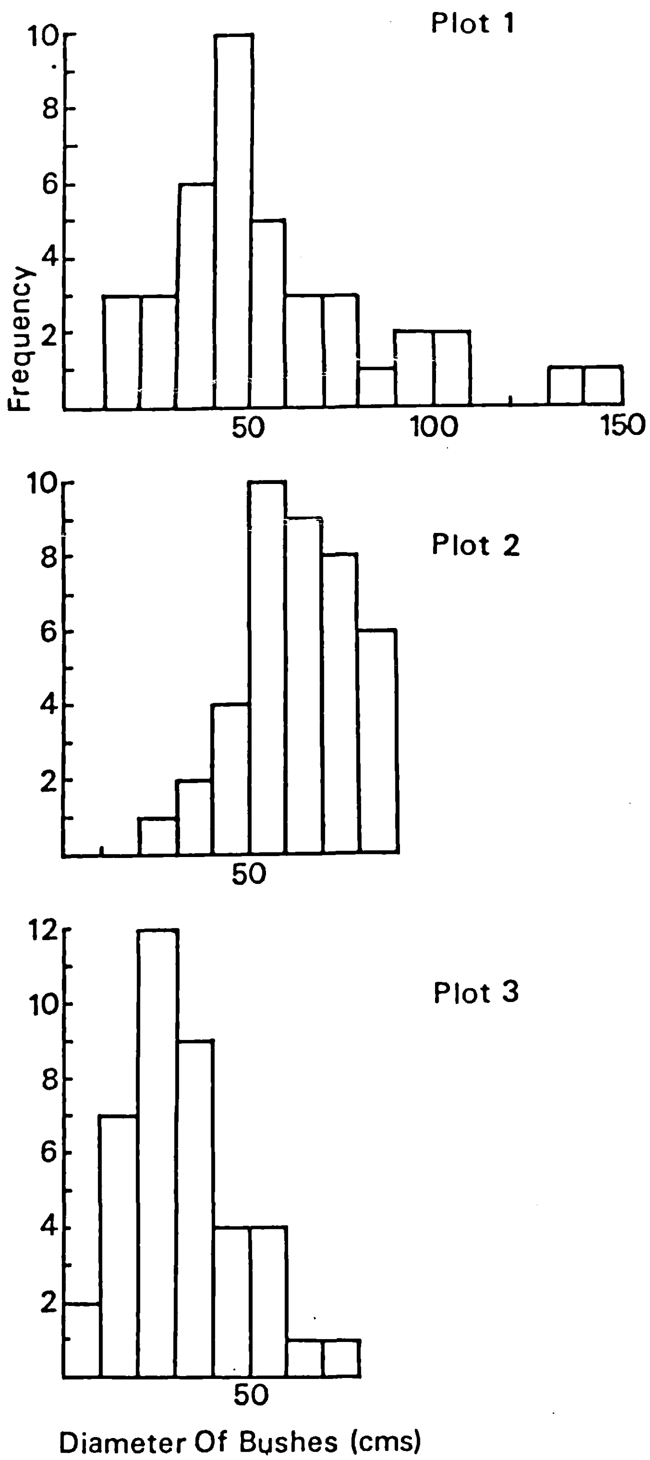


Figure 2·11 Histograms of the distance between bushes on plots 1, 2, and 3

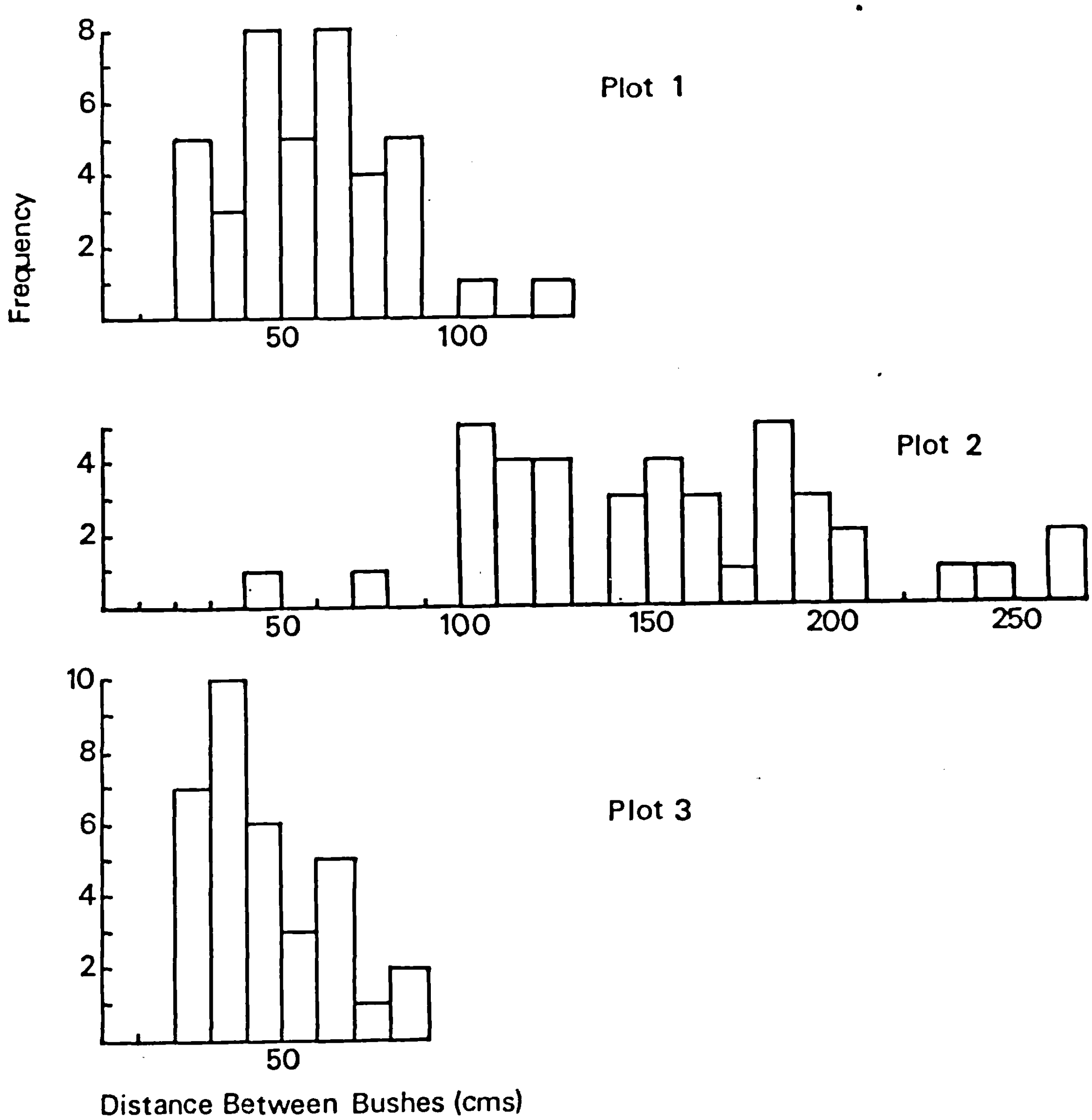


Table 2.9 Summary of average values of organic and soil moisture samples for vegetation plots 1, 2, and 3.

Plot	Vegetation Type	Organics > 2mm (grms)	Organics < 2mm (%)	Total Organics (%)	Soil Moisture (grms)
1	G umbellata	1.076	5.429	6.217	10.78
2	Artemisia	0.269	4.969	5.108	11.63
3	Mixture	0.919	3.929	4.833	20.11

Table 2.10 Results of between-plot variation

Variable		Plots			
		1 V 2	2 V 3	3 V 1	1&3 V 2
Org > 2	S.L.	-	-	NS	0.001
Org < 2	S.L.	0.01	0.01	-	-
Tot Org	S.L.	0.001	NS	0.01	-
Soil Moisture	S.L.	0.05	0.001	-	-

Table 2.11 Results of within-plot variation

Variable		Plots 1	2	3	Units
Org > 2	Bush	1.00	0.29	1.28	(grms)
	Open	0.86	0.27	0.71	(grms)
	S.L.	NS	NS	0.20	
Org < 2	Bush	5.66	5.38	5.46	(%)
	Open	4.98	4.85	3.00	(%)
	S.L.	0.02	0.10	0.01	
Tot Org	Bush	6.38	5.56	6.76	(%)
	Open	5.56	4.99	3.69	(%)
	S.L.	0.05	0.10	0.01	
Soil Moisture	Bush	11.79	12.41	21.30	(%)
	Open	9.79	11.38	18.63	(%)
	S.L.	0.01	0.05	0.01	

S.L. = significance level

Within-plot variations compared those surface samples taken under bushes and in the open. For the macro-organics significant variations between bush and non-bush locations only occur on plot 3 and at the 0.20 significance level (table 2.11). Micro-organic and total organic contents are all significantly greater under bushes by at least the 0.10 level. Regression analysis of the distribution of organics with distance from the bush was undertaken to see whether there was a decay function with distance from the bush. Table 2.12 shows that there is no significant relationship between any of the organic factors and distance from a bush. Multiple regression analysis incorporating bush diameter only improved the coefficient of determination values (r^2) slightly. Soil moisture values are also significantly higher under bushes than in open ground by the 0.05 level. Table 2.11 shows the mean and standard deviation for both 'under' and 'open' sites.

Generally the highest values of macro-organic content are for areas with greater vegetation cover, and the content of total organics seems related to the interaction between plant size and density. Within a plot there is a build up of organic matter under bushes (ie it is source-sensitive), but there is not a decay in organic matter content with distance from the bush. This suggests that there are two distinct features for shrub vegetation firstly the background level of organic matter, and secondly the level under bushes rather than a graded variation.

The poor relationship between macro-organics and location under bushes may reflect the effects of redistribution by surface flow, wind action, and termites. This experiment was undertaken three days after heavy rain, so the variation in macro-organics reflects time since the last redistributing event as well as the location of the source area. Orndoff and Lang (1981) found that there is considerable redistribution of leaf litter in a hardwood forest as a result of wind following almost uniform deposition in the autumn. After redistribution leaf litter is greater on moderate slopes than steep sloes. They also noticed that downslope movement of leaf litter is episodic, being related to strong winds. For example they found that 70% of marked leaves moved during 1 day in a seven day period in December 1977. Similar

Table 2.12 Summary of regression analysis. Figures in brackets are r^2 values for the multiple regression analysis of distance from bush and bush diameter against organic content.

r^2	Site 1	Site 2	Site 3
DIST V OGT2	0.041 (0.058)	0.029 (0.046)	0.033 (0.034)
DIST V OLT2	0.008 (0.236)	0.037 (0.096)	0.000 (0.151)
DIST V TOTORG	0.001 (0.201)	0.028 (0.091)	
DIST V IGLOSS			0.034 (0.171)

DIST = distance from bush (cms)
OGT2 = organic content greater than 2mm (grms)
OLT2 = organic content less than 2mm (%)
TOTORG = total organic content of sample (%)
IGLOSS = total organic organic content determined by loss on ignition (%)

quantitative work on litter redistribution by wind, water or fauna have not been found for semi-arid areas.

The greater soil moisture for plot 3 probably reflects drainage characteristics of marls and conglomerate soils, with the latter draining more rapidly. However the variation in soil moisture between plots 1 and 2 may reflect greater interception losses or lower evaporation rates on plot 1 due to the more dense vegetation cover. While there are many studies of interception losses in forest ecosystems (Jackson and Aldridge 1973, Fedorov and Ye Rogestskaya 1971, and Ford and Deans 1978), there are few studies in shrubland. Zinke (1967) lists some work on chaparral in the United States. Interception losses are presented in the form of a regression line

$$Lri = a \times Pr + b$$

where Lri is loss by interception of rainfall, Pr is the gross precipitation from rain and a and b are constants. Zinke quotes values for 'a' ranging from 0.0162 to 0.4000, and 'b' from 0.01 to 0.083 for various authors. For a storm of 23mm in Ugijar (see Chapter 3) this would lead to interception losses between 0.38 to 9.28 mm.

In other studies Villiers (1976 - English summary) notes that an appreciable quantity of rainfall was lost through evaporation from vegetation in mixed bushveld. Aldridge (1968) studying gorse (Ulex europaeus) found a linear relationship between throughfall and gross rainfall, but for storms less than 0.18mm there was no throughfall. Noble and Morgan (1983) measured throughfall and erosion under a solitary Brussel sprout plant in a laboratory experiment and found that plant canopies of 10 - 25% result in reductions of 10 - 25% in rainfall volume and 10 - 81% in rainfall energy, although no reduction was noted in the rate of splash detachment. Jan de Ploey (1982) has tried establishing stem flow equations for grasses.

Within each plot the greater soil moisture under bushes suggests that despite relatively higher interception and transpiration losses under plants, the amount of soil moisture is effectively controlled by direct evaporation losses. Shrubs offer enough shade to reduce such losses. Vegetation cover is analysed at a smaller scale for the erosion plots above the Gerlach troughs, and the discussion of the findings is given in Chapter 5 in relation to sediment yield.

The vegetation cover does affect the hydrological budget and erosion by protecting the soil. Despite some research, little is known about the interaction between vegetation, the hydrological regime and soil erosion. Although there is considerable scope for research in this area, no further work on vegetation was undertaken in this study, the remainder of the thesis concentrating on surface and throughflow rates.

Evaluation of Study Site With Respect To Ideal Conditions

It is evident that not all the ideal site conditions discussed earlier could be met. The main disadvantages are the presence of rills, vegetation cover, and the possibility of crusted and swelling clays in the marl.

The set of rills at the southern end of the slope runs through part of the instrumented area. On closer inspection, only one rill actually enters the study area, so that the influence of most of the rills was not significant because of their location. It would not have been possible to monitor the one rill within the logistic constraints of the project. The omission of monitoring the rills should, at the worst, lead to the underestimation of soil loss and values of concentrated surface discharge. However the location of access tubes and Gerlach troughs above the gully head may provide sufficient evidence to say whether the rills locally influence soil moisture values.

The second group of rills in the concave depression does not interfere with the monitored areas. However rills are, on this slope, related to man-made features ie the road and the path. This suggests that under normal conditions (as represented by areas above the path and away from the road) the slopes are stable with respect to rilling.

Pipes were not found on or in the hillslope. Although the lack of sightings does not mean there are no pipes, the combined evidence of factors associated with pipes may help the observation. Factors affecting piping are discussed by Jones (1971) amongst others where pipes are associated with:

1 Susceptibility to cracking in dry periods, high silt/clay content, and high percentage of montmorillonite
2 Periodic high intensity rainfall and devegetation
3 Biotic break up of soil and a relatively impermeable basal horizon
4 An erodible layer above this base, high exchangeable sodium and high base exchange capacity or high soluble salts
5 Steep hydraulic gradient

The chemical composition of the soils was not examined although sodic marls occur elsewhere. On the marl the main change in bulk density is within the top 20cms, and on the conglomerate between horizons 2 and 3. The marl does crack on drying (although it is not known how deep the cracks are) but the conglomerate soil does not. The evidence suggests that there are no pipes on this site.

Most of the possible sites around Ugijar have some sort of vegetation cover on them, even the dissected slopes, unless they have been ploughed. It is impossible to get unvegetated sites, so the importance of vegetation is accepted and then ignored! The discussion of crusted and swelling soils is deferred to chapter 4.

The hillslope is however susceptible to erosion with small gullies developing on the footslopes. Over much of the hillslope reticular flow is probably the dominant erosive force aided by steep slopes, long slope lengths, a scattered bush vegetation, and erodible soils. The slope form is simple for site 2, but slightly more complicated for site 1. The marl appears relatively homogeneous, but the conglomerate deposit has a more variable profile. The slope does not seem to have been cultivated recently, although judging by the sparsity of vegetation and furrow features on the soil it has probably been ploughed in the past. Overall the study site appears quite suitable for research on the hydrological parameters of gully catchments affecting gully growth.

CHAPTER THREE PRECIPITATION AND THE SOIL MOISTURE BUDGET

Precipitation affects the erosional system not only through its direct impact on the slope, but also indirectly through its influence on the soil moisture budget and plant growth. Erosion and transport by rainsplash and overland flow are affected by the seasonality of rainfall, rainfall intensity and duration, and the recurrence intervals of extreme as well as small storms. However the relationship between precipitation and erosion is blurred by the affect of antecendent conditions such that similar sized storms may have quite different impacts if the antecedent conditions are dissimilar. This introduces the importance of temporal data for rainfall. Precipitation is the main input to the soil moisture store, but this is countered by the evapotranspiration and seepage losses. Evaporation is particularly important in hot climates, so that an estimate of the soil moisture budget involves both precipitation and evapotranspiration factors. Thus this chapter is divided into three sections, the first considers the precipitation characteristics for the study area, the second describes the precipitation pattern for the study year, and the third tries to estimate the soil moisture deficit from precipitation and evapotranspiration data.

3.1 Precipitation Characteristics

The long term precipitation record is examined with two ideas in mind. Firstly it is necessary to establish whether there are any long term changes which may affect erosion on hillslopes, and secondly the long term data is required to describe some typical characteristics which are most important for erosion such as seasonality, intensity and recurrence intervals. Precipitation records of daily 24 hour intensity rainfall are available between 1942 and 1983 for Ugijar, which is the nearest station to the site (700m) both in terms of distance (4km) and altitude (559m). Records are also available for Mecina Bombaron (1200m) but this data set is only used to supplement the Ugijar data in the second section, as Mecina Bombaron is affected by orographic rainfall.

Total annual precipitation values for Ugijar from 1942 to 1983 are drawn in figure 3.1. The mean annual rainfall for the record is 374.3mm, but the standard deviation is 115.5 so that the annual rainfall may vary by about 30% on either side of the mean in any year. Such variability is common in south east Spain and increases eastward as the climate becomes drier. In Carboneras the annual variability reaches 40% (Geiger 1973). Despite this variability the annual rainfall appears to be normally distributed (figure 3.2).

There does not appear to be any trend in the annual rainfall over the last 40 years, although values seem to have fallen over the last ten years. This is confirmed by regression analysis of annual rainfall (y) against the time from the beginning of the record (x) which gives the equation:

$$y = 398.43 - 1.12x.$$

with a coefficient of correlation (r) equal to 0.11. A second type of pattern was examined by looking at the residuals which were calculated by subtracting the mean annual precipitation from each year's value. Figure 3.3 shows that there are irregular runs of wet and dry years with ten years of above average rainfall and eleven years of below average rainfall. For both above average and below average cases, the modal class of a length of run is one year, but the maximum nubmber of runs (so far) is three years during a wet sequence and five years during a dry sequence. Periods with three consecutive years with above average rainfall occurred in 1945-1948 and 1968-1971, whereas periods with three or more years of below average rainfall were 1952-1955, 1963-1968, and 1978-1983. The persistance of runs was tested with an autocorrelation function where the autocovariance at lag_l is:

$$C_l = \frac{1}{n-l}\sum_{k=1}^{n-1}(x_k - \bar{x})(x_{k+l} - \bar{x})$$

and

$$r_l = C_l / C_0$$

Table 3.1 shows that the autocorrelation at lag_0 is 1, but then falls quickly to 0.25 at lag_1, and -0.002 at lag_2. Thus statistically there is only a very small persistance between years in the annual rainfall for the long term record.

Figure 3·1 Annual precipitation at Ugijar 1942 to 1983

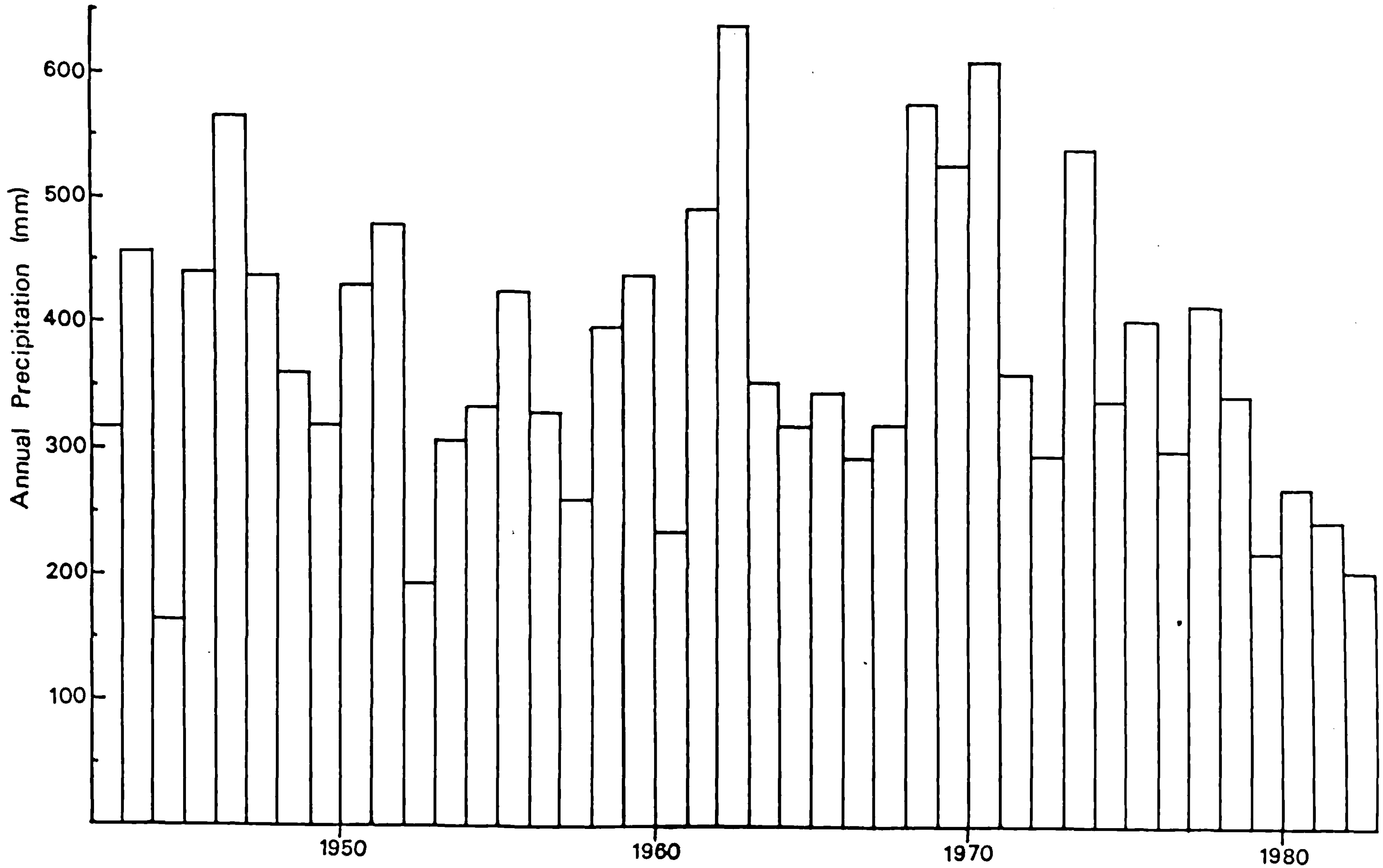


Figure 3·2 Frequency distribution of annual precipitation

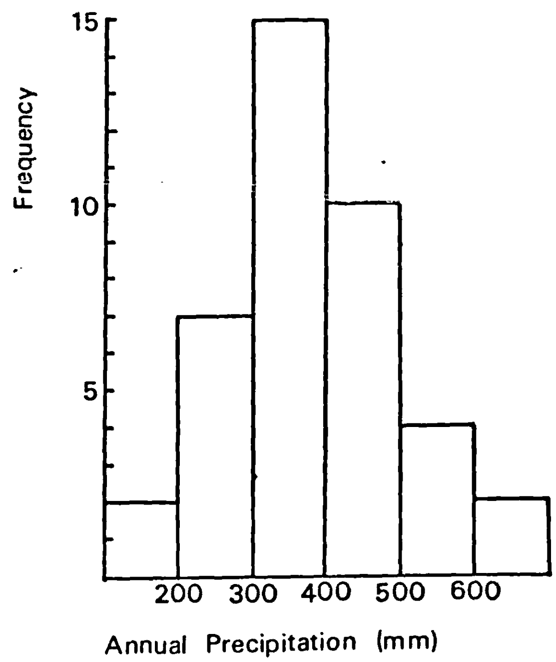


Figure 3·3 Residuals of annual precipitation 1942 to 1983

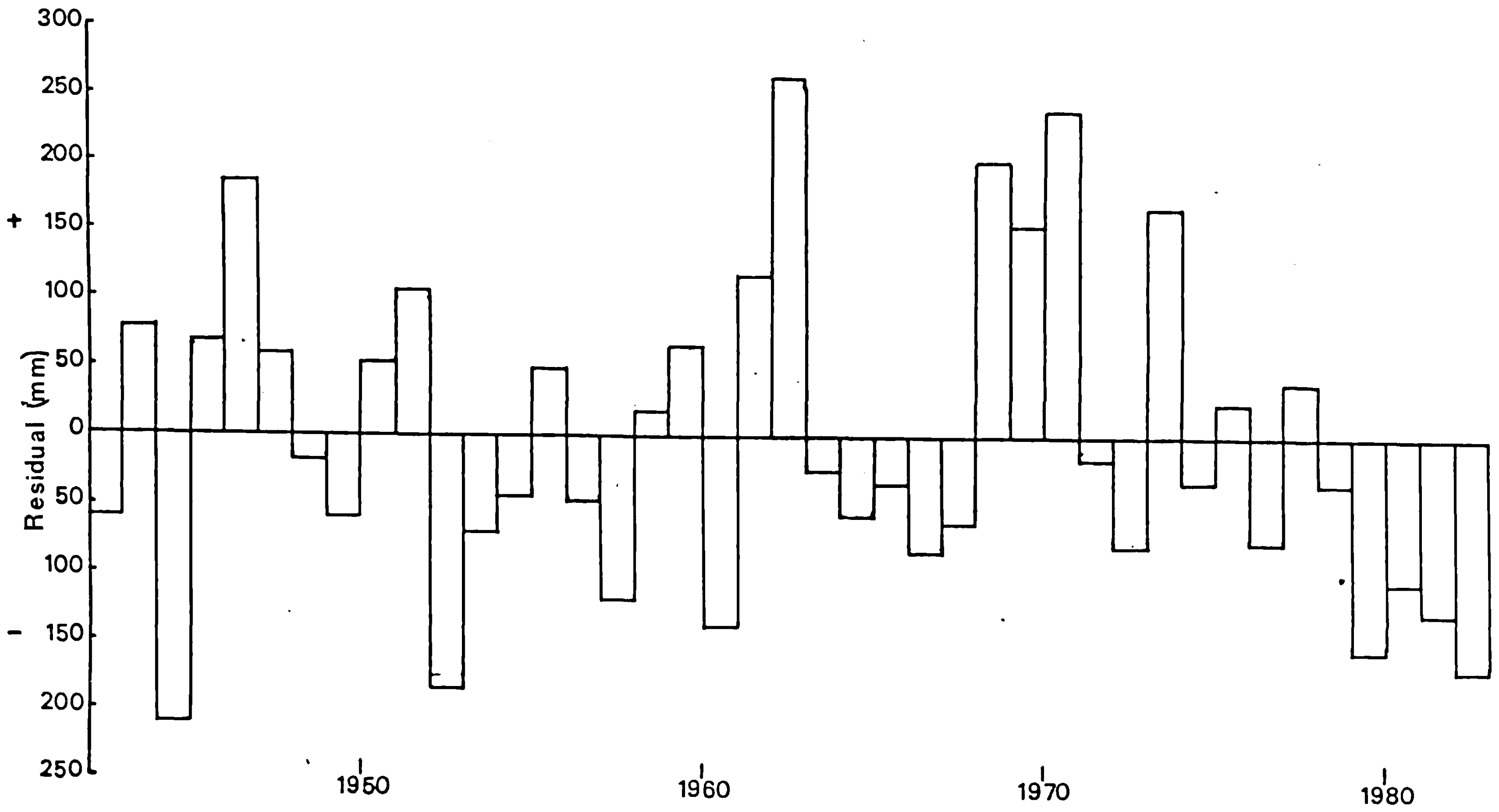


Table 3.1 Summary of autocorrelation analysis on annual data

$$C_l = \frac{1}{n-1} \sum_{k=1}^{n-1} (x_k - \bar{x})(x_{k+1} - x)$$

$$r_l = C_l / C_o$$

Number of observations	40
Mean	377.81
Standard deviation	116.24

Lag	r
Lag 0	1.0000
Lag 1	0.2500
Lag 2	-0.0022
Lag 3	-0.0408
Lag 4	-0.0913
Lag 5	0.0529
Lag 6	-0.1461

Further analysis of the records shows that despite no trends in the annual rainfall since 1942, there have been other types of change in the rainfall data. Figure 3.4 shows the number of rain days per year for Ugijar, 1942-1983. The negative trend seen in the diagram is confirmed by regression analysis which gives

$$y = 50.409 - 0.636x$$

and a coefficient of correlation (r) of -0.653, which is significant at the 0.01% level using the F test. The residuals (determined by subtracting the calculated value from the actual value of the number of rain days per year) again show an irregular sequence of above and below average number of rain days per year. Periods with above average number of rain days tend to be during years with above average annual rainfall and vice versa (figure 3.5). There are some exceptions, for example in 1973-74 there was an average number of rain days but an above average annual rainfall. This is due to an exceptionally large storm on 17-19 October (which accounted for 35.4% of the annual rainfall) plus a further four storms with rainfall over 20mm each. In 1943-44 the number of raindays was below average, but there were nine storms above 20mm increasing the annual precipitation value.

The average rainfall per rain day has increased between 1942 and 1983 (figure 3.6) reflecting the decrease in the number of rainy days. This is shown by the regression

$$y = 7.233 + 0.166x$$

where y is the rainfall per rain day and x is the time from the beginning of the record. The coefficient of correlation is 0.63 which is significant by the 0.01% level. This is substantiated by the 24 hour intensity data. For each year the daily rainfall was classified into three intensity classes (0.01 to 1.99mm, 2.00 to 9.99mm, and 10.00mm and above). The results shown in figure 3.7 indicate that there has also been a change in the relationship between relatively light and heavy intensity rainfall over the last 40 years with an increase in the proportion of relatively high intensity storms (ie greater than 10mm/24 hours) and a decrease in the lighter intensity storms.

All these changes have implications for soil erosion. There appear to be fewer days with rain, but also an increase in the amount of rainfall per rain day. It is well known that the

Figure 3·4 Number of rain days per year 1942 to 1983

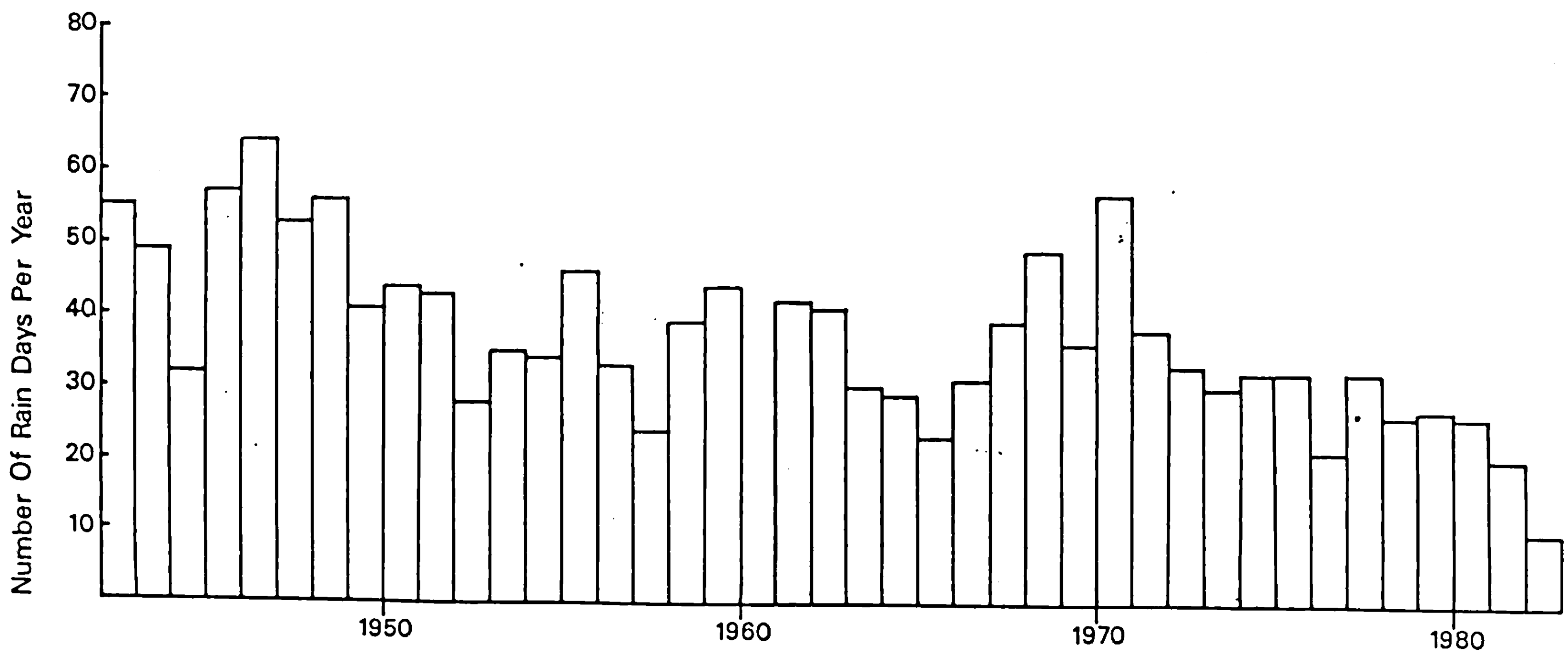
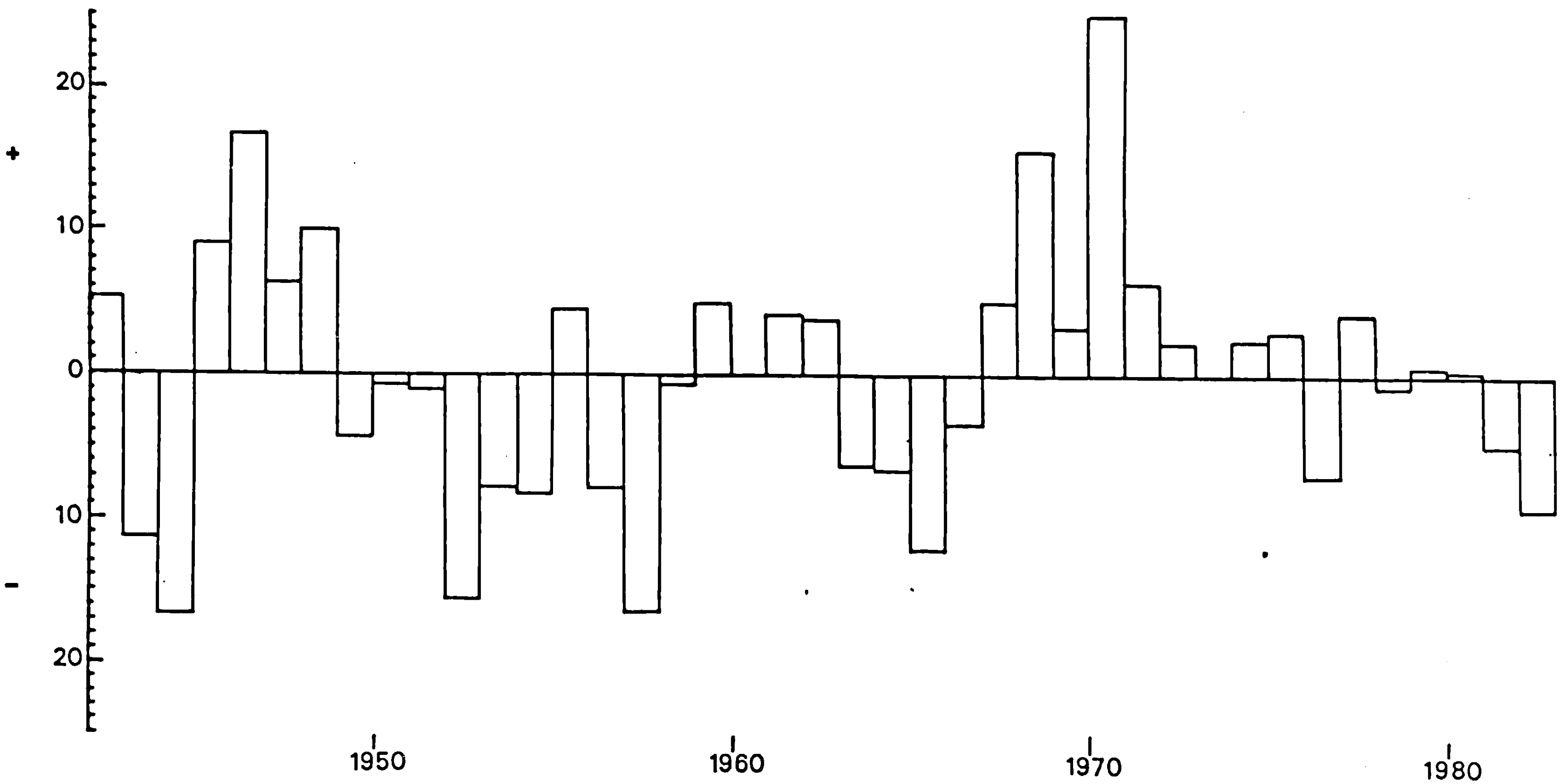


Figure 3·5 Residuals of the number of rain days per year 1942 to 1983



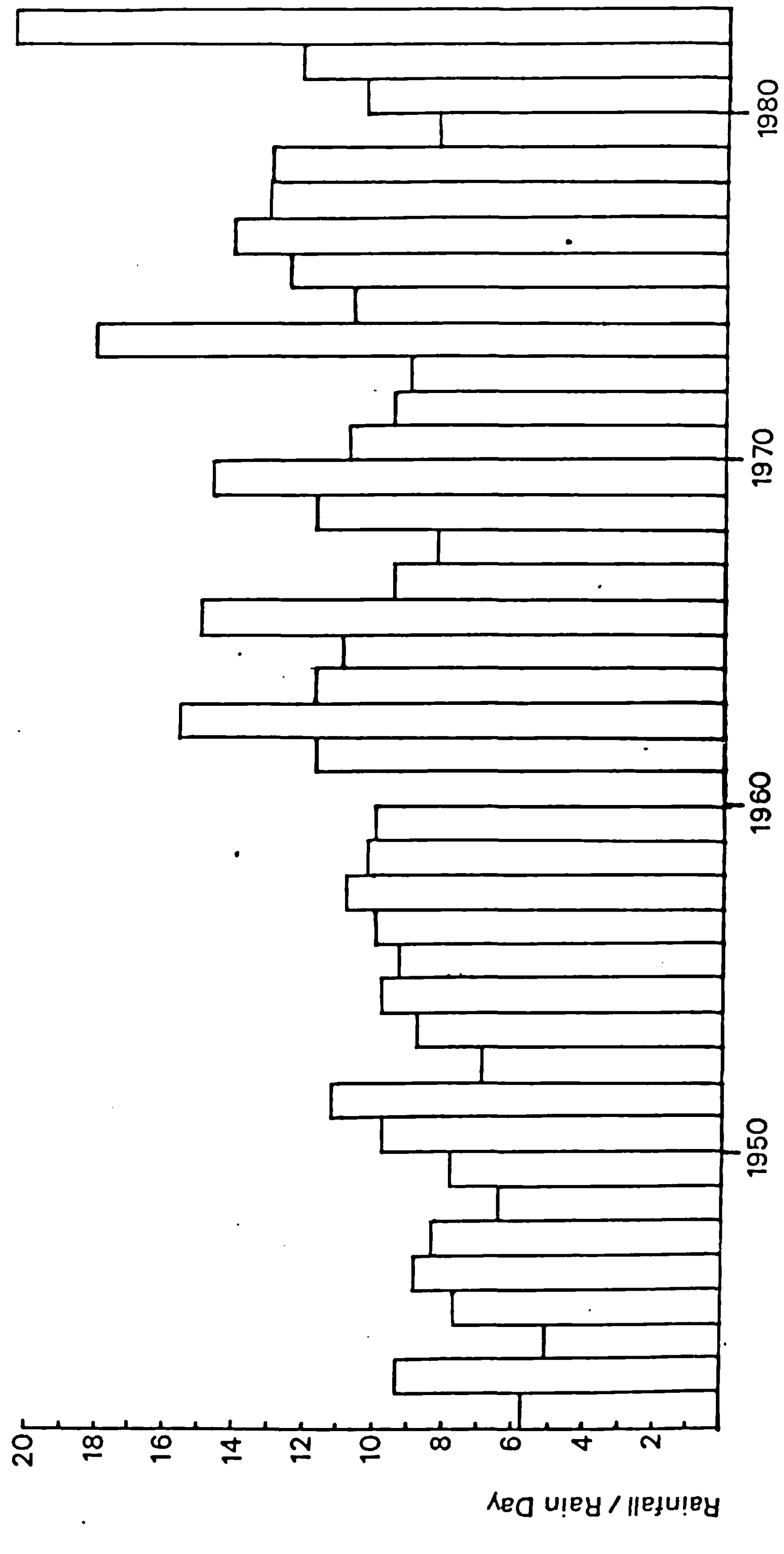
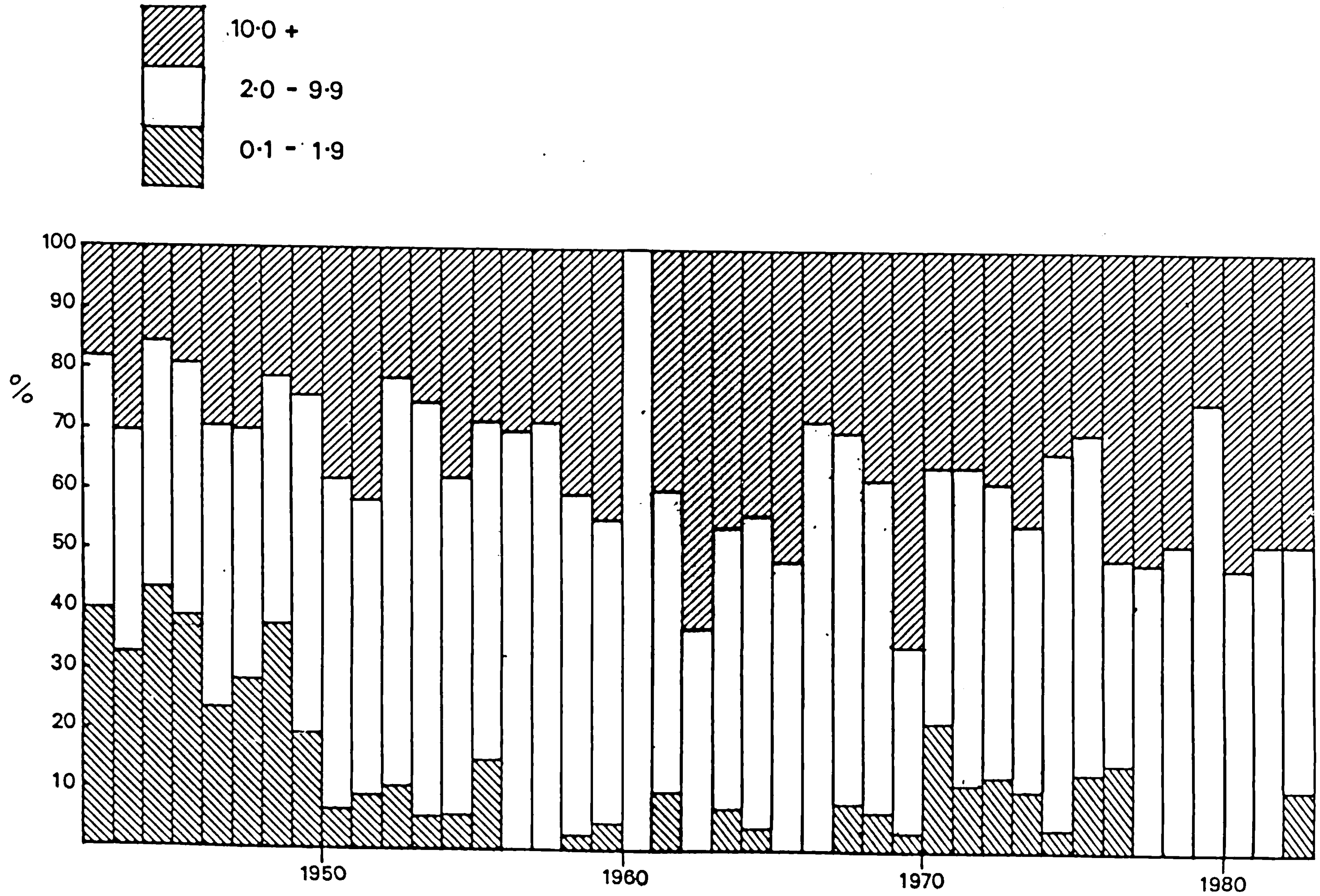


Figure 3·6 Rainfall per rain day 1942 to 1983

Figure 3·7 Light, medium, and heavy 24 hour rainfall intensity 1942 to 1983 (mm/24hrs)



heavier intensity storms cause more erosion whilst light intensity storms are important for replenishing the soil moisture store. Thus over the last forty years there may have been an increase in the erosivity of the rainfall.

The most recent run of dry years in south east Spain is particularly interesting as it means that the study period took place during an uncommonly dry period. Martinez et al. (1982) studied the synoptic data on a temporal and spatial basis in order to describe the development of the drought and its severity. The whole of Spain was classified into three groups:

1 Areas of long standing drought eg Teruel, Valencia and Almeria.
2 Areas of intermediate drought eg Madrid, Gerona.
3 Areas of short-term drought eg Badajoz, Santiago de Compostela.

By 1982 the whole of Spain was suffering from the effects of drought, but the longest affected region was the south. Martinez et al. suggest that the drought conditions developed in the south during the spring and summer of 1977 by a smooth transition from humid to dry conditions. Occasional heavy storms were not sufficient to reverse the overall trend, and the drought intensified over the next few years. They suggest that the drought was partly a result of a decrease in the frequency of rain-bearing formations, especially fronts. Between 1979-81 there was an overall decline in frontal situations of 6.3% in Mediterranean Spain.

A series of graphs from the Instituto Nacional de Meteorologia chart the intensification of the drought over Spain. Figure 3.8 shows the percentage of actual precipitation for the three years 1979-1982 of the "normal" precipitation based on data records from 1931-60. The first map for 1979-80 shows that rainfall in the Sierra Nevada was 75-100% of the normal. The figure falls to 50-65% of normal in 1980-81, and 60-75% for 1981-82. Figure 3.9 compares the difference between actual accumulated potential evapotranspiration rates for the norm (ETPNA) and those calculated for each year (ETPA). These show an increase in potential evapotranspiration rates up to 40-60mm by 1982 in the Ugijar area.

Figure 3·8 Percentage of precipitation with respect to the norm (1931 to 1960) in south east Spain

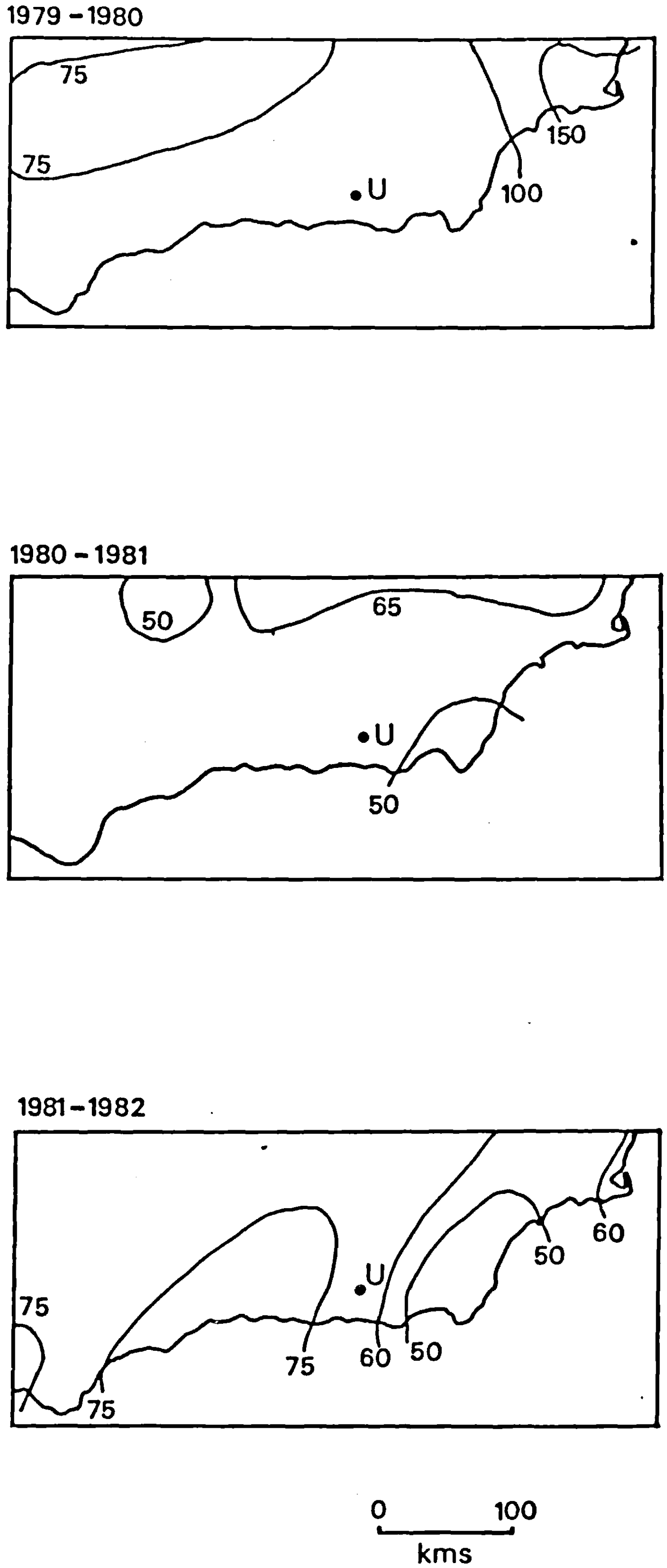
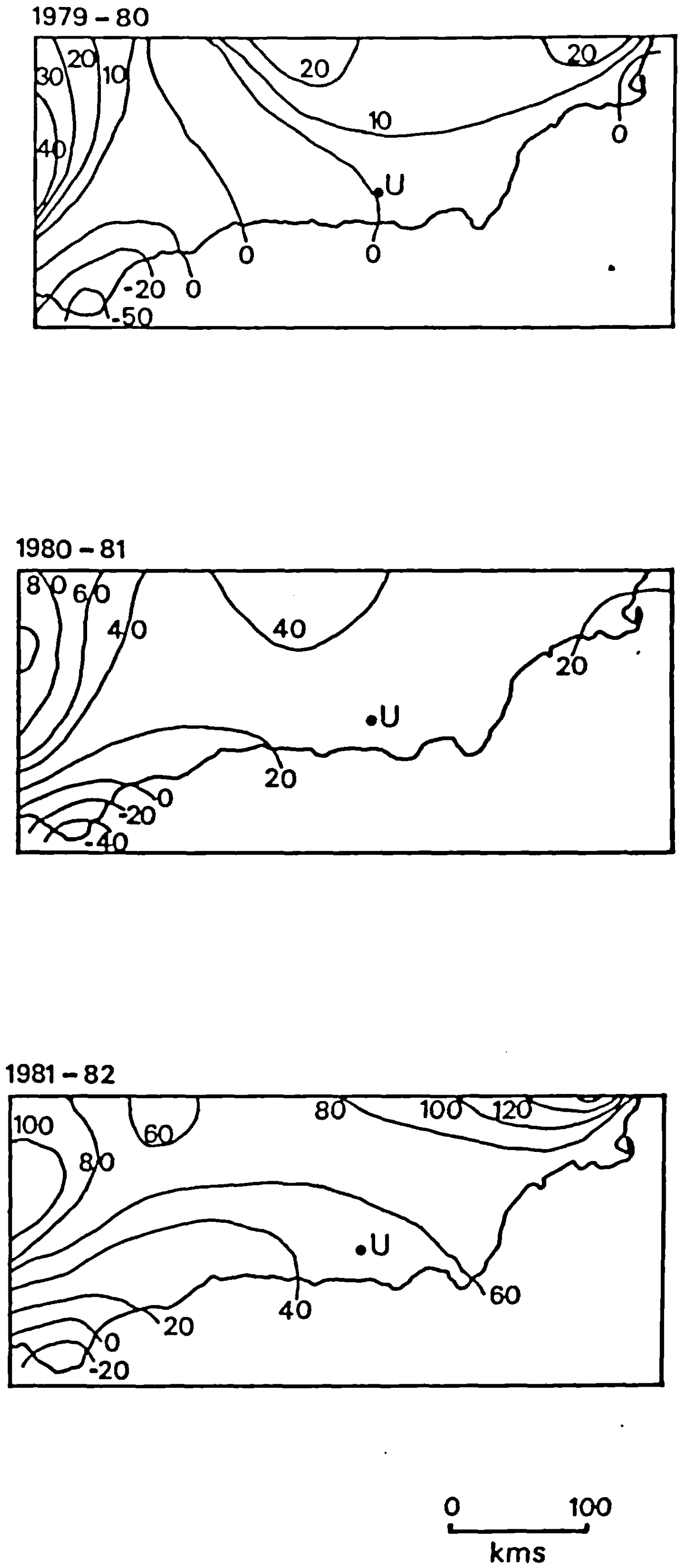


Figure 3·9 Difference between the actual accumulated potential evapotranspiration and the average for 1931 to 1960



The drought in south east Spain is very important for its duration and intensity. The sequence of dry years for Ugijar between 1978-83 is the second five year period with below average rainfall to have occurred since 1942, but it is more severe than the first with the annual rainfall sustained at c100-170mm below the average (figure 3.3). In 1982-83 there were only ten rain days, which was the main study period. A drought of this severity is unusual and will be reflected by the values measured for soil moisture content, herbaceous plant cover and soil loss. In the data analysis the effect of the drought on the magnitude of erosion processes is discussed more fully.

During the 'average' year there is a marked seasonality of rainfall. The rainfall maximum occurs in winter with two peaks (figure 3.10), the largest peak in December and a slight rise in April. Between October and April mean monthly rainfall values are above 40mm per month and decrease rapidly during late spring and summer. In 36 out of 40 years record, no measurable quantity of rain fell in July (table 3.2). The standard deviation of monthly rainfall is generally similar to the mean, indicating the great intermonthly variability of rainfall (table 3.2). Monthly rainfall patterns for any two years may not only be unlike each other, but also unlike the average pattern outside the broad generalisations of wet winter, dry summer.

On average 85.6% of the annual precipitation falls between October and April although this varies, and as table 3.3 shows one effect of the most recent drought is the fall in the proportion of summer precipitation. Furthermore the largest storm may occur in any month (although July is very unlikely) but is most likely to occur in October, November or December according to the frequency diagram in figure 3.11 Monthly rainfall may be dominated by one or two large storms, so that the rainfall is not spread evenly throughout the month, which may be important for plant growth and modelling soil moisture.

The seasonality of rainfall is a result of two dominating types of weather conditions. The summer weather is largely influenced by anticyclonic conditions over the Iberian peninsula caused by high temperatures which tend to deflect onshore winds. As a result

Figure 3·10 Mean monthly rainfall at Ugijar 1942 to 1983

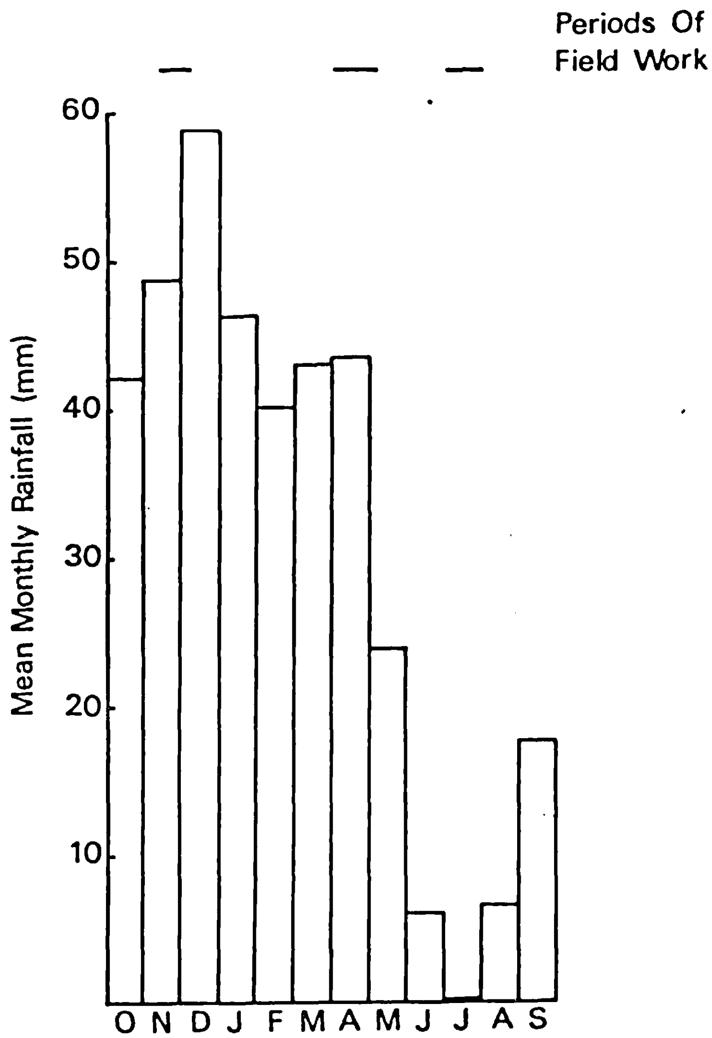


Table 3.2 Monthly rainfall at Ugijar 1942 to 1983

Year	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Total
1942-43	73.2	65.6	18.7	5.0	34.8	56.1	12.8	13.9	6.3	4.4	0.0	26.0	316.8
1943-44	14.1	8.0	171.9	0.0	77.9	4.2	0.0	67.3	0.2	0.0	0.5	112.0	456.1
1944-45	34.6	19.6	34.0	57.6	0.0	9.2	5.1	0.0	4.3	0.0	0.0	0.0	164.4
1945-46	11.4	59.4	109.9	30.9	27.6	71.5	109.1	19.5	0.5	0.0	0.0	0.0	439.8
1946-47	20.9	124.6	44.8	112.6	102.9	87.5	1.7	26.3	0.0	3.2	33.0	7.9	565.4
1947-48	22.8	20.0	75.4	103.0	37.3	8.6	106.2	46.6	10.3	0.0	0.0	7.2	437.4
1948-49	19.7	0.0	24.7	51.2	38.4	22.1	108.5	21.9	0.0	0.0	28.7	44.7	359.9
1949-50	0.0	64.6	41.7	56.4	0.0	24.2	53.2	29.3	0.0	0.0	4.5	44.3	318.2
1950-51	87.1	0.0	10.9	75.6	66.8	66.9	34.1	32.2	14.0	0.0	4.2	36.5	430.3
1951-52	38.8	137.6	26.3	39.4	15.2	30.1	112.0	25.0	0.0	0.0	58.5	0.0	482.9
1952-53	28.4	28.2	23.0	9.1	9.0	34.7	33.0	1.5	14.5	0.0	0.0	13.0	194.4
1953-54	87.7	12.4	86.5	8.5	26.0	52.1	25.0	0.0	0.0	0.0	0.0	9.5	307.7
1954-55	1.5	40.7	49.2	64.6	93.2	32.5	3.5	14.5	5.5	0.0	24.5	3.5	333.2
1955-56	110.9	82.3	25.2	27.7	19.7	58.7	63.5	0.0	0.0	0.0	0.0	38.0	426.0
1956-57	27.0	26.0	14.0	51.2	15.0	74.0	76.5	42.7	3.0	0.0	0.0	0.0	329.4
1957-58	48.0	40.5	85.0	26.0	10.5	38.0	2.5	7.5	0.0	0.0	0.0	0.0	258.0
1958-59	20.5	27.7	131.0	32.5	12.0	47.0	0.0	80.2	0.0	3.7	7.2	34.3	396.1
1959-60	38.5	10.5	46.2	22.5	102.4	119.0	68.0	19.5	0.0	0.0	0.0	12.5	439.1
1960-61													235.3
1961-62	3.0	144.0	131.0	16.0	5.0	63.0	82.0	48.0	2.0	0.0	0.0	0.0	494.0
1962-63	102.0	81.0	136.0	137.0	43.0	0.0	0.0	86.0	0.0	0.0	0.0	57.0	642.0
1963-64	0.0	43.0	186.0	0.0	56.0	20.0	3.0	24.0	16.0	0.0	0.0	6.0	354.0
1964-65	0.0	46.0	74.0	21.0	81.0	58.0	18.0	0.0	0.0	0.0	0.0	21.0	319.0
1965-66	98.5	64.0	20.5	5.0	93.5	0.0	26.0	12.5	7.0	0.0	0.0	20.5	347.5
1966-67	66.5	30.5	0.0	20.5	55.0	32.1	28.0	12.5	19.0	0.0	10.5	20.0	294.6
1967-68	20.5	72.0	5.5	22.5	109.5	44.5	17.5	13.5	8.5	0.0	6.0	1.5	321.5
1968-69	0.0	57.0	33.5	79.5	161.0	42.1	49.9	98.5	21.1	0.0	27.0	9.0	578.6
1969-70	107.5	136.0	16.1	205.8	0.0	44.6	16.8	5.5	0.0	0.0	0.0	0.0	532.3

Table 3.2 (cont)

Year	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Total
1970-71	9.5	2.5	73.9	59.6	0.0	86.4	240.8	46.2	17.3	0.0	6.2	71.5	613.9
1971-72	18.7	46.5	92.4	69.5	11.0	54.2	7.5	16.0	9.5	0.0	0.0	37.0	362.3
1972-73	96.7	58.5	3.5	29.2	23.5	48.0	8.5	24.3	5.3	0.0	0.0	0.0	297.5
1973-74	213.0	28.0	115.2	13.5	22.4	68.0	62.2	0.0	23.8	0.0	0.0	0.0	546.1
1974-75	46.6	0.0	0.0	20.0	39.9	105.5	59.5	43.6	8.5	0.0	0.0	17.9	341.5
1975-76	0.0	24.5	95.5	53.0	20.2	8.3	140.5	0.0	0.0	0.0	18.0	45.0	405.0
1976-77	51.0	0.0	202.5	33.6	0.0	0.0	10.5	0.0	0.0	1.3	1.5	0.0	300.4
1977-78	69.5	80.5	11.8	29.0	69.5	107.0	13.5	40.0	0.0	0.0	0.0	0.0	420.8
1978-79	4.5	36.0	74.0	177.0	19.3	25.9	0.0	7.5	0.0	0.0	0.0	0.0	344.2
1979-80	59.2	0.0	11.0	30.5	25.4	0.0	28.9	24.3	4.5	0.0	23.5	13.5	220.8
1980-81	4.5	74.0	0.0	0.0	13.5	32.0	92.0	4.5	45.0	0.0	4.5	0.0	270.0
1981-82	4.5	0.0	70.7	71.4	21.0	42.5	32.5	3.0	0.0	0.0	0.0	0.0	245.6
1982-83	27.5	156.2	0.0	0.0	16.0	2.5	0.0	0.0	0.0	0.0	3.5	0.0	205.7
Mean	42.2	48.7	59.3	46.7	39.4	43.0	43.8	23.9	6.2	0.3	6.5	17.8	374.3
Standard Deviation	44.5	43.2	54.7	46.5	38.0	31.3	50.3	25.1	9.3	1.0	12.5	24.1	115.5

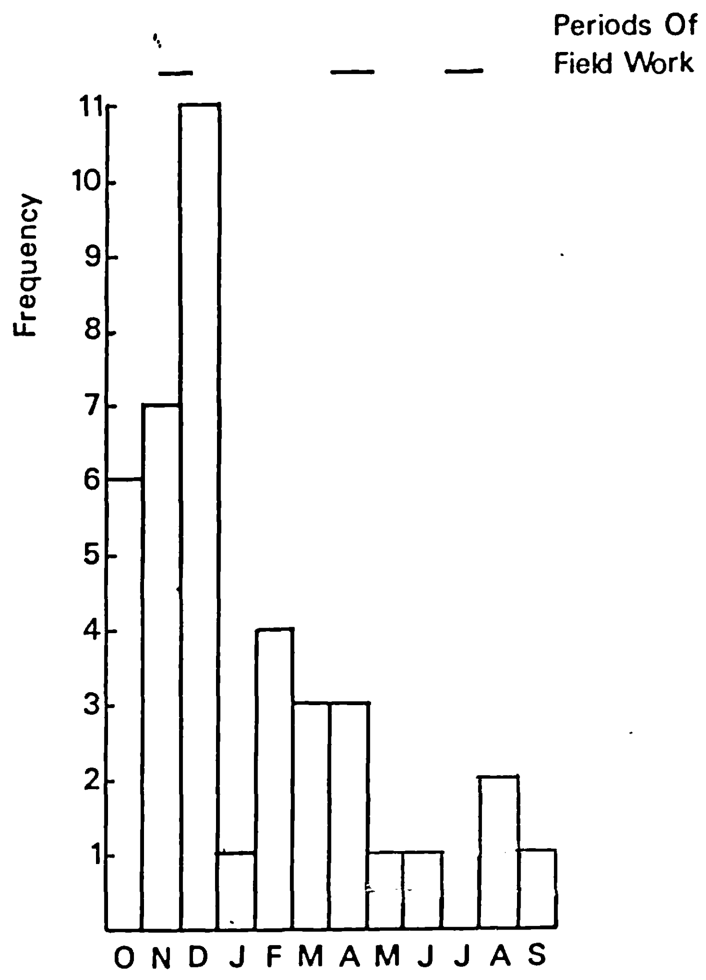
Table 3.3 Proportion of winter rainfall for the years 1979 to 1983 at Ugijar and Mecina Bombaron.

Town		1979/80	1980/81	1981/82	1982/83
Ugijar	Nov-April	71.5%	80.0%	98.8%	98.3%
M. B.	Nov-April	-	70.2%	94.2%	96.3%

Table 3.4 Frequency of duration of successive rainy days for a nine year record (Source: Thornes 1976)

Duration (days)	1	2	3	4	5	6	7	>7
Frequency	141	36	21	7	2	1	0	1

Figure 3·11 Frequency of monthly occurrence of the largest storm in each year 1942 to 1983



summer rainfall is normally convective. In winter cyclonic depressions from the Atlantic penetrate inland and together with orographic influences lead to precipitation. With variations in type of rainfall, one would expect variations in intensity, persistence of rainy days, and storm durations between the two main seasons.

For a simulation of rainfall Scoging (1976) not only distinguishes between summer and winter periods, but divides the winter section into three subgroups, October-November, December-February, and March-June with summer represented by July-September. Scoging found frequency distributions of duration are negatively skewed for all seasons, with 67.8% of storms occurring on single days (table 3.4). The dependency between yields for successive events was also tested by Scoging using autocorrelation analysis, and was found to be insignificant. Daily rainfall yields are therefore assumed to be independent of each other.

The persistence of wet or dry days was also tested by Scoging and simulated using Markov chain analysis. In winter there is a tendency for wet days to be followed by either a wet or dry day, but in summer dry to dry transitions, and to a lesser extent wet to dry transitions dominate, emphasising the infrequency of rainy days.

Rainfall intensities for Ugijar (based on 24 hour periods) for a given year have a negative, exponential distribution (they appear linear on semi log pager in figure 3.12). However care is needed with such descriptions if there is a long term trend in the data of a decrease in low magnitude events. By the early 1980's, figure 3.7 suggests 50% of the rainfall was less than 10mm/24 hours intensity. Table 3.5 shows that in the last ten years the greater proportion of high intensity storms (ie over 10mm) occur during the winter, although the highest intensity storm may occur during the summer eg in 1980-81 and 1979-80. Medium intensity storms occurring in the summer, although small in number, are very important for erosional processes as they occur when vegetation cover is low and the soils have been broken up by mechanical disturbance.

Figure 3·12 Frequency of 24 hour rainfall for 1942 to 1983

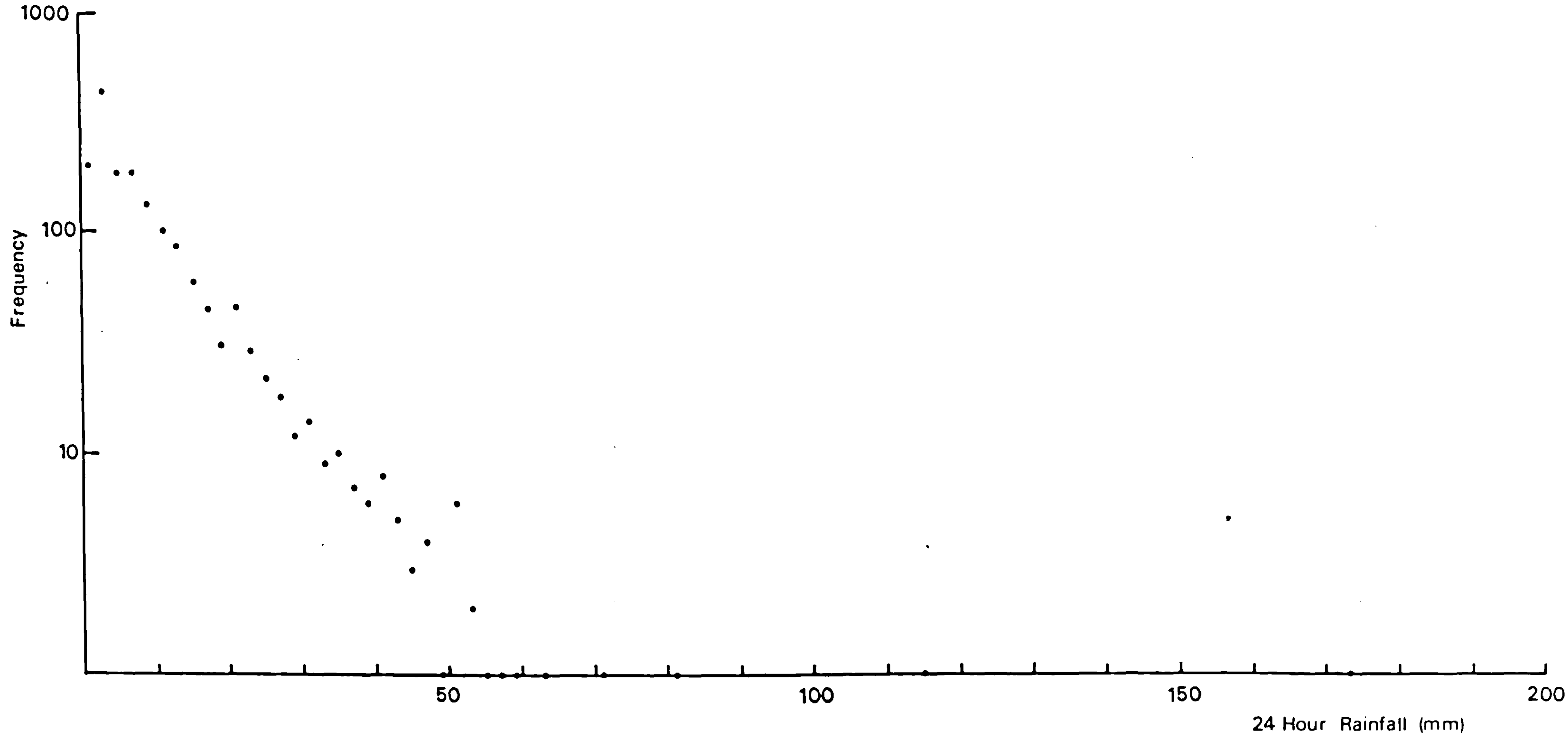


Table 3.5 Analysis of storms greater than 10 mm/24 hours at Ugijar 1972 to 1983

Year	October-April			May-September			Total
	%ppt	N	Lgest (mm)	%ppt	N	Lgest (mm)	Storms
1982/83	100.0	5	70.0	0.0	0	-	5
1981/82	100.0	10	32.7	0.0	0	-	10
1980/81	85.7	12	25.5	14.3	2	28.0	14
1979/80	57.1	4	19.5	42.9	3	23.5	7
1978/79	100.0	13	41.0	0.0	0	-	13
1977/78	87.5	14	42.0	12.5	2	25.0	16
1976/77	100.0	11	115.0	0.0	0	-	11
1975/76	80.0	8	53.5	20.0	2	38.0	10
1974/75	100.0	11	48.0	0.0	0	-	11
1973/74	92.9	13	172.0	7.1	1	15.3	14
1972/73	83.3	10	25.5	16.7	2	11.5	12

%ppt = percent of the annual precipitation falling in that season

N = number of storms greater than 10 mm in that season

Lgest = the rainfall total for the largest storm occuring in that season

Total = total number of storms greater than 10 mm occuring in year

All the storms recorded between 1940-83 are used to construct a curve of the percentage of storms for a given frequency shown in figure 3.13, assuming each rainy day represents a separate storm event (Shaw 1983). Approximately 50% of all storms exceeded 9.0mm/24 hours, and 10% of all storms were greater than 24mm/24 hours. Below 10% the size of storms increases rapidly.

The largest storm on record occurred on 17-19 October 1973, and was widespread over south east Spain causing considerable flooding, and flood related damage. A total of 198.0mm of rain fell at Ugijar, with 175.0mm on 18 October alone. At nearby towns rainfall totals were 240.1mm at Mecina Bombaron, 233.6mm at Cadiar, and 196.0mm at Bayarcal. Return periods for the annual maximum 24 hour storm have been calculated by Heras (1973) and presented in tables 3.6 and 3.7 for Ugijar and Mecina Bombaron. Figure 3.14 shows isolines of expected intensities over south east Spain for the 100 and 500 return periods from Heras, and the distribution of rainfall intensities for the October storm. According to these calculations the storm with a return period of 1000 years at Ugijar has 89.5mm of rain, which is almost half the amount of rainfall falling on 18 October at Ugijar (table 3.6). The data base for the return periods calculated by Heras at Ugijar is short - only 28 years - and the most extreme event is 59mm (table 3.6). The return periods are recalculated for the 40 year series which now includes storms of 172.0mm, 115.0mm, 80.1mm and 70.0mm / 24 hours. The Gumbel distribution for extreme events is fitted using the equation:

$$F(x) = \exp(-e^{-b(x-a)})$$

where F(x) is the probability of an annual maximum Q being less than or equal to x, and a and b are two parameters:

$$a = \mu_e - \gamma / b \text{ , where } \gamma = 0.5772$$

and

$$b = \pi / \sigma_e . \sqrt{6}$$

The values of μ_e and σ_e relate to the whole population and are estimated by using the sample mean ($\hat{\mu}_e$) and sample variance ($\hat{\sigma}_e^2$) (Shaw 1983). The calculations are presented in table 3.8 and the new distribution is plotted in figure 3.15. The new estimate for the return period for the storm of 18 October 1973 is 1000years. The storm certainly had a return period in excess of 500 years, but Thornes (1976) points out that larger storms

Figure 3·13 Percentage of 24 hour rainfall intensity for all storms between 1942 and 1983

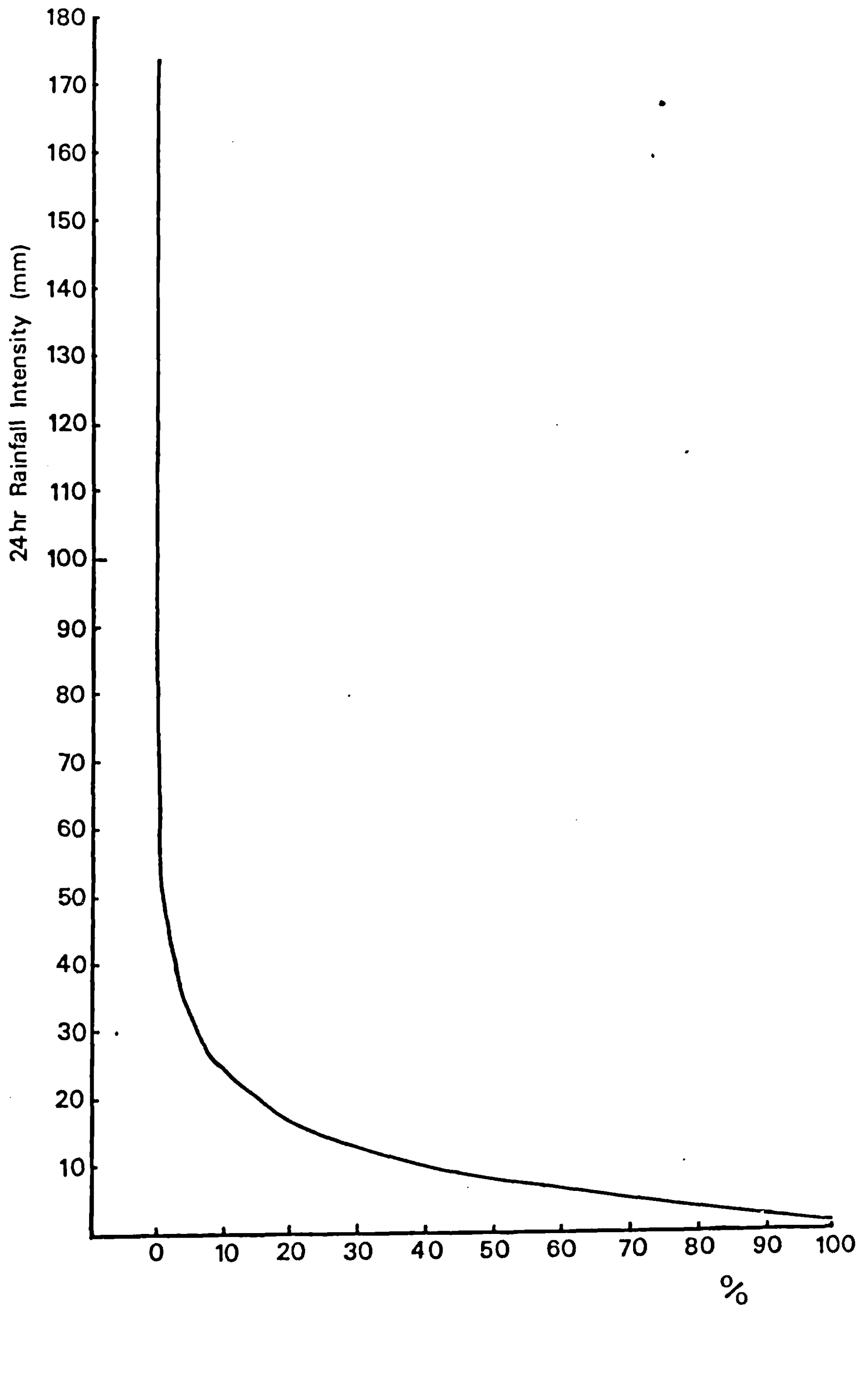


Table 3.6 Return periods for storms at Ugijar (Source: Heras 1973)

Prec Max 24h	Frequency
59.00	0.018
57.00	0.054
54.00	0.089
52.00	0.125
51.00	0.161
51.00	0.196
51.00	0.232
48.00	0.268
46.00	0.304
45.00	0.339
44.00	0.375
43.00	0.411
40.00	0.446
40.00	0.482
39.00	0.518
39.00	0.554
38.00	0.589
36.00	0.625
35.00	0.661
35.00	0.696
34.00	0.732
32.00	0.768
30.00	0.804
30.00	0.839
28.00	0.875
28.00	0.911
23.00	0.946
22.00	0.982

Return Period	Fx	Prec Max 24hrs
	0.100	29.41
	0.200	32.19
	0.300	34.44
	0.400	36.56
	0.500	38.72
	0.600	41.09
	0.700	43.88
5	0.800	47.52
10	0.900	53.34
25	0.960	60.70
50	0.980	66.15
100	0.990	71.58
500	0.998	84.09
1000	0.999	89.47

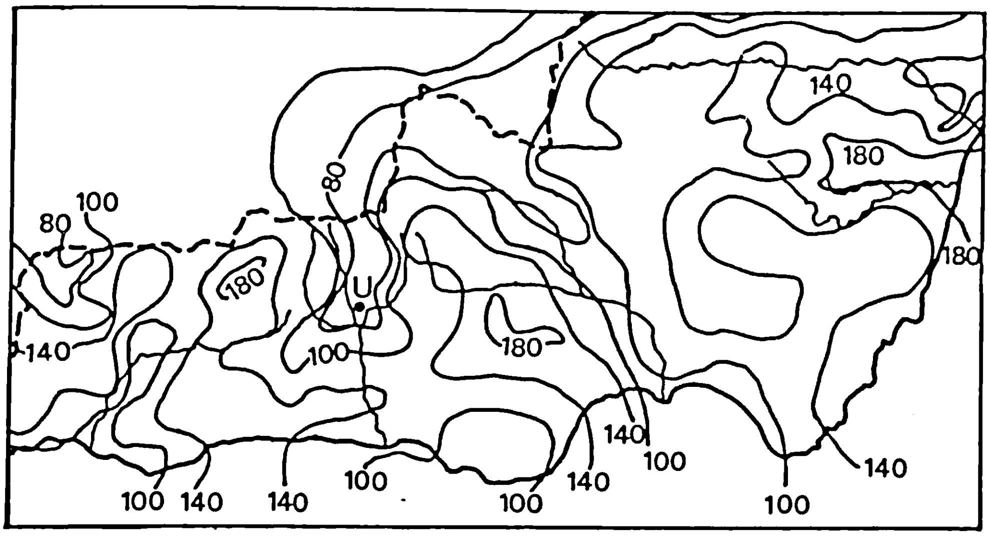
Table 3.7 Return periods for storms at Mecina Bombaron (Source: Heras 1973)

Prec Max 24h	Frequency
113.00	0.026
94.00	0.079
75.00	0.132
74.00	0.184
74.00	0.237
73.00	0.289
72.00	0.342
69.00	0.395
67.00	0.447
64.00	0.500
59.00	0.553
56.00	0.605
52.00	0.658
50.00	0.711
48.00	0.763
48.00	0.816
30.00	0.868
28.00	0.921
27.00	0.974

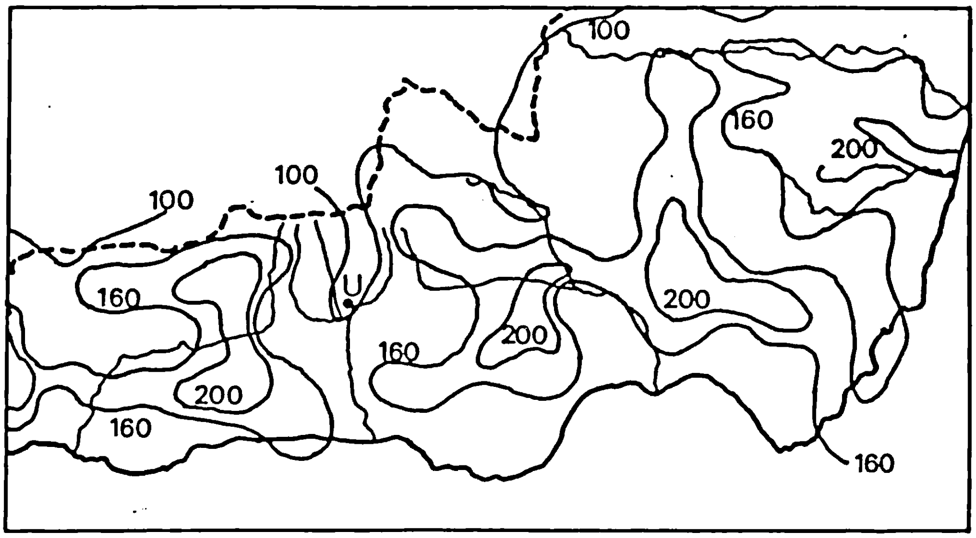
Return Period	Fx	Prec Max 24h
	0.100	38.51
	0.200	44.40
	0.300	49.18
	0.400	53.68
	0.500	58.27
	0.600	63.29
	0.700	69.20
5	0.800	76.92
10	0.900	89.27
25	0.960	104.88
50	0.980	116.46
100	0.990	127.97
500	0.998	154.50
1000	0.999	165.93

Figure 3·14 Isohytes of storm intensities with return periods of 100 and 500 years, and rainfall isohytes for 18 and 19 October 1973 for south east Spain (mm/24 hrs)

100 Year Return Period



500 Year Return Period



Rainfall Isohytes For 18 & 19 October 1973

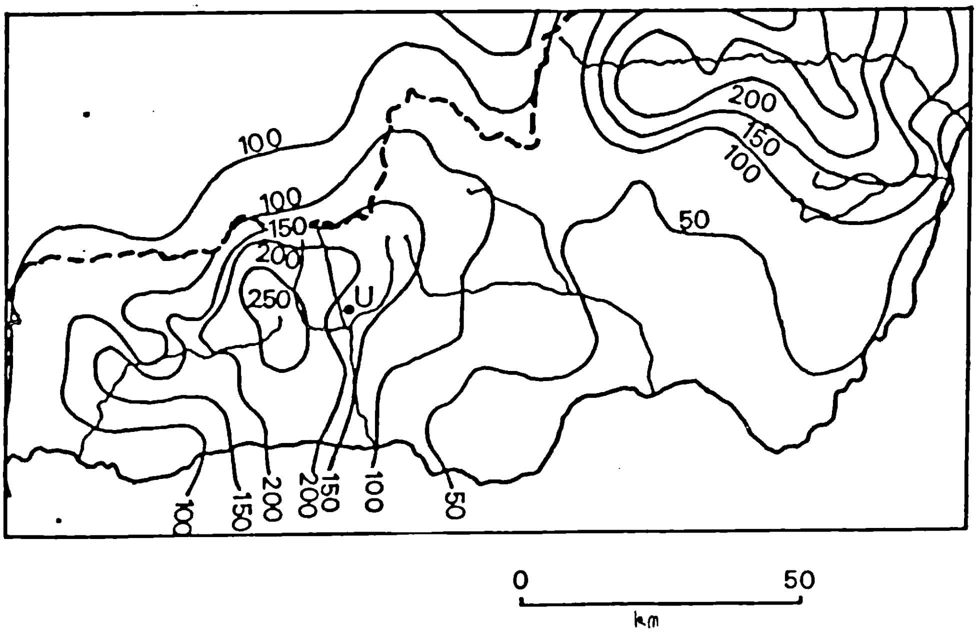


Table 3.8 Calculation of the Gumbel Extreme Distribution

Example 1 (Heras 1973)

Number of observations	28
Mean	40.357
Standard deviation	10.133
a	35.812
b	0.127

Intensity	F(x)
30	0.123
50	0.848
70	0.987
80	0.996

Example 2

Number of observations	40
Mean	45.768
Standard deviation	26.948
a	33.642
b	0.048

Intensity	F(x)
30	0.304
50	0.632
70	0.838
170	0.998

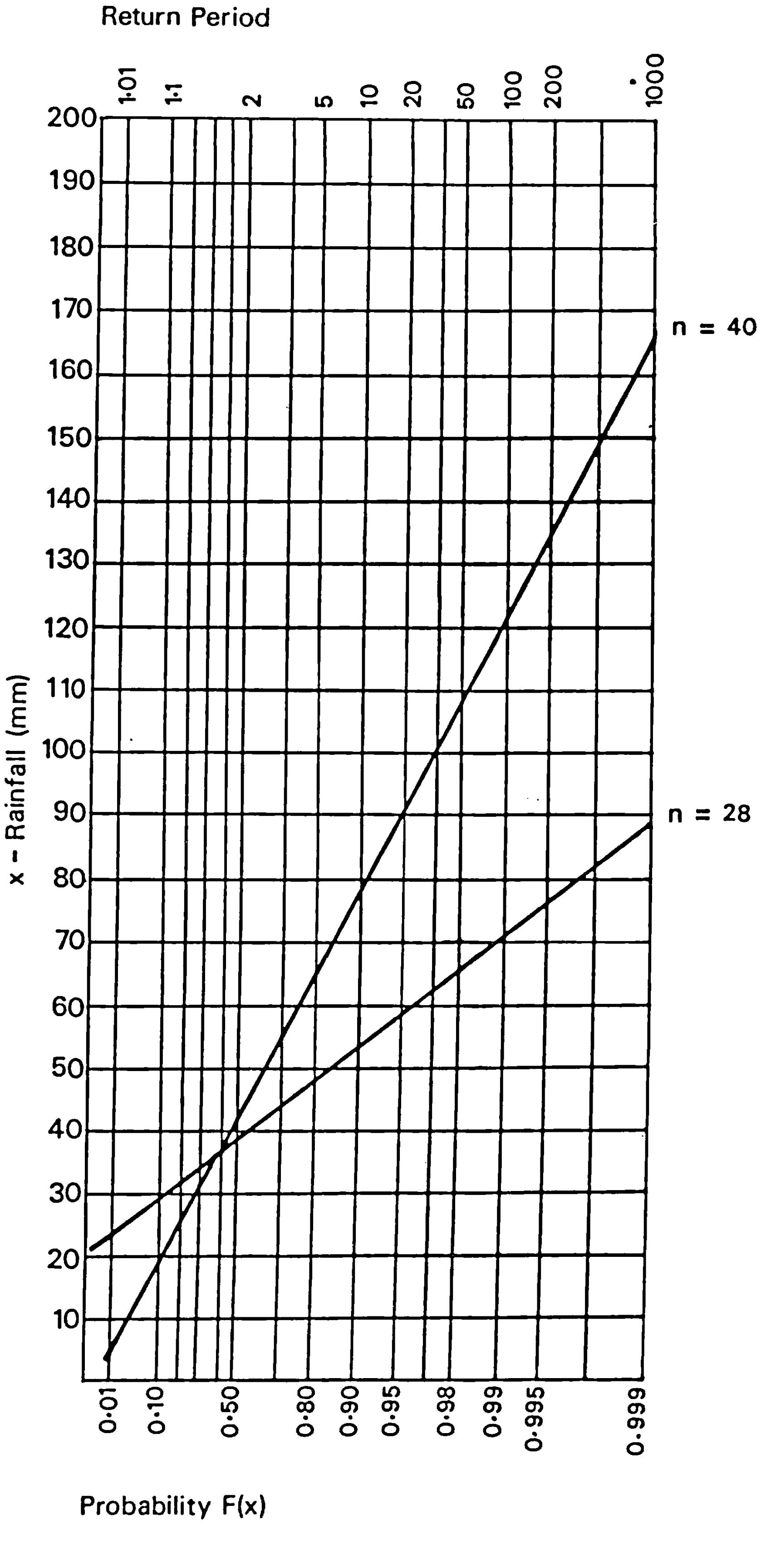
where

$a = \mu_e - \gamma/b$ $\gamma = 0.5772$

$b = \pi/\sigma_e\sqrt{6}$

$F(x) = \exp(-e^{-b(x-a)})$

Figure 3·15 Gumbel plot of the annual maximum series to determine return periods for extreme storms



occurred during the last century. Although the second, extended data set gives a more sensible estimate of the return periods, it is still too short to provide really meaningful return periods for extreme events.

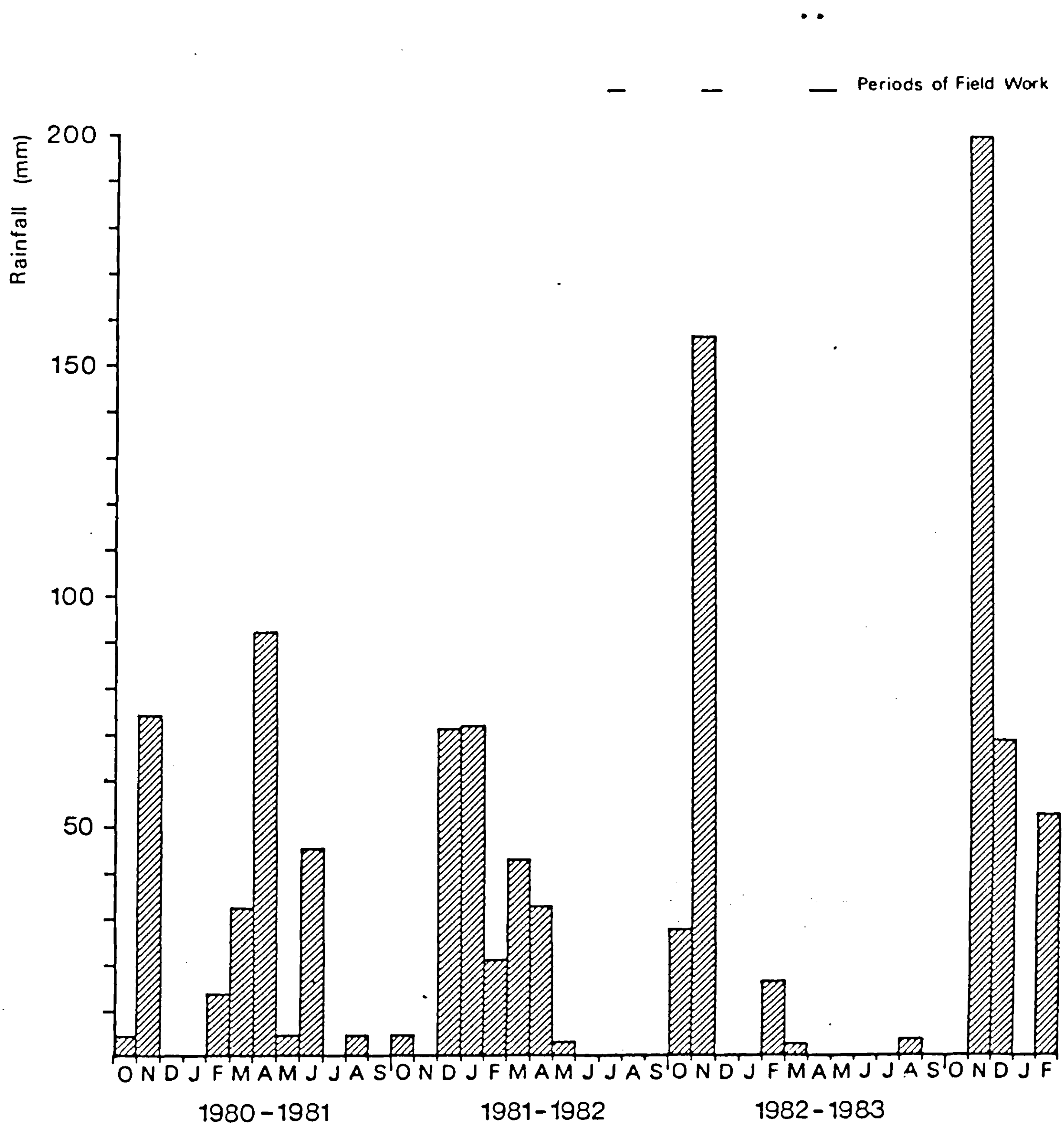
Storm-based rather than 24-hour based data on intensities are rare for south east Spain. However Scoging (pers. comm.) has analysed 82 storm profiles at sites around Ugijar. There is a skewed distribution of storm lengths with 86% of storms lasting 1 hour, and 50% lasting only 10 minutes. Maximum 10 minute intensities are also skewed with 63.6% less than 0.5mm. However the tail of the distribution is important with 10% of the 10 minute intensity storms exceeding 2mm.

The overall impact of the precipitation record points to a number of important characteristics with respect to erosion. All types of data at different timescales are highly variable. The total annual rainfall for the record is normally distributed, but most other data sets have a negative exponential distribution with a large number of small magnitude events and one or two high magnitude events. The extreme storms are very important because, although they come infrequently, they are responsible for a considerable amount of damage when they do occur. The rainfall is seasonal with a summer drought occurring annually. Furthermore this area is prone to a sequence of dry years which may lead to quite severe droughts.

3.2 Precipitation Patterns for the Study Period

Figure 3.16 shows the monthly precipitation data for Ugijar from October 1980 to February 1984 with the field trips in July 1982, November/December 1982, and April 1983 marked on (daily data for the same period is given in Appendix 1). The main recharging of moisture prior to the study period was the winter of 1981/82 which was relatively dry with 245.6mm of rain. Maximum rainfall was in December (70.7mm) and January (71.4mm) but following this, rainfall was more sporadic. The last relatively large rainfall was on 31 March (32.0mm) after which the amount of rainfall decreased rapidly. The last storm before the field investigations started

Figure 3·16
MONTHLY RAINFALL TOTALS AT UGIJAR
(October 1980-February 1984)



(Source: Comisaria De Aguas Del Sur De España) occurred on 10 May (3.0mm). A similar pattern is found at Mecina Bombaron although rainfall values are higher due to the orographic influence. Figure 3.17 for Mecina Bombaron shows that the wettest months were again December (167.8mm) and January (148.6mm), and the last relatively large storm was on 15 April (45mm).

Taking the Ugijar data, by the time of the first soil moisture readings on 16 July 1982, there had been 35 consecutive days without rain, and 54 days since the last storm greater than 10mm/24 hours. Together with the temperature regime, one may expect high potential evapotranspiration rates (section 3.3) and dry soils (Chapter 6).

No rain was recorded at Ugijar or Mecina Bombaron during the first study period (12 July - 1 August 1982), but on the night of 18/19 July rain did fall in Ugijar (pers. obs.). The amount was small and did not make much impact on the field site but the amount should have been recorded at the rainfall station in Ugijar, especially as measurements as fine as 0.2mm are noted in the records. Similar discrepancies were also found by Scoging (pers. comm.). Scoging used an automatic raingauge on her sites and was able to assess the accuracy of the data itself, but that was not possible in the present study. An attempt was made to measure the rainfall on site, however because of a malfunction of the instrument, and the very small number of rainfall events, recourse was made to the data from adjacent stations. Thus there are not only problems in interpreting processes based on rainfall recorded 4.5km away in Ugijar, but also inaccuracies in the precipitation record itself. The rainfall data is assumed to be correct, although it is acknowledged that there are unknown operating errors.

Between the July and November field sessions several rainy days were recorded at Ugijar and Mecina Bombaron (table 3.9). Hourly data from Mecina Bombaron suggests this reflects the passage of only three storms. The storm on 19-21 October was quite modest with 12mm falling on 20th and 15.5mm on 21st October in Ugijar, and 9mm and 16mm on 19th and 20th October at Mecina Bombaron. The storm on 1 November 1982 was very small, with only 1.2mm of rain falling at Ugijar, and a total of 13.2mm at Mecina Bombaron

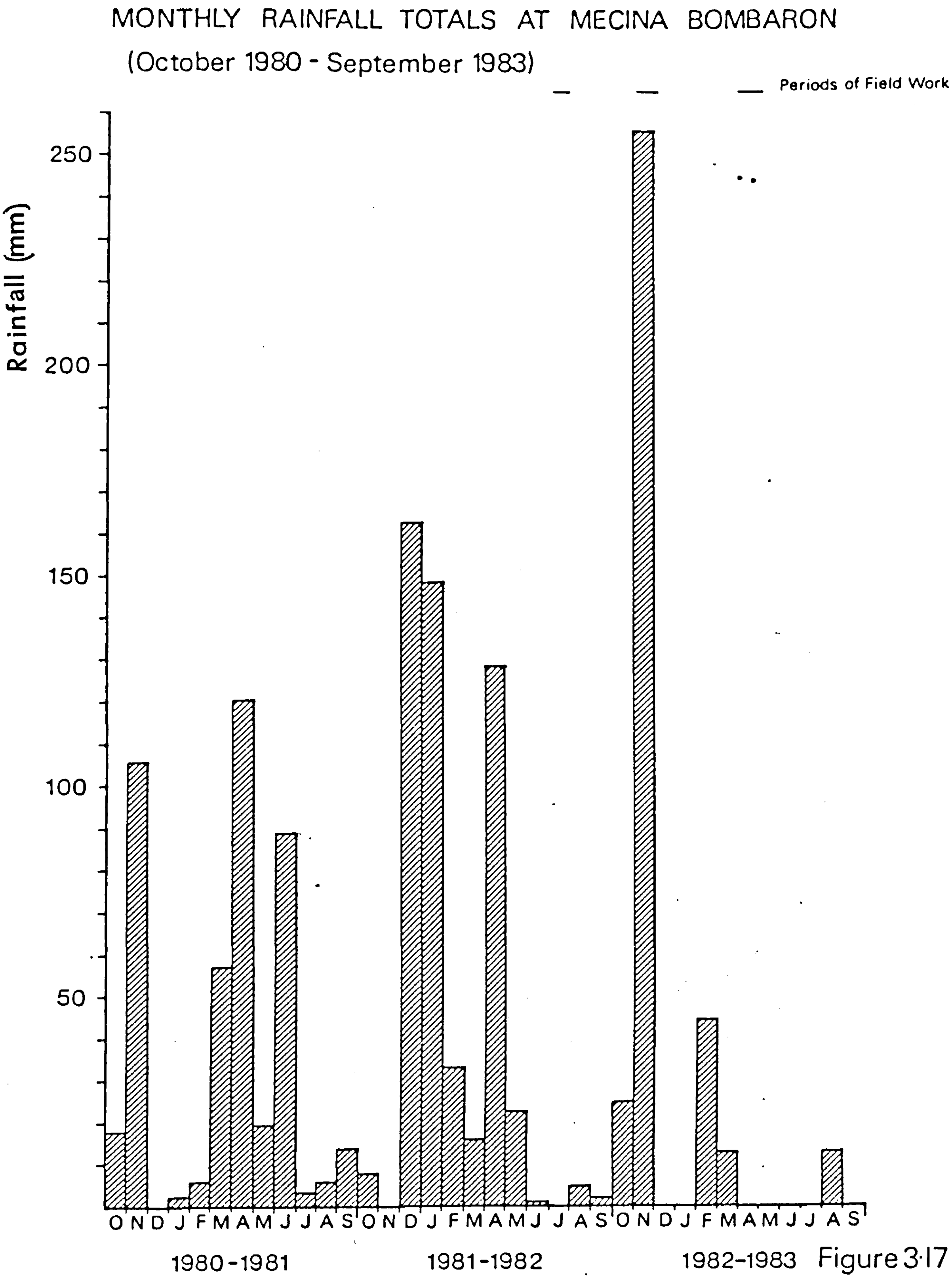

MONTHLY RAINFALL TOTALS AT MECINA BOMBARON
(October 1980 - September 1983)
Periods of Field Work
Rainfall (mm)
250
200
150
100
50
O N D J F M A M J J A S O N D J F M A M J J A S O N D J F M A M J J A S
1980-1981
1981-1982
1982-1983

Figure 3·17

Table 3.9 Rainfall occurrences between 1 August and 9 November 1982 from monthly data

DATE	UGIJAR mm	MECINA BOMBARON mm
19 October	00.0	9.0
20 October	12.0	16.0
21 October	15.5	000.0
1 November	1.2	8.0
2 November	00.0	5.2
6 November	62.0	210.0
7 November	70.0	000.0

over 1st and 2nd November.

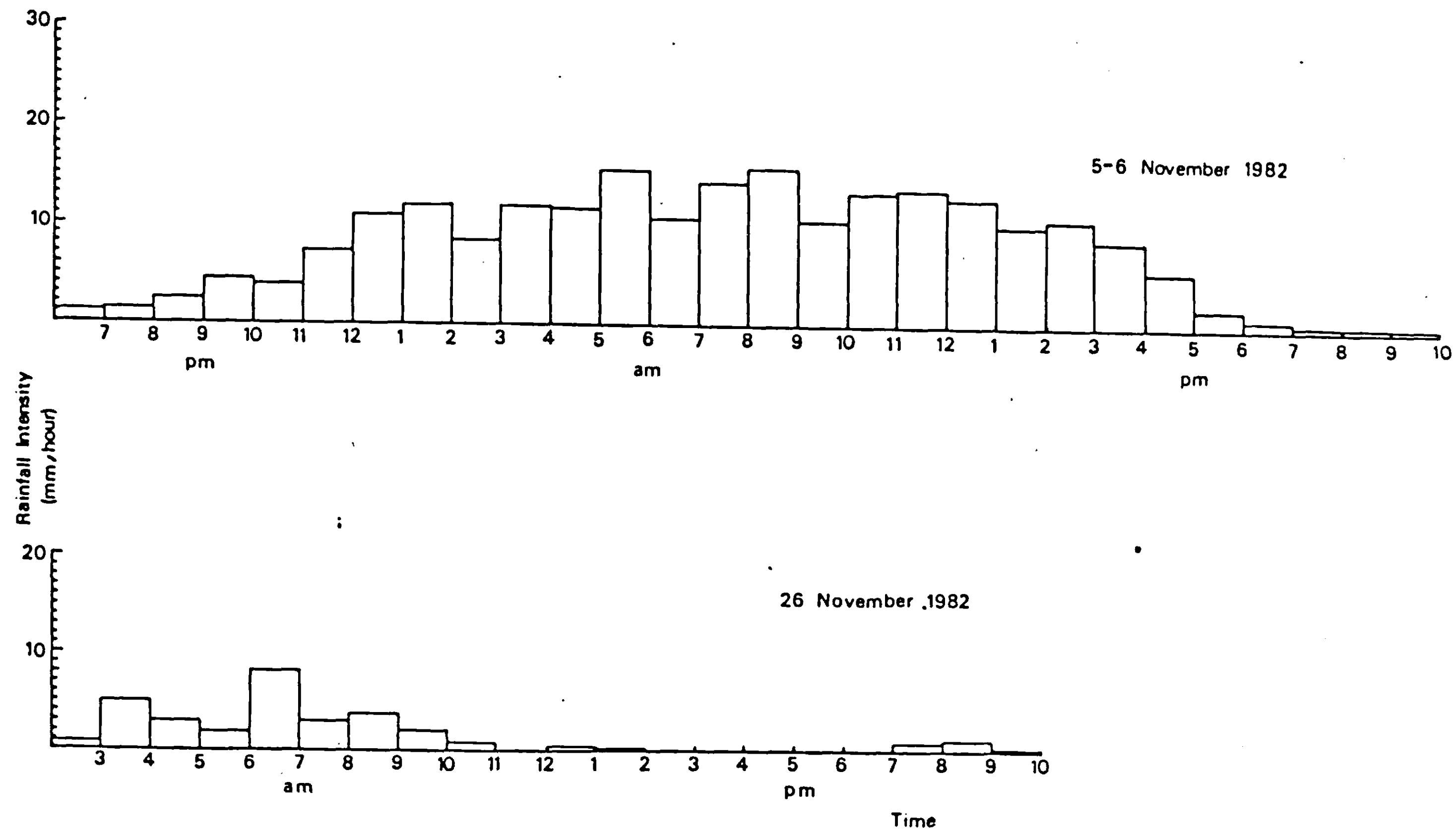
The storm on 6-7 November overshadows both previous events. Total rainfalls of 210.0mm and 132.0mm fell at Mecina Bombaron and Ugijar respectively. For Ugijar this represented 84.5% of the rain in November and 64.2% of the annual precipitation for that year. The hourly data from Mecina Bombaron shows this was a prolonged storm with intensities reaching 15mm/hour (figure 3.18a). Here, in Mecina Bombaron, a total of 228.2mm fell in 38 hours, and 190.3mm within 16 hours. Although the daily data shows all the rain falling on one day, the hourly data shows the storm starting after c6.00pm on 6 November and continuing until c5.00am on the morning of 8 November - another discrepancy.

At Ugjiar 62mm is recorded for 6 November and 70.0mm on 7 November. These have return periods of 4.4 years and 6.2 years respectively as separate storms. However if it is assumed that both are part of the same storm and artificially divided then the subtotals added are 132.0mm which would have an expected return period of 100 years. Certainly the probability of having two successive days of high rainfall, or one large storm of 132.0mm makes this a fairly rare event.

During the second field trip between 13 November - 9 December 1982 there was only one storm. This is recorded as occurring on 26 November in Ugijar with 23mm of rain, and on 25-26 November at Mecina Bombaron with 27.4mm and 5.0mm respectively. The data for Mecina Bombaron shows that the rain here started after 2 am on the morning of 26 November (rainfall before 8.00am is considered part of the previous day's account). The two totals have been summed to give 32.4mm on 26 November for Mecina Bombaron.

A histogram for the storm trace at Mecina Bombaron is shown in figure 3.18b. Rainfall was heaviest between 6.00-7.00 am and continued more gently from 12.00-2.00pm (this hides shorter term high intensity). Field observations show that the rain was very fine and beginning to stop around 11.00 am on the site, which accords quite well with the records at Mecina Bombaron. Rain continued again in the evening from 7.00-10.00pm according to the records at Mecina Bombaron, but this pattern is unsubstantiated at

Figure 3·18 Hourly storm traces for (a) 5-6 November 1982 and (b) 26 November 1982, at Mecina Bombaron



Ugijar. Maximum intensities were 8.0mm/hour, 2 hours had rainfall greater than 4mm/hour, and 7 hours had intensities less than 2mm/hour out of a total 14 hours. A storm of 23mm at Ugijar was exceeded by 11% of storms in the forty year record (figure 3.13) and has a return period of 1.2 years using the corrected 24hour maximum distribution (figure 3.15).

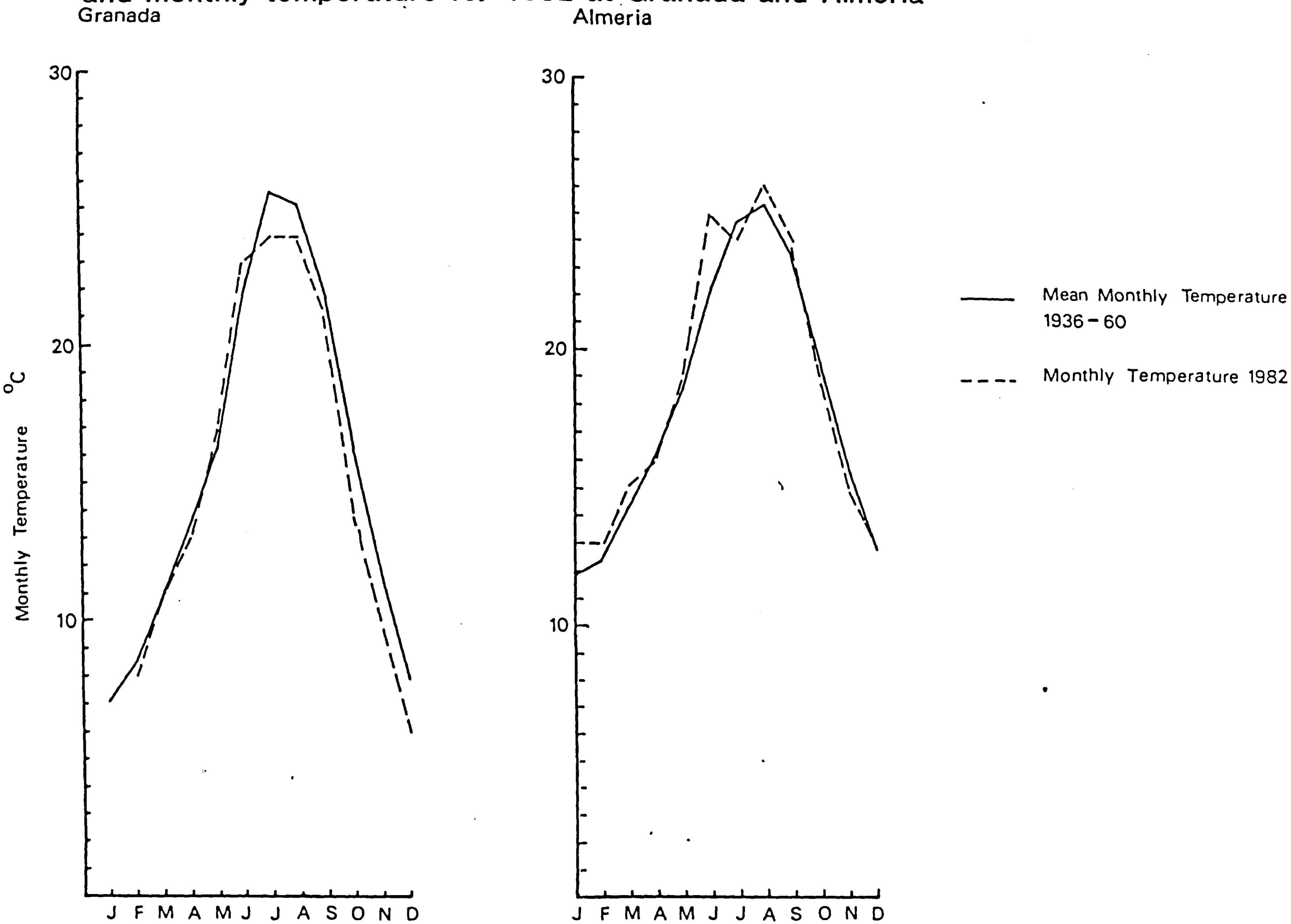
The winter of 1982-83 was markedly dry with a total of 18.5mm falling on three occasions between 9 December and 6 April at Ugijar. From 16 November to the next rainy day on 12 February there were 11 consecutive weeks of dry weather. At Ugijar the 1982-83 hydrological year was the third driest on record despite the heavy storm on 6 November. The winter was followed by a dry spring and summer with no rain between 22 March and 30 August - 157 consecutive days without rain. This would undoubtedly intensify the water stress conditions of the drought. On 30 August only 3.5mm of rainfall is recorded for Ugijar, and dry weather continued for another 62 days till 3 November.

The autumn of 1983 was a significantly wetter period breaking the drought spell. The Gerlach troughs remained installed until 20 December 1983, although other monitoring operations had ceased in the previous spring. Between 3 November and 20 December some 15 rainy days with 268.2mm of rain occurred - more rain in two months than in the previous twelve. The records for Mecina Bombaron only continue to September 1983, but they mirror the data from Ugijar, with relatively low rainfall values for the winter of 1982/3 and a dry spring and summer.

3.3 Soil Moisture Deficit

Temperature, like precipitation, varies considerably throughout the year. Temperature data is not available for Ugijar, and the two nearest stations are Granada and Almeria. Figure 3.19 shows the mean monthly temperatures for 1936-60 and temperatures for 1982 at Almeria and Granada. Both graphs peak with mean maximum temperatures of 25.4°C in August at Almeria and 25.7°C in July at Granada. Mean January temperatures are lower at Granada (7.9°C) than Almeria (11.8°C) reflecting the influence of

Figure 3·19 Mean monthly temperature for 1936 to 1960 and monthly temperature for 1982 at Granada and Almeria



continentality and elevation. The small variation between the mean and 1982 values show that temperature values are less variable than rainfall.

A combination of the temperature and precipitation data suggests that there will be seasonal water stress as a result of low rainfall and high temperatures in the summer which will restrict plant growth and soil moisture movement. There are two ways of looking at this problem, firstly to model the soil moisture budget and secondly to monitor the soil moisture itself. The latter is very time consuming and applicable only to small areas due to variations in precipitation, soil type and plant cover, however some values for soil moisture content at different seasons are described in Chapter 6.

There are a number of models available for examining soil moisture conditions over a watershed. Just one example is SPUR (Wight 1983), a rangeland simulation model with five components: climate, hydrology, plant, animal and economic. In the hydrology component the water balance in the soil is estimated by:

$$SW = SWo + P - Q - ET - PL - QR$$

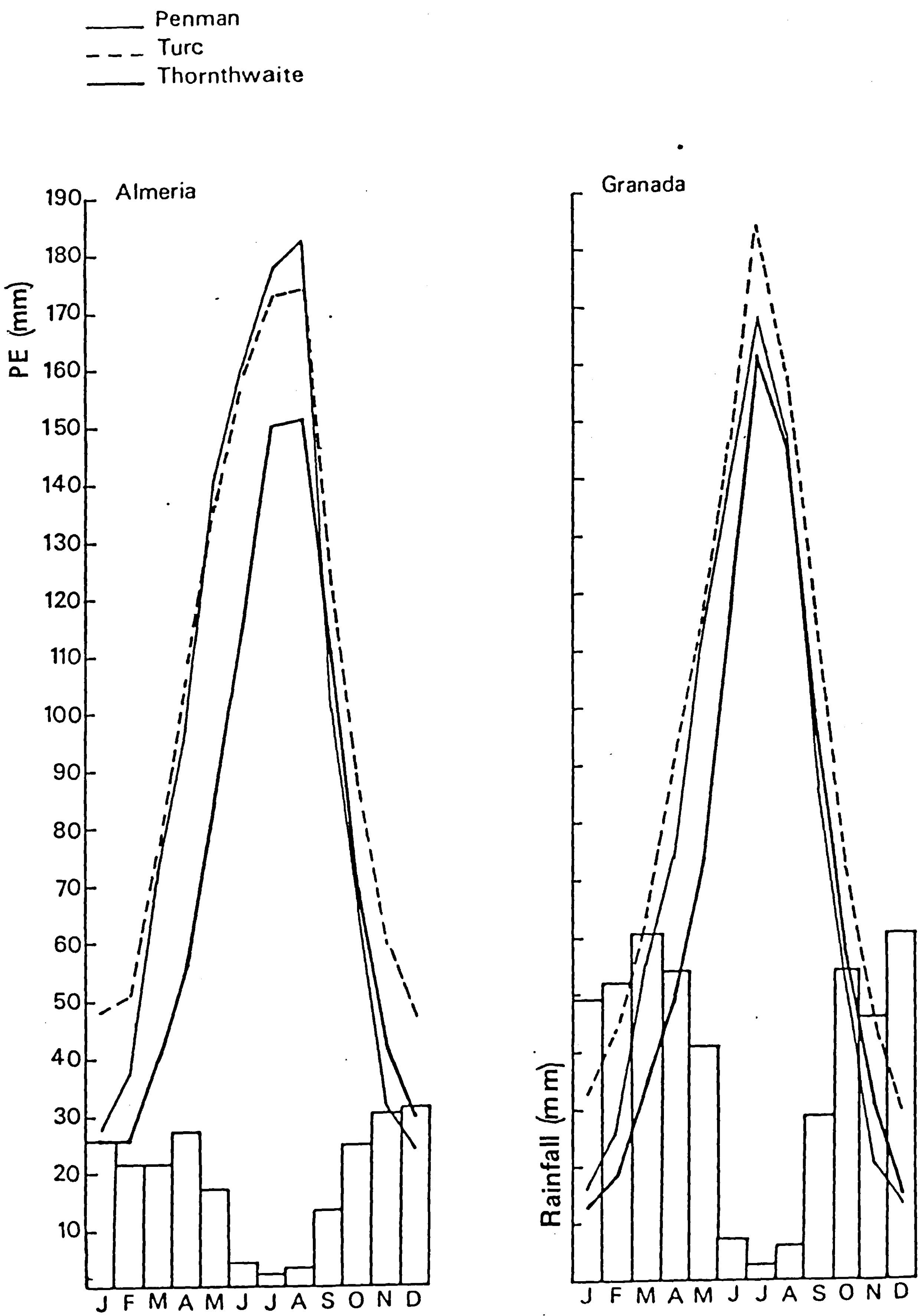
SW = current soil water content (in)
SWo = initial soil water content (in)
P = cumulative rainfall (in)
Q = cumulative amount of surface runoff (in)
ET = cumulative amount of evapotranspiration (in)
PL = cumulative amount of percolation loss to ground water storage (in)
QR = cumulative amount of return flow (in).

Complex watersheds are divided into subgroups to reflect variations in soil, vegetation, topography etc. For each subgroup the runoff is calculated and routed to the outlet of the drainage basin. The total storage, field capacities and initial storage in the soil layers are set by the soil characteristics. Within this framework the soil moisture content is calculated on a daily basis on gains from precipitation and losses by percolation and evapotranspiration. The percolation algorithm combines a storage routing model with a crack flow model to predict flow through the root zone.

The evapotranspiration component calculates both the potential and actual values. Potential evapotranspiration is based on the slope of the saturation vapour pressure curve at the mean air temperature, the net solar radiation, and a psychometric constant. The actual evapotranspiration is the summation of soil evaporation and plant transpiration estimates. The soil evaporation is calculated in two parts, for the wet and dry conditions. In the first place soil evaporation is approximated by the potential rate, which is related to either the leaf area index of the vegetation or a mulch cover factor. In the dry case the rate is estimated from the transmission characteristics of the soils. Transpiration rates are also separated according to whether the moisture is limiting or not. In the unlimited case the transpiration rate is either (i) the product of the potential evaporation and leaf area index divided by a constant, or (ii) the potential evaporation minus the actual soil evaporation for the limiting condition, depending on whether the leaf area index is greater or less than three. When soil moisture is limiting the transpiration rate is the product of the potential transpiration and the current soil moisture in the root zone divided by a quarter of the total soil water storage capacity. The computed evapotranspiration for a day is then distributed in the soil layers based on the rooting depth.

This type of modelling requires considerable data input which are not available for many catchments and incorporates site/climate/vegetation specific constants. At a lower level of modelling potential evapotranspiration and soil moisture deficits can be calculated from simple, if less accurate, standard equations which provide an upper maximum for evapotranspiration assuming no losses of water (for example from runoff) and optimum operating conditions. Examples of empirical formulae for PE include the Thornthwaite, Turc and Penman methods which have been calculated for a number of stations in Spain and are presented by Castillo and Ortiz (1965). Values for Granada and Almeria are plotted in figure 3.20 for comparison. At Granada the Thornthwaite method estimates the lowest rates of PE between January and August, and thereafter records similar values to the Penman method . The Turc method gives the highest values of PE for all months, and is relatively higher than the other methods for the winter months November to February, and for July. A similar pattern emerges for Almeria where the Turc method tends to

Figure 3·20 Comparison of potential evapotranspiration calculated by the Penman, Turc and Thornthwaite methods for Almeria and Granada



give the highest results. Between March and August the Penman and Turc methods are similar, but in the autumn the Penman and Thornthwaite methods are similar. Castillo and Ortiz (1965) suggest that the Penman formula gives the best results in those areas where direct measurements of evapotranspiration are not available. This equation requires a fair amount of data only obtainable from a small number of meteorological stations in Spain. The simplest is the Thornthwaite method which is based on the mean monthly temperature, and an estimate of hours of sunlight according to latitude. It was developed in the eastern U.S.A. and should be used only in areas with similar climates although it has been applied worldwide (Shaw 1983). It has been criticised mainly for being too heavily dependent on temperature and empirical. Figure 3.20 also indicates the relationship between PE and precipitation. The total mean annual PE using the Thornthwaite method is 818.12mm at Granada and 890.79mm for Almeria, and total mean annual precipitation is 473mm and 234mm respectively. Thus PE may be between 1.7 to 3.8 times greater than the actual rainfall in south east Spain.

The potential soil moisture deficit can now be estimated by subtracting the monthly potential evaporation from the monthly rainfall (Shaw 1983). There are several drawbacks to the method. The accumulation of data to the monthly basis smoothes out the effect of storms so that the impact of one or two large storms is spread out for the whole month. This will lead to inaccuracies in the data. Also a simple deducting model such as this will always ensure a soil moisture deficit by the end of the year as PE is so much greater than the annual rainfall. If the procedure is continued for a second year the deficit will be cumulative. There is, in fact a limit to evaporation from the soil when the soils become dry. Hillel (1982) describes three phases of evaporation from a bare soil surface in the absence of a water table:

1. There is an initial constant-rate stage when the soil is wet and capable of supplying enough water to meet the evaporative demand. In this case the evaporation rate is limited by the meteorological conditions.
2. Next is an intermediate falling-rate stage when the evaporation rate falls below the potential rate. Evaporation in this stage is controlled by the transmission character of the soil.

3 Finally there is a residual slow-rate stage which may persist at an almost constant rate for several days, weeks or even months. Here transmission occurs primarily through vapour diffusion.

Despite these problems this model is used to estimate soil moisture deficits at Ugijar and look at the effect of the drought. The problem is still hampered by the lack of temperature data for Ugijar, the nearest stations with temperature data being Granada and Almeria. The temperature regime at Ugijar is more likely to resemble that of Granada than Almeria. Granada and Ugijar lie at 570m and 559m respectively, whereas Almeria is 21m above sea level, and like Granada, Ugjiar lies inland and is surrounded by mountains, albeit in a much smaller basin. Values for the potential soil moisture deficit are then calculated using precipitation data for Ugijar for the average precipitation record (1942-1983) and the average temperature data for Granada (based on records from 1936-1960), and separately for the years 1981, 1982, and 1983 using the precipitation data from Ugijar and temperature data from Granada. These data are presented in table 3.10 a, b, c, and d. In the 'average' year the soil moisture deficit begins in June and rises to 502mm by October before decreasing. In 1981 and 1983 there was a soil moisture deficit for almost the whole year, and in 1983 the maximum deficit reached 715mm by October. The effect of the drought seems to have increased the soil moisture deficit, and made it start earlier in the year.

The difference between the study year and average conditions is emphasised by some maps produced by the Instituto Nacional de Meteorologia. Data collected from 1 September to 10 July 1983 on precipitation, evaporation, soil moisture reserves and deficits are compared to the average conditions (1931-60) for Granada and Almeria. These are shown in table 3.11. Actual and average evapotranspiration rates are similar, but rainfall is between 1/3 and 1/2 of the norm. Reserves of soil moisture are zero as expected, but deficits of soil moisture are 1.2 to 2.0 times greater than normal for the hydrological year due to the lack of rainfall. The evapotranspiration and deficit data is mapped for south east Spain in figure 3.21 to show the regional distribution of the drought intensifying in the most south easterly part of the country.

Table 3.10 Calculations of soil moisture deficit at Ugijar

(a) Estimate of the SMD using average rainfall from Ugijar (1942-83) and monthly temperature data from Granada (1931-60).

Month	Rain (mm)	PE (mm)	R-PE (mm)	Pot SMD (mm)
Jan	46.55	13.48	33.07	33.07
Feb	40.11	19.75	20.36	53.43
Mar	43.03	32.07	10.96	64.39
Apr	43.56	50.29	-6.73	57.66
May	23.65	73.07	-49.42	8.24
Jun	6.15	124.34	-118.19	-109.95
Jul	0.32	155.96	-155.64	-265.59
Aug	6.55	142.01	-135.46	-401.05
Sep	17.75	102.08	-84.33	-485.38
Oct	42.18	58.78	-16.60	-501.98
Nov	48.70	30.86	17.84	-484.14
Dec	58.84	15.43	43.41	-440.73

(b) Calculations for SMD at Ugijar for the calendar year 1981.

Month	Rain (mm)	PE (mm)	R-PE (mm)	Pot SMD (mm)
Jan	0.0	8.30	-8.30	-8.30
Feb	13.5	15.37	-1.87	-10.17
Mar	32.0	43.93	-11.93	-22.10
Apr	92.0	43.08	48.92	26.82
May	4.5	72.40	-67.90	-41.08
Jun	45.0	132.92	-87.92	-129.00
Jul	0.0	139.74	-139.74	-268.74
Aug	4.5	130.50	-126.00	-394.74
Sep	0.0	90.37	-90.37	-485.11
Oct	4.5	63.39	-58.89	-544.00
Nov	0.0	33.29	-33.29	-577.29
Dec	70.7	20.00	50.70	-526.59

Table 3.10 (Cont.)

(c) Calculations for SMD at Ugijar for the calendar year 1982

Month	Rain (mm)	PE (mm)	R-PE (mm)	Pot SMD (mm)
Jan	71.4	17.81	53.59	53.59
Feb	21.0	19.53	1.47	55.06
Mar	42.5	34.70	7.80	62.86
Apr	32.5	49.86	-17.36	45.50
May	3.0	80.85	-77.86	-32.35
Jun	0.0	134.24	-134.24	-166.59
Jul	0.0	141.00	-141.00	-307.59
Aug	0.0	131.68	-131.68	-439.27
Sep	0.0	98.66	-98.66	-537.93
Oct	27.5	47.76	-20.26	-558.19
Nov	156.2	25.72	130.48	-427.71
Dec	0.0	11.02	-11.02	-438.73

(d) Calculations for SMD at Ugijar for the calendar year 1983

Month	Rain (mm)	PE (mm)	R-PE (mm)	Pot SMD (mm)
Jan	0.0	11.00	-11.00	-11.00
Feb	16.0	12.06	3.94	-7.06
Mar	2.5	38.69	-36.19	-43.25
Apr	0.0	48.70	-48.70	-91.95
May	0.0	65.31	-65.31	-157.26
Jun	0.0	132.62	-132.62	-289.88
Jul	0.0	139.42	-139.42	-429.30
Aug	3.5	113.67	-110.17	-539.47
Sep	0.0	112.13	-112.13	-651.60
Oct	0.0	63.25	-63.25	-714.85
Nov	199.7	37.63	162.07	-552.78
Dec	68.5	16.61	51.89	-500.89

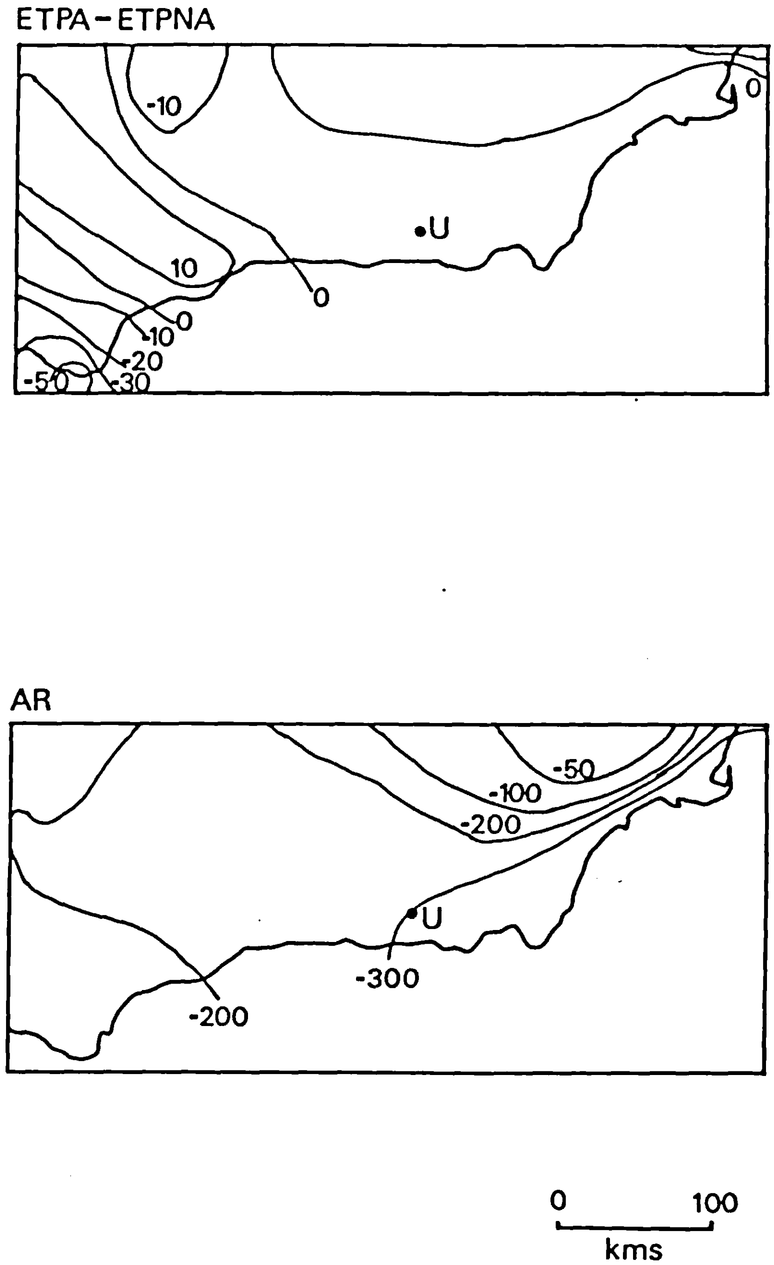
Table 3.11 Comparisons of evapotranspiration, rainfall, soil moisture reserve, and soil moisture deficit to the "norm" from 1 September 1982 to 10 July 1983

	GRANADA	ALMERIA
ETPA	557.7	669.9
ETPNA (1931-60)	563.4	674.0
PA	207.6	103.3
PNA (1931-60)	391.3	224.6
R-R	0	0
R-N (1931-60)	0	0
D-R	218.8	361.8
D-N (1931-60)	110.2	290.5

Source: Instituto Nacional De Meteorologia

ETPA = Actual potential evapotranspiration
ETPNA = Mean potential evapotranspiration (1931-60)
PA = Actual precipitation
PNA = Mean precipitation (1931-60)
R-R = Actual soil moisture reserve
R-N = Mean soil moisture reserve (1931-60)
D-R = Actual soil moisture deficit
D-N = Mean soil moisture deficit (1931-60)

Figure 3·21 Potential evaporation and soil moisture deficit for south east Spain from 1 September 1982 to 10 July 1983 compared to average values from 1931 to 1960



It is evident that the climate of south east Spain is variable at all the temporal scales. Superimposed on this, the annual seasonality of rainfall and temperature leads to summer soil moisture deficit, which may be intensified during a run of dry years. This will affect the erosional processes in a number of ways. Firstly one may expect low values for soil moisture and throughflow values. This will have implications for vegetation growth and cover which indirectly affects erosion. The occurences of erosion-causing storm events will be below normal possibly underestimating erosion rates. However long periods of mechanical weathering between storms may increase relative sediment yields per storm. As the study period took place during a particularly dry year, the number of erosional storms was severely limited, and the processes which occurred did so under very dry conditions. This may affect the typicality of the results, however the analysis of surface and subsurface hydrology offers actual measurements for water stress conditions important for agriculture and erosional studies.

CHAPTER 4 INFILTRATION AND MODELS OF OVERLAND FLOW

4.1 Infiltration: Research Problems and Results

Infiltration characteristics are used to describe the hydrological response to rainfall for different soils and topographic locations. In particular they determine variations of amount and location of surface flow production for different storms, which in turn affect the amount of water available for gully head erosion. The more important infiltration characteristics are the shape of the infiltration curve, the final infiltration rate, the storage capacity, the sorptivity values, and the time to ponding. These data are studied for both between and within site variation. Firstly variations in infiltration parameters are established for the two soils to compare the response of both. Secondly, for each soil infiltration differences over the slope which reflect the within site variability are studied with reference to key locations alongside Gerlach troughs measuring surface discharge. However before commencing on this is it important to establish the possible sources of variation in the data.

Sources Of Variation For Infiltrometer Parameters

There are two main sources of variation for infiltrometer readings. The first group is associated with geomorphological factors such as soil type and variability, antecedent soil moisture conditions, and long term changes in the properties of the soil surface due to, for example, armouring, vegetation cover, and catena development. The second group is variability due to instrumentation.

Firstly geomorphological factors affecting infiltration rates have been described by various authors (Horton 1945, Parr and Bertrand 1960, Musgrave and Holtan 1964, Chorley 1978 and Dunne 1978). These are classified into three groups by Knapp (1978) namely factors affecting:

1 the quantity and characteristics of water input,
2 the nature of the soil surface, and
3 the ability of the soil to conduct water away from the soil surface.

The rate of infiltration is affected by a number of rainfall characteristics such as intensity, storm duration, and the distribution of drop sizes. The rainfall intensity has received a lot of attention. For a given intensity there is a time from the beginning of the storm in which no runoff occurs as the soil is unsaturated, and the infiltration rate is set by the rainfall. When the surface soil is saturated the infiltration rate begins to fall and the rate is profile controlled. The set of curves produced for different intensities has been called the infiltration envelope by Smith (1972). The storm duration is important for storage type models as the available storage may fill and lead to overland flow, whilst compaction of the soil surface is related to the size distribution of drop sizes.

The nature of the soil surface affects rainwater entry rates through the size, number and connectedness of pore openings. (Dixon 1971, Knapp 1978) and potential changes in these variables caused by crusting, swelling, shrinking, compaction, and the inwashing of fines. These are, in turn, affected by soil type, land use practices and vegetation cover (Horton 1945).

The development of crusts during storms and their effectiveness in reducing rates of infiltration are well documented. Various authors have described crust morphology (McIntyre 1958, Tackett and Pearson 1965, and Farres 1978). For example McIntyre describes two components, firstly a skin seal due to compaction and secondly a washed-in zone of decreased porosity. Farres (1978) describes the vertical and areal development of crusts in laboratory conditions. The thickness of crusts may be self limiting by the processes of crustal development which protects the soil beneath from raindrop impact. Crusts developed during storms by raindrop activity which breaks down aggregates whose particles segregate and orientate themselves in a fine layer. Crust strength increases with drying but excessive drying induces cracking. Tackett and Pearson (1965) illustrate the differences between compacted and crusted soils, the latter having a much denser surface layer 1-3mm thick underlain by a more porous structure, the surface of which is coated with a thin skin of very well orientated clay. However they note that the crust strength of reconstituted soil is much greater than the original soil.

The effect of crusting on infiltration is to reduce it. Tackett and Pearson (1965) suggested that permeabilities below crusts are five times greater than permeabilities of the crust. McIntyre (1958) noted that if the crust was not complete, was perforated or cracked the rate of flow was high and the permeability of the surface of the same order of magnitude as the underlying soil. However once the crust is formed fully it can reduce infiltration by 200 times for the washed in region and 2000 times for the skin seal for a fine sandy loam. These results however are estimated in the laboratory for very small areas, and over larger areas the crustal development may be quite variable, for example the effect of crusts is reduced around pebbles and water stable aggregates protruding above the surface. Field evidence of the effectiveness of crusted soils is given by Imeson (1983) who found markedly lower infiltration rates from sprinkler experiments on crusted than non-crusted soils.

When describing the infiltration data for the Spanish sites it is important to keep in mind the strong crusting tendency of the marl soil. Observations suggest that the marl aggregates do break down readily during rainfall, but the crusts themselves are a result of the drying after the storm rather than crustal development during the storm. Further implications are discussed at the end of the chapter.

The ability of the soil to conduct water away from the surface depends on a variety of physical, chemical, biotic and temporal factors. Soil structure reflects the porosity and permeability of the soil. These vary between soils and with depth, but may also change with time by shrinking and swelling or the action of dispersive chemicals in the soils. Biotic structures such as roots, humus and worm action increase soil water movement in macropores and the water holding capacity of the soil. It is well known that small scale changes in soil structure will lead to infiltration variations within "homogeneous" soil units (the variability of soil structure is discussed in Chapter 6). Such variation is documented in the case of infiltration by Sharma, Gander and Hunt (1980). Measuring infiltration characteristics over a watershed 9.6ha they found no obvious pattern in the distribution of infiltration parameters with respect to soil type or position in the watershed. The within-soil variability was such that the different soils were hydrologically similar.

The antecedent soil moisture conditions affect infiltration with lower initial rates for wetter soils. As a result infiltration will vary between storms and seasonally (Bertoni et al 1958). On a longer timescale changes in the soil or rainfall characteristics will alter the infiltration characteristics.

The second main source of variability is due to instrumention. There are two main types of techniques for measuring infiltration, rainfall simulators and infiltrometers, which are described by several authors (Parr and Bertrand 1960, Musgrave and Holtan 1964, and Hills 1970). Both techniques measure infiltration relatively but direct comparison between the two methods is difficult with rainfall simulators often giving results an order of magnitude lower than infiltrometers. However Scoging (1982) found that despite the difference in values, the <u>pattern</u> of areas with higher and lower infiltration rates were the same for both methods.

The most accurate way of measuring infiltration is to use rainfall simulators on bounded plots at suitable intensities, then the infiltration is the deficit between water applied and the amount of water flowing off the soil surface. There are several disadvantages to this method. Care is needed to simulate natural raindrop distributions and intensities otherwise excessive compacting, crusting, or runoff may occur (Bork and Rohdenburg 1981). Lateral seepage is a problem but can be reduced by having a wetted buffer zone and relatively large plot areas to minimize lateral losses in comparison with vertical losses. Sprinklers are susceptible to wind which blows the water drops out of the controlled area. Even without wind effects it is difficult to ensure an even cover of raindrops over the plot. Infiltration estimates are affected by delivery problems as the infiltration is aggregated over a large area. Sprinklers are not very portable, they require large volumes of water (which may be restricting in many field locations), and they are expensive.

Infiltrometers measure the infiltration rate over a small area with ponded conditions. The water head may be constant or falling. Hills (1970) lists the main disadvantages as:

1 Disturbance to the soil during emplacement
2 Lateral flow of water underneath cylinders
3 Water seepage between cylinder and soil interface
4 Entrapped air reducing infiltration
5 Effect of temperature of soil and water
6 Representative effect of raindrop impact.

Disturbance of the soil structure during emplacement usually increases infiltration by creating cracks and macropores which are particularly significant on crusted and stony soils. This disturbance can be minimised using thin gauge steel infiltrometers inserted carefully using a spirit level. The infiltrometers may be single or double ringed, the latter reducing the effect of lateral water seepage as an outer zone of soil is kept wet.

The effect of entrapped air in the soil is reduced by shallow insertion of the infiltrometers with small areal extent. The effects of temperature of soil and water, and head of ponded water are thought to be negligible (Hills 1970). Philip (1958) suggests that initially for small values of head, an increase in head raises the infiltration rate by about 2% per centimetre, but with time the effect of head diminishes until it is negligible. Despite these problems infiltrometers are used frequently and were used here for their logistical advantages:

1 They are easily portable, installed and operated.
2 They need relatively small volumes of water.
3 They measure relative infiltration characteristics.
4 They are suitable over the range 30-500mm/hour (Hills 1970).
5 They can produce replicable results.
6 They are inexpensive.

Infiltrometer measurements themselves are known to be variable. Hills (1970) found the range of infiltration in oak woodland varied between 0 and 75cms/hour, with a mean of 6.8cms/hour. Knapp (1978) reported that Burgy and Luthin (1957) suggested that six cylinder infiltrometer measurements can come within 30% of the mean value when no restricting layers in the soil are present. Also Slater (1957) suggested 15 replications of cylinder infiltrometers are required to obtain the accuracy of one sprinkler infiltrometer measurement. Despite the view that infiltrometers tend to overestimate infiltration rates, they still give reliable data on relative rates, storage, and sorptivity.

Experimental Design

Infiltration experiments were carried out on both the marl and conglomerate soil over three field trips. At first a falling head infiltrometer was used however this suffered from several drawbacks, the worst being that the readings could only be taken for 1cm falls (ie about every 5 minutes) and so this method was relatively insensitive at the early part of the curve. This was modified by introducing a drip feed mechanism to maintain a small but constant head of water. Initially the modification was not thought severe enough to inhibit comparisons between results because it should not affect the values for infiltration rate and storage, but only change the shape of the curve. The range of head for the falling head infiltrometer used in the study varied between 2 to 7cms. For the constant head infiltrometer the head was always below 1cm.

Two experiments were undertaken involving a total of 31 infiltration runs. Firstly infiltration characteristics were measured downslope alongside Gerlach troughs to compare infiltration parameters with discharge measurements, and topographic position. In the autumn of 1982 and spring 1983 a total of 13 infiltration runs were conducted using the falling head type infiltrometer. The autumn experiments were done between 8 and 15 days after the storm on 26 November, when the soil was still damp. The spring experiments were done in drier conditions. Unfortunately soil moisture samples were not taken to quantify the soil moisture conditions.

Secondly in March 1984, a further 18 infiltration experiments were performed to examine spatial distributions of infiltration within small areas, and between soils. Three groups of six infiltration experiments were conducted on the perimeter of a circle with a 1m radius centred on neutron probe tubes 4 and 5 on site 1, and tube 2 on site 2.

Each experiment was conducted for 1 hour in the knowledge that the majority of storms are of shorter duration (Chapter 3 page 125). For the falling head infiltrometer readings were taken about every 5 minutes, but for the constant head devices, readings were taken more frequently, often at less than one minute intervals in the early stages.

Results

A summary of the results for infiltration rates, total storage and sorptivity values is shown in table 4.1. This table subdivides the data according to the type of infiltrometer used,.but not according to the antecedent soil moisture conditions. Several points can be drawn from this table.

1 There are differences between the results for the constant and falling head infiltrometers. For both lithologies, infiltration rates and storage are significantly higher using the constant head device by at least the 0.20 significance level, using a difference of means test.

2 The values obtained for infiltration rates, storage, and sorptivity are high.

3 The amount of variation indicated by the standard deviations is high. Table 4.2 shows the estimated size of sample populations required to get infiltration rates within 10% of the mean at the 0.05 significance level. This table brings out the greater variability found with the constant head infiltrometers, but in both cases the required sample populations are unmanageably large.

4 The sorptivity values for a given lithology are not significantly different between instruments, suggesting that the sorptivity values are less sensitive to type of instrument.

5 For a given type of instrument there is no significant difference for the infiltration parameters measured between marl and conglomerate soil.

The modification made to the infiltrometer should only have improved the resolution of the measurement, and not changed the nature of the data. It may be that the variations between the

Table 4.1 Summary table of infiltration characteristics

Infiltration Rates (cms/hr)

		Total	Falling Head	Constant Head
MARL	- mean	25.57	17.38	28.30
	stan.dev.	16.83	6.56	18.50
	number	16	4	12
CONG.	- mean	21.12	17.38	27.85
	stan.dev.	10.18	9.21	8.91
	number	14	9	5

Storage Values (cms)

		Total	Falling Head	Constant Head
MARL	- mean	12.57	2.37	15.97
	stan.dev.	12.33	1.45	12.50
	number	16	4	12
CONG.	- mean	6.18	2.89	12.11
	stan.dev.	7.60	2.29	10.44
	number	14	9	5

Sorptivity Values

		Total	Falling Head	Constant Head
MARL	- mean	4.16	4.56	2.94
	stan.dev.	2.71	3.01	1.66
	number	16	4	12
CONG.	- mean	2.90	2.98	2.87
	stan.dev.	1.37	1.02	1.26
	number	14	9	5

Table 4.2 Estimates of sample population sizes required for infiltration variables

A standard test is used to determine sample sizes required to get values within 5% and 10% of the mean at the 95% confidence level

Use the formula

$$n = \left(\frac{S \times t(_{ns-1})}{d}\right)^2$$

where n = the required sample size
S = the standard deviation of the present sample
$t(_{ns-1})$ = the value for student's 't' for ns-1 number of samples at the 95% confidence level
ns = number of samples in the pilot study
d = the maximum tolerable sampling error

1 Infiltration rates on marl soil

(a) Answer within 5% of mean
S = 16.83
ns = 16
$t(_{ns-1}) = 2.131$
d = 1.28

$$\left(\frac{2.131 \times 16.83}{1.28}\right)^2 = 785.1$$

(b) Answer within 10% of mean
S = 16.83
ns = 16
$t(_{ns-1}) = 2.131$
d = 2.13

$$\left(\frac{2.131 \times 16.83}{2.56}\right)^2 = 196.3$$

2 Infiltration rates on conglomerate soil

(a) Answer within 5% of mean
S = 10.18
ns = 14
$t(_{ns-1}) = 2.16$
d = 1.06

$$\left(\frac{2.16 \times 10.18}{1.06}\right)^2 = 430.3$$

(b) Answer within 10% of mean
S = 10.18
ns = 14
$t(_{ns-1}) = 2.16$
d = 2.11

$$\left(\frac{2.16 \times 10.18}{2.11}\right)^2 = 108.6$$

falling and constant head infiltrometers reflect different antecedent soil moisture conditions, however the two sets of data are discussed separately.

The falling head infiltrometer was used alongside the runoff troughs to compliment measurements of spatial variations in overland flow production. On site 1 there is a gradual change in soil type from conglomerate to marl, with troughs 8 and 9 on the conglomerate and the remainder on the marl (figure 2.5). The infiltration rates (A.I.) and storage values (A.S.) shown in table 4.3 do not indicate any trend or change in the data downslope. This is surprising considering firstly the change in soil cover and secondly Scoging's (1982) findings of a decrease in infiltration downslope on marl.

Analysis on site 2 is difficult because three readings were taken between 4-9 December 1982 and four between 12-14 April 1983 so that both sets of data have different antecedent soil moisture conditions although this should affect the storage values and not the final infiltration rate. The December measurements took place between 8 and 15 days after the storm of 26 November with 23mm of rainfall. These give final infiltrabilities of 14.17cms/hour and 10.59cms/hour at five and ten metres from the watershed (S2T5m and S2T10m) but 40.64cms/hr towards the lower part of the slope near trough 2. The results for S2T2 are misleading because the test was only run for 32 minutes rather than 1 hour. Even so infiltration seems to be greater towards the gully head on the conglomerate soil. This is supported to a small extent by the four April values (S2T6, S2T5, S2T4, and S2T3). These were measured 21-23 days after a 2.3mm storm and show a slight increase in final infiltration from 16.84-18.38cms/hour. This suggestion that infiltration increases downslope on the conglomerate soil is corroborated by the sediment yield data (Chapter 5) and Scoging's (1982) data of infiltration rates on coarse-grained soils near Ugijar.

Small scale variability is analysed using data from the constant head infiltrometer for three locations, two on marl and one on the conglomeratic soil (table 4.4). An analysis of variance test is used to examine the null hyposthesis that there is no significant

<u>Table 4.3</u> Actual and calculated results for downslope infiltration runs using a falling head infiltrometer.

Location	Model	r^2	a	b	F	A.I.	C.I.	A.S.	C.S.	Sig	Sorp
						cms/hr	cms/hr	cms	cms	%	
S1T9*	1	0.203	0.154	0.026	2.03	10.45	10.32	0.00	0.18	NS	1.77
S1T8*	2	0.877	0.243	0.492	99.91	15.31	15.10	1.68	7.52	99	2.43
S1T6	1	0.479	0.645	-0.197	32.15	18.28	17.29	3.72	5.82	99	3.38
S1T5	1	0.244	0.318	-0.074	11.96	13.43	14.12	1.57	1.71	99	2.39
S1T4	1	0.914	0.619	-0.329	127.24	11.56	9.67	3.45	3.88	99	1.92
S1T1	1	0.099	0.485	-0.024	4.08	26.26	26.34	0.74	1.02	NS	4.21
S2T5m	2	0.903	0.232	0.464	83.94	14.17	14.40	1.78	8.76	99	2.40
S2T10m	1	0.775	0.382	-0.196	34.36	10.59	10.27	2.06	2.70	99	1.78
S2T6	2	0.901	0.207	0.277	281.09	16.84	16.49	3.16	3.05	99	2.77
S2T5	2	0.879	0.283	0.281	276.32	17.81	17.27	1.19	3.26	99	2.93
S2T4	2	0.902	0.210	0.433	303.13	12.19	13.02	4.81	5.03	99	2.57
S2T3	2	0.877	0.306	0.346	286.76	18.38	18.68	3.62	3.75	99	3.22
S2T2	1	0.395	1.097	-0.100	7.18	40.64	43.69	7.68	4.19	NS	5.97

r^2 - coefficient of correlation
a and b - parameters of models 1 and 2
F - value for the 'F' test
A.I. - actual infiltration
C.I. - calculated infiltration
A.S. - actual storage
Sig - significance level
Sorp - sorptivity
* samples on site 1 but on conglomeratic soil (figure 2.5)

C.S. - calculated storage

Table 4.4 Actual and calculated results for infiltration variability tests using a constant head infiltrometer.

Location	Model	r^2	a	b	F	A.I.	C.I.	A.S.	C.S.	Sig	Sorp
						cms/h	cms/h	cms	cms	%	
Marl											
T5R1	2	0.649	0.331	1.696	105.56	23.85	21.53	5.19	22.01	99	2.25
R2	1	0.840	4.515	-0.642	367.79	20.29	19.59	26.98	35.03	99	6.63
R3	2	0.807	0.134	1.890	167.72	4.78	9.91	10.23	21.83	99	2.27
R4	2	0.562	1.174	3.216	86.07	56.07	73.65	27.23	37.79	99	7.56
R5	2	0.824	0.418	2.377	275.85	25.44	27.43	11.02	20.96	99	3.18
R6	1	0.634	6.296	-0.484	166.31	52.06	52.09	39.96	56.34	99	10.23
T4D1	1	0.746	3.597	-0.369	220.32	40.47	47.61	20.33	30.12	99	7.08
D2	1	0.676	5.692	-0.470	60.63	53.77	49.91	28.12	52.92	99	7.27
D3	2	0.573	0.230	0.766	49.30	11.15	14.56	2.80	5.73	99	1.24
D4	2	0.676	0.444	1.549	73.64	28.61	28.21	0.69	19.24	99	2.64
D5	2	0.570	0.494	1.060	102.21	19.59	30.72	15.26	11.61	99	3.19
D6	2	0.930	0.040	1.101	573.50	3.50	3.50	3.87	8.39	99	1.18
Conglomerate											
T2A1	2	0.755	0.262	1.383	129.10	17.61	17.07	2.16	9.46	99	1.35
A2	2	0.764	0.647	1.242	61.87	27.84	40.09	8.79	7.81	99	3.29
A3	2	0.525	1.166	1.534	165.95	39.09	71.47	26.68	13.60	99	4.98
A4	1	0.590	1.559	-0.415	6.11	20.80	17.11	3.95	10.69	99	2.28
A5	2	0.661	0.730	2.543	132.16	33.90	46.33	18.96	24.11	99	4.66
A6	1	0.320	0.582	-0.521	165.39	-	4.14	0.00	4.99	95	1.07

difference in infiltration and storage values between each site. Table 4.5 shows that the null hypothesis is accepted so that no distinction is made between the soil types for this data set. This is surprising considering the very different observed soil structures and profiles described in Chapter 2. The standard deviations for all the parameters are high showing there is considerable variation in values within groups (table 4.5).

A large number of infiltration experiments have been undertaken on marls and other soil types in southeast Spain (summarised in table 4.6). Thornes (1976) found infiltration rates on decalcified marls of 0.02 to 0.06 cms/hr and rates on unweathered marls of 0.3-1.6 cms/hour using a ring infiltrometer. Scoging (1982) measured infiltration rates using both sprinklers and infiltrometers for several soil types. Her infiltrometer values for infiltration rates are nearer to those reported here with means of 10.08 cms/hr on marl and 39.3 cms/hr on sandy soil, but the sprinkler values are an order of magnitude less with mean values of 1.00 cms/hr on marl and 2.16 cms/hr on sandy soils. Furthermore the variability within soils was greater for the sandy soil. Harvey (1982) measured cumulative infiltration rates on intensively gullied marls and using his final infiltrabilities for a 30 minute experiment, I estimate storage values of 0 to 3.7cms only for the first 30 minutes and infiltration rates of 2.4 to 22.2 cms/hr on marls during very dry conditions. Thornes and Gilman (1983) analysed infiltration characteristics for 59 samples taken on cleared marl bedrock throughout south east Spain. They found that the average storage value is 2.85 cms and the final infiltrability is 0.203 cms/hr however they too note the large standard deviations.

Recent studies have been made of infiltration characteristics within a small marl depression in the Rio Mula basin, Murcia, to examine variations in infiltration with surface conditions (López Bérmudez 1985) using both cylinder infiltrometers and sprinklers. The first experiment uses the constant head infiltrometer, and some preliminary results given in table 4.7 show that there is a marked progression in infiltration rates from low values on unweathered marl through weathered, stone covered and vegetated marl, which indicates the very wide range of infiltration rates possible

Table 4.5 Analysis of variance between infiltration rates and storage on 2 marl and 1 conglomeratic soil lithology

Infiltration Rates

	Marl 1	Marl 2	Congl. Soil
n	6	6	5
$\bar{X}$	30.42	26.18	27.85
sd	19.77	18.74	8.91
ΣX	182.49	157.09	139.24
ΣX^2	7504.64	5867.91	4195.06

TSS = 17567.61 - 13486.39 = 4081.22	17		
SS = 13540.87 - 13486.39 = 54.48	2	27.24	
SS = TSS - SS = 4026.74	14	287.62	0.095

Calculated F values = 0.095 is not significant by 95% where $F_{2,14}$ = 3.74.

Storage Values

	Marl 1	Marl 2	Congl. Soil
n	6	6	5
$\bar{X}$	17.81	15.64	11.74
sd	11.02	15.33	9.74
ΣX	106.83	93.81	58.71
ΣX^2	2508.89	2641.68	1069.16

TSS = 6219.73 - 3956.61 = 2263.12			
SS = 4058.20 - 3956.61 = 101.59	2	50.80	
SS = TSS - SS = 2161.53	14	154.39	0.33

Calculated F value = 0.33. In tables $F_{2,14}$ = 3.74. There is no significant difference in storage values between location at the 95% significance level.

Table 4.6 Infiltration rates from other authors

South East Spain

Author	Lithology	Method	Rate cms/hr
Thornes (1976)	Decalcified Marl	Cylinder inf.	0.02-0.06
	Marl	Cylinder inf.	0.30-1.62
	Sands and gravels	Cylinder inf.	3.78-6.13
Scoging (1982)	Marl	Sprinkler	$\bar{x}$ 0.996
		Cylinder inf.	2.94-18.06
	Sandy soil	Sprinkler	$\bar{x}$ 2.16
		Cylinder inf.	26.52-71.22
Thornes and Gilman (1983)	Marl	Cylinder inf.	$\bar{x}$ 0.203
General			
Hills (1970)	Oak woodland	Cylinder inf.	0.0-75.0 $\bar{x}$ 6.8
	Pasture	Cylinder inf.	3.8-18.3 $\bar{x}$ 7.2
	Cultivated land	Cylinder inf.	0.0-92.5 $\bar{x}$ 1.2
Kirkby (1969)	Clay		0.0-0.4
	Silt		0.2-0.8
	Sand		0.3-1.2
Yair and Klein (1973) Israel	Mixed coarse debris	Cylinder inf.	
		Slopes	3.6-18
		Channel	230.4

Table 4.7 Infiltration and storage values measured on marls with different surficial cover

Surface Condition		Infiltration Rate (cms/hr)	Storage (cms)
Unweathered marl	mean	4.84	3.599
	s.d	2.06	2.435
Weathered marl	mean	8.33	3.940.
	s.d.	4.73	3.002
Stone-covered marl	mean	21.32	4.592
	s.d.	4.29	2.487
Vegetated marl	mean	62.18	10.815
	s.d.	32.16	19.726

Table 4.8 Effects on predictions of infiltration rate and storage following the 'smoothing' of the data.

Location	M	r^2	a	b	F	Inf Rate cms/hr	Storage cms
Marl							
T5 R4	2	0.672	1.199	2.575	49.16	74.51	30.94
R6	1	0.926	5.818	-0.421	211.28	62.30	49.62
T4 D5	2	0.859	0.444	1.952	133.77	28.59	27.32
Conglomerate Soil							
T2 A3	2	0.818	0.998	2.751	116.66	62.63	23.61
A4	2	0.986	0.304	1.518	1025.89	19.73	8.31
A5	2	0.939	0.609	3.660	279.06	40.22	34.94

within a few square metres depending on the nature of the soil surface. A second experiment using a sprinkler infiltrometer gives infiltration rates on the unvegetated marl of 1.27 to 1.46 cms/hour, but on vegetated marl a mean rate of 13.74 cms/hour with a range of 15-29 cms/hour. These values suggest that the main source of variation for infiltration is the presence or absence of vegetation. Furthermore these results are compatible with those measured on the study site with a cylinder infiltrometer, assuming the latter include some infiltration rates measured around small plants.

All these data, although for the same lithology, reflect spatial variations in the marl as well as differing techniques and experimental design (either in terms of instrument or preparation of the ground surface). The most comparable sites are the marl slope studied by Scoging and the other studied by myself which are within a few kilometres of each other. However even here there are notable differences in the soil profile of both sites. On Scoging's site (pers. comm.) the upper horizon of 'soil' extends to 46cms with a mean bulk density of 1.74, below which is poorly differentiated marl bedrock. My site, by contrast, has a more permeable horizon to 15cms with bulk densities of 1.4 lying on top of bedrock with a typical bulk density of 1.6 to 1.8. This probably shows the difference between my site which has been ploughed in the past, and Scoging's site which is 'natural'.

One possible check on the final infiltrability after one hour is to compare it to the saturated hydraulic conductivity of the soil, as the infiltration rate falls towards this value. The saturated hydraulic conductivities of the marl and conglomerate soil were determined in the laboratory (see chapter 6). The values for marl were taken for the bedrock and range between 0.235 to 1.007 cms/hour which are similar to the infiltration rates measured by Scoging (1982) on the soil surface with a sprinkler, and those measured by Thornes (1976) on cleared bedrock using an infiltrometer, but not for the final infiltrabilities measured in this project. The saturated conductivities measured on the conglomerate soils vary between 4.6 to 34.8 cms/hour. This range is sufficient to incorporate final infiltration rates measured by both sprinkler and cylinder infiltrometers on this soil type.

This suggests that the infiltration characteristics measured on the conglomerate may approach the true values. On the marl the effect of cultivation has lead to the development of a highly permeable top horizon as reflected by the much higher infiltration rates than expected. One question to come back to later is the effectiveness of the crust for inhibiting infiltration into this porous horizon and the implications of measuring infiltration with an infiltrometer which has to be inserted into the crust.

Models Of Infiltration

A number of physically and empirically based models of infiltration have been developed by various authors. One of the earliest physical models is the Green and Ampt (1911) equation:

$$i = K\,(H + Z_f + P_f/Z_f)$$

where K = saturated hydraulic conductivity, H = depth of ponded water, Z_f = vertical depth of saturated zone, and P_f = the capillary pressure of the wetting front. This equation has been shortened to:

$$i = A + B\,/\,Z_f \quad \text{and if } Z_f \propto t,\; i = A + \frac{B}{kt} \quad \text{or} \quad A + \frac{B'}{t}$$

where t is the time from the beginning of the run, and A and B are constants. This model was not employed for a long time due to the difficulty measuring the variables, particularly P_f, although it is now more commonly used.

Philip (1957) discussed at length the theory of infiltration and developed a model:

$$i = \tfrac{1}{2}\, s\, t^{-\frac{1}{2}} + A$$

where s equals the sorptivity value, t is the time, and A is the saturated hydraulic conductivity.

note that $t = 0$, $i \rightarrow \infty$ and

$t \rightarrow \infty$, $i \rightarrow A$

One of the earliest empirical models is by Horton (1945) where the instantaneous infiltration rate at time t :

$$(f) = f_c + (f_o - f_c)\, e^{-ct}$$

$(f) \rightarrow f_c$ as $t \rightarrow \infty$ and

$(f) = f_o$ at t_o

where f_o = minimum infiltration capacity at t=0, f_c = minimum infiltration capacity at $t \to \infty$, c = soil constant. This model has been widely used because it is easy to apply, even though it assumes unimpeded water movement, has no physical basis and is only suited to ponded infiltration (Knapp 1978).

Another empirical model was suggested by Kostiakov (1932) which simply takes the form

$$i = A + Bt^m$$

where a, b and m are characteristics of the soil. If m equals $-\frac{1}{2}$ we have the Philips curve. With m equaling -1 we have

$$i = a + (b/t)$$

which has been found to give reasonably good agreement to semi-arid soils and in which a and b are easily determined in the field (Scoging and Thornes 1980).

Two infiltration models were fitted to the data set to determine whether either described the observed infiltration curves or gave good estimates of storage and final infiltration. These were:

Model 1 $\quad i = A_1 \times t^{-B_1}$ and

Model 2 $\quad i = A_2 + (B_2/t)$

where i = infiltration rate (cms/min), t = time from the beginning of the experiment in minutes, and A and B are parameters describing the curve. Model 1 is based on the Kostiakov equation and model 2 on the Green and Ampt equation. The difference between models 1 and 2 is that whilst the infiltration rate falls asympotically to zero in model 1, it falls to a parameter A in model 2 which represents the final infiltrability. A computer program fitting the models calculates the correlation of determination (r^2), F values to test the significance of the fit, A, B, the infiltration rate after one hour (cms/hour), and the storage (cms). These data are given in tables 4.3 and 4.4. The significance of the fitted curves is tested for all samples using the F distribution. Of the 31 best fitted models, 3 relationships are not significant by 0.05% level. Of the remaining 28, 27 are significant by the 0.01% level. The r^2 values calculated using all the data points for each curve show 25 cases have coefficients greater than 0.5, of which 14 have coefficients greater than 0.75. Thirteen of the curves are best fitted by model 1 and 18 by model 2. On the marl each model fits eight cases. On the conglomerate soil model 2 provides the best fit for ten of the 15 examples.

A better test of the aptness of the models is to compare calculated and actual values of infiltration rates and storage shown in tables 4.3 and 4.4. The relative success of the models is assessed by the percentage difference of the calculated to the actual values where:

$$\% \text{ difference} = \frac{\text{Actual value} - \text{Calculated value}}{\text{Actual value}} .$$

For the falling head infiltrometer data the percentage difference is 4.25% (σ = is 4.27) between the actual and calculated final infiltration rate, and 93.2% (σ = 137.7) for actual and calculated storage values. For the constant head device the mean difference between infiltration rates is 26.9% (σ = 31.0) and for storage values a mean of 253.1% (σ = 634.9). Thus the models estimate the infiltration rates after one hour better than the storage values, and data from the falling head mechanism better from the constant head mechanism.

Some infiltration curves for both types of infiltrometers representing models 1 and 2 are drawn in a series of figures 4.1, 2, 3, and 4. These show, as expected that the curves for the falling head infiltrometer tend to be very smooth whilst those for the constant head infiltrometer have a more pronounced fall in the early stages of the curve, but are also very spikey. This is due to three possible causes.

1 the difficulty of maintaining a constant head manually, although this improves with practise,
2 the small time intervals between measurements,
3 some variability in the actual infiltration rate.

One would expect that as some of the irregularity is due to operational error, that the curves would normally be smoother. Although the curves from the constant head infiltrometer may be more variable than those from the falling head infiltrometer as the former are more sensitive to actual variations in the infiltration rate. One way of "smoothing" the curves is to calculate the infiltration rate for longer time periods. This procedure was done for six of the worst fits numbered R4, R6, D5, A3, A4 and A5. The models now fit the curves better as the amount of variation has been decreased, showing that the shape of the

Figure 4·1 Infiltration curves for model 1 using the falling head infiltrometer

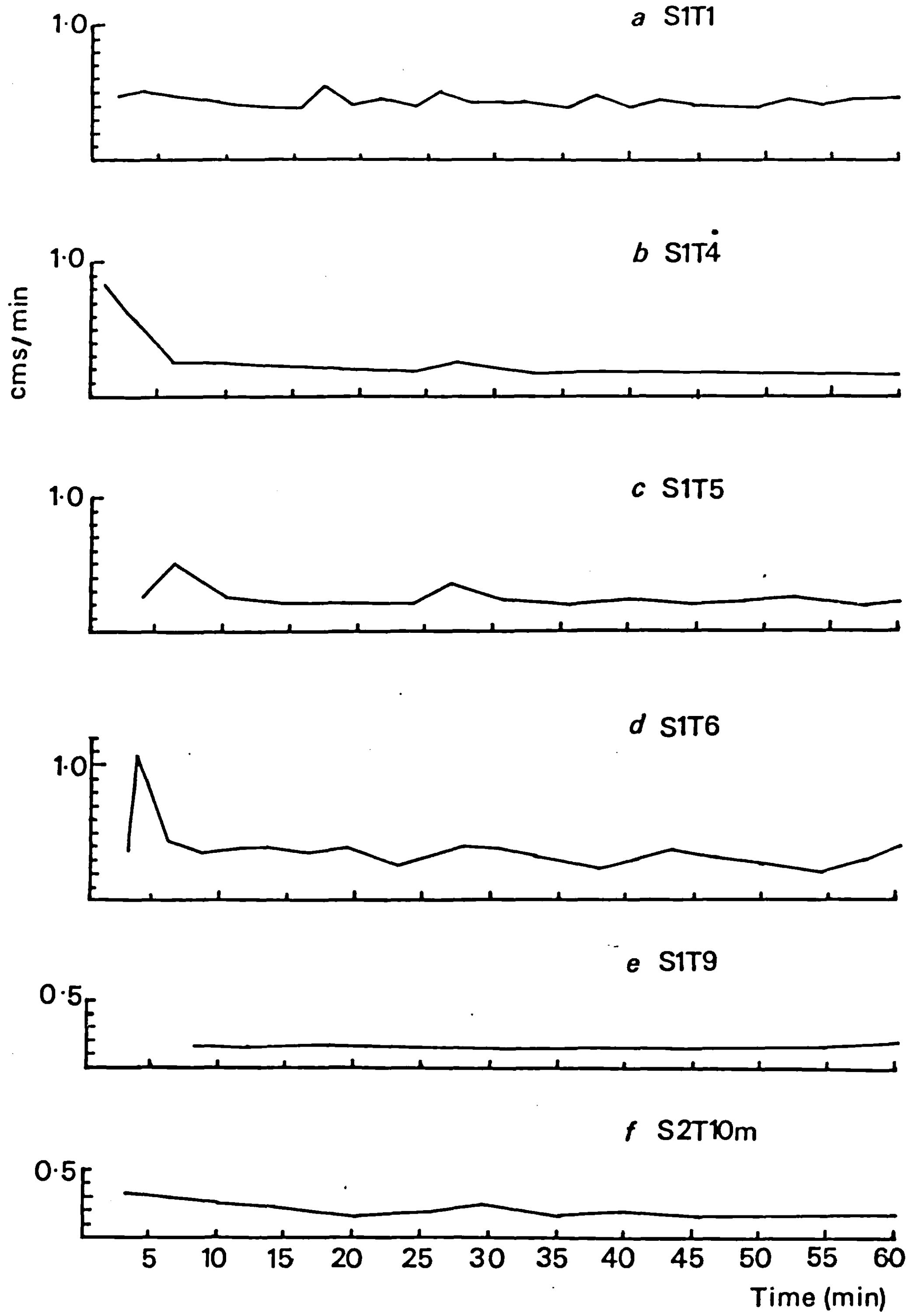


Figure 4·2 Infiltration curves for model 2 using the falling head infiltrometer

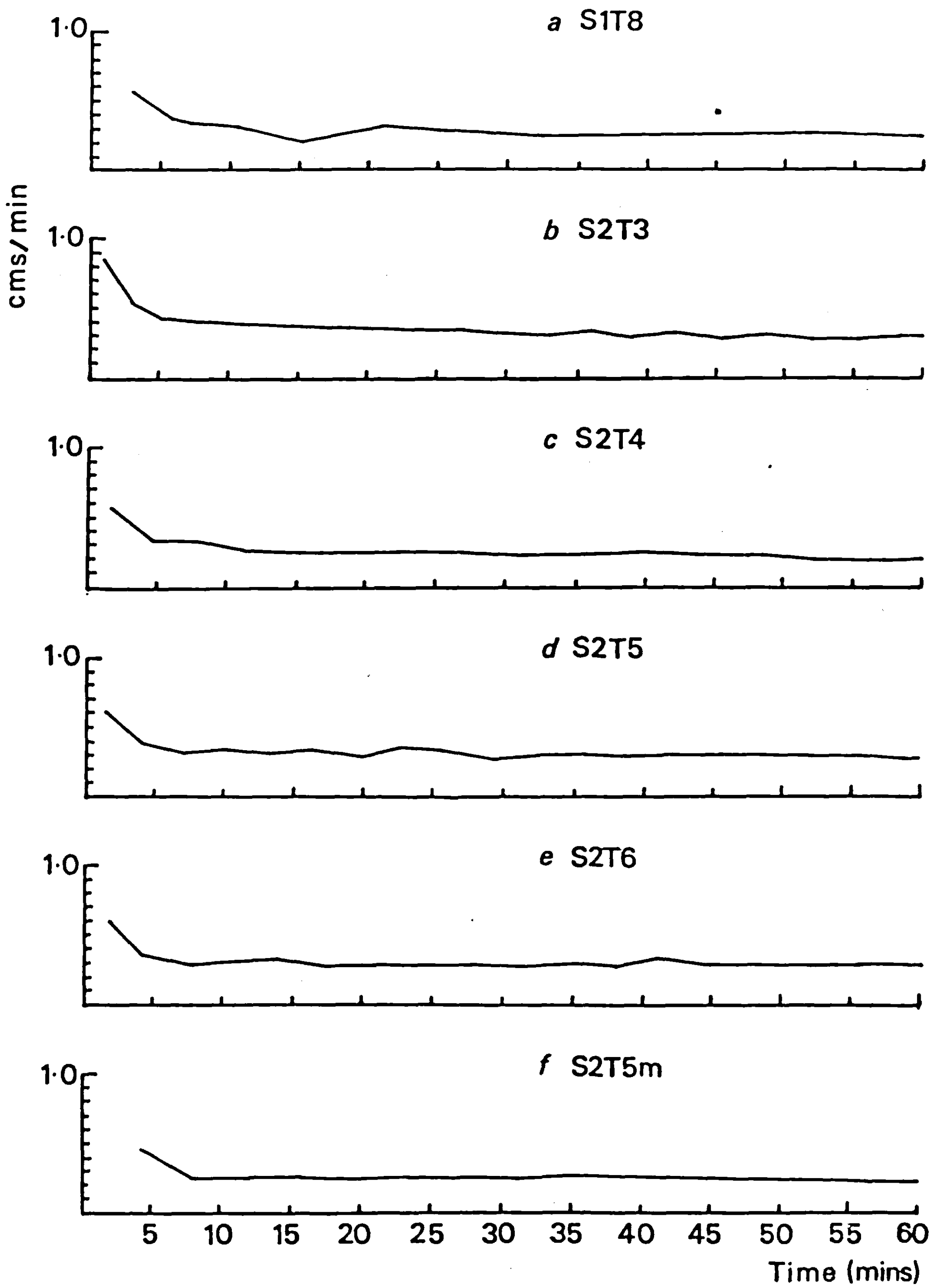
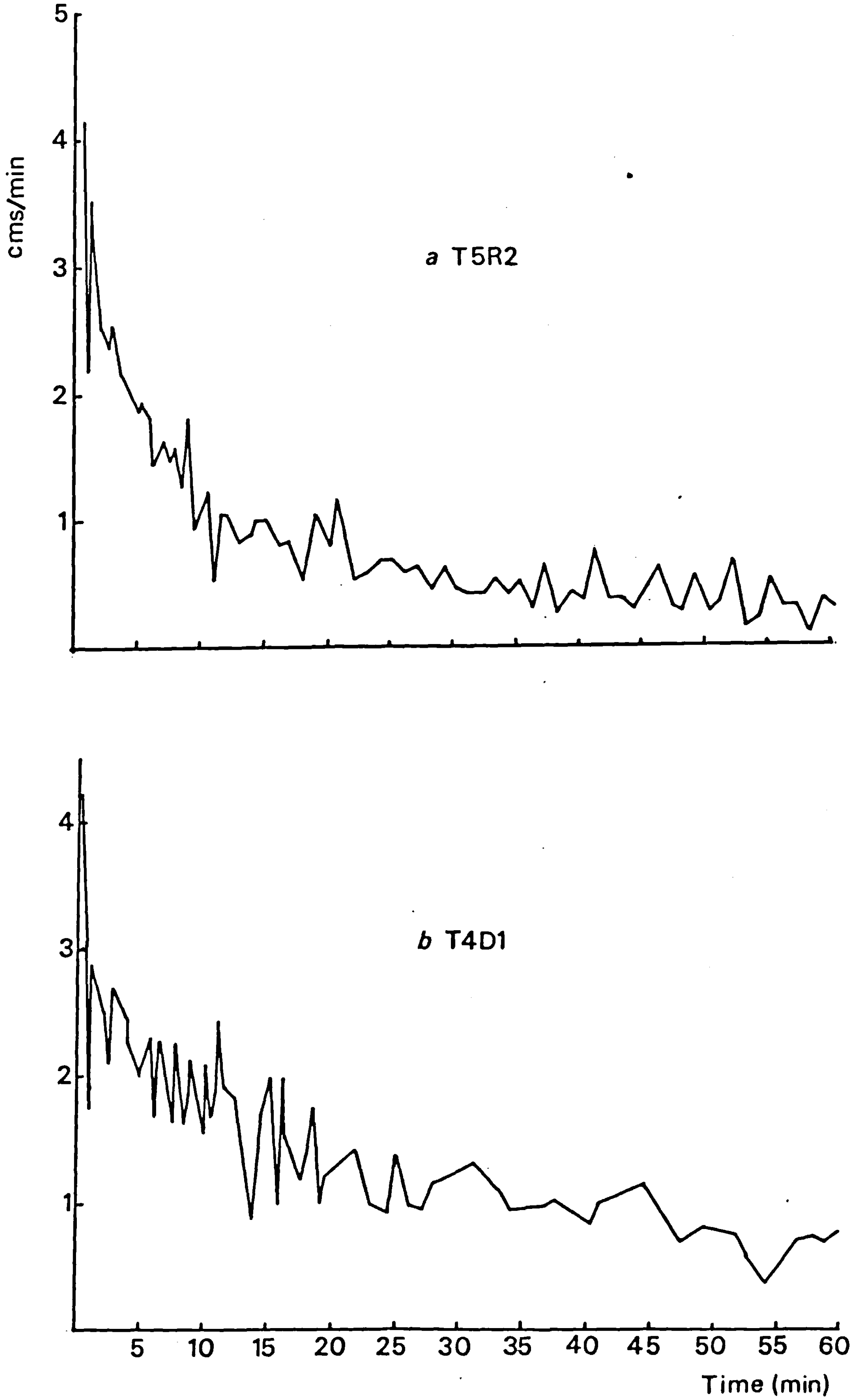


Figure 4·3 Infiltration curves for model 1 using the constant head infiltrometer



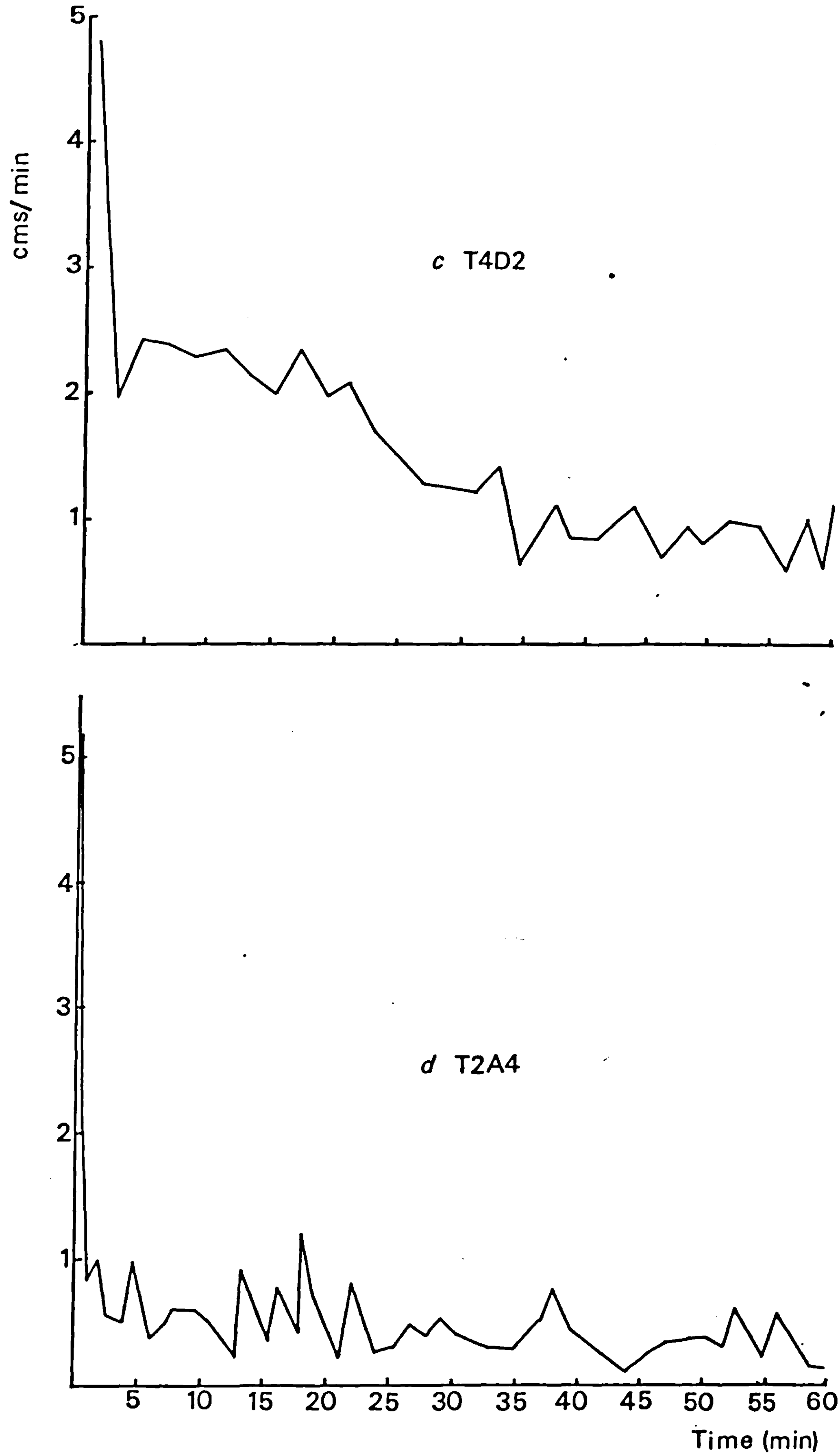
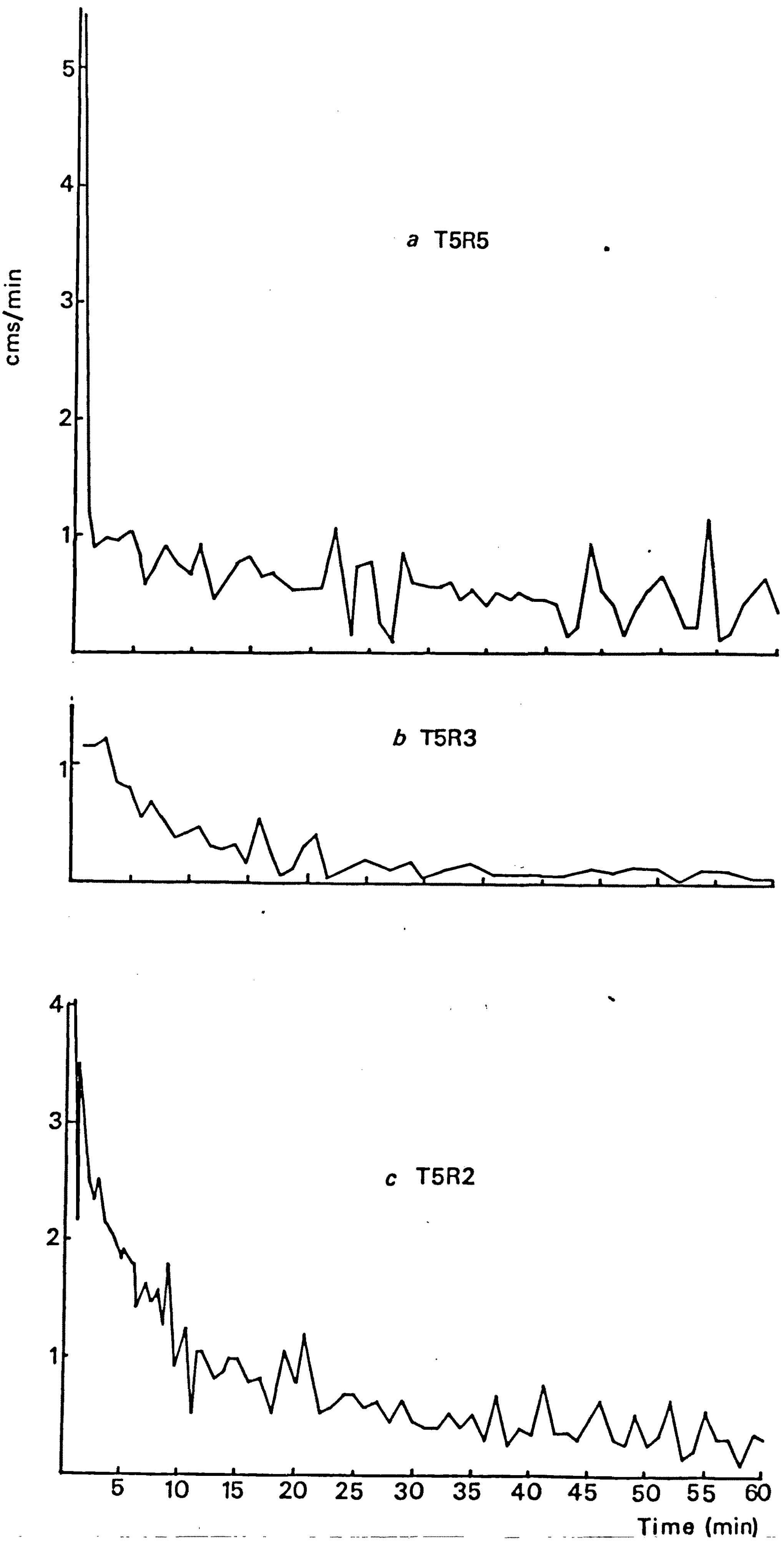
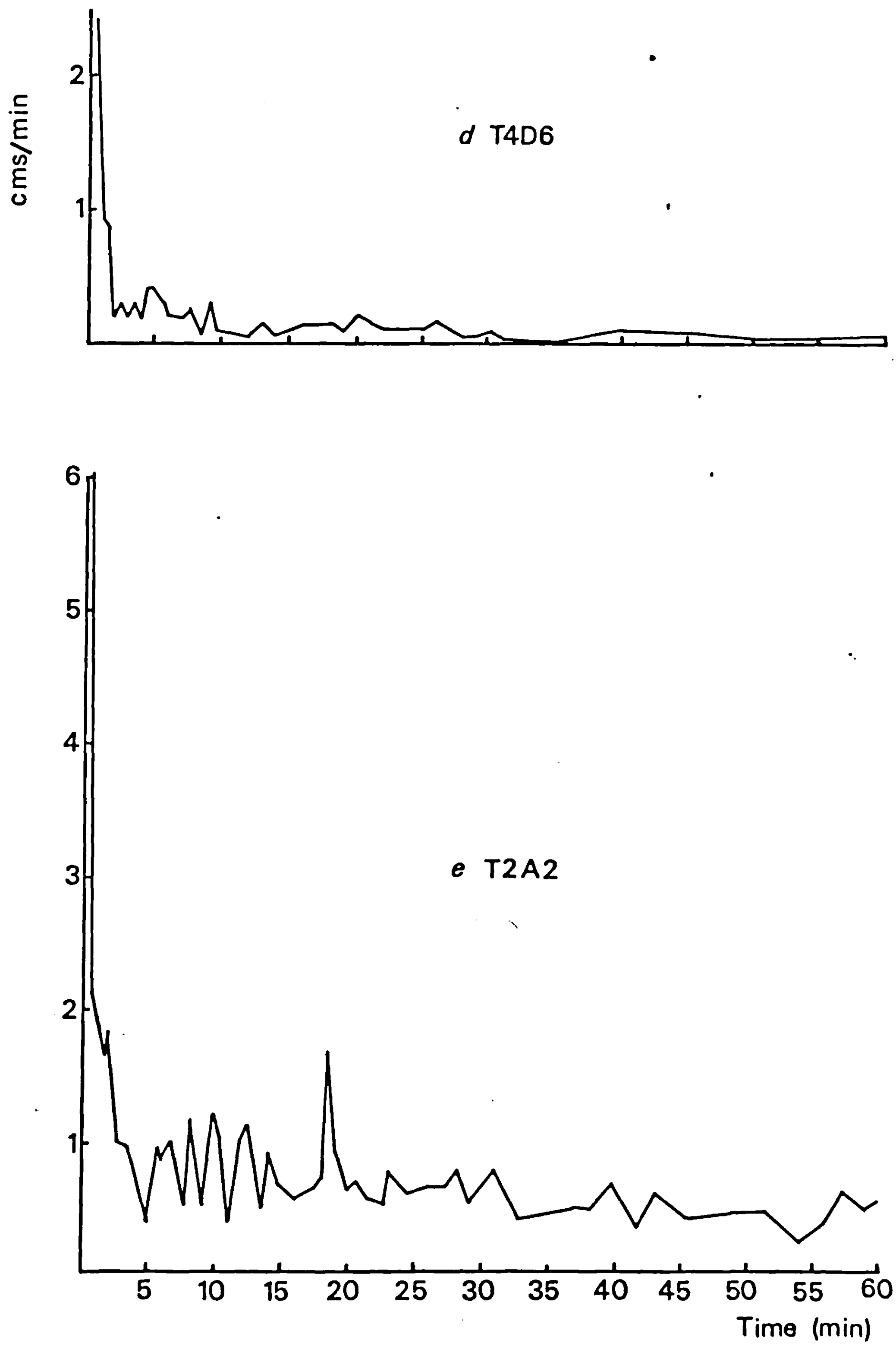

cms/min
5
4
3
2
1
c T4D2
5
4
3
2
1
d T2A4
5
10
15
20
25
30
35
40
45
50
55
60
Time (min)

Figure 4·4 Infiltration curves for model 2 using the constant head infiltrometer



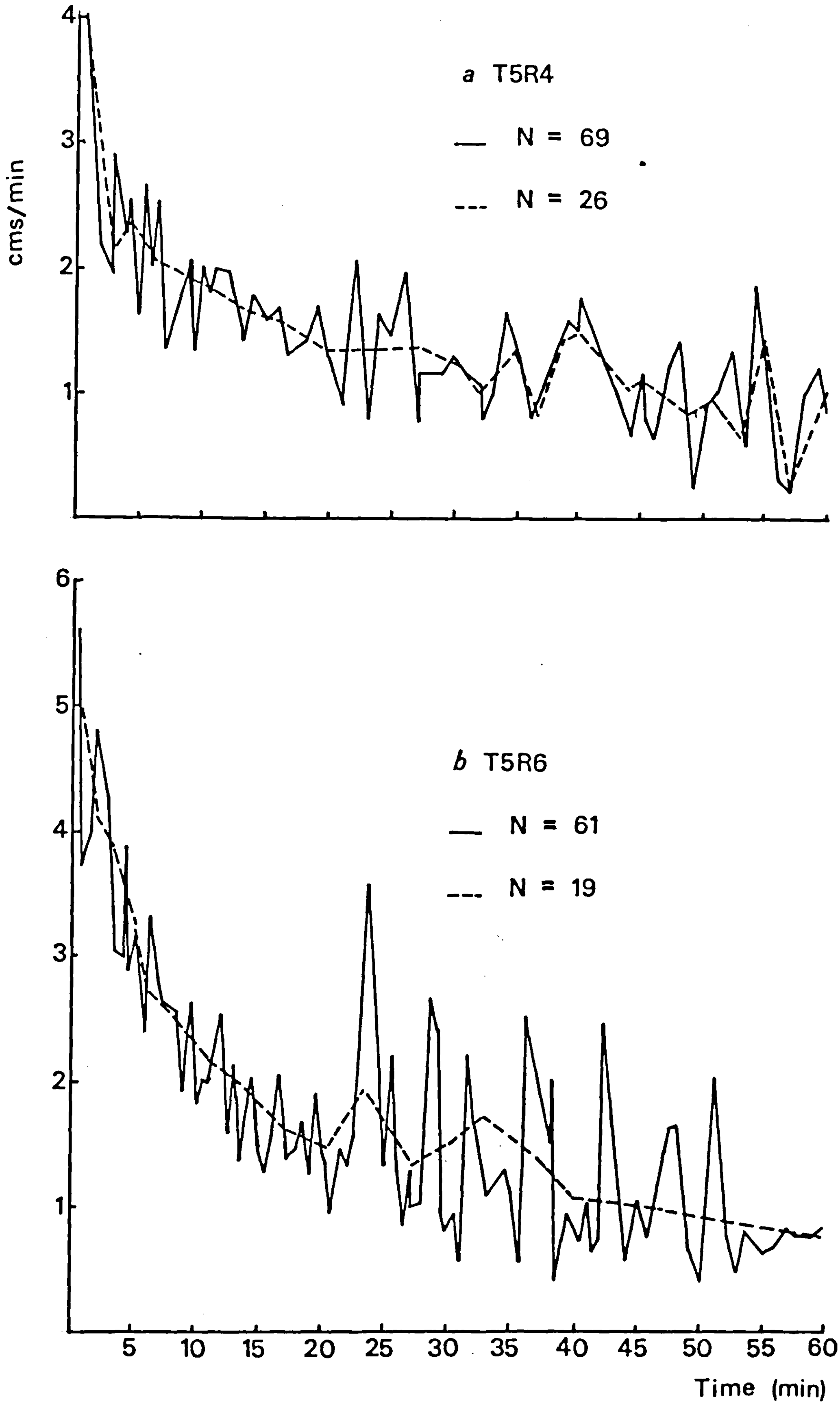

cms/min
2
1
d T4D6
6
5
4
3
2
1
e T2A2
5 10 15 20 25 30 35 40 45 50 55 60
Time (min)

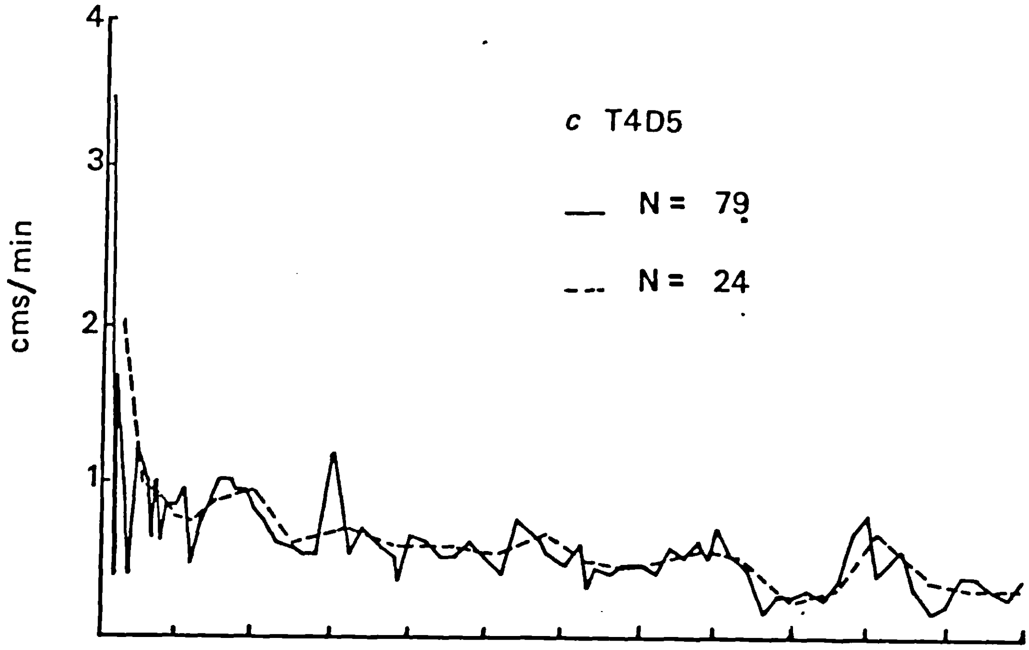
model curve is similar to the smoothed data curve as shown in figure 4.5 and table 4.8. Values for R^2 rise for example from 0.634 to 0.926 for R6, and 0.590 to 0.986 for A4. Calculated values for storage and infiltration rates are improved for four out of six cases, but the difference between actual and calculated values is still fairly high, so the smoothing procedure does not improve the predictive performance of the models.

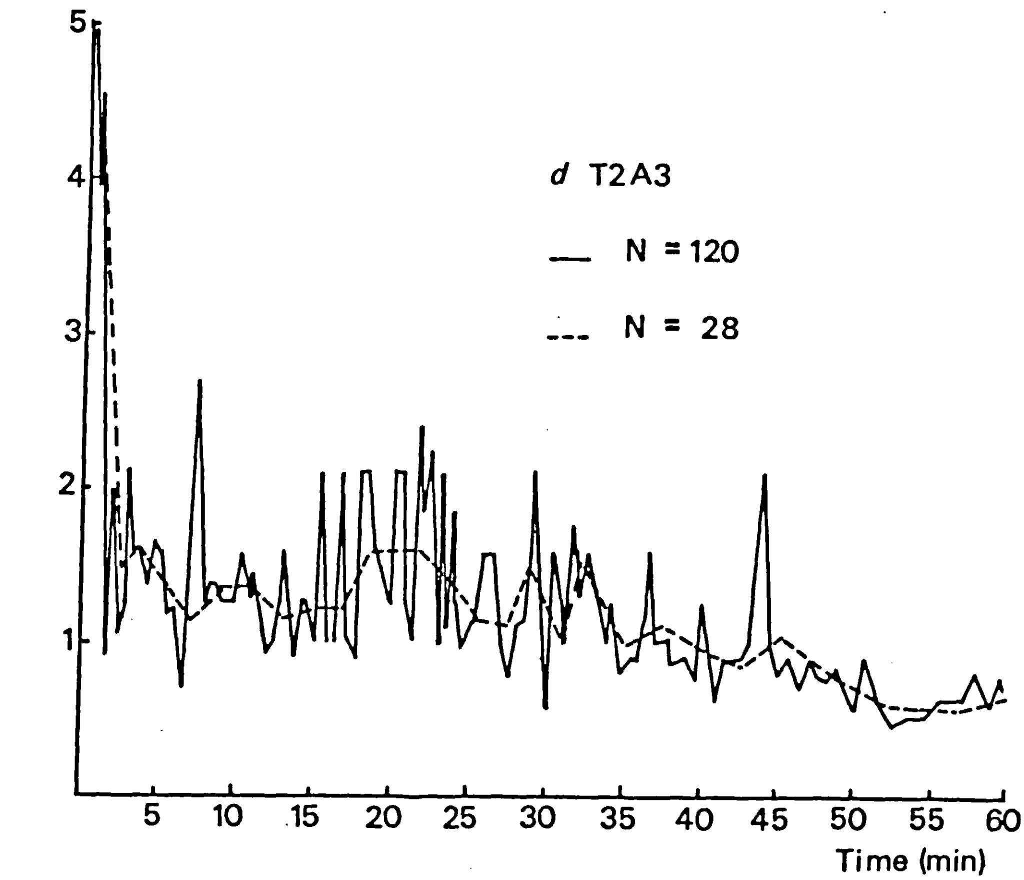
The models have shown that the infiltration curves do conform to standard shapes, and that they can be used to estimate the infiltration rate after one hour and the storage (to a lesser extent). Model 1 is surprisingly successful considering that Scoging and Thornes (1980) found that model 2 forms dominated infiltration curves on the same lithologies elsewhere. However the models do not show any differences between the soil types.

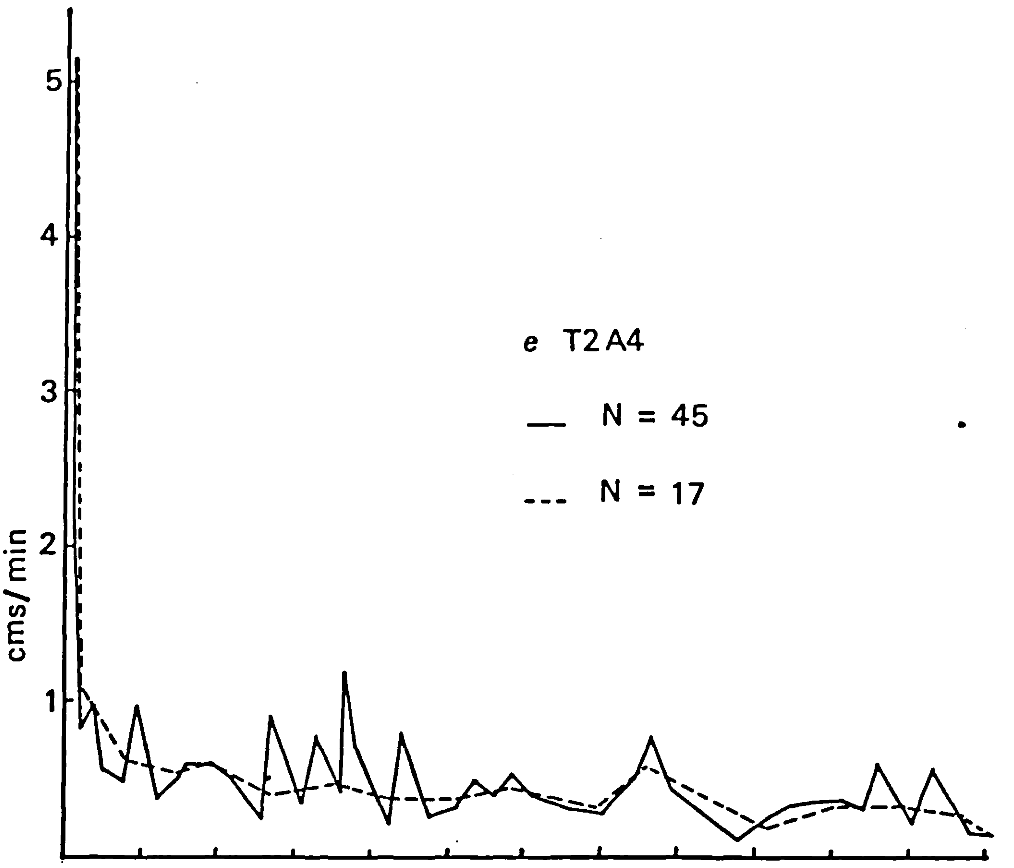
Sorptivity values can be used to calculate the final infiltration rate according to the Philips equation (Dunin 1976). The sorptivity value is derived from a plot of the square root of time in minutes (x axis) against cumulative infiltration (y axis) (figure 4.6). The sorpitivity is the gradient of the resulting straight line in the initial part of the curve. This represents one dimensional flow within the confines of the infiltration ring as the wetting front descends from the soil surface towards the bottom of the infiltrometer. Final infiltrabilities after one hour are calculated for the conglomerate soil for both types of instrument, assuming that the saturated hydraulic conductivity varies between 0.0548 and 0.5802 cms/hr (Chapter 6). The results, shown in table 4.9, indicate that depending on the value for the saturated hydraulic conductivity the final infiltration can vary between 5.89 to 50.77 cms/hr. The final infiltration rates were not calculated for the marl as the saturated hydraulic conductivities were for horizon 2 and the sorptivity values were calculated for horizon 1.

Figure 4.5 To show the effect of smoothing on the infiltration curves measured using the constant head infiltrometer

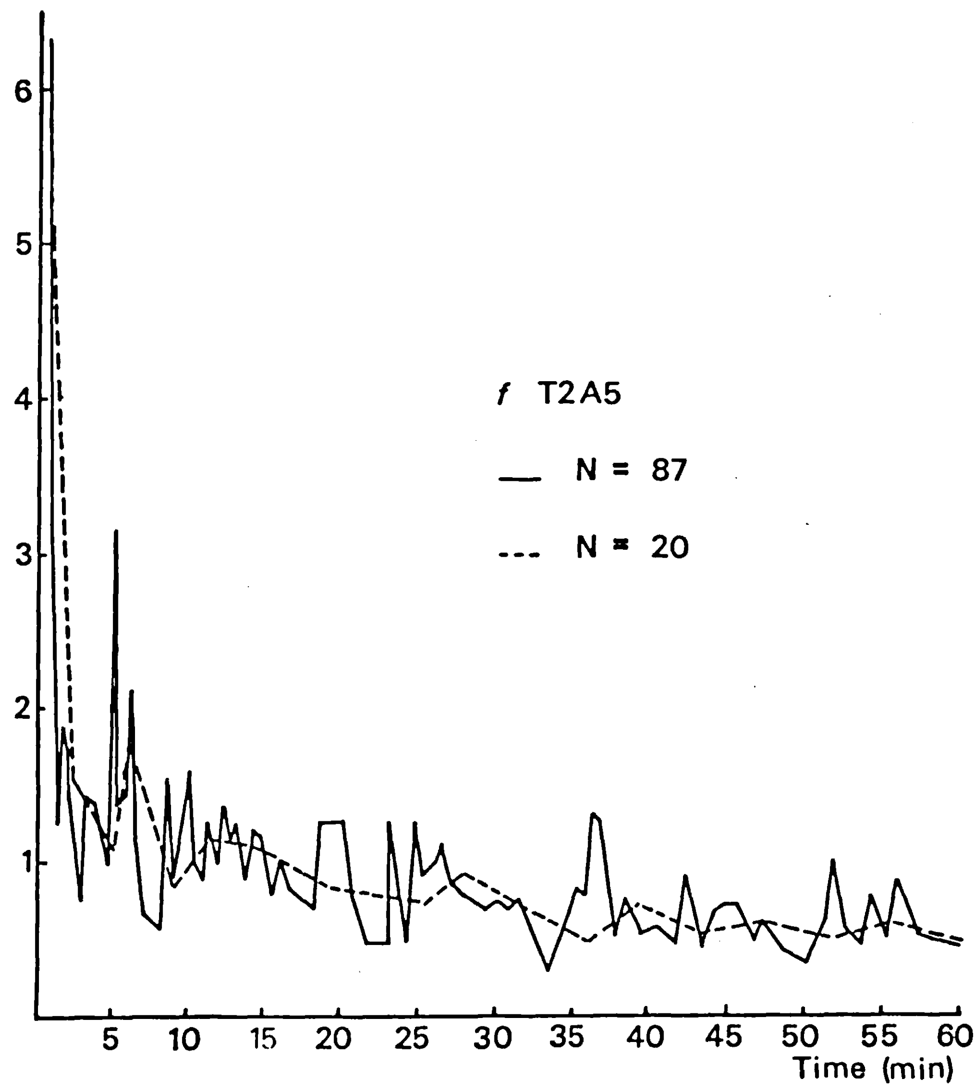



c T4D5
N = 79
N = 24
cms/min
1
2
3
4


d T2A3
N = 120
N = 28
1
2
3
4
5
5
10
15
20
25
30
35
40
45
50
55
60
Time (min)



cms/min
5
4
3
2
1
e T2A4
N = 45
N = 17



6
5
4
3
2
1
f T2A5
N = 87
N = 20
5
10
15
20
25
30
35
40
45
50
55
60
Time (min)

Figure 4·6 A plot of cumulative infiltration against time

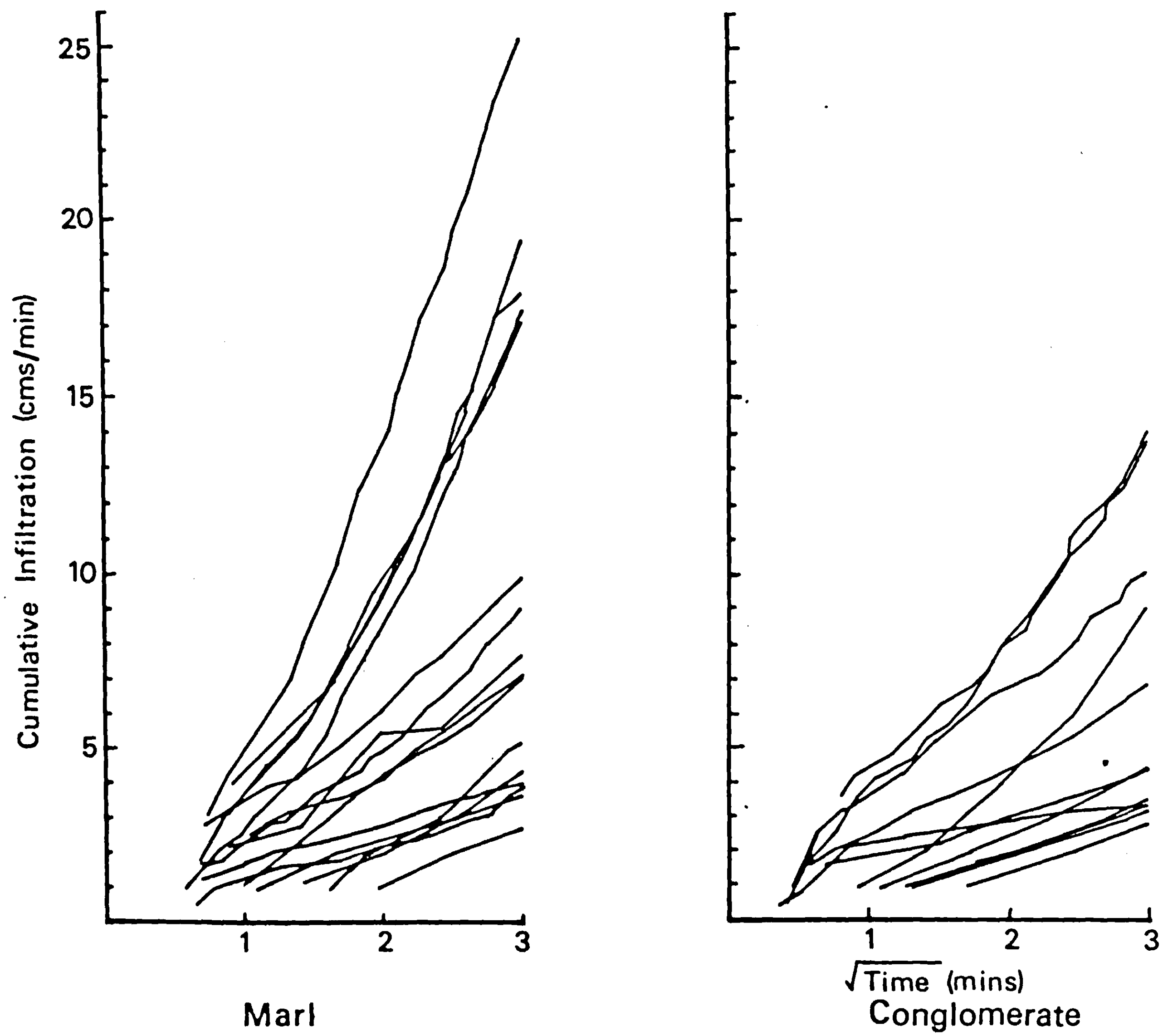


Table 4.9 Infiltration rates estimated for conglomerates using the Philips equation for the maximum and minimum value for the saturated hydraulic conductivity measured on the conglomerate

Maximum Ksat on conglomerate = 0.5880 cms/min
Minimum Ksat on conglomerate = 0.0548 cms/min

Number	Sorptivity	Infiltration Rate cms/hour
T2A1	1.350	8.52 - 40.04
T2A2	3.290	16.03 - 47.56
T2A3	4.121	19.25 - 50.77
T2A4	1.958	10.87 - 42.40
T2A5	3.602	17.23 - 48.75
T2A6	0.673	5.89 - 37.42
Mean	2.499	
σ	1.373	

Philips equation

$$i = \frac{1}{2} s t^{-\frac{1}{2}} + A$$

4.2 Infiltration and Implications For Overland Flow Production

The aims here are to go from the infiltration data to estimating overland flow production, and to compare this with measured values. As only one storm was measured during the field study, data on overland discharge is limited, increasing the dependence on models.

One of the simplest models of overland flow production is the precipitation excess model where runoff equals precipitation minus the infiltration (Horton 1945). An alternative is a storage based model where runoff commences once the available subsurface storage is filled (Kirkby 1978, Thornes and Gilman 1983). In semi-arid areas Horton's model is widely accepted, as rainfall intensities are assumed to be greater than infiltration rates. However storage based models are appropriate where storage capacities are low (despite high infiltration rates) and for long duration storms.

Figure 4.7 shows the storm intensity traces for 5/6 November and 26 November 1982 from the station at Mecina Bombaron with some infiltration intensities marked on. The infiltration rates measured on the slopes of c20 cms/hr (200 mm/hour) are far higher than the rainfall intensities, and would not produce surface flow from a simple rainfall excess model (even if the hourly rainfall intensity masks very short duration high intensity rainfall), unless the storm is long enough to fill up the available storage in the soil.

It is assumed for the time being that infiltration rates on the marl and conglomerate soil are more in line with the sprinkler measurements by Scoging (1982), or infiltrometer measurements by Thornes (1976). The amount of rainfall excess (cms) is calculated for both storms and soil types (table 4.10). On 5/6 November between 3.3 to 12.5 cms depth (or 33 to 125 l/m^2) runoff would have been produced on the marl during the entire storm depending on the infiltration rate chosen and up to 0.4cms (or 4 l/m^2) on the conglomerate soil. The same model calculates that on 26 November about 4 l/m^2 of runoff is produced on the marl with no runoff on the conglomeratic soil. Quantities of discharge

Figure 4·7 To show the relationship between the storms on 5-6 November and 26 November, and infiltration rates for marl and conglomerate measured by Thornes (1976) and Scoging (1982)

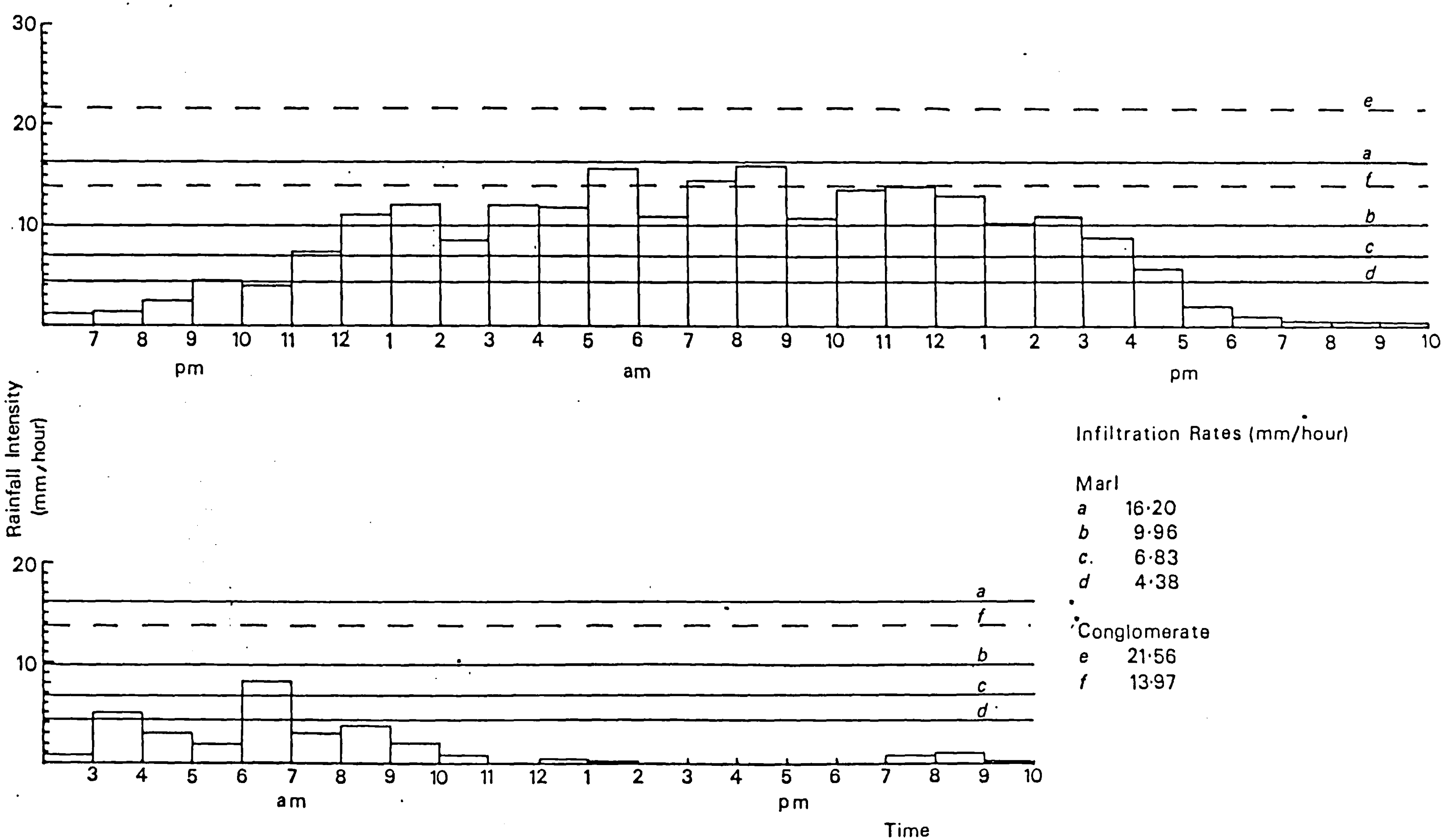


Table 4.10 Rainfall excess for two storms at Mecina Bombaron calculated by the difference method after Horton using different final infiltrabilities from Thornes (1976) and Scoging (1982)

Inf	5/6 November 1982		26 November 1982	
cms/hr	cms	$1/m^2$	cms	$1/m^2$
Marl				
1.620^t	0.000	0.00	0.000	0.00
0.996^s	3.312	33.12	0.000	0.00
0.683^s	8.155	81.16	0.370	3.70
0.438^t	12.498	124.98	0.444	4.44
Conglomeratic Soil				
2.156^s	0.000	0.00	0.000	0.00
1.397^s	0.389	3.89	0.000	0.00

t = Thornes (1976)
s = Scoging (1982)

Table 4.11 Rainfall excess for two storm intensities on two lithologies from Thornes and Gilman (1983) for a one hour period.

Lithology	Storm Intensity 6.0 cms/hr	Storm Intensity 12.0 cms/hr
Marl	3.165	9.150
Conglomerate	0.360	6.360

measured on the slope are only available for 26 November and give mean values of 6.21 litres on marl (σ = 1.60), and 2.86 litres (σ = 0.83) on the conglomeratic soil (Chapter 5). This gives a range of 4.6 to 7.8 litres on the marl and 2.0 to 3.7 litres on the conglomerate for one standard deviation either side of the mean. This suggests that the estimates for the marl appear appropriate, however it is recalled that the runoff measured on the site is for undetermined areas, whereas the model results are for a metre squared. The infiltration rates on the conglomerate are too high to obtain surface flow (even when allowing for the deviation in the infiltrometer values), by a simple excess model. This is reinforced by the fact that the rainfall intensities measured are likely to overestimate values on the site as Mecina Bombaron lies 500m above the study site and rainfall parameters reflect the orographic influence.

Thornes and Gilman (1983) calculate runoff from rainfall excess based on an hour long storm for 2 intensities 6 cms/hour and 12 cms/hour (table 4.11). The storm intensity of 6.0cms/hour has a return period of 100 years in Almeria. Both this and the previous table bring out variations between soil types and the effects of different magnitude events based on their infiltration data.

Annual overland flow estimates are published in Thornes (1976) for the Ugijar area (table 4.12) using Kirkby's model for annual overland flow (q):

$$q = R \times e^{-rc/ro}$$

where R = total rainfall, ro = mean annual rainfall/rainday and rc = amount of daily rainfall lost to run off. The value rc reflects not only different infiltration losses with varying soils, but also the seasonal effects of water subtraction for irrigation, evaporation and vegetation cover. Combined, these show that overland flow values vary considerably with season.

Run off coefficients indicate the proportion of rainfall that becomes runoff and are published for Spanish catchments by the Ministerio de Obras Publicas. Some of these values for catchments in south east Spain are given in table 4.13, together with the main geology in the catchment. The Rambla de Algeciras and Rio Mula (both in the Province of Murcia), and the Rio Jauto (Province

Table 4.12 Annual overland flow estimated from $R \times e^{-rc/ro}$ where R = total rainfall, ro = mean rainfall/rainday and rc = amount of daily rainfall not available as runoff. (Thornes 1976)

Period	rc				
	10	20	30	40	50
Year	386.5	183.0	86.7	41.0	19.4
Winter	363.6	171.4	80.8	38.1	17.9
Jul-Aug-Sept	19.0	9.0	4.2	2.0	1.0
Winter Year	484.8	228.5	107.5	50.8	23.9
Summer Year	76.0	36.0	16.8	8.0	3.6

Table 4.13 Runoff coefficients and estimates of overland flow

(a) Runoff coefficients for some catchments in south east Spain

Catchment	Lithology	Area	Record Period	RO Coeff.
Rio Mula	Marl	156km^2	1943-65	0.10
Rambla de Algeciras	Marl	52km^2	1943-65	0.09
Rio Jauto	Marl	68km^2	1943-69.	0.05
Rio Ugijar	Mixed	120km^2	1943-69	0.14

(b) Estimates of overland flow on marl using the runoff coefficients.

	Ugijar litres/m^2	Mecina Bombaron litres/m^2
26 November	23 mm rain	27.4 mm rain
0.05	1.15	1.37
0.09	2.07	2.47
0.10	2.30	2.74
0.14	3.22	3.84
5/6 November	132.0 mm rain	210.0 mm rain
0.05	6.60	10.50
0.09	11.88	18.90
0.10	13.20	21.00
0.14	18.48	29.40

of Almeria) lie mainly in marl, with runoff coefficients between 0.05 to 0.10. The catchment area of the Rio Ugijar above Las Tosquillas only flows through a small section of marl at the lower end, with most of the basin on quartzite and metamorphic rocks in the Sierra Nevada (Chapter 2), and has a runoff coefficient of 0.14. These coefficients are intended for application at the basin scale and even then are unreliable for runoff prediction as rainfall is variable, and both the rainfall and runoff records are short (perhaps 20 years). Furthermore the coefficients are related to the drainage basin size, and are generally smaller for larger catchments as these have an increased capacity for storage, evaporation, the proportion of producing areas changes, and the effects of vegetation become greater.

Nevertheless the coefficients suggest that 5 to 10% of rainfall becomes runoff over large areas on marl. This would imply total runoff on marl slopes of 1.15 to 2.30 litres/m^2 for the 23mm storm on 26 November, which is in the same order of magnitude as the rainfall excess model and the measured quantities. For the larger storm on 5/6 November (with 132mm at Ugijar), the runoff coefficients produce 6.6 to 13.2 litres/m^2 on marl slopes, which falls below the lowest estimates for rainfall excess (table 4.10). Nevertheless these values compare well considering the logistic drawbacks using runoff coefficients.

4.3 Summary

The infiltration data collected on the site are not significantly different between the soils, although there is a distinction between different shapes of infiltration curve through the analysis using the models. This was counter intuitive, but on reflection the very high rates on marls are probably associated with infiltration through a cracked crust caused by the insertion of the infiltrometer into a layer some 10-15cms deep of loose permeable aggregates. A key issue therefore is the effectiveness and stability of the crust. Firstly areas of uncracked crust may be expected to reduce infiltration rates and consequently increase runoff values on the marl. Following a dry spell, the first part of the storm would fall on a cracked crusted surface with high infiltration rates resulting. During the storm the cracks may close up with the redevelopment of a new crust and the impact of swelling clays. Secondly the survival time of crusts in rainstorms of different intensities and duration is important as the gradual breakdown of the crust in very intense storms (rather than crustal development) may change infiltration rates from relatively low to quite high values.

Unfortunately the effectiveness of crusting on infiltration rates could not be examined within the scope of this work. However judging by discharges measured on the slopes following the storm on 26 November (assuming comparable rainfall parameters, catchment areas, slopes and vegetation cover) the difference between runoff measured on marls and the conglomerate soils reflects the impact of crusting such that runoff is 2-3 times greater on the marl. In view of this, the infiltration measurements by sprinkler systems (Scoging 1982), final infiltration rates for marl bedrock (Thornes 1976), or saturated hydraulic conductivities (Chapter 6) may provide a better insight into the hydrologic response of the two soils.

CHAPTER 5 SURFACE FLOW AND SEDIMENT TRANSPORT

An assessment of the amount, variability and factors affecting both surface discharge production and sediment detachment and transport is an integral part of the project. As surface wash is one of the main causes of gully head recession, measurements of surface wash, spatial variations in production (in particular above the gully head) and the competence of flow (as reflected indirectly by sediment transport) will have a bearing on the headward recession of gullies. The characteristics of sediment transported by flows indicate the relative susceptibility of different lithologies to erosion by surface wash. Theoretical considerations have shown two things. Firstly the relationship between potential sediment transport and actual rates of sediment supply and removal affect the stability of the gully (Carson and Kirkby 1972, and Smith and Bretherton 1972). Secondly the strength of the soil material affects the morphology of the gully and possibly the network and drainage density of gullies (Thornes 1984). Measurements of sediment transported also indicate the rates of denudation which reflect the significance of the soil erosion problem, and can be compared to model values for example from the Universal Soil Loss Equation. Over the long term the patterns of erosion and deposition, and changes in the slope characteristics will affect future discharge, and erosion values, and subsequently future development of the gullies.

5.1 Experimental Design

Discharge and sediment transport were measured using Gerlach troughs connected to water collecting barrels and their use has been described by various authors (De Ploey and Gabriels 1980, McGregor 1980, Morgan 1980, Le Roux and Roos 1982, and Van Asch 1983). In the Easter field trip (March 1982) tipping bucket mechanisms replaced the water bottles and were wired to a data logger to monitor the overland flow hydrographs, but the lack of storms precluded their use.

The troughs were arranged downslope en echelon (figure 2.5) to minimize interference between sites for the unbounded case (De Ploey and Gabriels 1980). On site 1 ten troughs were used at approximately 20m intervals, and six on site 2 at 10m intervals. On site 1, two troughs were placed just above the gully head to increase data density in this important area. Le Roux and Roos (1982) tried to estimate the ability of the troughs to replicate results. They used six pairs of Gerlach troughs en echelon downslope, and found that there was no significant difference in sediment trapped for the two troughs for five out of six pairs. This suggests that Gerlach troughs do replicate sediment yields for given locations.

Gerlach troughs (Gerlach 1967, De Ploey and Gabriels 1980) are designed to measure surface flow only, and minimise the effect of splash transportation. Great care needs to be taken during their installation, otherwise disturbance causes accelerated erosion around the trough. Some discussion has centred on whether Gerlach troughs should be bounded or not. Three questions need to be asked:

1 What are the dominant processes on the slope?
2 How is the data to be expressed?
3 What affect do the Gerlach troughs have on surface wash?

Firstly it is important to consider whether the erosion of the soil and sediment transport are related to meso-scale factors affecting the whole slope such as distance from divide, soil type, or the overall vegetation cover, or micro-scale factors affecting the local area such as local vegetation cover, discharge, or slope. To capture the influence of the meso-scale factors the contributing area of the troughs should extend to the watershed and may be bounded or otherwise. However to examine local factors affecting discharge and sediment transport, the contributing area of the troughs should be limited to the local environment, which necessitates the use of boundaries (Pearce 1976).

Secondly data on sediment detachment and erosion are often expressed as weight/area/time or depth eroded/area/time in order to estimate denudation rates. In either case the contributing area must be known. This can be done by careful surveying (Le

Roux and Roos 1982) or by assuming the average length of overland flow for unbounded areas, but both methods are difficult to employ. Surveying alone is often insufficient to determine catchment areas on a morphological basis (Morgan 1980). On long slopes the length of overland flow will not necessarily equal the distance from divide and furthermore estimates of the average length of overland flow are not only difficult to make but vary with storms, infiltration and runoff characteristics and the antecendent conditions. The most accurate estimate of catchment area is from bounded troughs.

Finally the presence of the troughs and borders will interrupt the slope processes and may enhance or subdue their effect. The standard method for bounding troughs (particularly for long term projects) is to dig a ditch around the plot, insert sheet metal walls, and backfill. This reduces water seepage and sediment inputs and outputs, but disturbs the soil which needs time to "settle down". Once installed the presence of the borders may be sufficient to induce localised scour or deposition so that some instrumental error is introduced to the experiment.

In order to compare the difference between slope and local factors affecting run off and sediment transport, both unbounded and bounded troughs were used. The troughs were unbounded for periods 1, 2, and 3, and bounded for periods 4 and 5. As the study period was short the boundaries employed had to minimise disturbance to the soil because there would not be enough time for the soil to settle. For this project the walls were made with stones and cement along the upper boundary only. These successfully impeded sediment from upslope, but allowed water to drain through the soil under the walls. The error involved with increased water input from seepage was considered to be less important than increased sediment yields from inserting the boundaries into the soil. Lateral walls were not built to reduce the boundary effects. The drainage areas of the erosion plots were estimated as the trough width (0.5m) multiplied by the distance to the upper boundary (3.0m) giving $1.5m^2$. This assumes that within the erosion plot the flow lines of water contributing to the troughs are 0.5m wide and the troughs catch all the water flowing from immediately upslope.

Photograph 5.1 shows the runoff plot on the conglomerate soil for trough 9 on site 1. This shows in the foreground the effectiveness of the upper boundary for halting sediment movement. Note also that the material which has piled up against the wall appears finer than the surrounding soil. Photograph 5.2 is a close up view of the same trough. This (taken in September 1984) gives a good impression of the soil surface with relatively little living vegetation but plenty of litter lying on a very stony soil. Photograph 5.3 shows Gerlach trough number 2 on the marl at the foot of site 1 (see figure 2.5). This illustrates the difference in soil type with an almost stone free surface, but with a lot of litter and lichens.

Gerlach troughs are designed to catch water and sediment from overland flow as entry is by a 1 to 2 cm gap upslope between the lip of the trough and the lid only. The occasions when overland flow occurred on the site were never observed so it is impossible to know whether the quantities of sediment and water caught were representative or not. The lack of scour or depositional features around the troughs suggest that they were successful in catching the sediment supplied to them. It is subsequently assumed that the instrumental errors are negligible compared to the sediment load caught so that the quantities collected approach absolute values. Also it is assumed that as the experimental design for all the troughs was the same, differences in sediment and discharge between troughs represent differences in properties of the contributing area rather than instrumental variation.

The sediment was collected from the troughs and the water discharge from both the troughs and water barrels at the beginning of each field session and after rain. The outlet pipe from the trough to the water barrel was fitted with a fine gauze, which prevented most sediment going into the barrrel. However some very fine material did get through but this was not retrieved because of the problems with allowing the fines to settle or be filtered off in difficult field conditions. This represents a second source of error for underestimating sediment yield.

The sediment caught in the troughs was dried and weighed, with corrections made for organic content by ignition loss tests, and

Photograph 5·1 Runoff plot and Gerlach trough on the conglomerate
S1T9

Photograph 5·2 Gerlach trough on the conglomerate (S1T9)

Photograph 5·3 Gerlach trough on the marl (S1T2)

further corrections for $CaCO_3$ content on the marl samples (Black 1965). Particle size analysis was undertaken for the coarse material greater than 2mm only. Supplementary soil data was collected from the soil surface and within the soil to compare the sediment transported, the sediment left behind and the original sediment matrix in the soil. Surface soil samples 4cms deep were cut in a square and scooped up, whereas depth samples with a volume of 1000cc were taken with a bulk density ring.

The Gerlach troughs and erosion plots were deliberately sited in locations which afforded as much overland flow as possible. Thus in the more densely vegetated areas, for example, the troughs were placed below tracts of relatively open ground and are biased towards overland flow conditions.

Some characteristics of the plots were measured to determine some of the "local" factors affecting sediment yield and discharge. In particular the plot slope, distance from divide, and vegetation cover were measured. In the latter case the effect of vegetation was measured by two indices:

1 The cover area of bushes
2 The percent of bare ground.

The bush cover was determined using tapes and a quadrat and the results are drawn in figure 5.1 and 5.2. The amount of cover is important because it affects the interception and storage capacity of the canopy and the consumptive use of soil water. The percent bare ground was estimated using a quadrat only and excluded areas of litter, plant cover above c3cms (but not distinguishing plant species), lichen and large stones, but bare ground under bushes, for example, is recorded. This presents a slightly different picture to the first and reflects the available source area for sediment entrainment by overland flow. Patterns of bare ground are shown in figure 5.3 and 5.4 for erosion plots on sites 1 and 2 and it is this value which is subsequently used in the analysis relating vegetation to sediment entrained. The characteristics of the erosion plots are shown in table 5.1.

An important handicap to monitoring soil erosion in semi-arid areas is that there are few storms per year, and even fewer events

Figure 5·1 Bush cover on the run off plots for site 1.

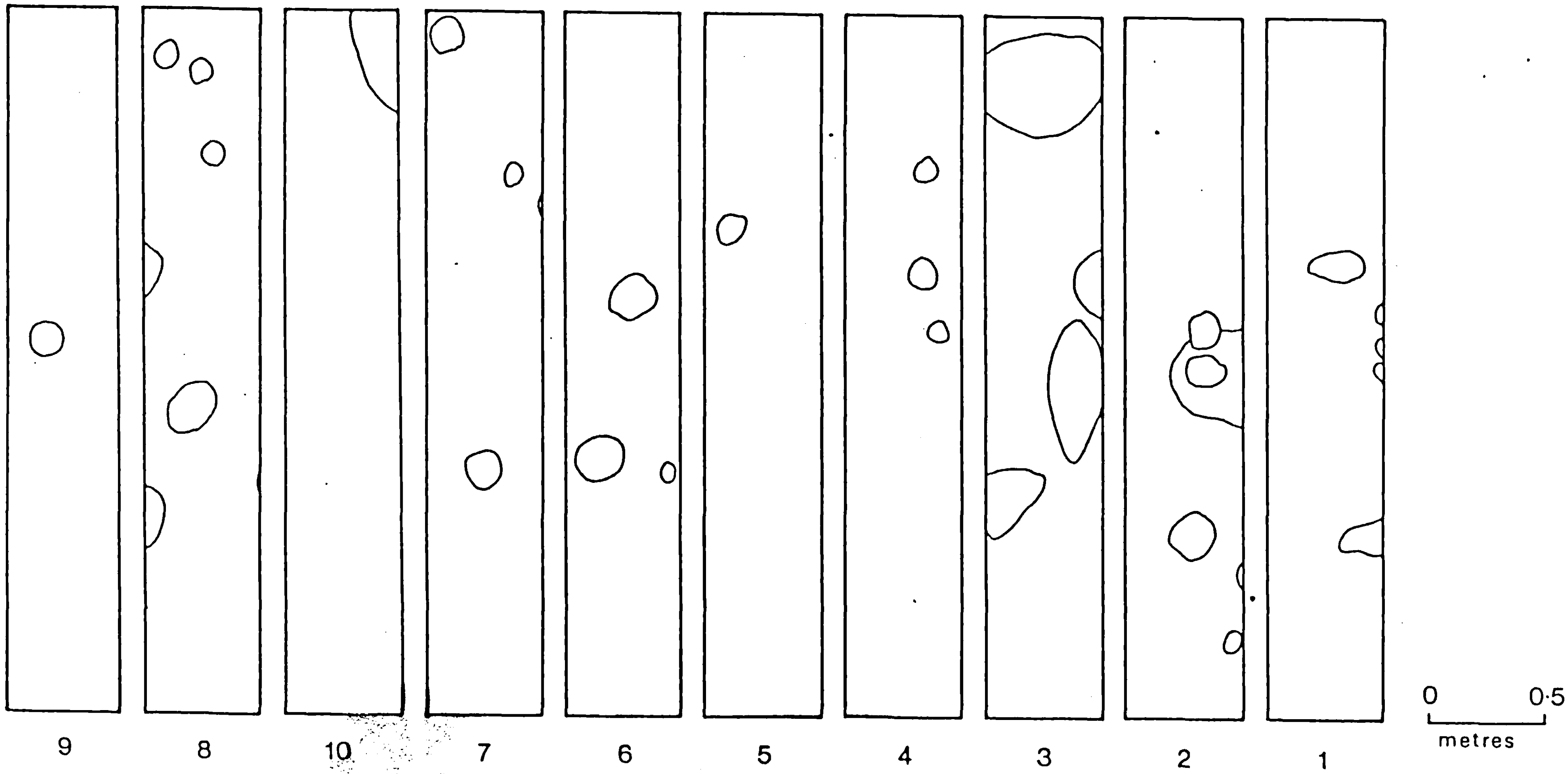


Figure 5·2 Bush cover on the run off plots for site 2

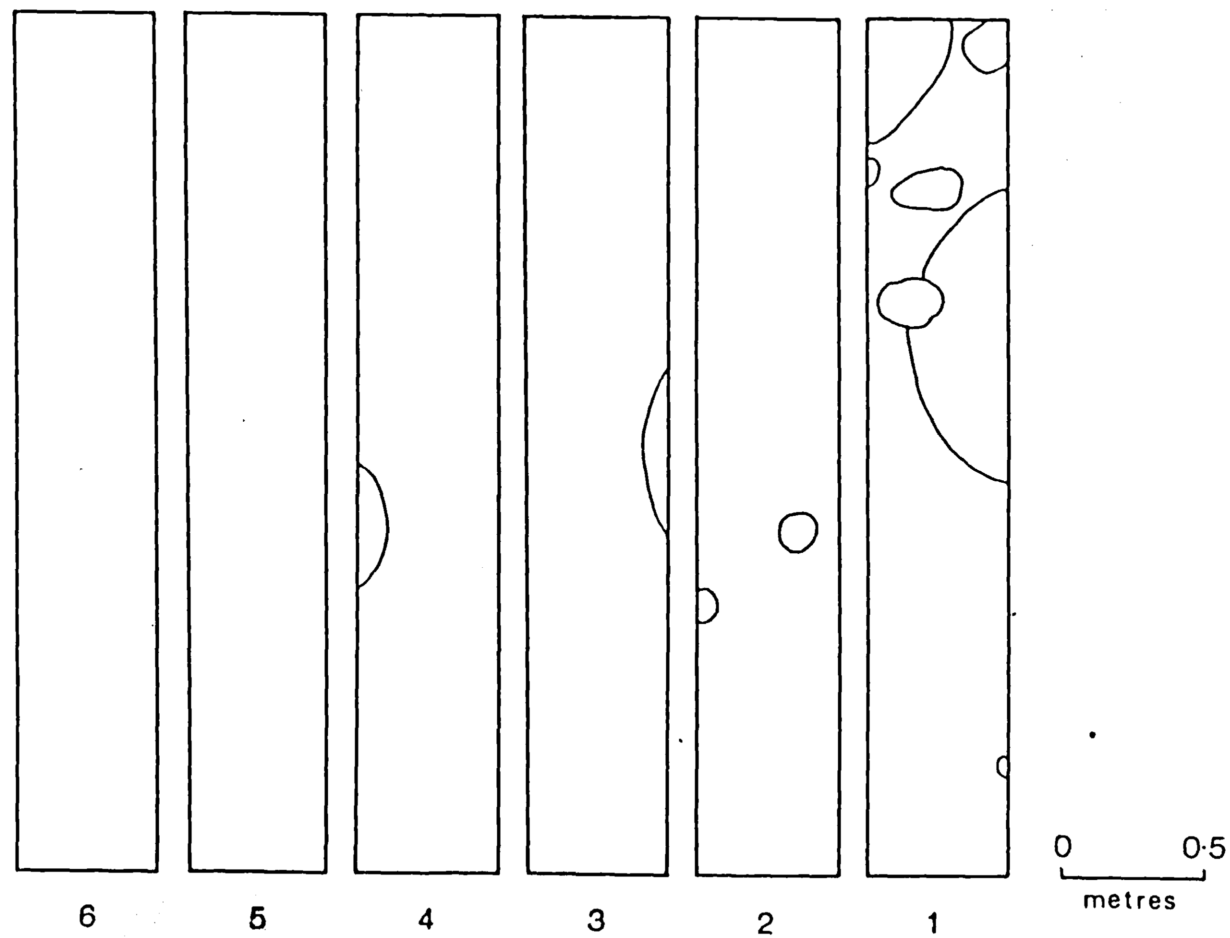


Figure 5·3 Estimate of the percent bare ground on the run off plots for site 1.

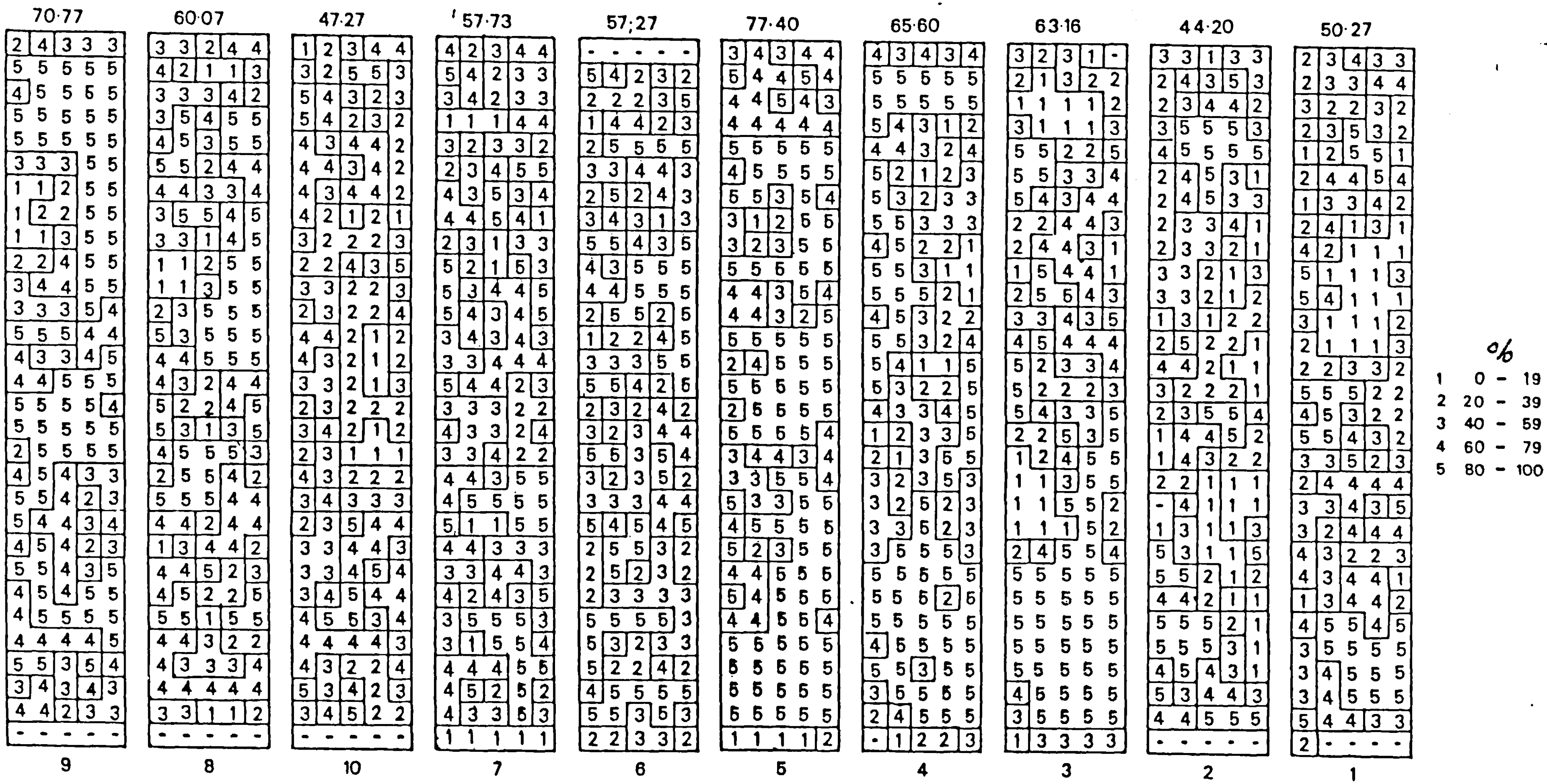


Figure 5·4 Estimate of the percent bare ground on the run off plots for site 2

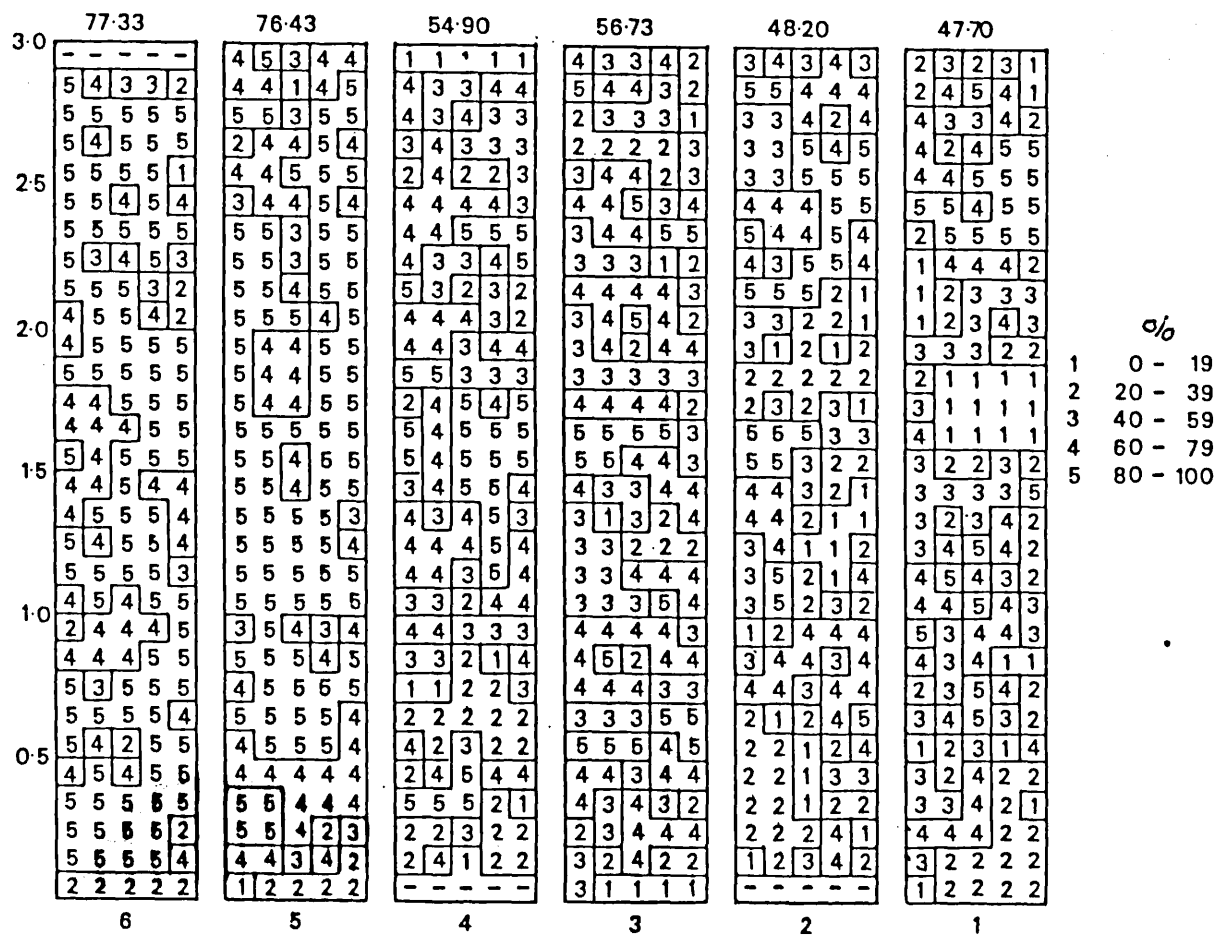


Table 5.1 Characteristics of the Erosion Plots

T R O U G H S

	Downslope									Upslope
SITE 1										
	1	2	3	4	5	6	7	8	9	10
S	15°10	13°50	17°20	22°00	28°30	26°10	19°40	19°10	14°20	19°20
%BG	50.27	44.20	63.17	65.60	77.60	57.27	57.73	60.07	70.77	47.27
SITE 2										
	1	2	3	4	5	6				
S	28°00	23°30	25°30	24°50	20°40	14°10				
%BG	47.70	48.20	56.73	54.90	76.43	77.33				

S = slope of erosion plot
%BG = the percent of bare ground in the plot

large enough to produce overland flow. For example Yair and Klein (1973) studied erosion processes near Elat, Israel where during the year 1970/71 there were 10 days with rain totaling to 28mm of which five produced overland flow and only one produced channel flow. On this basis they suggested that there is a precipitation threshold for overland flow production in Israel of 3mm per day, or 1mm intensity/3mins. In order to collect as much storm data as possible the Gerlach troughs were left installed from July 1982 (November 1982 on conglomerate soil) to 20 December 1983.

5.2 Results of Discharge and Sediment Data

Despite the limitations of the data, a number of aspects can be examined. Firstly some general comments are made on the characteristics of the data set. Secondly patterns of sediment supply are related to governing parameters by the use of some simple models. Thirdly long term changes in sediment transport are considered, and finally the relationship of surface wash to gully head migration is assessed.

Only one data set for overland flow is available following the storm on 26 November, and discharge values are given in table 5.2. On the marl, values for discharge range from 3.71 - 9.18 litres (for troughs 1 to 10, excluding 8 and 9). The mean value for discharge on the marl is 6.21 litres (standard deviation 1.60). The highest value is for trough 6, which approaches the marl/conglomerate soil boundary. On site 2 the mean discharge is 2.86 litres (standard deviation 0.83). A difference of means test shows that the discharge on site 2 is significantly different from site 1, and in fact the mean is 46% lower on site 2. The two sites most probably received similar quantities and intensities of rainfall, and have a similar range of slope angles and vegetation cover. The difference in overland flow measured on the two lithologies may be due to differences in the infiltration characteristics, although this was not brought out by the infiltration experiments. With runoff about twice as high on the marl, this may reflect the effectiveness of the surface crust on the marl for overland flow generation. On site 2 the discharge decreases downslope, but there is no clear pattern on site 1. This

Table 5.2 Measured volumes of overland flow for unknown catchment areas - 26 November 1982

Trough Number	Site 1 Marl litres	Site 2 Conglomerate litres
1	6.56	1.25
2	4.67	2.84
3	6.68	3.17
4	3.71	3.40
5	6.27	3.51
6	9.18	2.99
7	6.43	
10	6.15	
Mean	6.21	2.86
Standard deviation	1.60	0.83

observation for site 2 fits the observed increase in final infiltration downslope.

Absolute denudation rates are estimated for the year 1983 only using data from the bounded plots for periods 4 and 5 that is 5 April to 20 December 1983. As the winter sediment for November 1982 to March 1983 is so small (ranging from 2.43g - 8.22g on marl and 3.03g - 17.07g on the conglomerate soil) an estimate of annual soil loss excluding this data from unbounded plots will not significantly alter the values. Mean annual soil loss for the year 1983 on the two sites is 1.68 kg m^{-2} yr^{-1} on the marl (with a standard deviation of 1.23), and 1.08 kg m^{-2} yr^{-1} for the conglomerate soil (standard deviation is 0.59). This represents 0.45 to 2.91 kg m^{-2} yr^{-1} soil loss on the marl and 0.49 to 1.67 kg m^{-2} yr^{-1} on the conglomerate. Despite the difference in the means, the range of variation is such that there is no significant difference between the sample populations. To compare with the Universal Soil Loss Equation estimates of up to 200 t ha^{-1} yr^{-1} (Chapter 2), this results in mean values of 16.8 and 10.8 t ha^{-1} yr^{-1} for marl and conglomerate soils respectively. The Universal Soil Loss Equation has grossly overestimated erosion rates on these hillslopes with steep slopes, poor vegetation cover, and erodible soils. Even so, mean soil losses of 1.68 and 1.08 kg m^{-2} yr^{-1} on marl and conglomerate respectively (with maximum values of 3.6 and 2.9 kg m^{-2} yr^{-1}) are high, approaching the accelerated rates suggested by Young (1969) (see section 1.1) of 4.5 to 45 kg m^{-2} yr^{-1}, showing that the area is suffering from quite rapid soil loss even in relatively dry years (290.2mm of rain for calendar year 1983).

Table 5.3 lists erosion rates by surface wash for a variety of environments. The rates measured by the author are for small plots for only one year, and it is well known that plot rates tend to overestimate erosion. The rate obtained by Scoging on the marl is about twice the value measured here, though Scoging's study period was considerably wetter with 405 mm of rain between October 1975 and September 1976. Young (1974) quotes a range of erosion rates. Those for semi-arid areas are in the same order of magnitude as the Spanish ones, whereas those for Mediterranean

Table 5.3 Slope denudation rates for surface wash

Area		Slope	kg. m^{-2} yr^{-1}	cms yr
SE Spain	Marl		1.680	0.120
	Conglomerate		1.087	0.068
From Scoging (1982)				
Semi-arid, S.E. Spain, marl			4.400	0.275
From Young (1974)				
Semi-arid				
Colorado				0.20
New Mexico, U.S.A.				0.64 to 0.82
New Mexico, Wyoming, U.S.A.				0.76 to 1.17
Mediterranean				
S. France				0.007 to 0.009
Temperate Continental				
New Jersey, U.S.A., badlands				23.0
Alberta, Canada, badlands				0.09
Crimea, U.S.S.R., badlands				10.0
From Morgan (1980)				
Humid tropics, bare soil				10.0 to 17
Savanna, bare soil				1.3 to 1.8
Temperate, bare soil				1.0

climates and badlands in temperate continental environments have much lower and higher values. This may represent greater vegetation cover for the French sites, and greater rainfall for the Crimea and New Jersey sites.

Table 5.4 breaks the erosion data down to compare between lithologies and periods. Only one period, number 2, was for a single storm, the remaining data is for cumulative sediment deposition over long time spans which may reflect sediment yields from one or more runoff events. This brings out a number of interesting points.

Firstly as a general rule the greater the amount of precipitation, the more sediment is eroded. This relationship cannot be stretched too far because the precipitation data is for Ugijar 4.5km away from the site, the number of observations was small, the precipitation is expressed as the cumulative amount during a period rather than amount per storm, and the antecendent conditions were variable. Period 4 is the exception to this rule as a mean of 2059.0 grms on the marl and 1369.8 grms of conglomerate soil were transported for 3.5 mm of rainfall. In this case rainfall occurred on 30 August 1983 after some 157 days of dry weather, during which mechanical weathering of the soil could have produced a large amount of loose debris ready for entrainment. Another explanation is that the storm was more intense at the site than at Ugijar, or there were other rainfall events during the period.

Secondly in three out of four cases the mean soil loss was greater on the marl. For periods 2 and 5 the difference is statistically significant, but not for period 4. This may suggest that generally the marl is more susceptible to erosion by surface wash, but for large storms (or at least for large accumulations of sediment) the difference between lithologies is reduced, possibly due to the crossing of an erodibility threshold on the conglomerate. In period 3 the sediment caught is greater on the conglomerate soil. The examination of the data shows that the sediment weights for period 3 are very small, and on site 2 are biased by the presence of two stones in troughs 2 and 4. These may have been close to the troughs and required little energy to

Table 5.4 Precipitation and sediment data for sites 1 and 2

	Rainfall at Ugijar mm	Number of Rain Days		Sediment Transport Marl grms	Conglomerate grms
1 Aug-15 Nov 82	160.7	5	X	279.6	-
			sd	97.8	-
26 November 82	23.0	1	X	33.7	11.6
			sd	16.6	18.1
27 Nov-4 Apr 82/3	18.5	3	X	3.8	7.7
			sd	1.1	6.2
5 Apr-28 Sep 83	3.5	1	X	2059.0	1369.8
			sd	1538.9	681.8
29 Sep-20 Dec 83	249.2	14	X	460.5	244.4
			sd	341.5	115.7

1983 soil loss

1.68kg m^{-2} yr^{-1} (standard deviation 1.23) - marl
1.08kg m^{-2} yr^{-1} (standard deviation 0.59) - conglomerate

move them in. When these are removed from the data, there is no significant difference between the average quantity of sediment caught in the troughs on site 1 and site 2. At such small sample weights the data are easily biased by one or two stones.

Thirdly the table shows that for each period and lithology there is a considerable amount of within site variation indicated by the standard deviations. Tables 5.5 and 5.6 give the quantities of sediment caught in all the troughs and the percentage of particles greater than 2mm for both the total deposit and the deposit corrected for organic matter. These show the range of sediment transport on the slopes for the different periods. This variability is examined to determine firstly consistent patterns of erosion as represented by sediment caught in the troughs and secondly to isolate factors affecting erosion loss. Both have an affect on gully head growth in terms of sediment supply. This is done by using four simple sediment transport models and four single variable parameters (table 5.7) and comparing expected rankings of troughs (according to the highest sediment content expectd within the trough) with the observed rankings of troughs (according to actual weights of sediment caught). There are three sets of data, these are firstly all the troughs on marl (8 troughs on site 1 numbered 1 to 7 and 10), all data on the conglomerate soil (8 troughs altogether, numbered 8 and 9 on site 1 and 1 to 6 on site 2), and all data on site 2 (6 troughs). The observed values of sediment accumulation in the troughs are ranked in order of magnitude starting with the highest. These are compared with the expected rankings for the different models where again the value of the model for each trough is ranked. Spearman's coefficient of rank correlation is used, namely:

$$R_s = 1 - \sqrt{\frac{6\ \Sigma di^2}{N3 - N}}$$

where Σdi^2 = sum of the difference between each pair of ranked scores squared, and N = number of observations. The significance of the correlation is tested using Student's 't' distribution with N-2 degrees of freedom where:

$$t = r_s \sqrt{\frac{N-2}{1 - r_s^2}}$$

The value for R_s ranges between -1.0 and +1.0, and is 0 when the variables are unrelated. The test is nonparametric so it does not

Table 5.5 Summary of sediment data for site 1

Period		1	2	3	4	5	6	7	8	9	10
1	a	288.3	198.7	167.0	450.2	357.9	267.2	227.9	65.9	51.9	-
	b	(267.5)	(185.0)	(156.5)	(431.6)	(327.8)	(252.6)	(214.2)	(62.8)	(49.9)	-
	c	2.2	2.9	2.2	3.4	1.0	1.0	9.2	38.1	32.2	-
	d	(1.8)	(2.8)	(1.6)	(3.4)	(1.0)	(1.0)	(9.4)	(38.4)	(32.9)	-
2	a	55.0	24.4	46.6	41.2	29.1	14.8	48.4	54.2	17.2	9.7
	b	(51.1)	(17.6)	(42.9)	(39.3)	(28.0)	(14.4)	(44.4)	(51.0)	(16.4)	(9.5)
	c	2.4	22.1	1.3	4.6	1.7	2.0	1.9	10.7	12.2	13.4
	d	(2.2)	(1.1)	(0.70)	(4.8)	(1.4)	(2.1)	(1.1)	(11.2)	(12.8)	(12.6)
3	a	3.2	5.4	3.9	4.4	5.3	3.1	2.9	8.2	2.6	2.4
	b	(2.9)	(4.8)	(3.1)	(3.9)	(5.0)	(2.9)	(2.6)	(7.7)	(2.4)	(2.2)
	c	43.5	37.9	27.7	16.5	8.7	17.9	18.3	55.0	23.4	37.0
	d	(34.5)	(8.4)	(3.7)	(3.8)	(3.0)	(15.5)	(13.6)	(54.1)	(21.0)	(33.6)
4	a	752.0	295.4	787.1	2760.3	1775.9	4273.8	4179.3	1627.2	1121.0	1647.7
	b	(673.0	(259.3)	(740.6)	(2642.9)	(1719.5)	(4140.8)	(3992.2)	(1611.5)	(1060.7)	(1579.2)
	c	8.7	6.4	5.8	10.0	5.1	11.8	21.9	31.9	34.7	9.4
	d	(6.4)	(4.8)	(5.1)	(10.3)	(4.7)	(12.0)	(22.7)	(32.0)	(35.8)	(9.2)
5	a	97.8	75.6	409.7	496.7	534.6	1138.7	648.6	241.7	133.6	282.2
	b										
	c	5.6	6.5	6.1	6.2	5.0	4.7	15.6	49.2	16.2	12.5
	d										

a = weight of the debris in the trough (grms)
b = weight of the debris corrected for the organic content (grms)
c = percent of the debris greater than 2mm
d = percent of the debris greater than 2mm corrected for the organic content

Table 5.6 Summary of sediment data for site 2

Period		1	2	3	4	5	6
2	a	0.3	1.9	1.2	3.2	5.1	9.6
	b	(0.2)	(1.8)	(1.2)	(2.9)	(4.8)	(9.0)
	c	0.0	20.2	28.5	23.8	17.1	11.5
	d	(0.0)	(17.1)	(16.4)	(23.8)	(9.4)	(12.3)
3	a	2.0	17.1	3.0	16.3	9.5	3.1
	b	(1.3)	(15.7)	(2.3)	(14.7)	(8.7)	(2.6)
	c	36.2	44.9	31.7	78.6	17.5	27.2
	d	(7.5)	(39.4)	(23.6)	(83.5)	(16.2)	(19.2)
4	a	843.1	1461.8	1634.3	2532.3	1546.9	192.1
	b	(811.6)	(1401.7)	(1574.9)	(2432.5)	(1495.3)	(184.5)
	c	36.2	37.7	31.7	31.0	24.3	51.4
	d	(37.2)	(38.4)	(32.8)	(31.9)	(24.9)	(53.3)
5	a	70.6	255.9	275.5	395.5	398.3	184.1
	c	36.9	25.3	33.9	21.5	25.8	42.8

a = weight of the debris in the trough (grms)
b = weight of the debris corrected for the organic content (grms)
c = percent of the debris greater than 2mm
d = percent of the debris greater than 2mm corrected for the organic content

Table 5.7 Models of sediment transport

1	$S_t = f(X^{1.6}.S^{1.3})$	X = horizontal distance from divide (metres) S = slope (tan)
2	$S_t = f(S)$	S = slope (degrees)
3	$S_t = f(X)$	
4	$S_t = f(\%BG)$	%BG = percent bare ground for the erosion plot
5	$S_t = f(Q)$	Q = discharge (litres)
6	$S_t = f(Q^2.S^{1.66})$	S = slope (tan)
7	$S_t = f(\%BG.S)$	S = slope (degrees)
8	$S_t = f(\%BG/S)$	S = slope (degrees)

depend on normality assumptions of the distribution of the data. There must be at least five pairs of observations to establish significance at a generally meaningful level (Norcliffe 1977), so the data sets presented here are just above the minimum requirements. For each data set the null hypothesis is that there is no significant difference in the ranking of sediment quantities for different locations according to the sediment transport model and the observed sediment quantities caught in the Gerlach troughs. The results of the statistical analysis are shown in table 5.8 which gives the r_s values and indicates the level of significance for some of the values namely:

1 Values significant at 0.20 (one underscore)

2 Values significant at 0.05 (two underscores)

3 Values significant by 0.01 (three underscores).

The first four tests compare the viability of a general slope wash model to describe patterns of sediment transport with local parameters of the erosion plots to determine the relative importance of largescale and local factors accounting for soil loss. The first model is:

$$S_t = f\ (x^{1.6}.s^{1.3})$$

where x is the horizontal distance from divide (m), s is the slope (tan), and the exponents are taken for the case of surface wash (Kirkby 1971). This model is evaluated for periods 1-3 only, being the unbounded conditions.

Table 5.8a for the marl shows that for periods 1 and 3 this model does produce a significantly similar ranking (at 0.05 and 0.20 significance levels) between observed and expected values, with r_s = 0.714 and 0.619 respectively, with an insignificant explanation for period 2 (r_s= 0.357). Data from site 2 for period 1 is lacking because the troughs were not installed until November. However, when data from all the conglomerate sites are examined (table 5.8b) the correlations are insignificant for periods 2 and 3 at r_s = -0.048 and -0.190. On site 2 only (table 5.8c) the rankings are negatively correlated significantly so that sediment yield decreases with increasing slope and distance. For period 2 this is particularly marked with r_s = -0.943.

Table 5.8 Correlation values for models

(a) Marl Site - Troughs 1-7, and 10 (n=8)

Model	Period 1	Period 2	Period 3	Period 4	Period 5
$X^{1.6}S^{1.3}$	+0.714	+0.357	+0.619	-	-
S	+0.571	-0.238	-0.095	+0.833	+0.857
X	-0.107	+0.452	+0.619	-	-
%BG	+0.464	+0.310	+0.238	+0.476	+0.595
Q	-	+0.214	-	-	-
$Q^2S^{1.66}$	-	+0.048	-	-	-
%BG x S	+0.571	-0.119	0.000	+0.810	+0.881
%BG / S	-0.429	+0.690	+0.429	-0.690	-0.595

(b) Conglomeratic soil - Site 1 Troughs 8 & 9, Site 2 Troughs 1-6 (n=8)

Model	Period 1	Period 2	Period 3	Period 4	Period 5
$X^{1.6}S^{1.3}$	-	-0.048	-0.190	-	-
S	-	-0.881	-0.095	+0.381	+0.143
X	-	+0.190	-0.310	-	-
%BG	-	+0.714	-0.024	-0.214	+0.143
Q	-	+0.786	-	-	-
$Q^2S^{1.66}$	-	+0.619			
%BG x S	-	-0.452	+0.190	+0.619	+0.714
%BG / S	-	+0.763	-0.119	-0.262	+0.060

(c) Conglomeratic soil - Site 2 Troughs 1-6 (n=6)

Model	Period 1	Period 2	Period 3	Period 4	Period 5
$X^{1.6}S^{1.3}$	-	-0.943	-0.200	-	-
S	-	-0.943	-0.486	+0.314	-0.257
X	-	-0.943	-0.200	-	-
%BG	-	+0.829	+0.029	-0.086	+0.371
Q	-	+0.543	-	-	-
$Q^2S^{1.66}$	-	+0.143	-	-	-
%BG x S	-	-0.143	-0.086	+0.714	+0.771
%BG / S	-	+0.829	+0.029	-0.086	+0.371

The components of this model (slope and distance from divide) are examined separately to determine their relative importance, along with two other local variables discharge, and the percent bare ground.

On the marl the single variable correlating most highly with sediment is slope, with r_s values 0.571, 0.833 and 0.857 for periods 1, 4 and 5. Distance from divide is only significantly correlated in period 3 with r_s equal to 0.619. The amount of bare ground just above the troughs is only significantly correlated once with r_s equal to 0.595 (at the 0.20 significance level). Surprisingly the discharge collected for period 2 in the Gerlach troughs does not correlate with the quantity of sediment caught. This suggests there were either substantial sampling errors, or the factors associated with sediment detachment and transport are dependent on other variables.

Taking all conglomerate soil samples, table 5.8b shows that although the $x^n s^m$ model offers no correlation with sediment, slope, bare ground and discharge all correlate highly with r_s values of -0.881, 0.714, and 0.786 respectively for period 2.

On site 2 (table 5.8c), slope, distance from divide, and the percent of bare ground all correlate significantly for period 2 with r_s values of -0.943, -0.943, and 0.829. Again slope is negatively related to sediment trapped in the troughs, as well as distance from divide, and these two values taken together account for the highly correlated negative relationship with the slope wash model $x^{1.6}s^{1.3}$.

The poor correlation between discharge and sediment is very surprising considering the marked attention to discharge as an important erosion parameter in many field experiments and models. Unfortunately the data on discharge is severely limited to one storm, which may exhibit unusual or random influences or sampling errors. However the relationship between discharge and sediment is developed by employing the standard sediment discharge equation

$$S_T = K.Q^2.S^{1.66}$$

where S_T is the sediment transport, K is a constant, and Q and S are slope and discharge. Despite the apparent significance of slope angle on the marl this model does not account for the distribution of sediment caught in the traps on site 1 with r_s = 0.048. Similarly it is not significant on site 2 either with r_s = 0.143 (table 5.8c) although when taking account of all the conglomerate samples (table 5.8b) the correlation is improved to 0.619.

Finally the affects of the percent bare ground and slope are crudely amalgamated to examine the effect of this combination of parameters on sediment yield. As slope and the area of bare ground increases sediment yield would be expected to increase (table 5.8a). On the marl BG x S correlates highly for periods 4 and 5, but judging by the ranking produced by this model, this is largely a reflection of the dominance of slope. On the conglomerate slopes (table 5.8 b and c) this is the only model to produce correlations significant by the 0.20 significance level for periods 4 and 5. However the model BG/S gives a high correlation of 0.829 for period 2, and on site 1 (figure 5.8a) gives significant negative correlations for periods 2, 4, and 5.

To summarise it is evident that no one model correlates for each time period, reflecting not only the variation in sediment yield with location but also the varying loci of sediment transport for different occasions. For some periods there is little correlation between actual and expected amounts of sediment transported. On the marl the only model to correlate with the distribution of sediment yield for the storm event of 26 November (period 2) is %BG / S. On the conglomerate soil there are no significant correlations for period 3.

On the marl (table 5.8a) the highest correlations are for periods 4 and 5. Slope is the single dominating parameter correlated positively with sediment and its strength probably accounts for the higher correlations for the compound models which have slope as a variable. Discharge appears unrelated to sediment transport, although data are limited. For large magnitude events (or at least large accumulations of sediment) there is a better agreement between observed and expected patterns of sediment transport. Table 5.4 shows that periods 2 and 3 relate to times with relatively low sediment movement, whereas periods 1, 4 and 5 had high sediment contents.

On the conglomerate soil a different picture emerges where the higher correlations occur for period 2. Although discharge is still not significantly correlated with sediment, r_s is higher at 0.543 (table 5.8c). The amount of sediment transported in period 2 decreases downslope possibly due to a decrease in the amount of bare ground despite an increase in slope (table 5.1). In period 3, table 5.8b shows that single large stones in troughs 2 and 4 account for a large proportion of the weight of that sediment. At such low sediment levels and small sample numbers, the chance occasion of a stone rolling into the trough can significantly alter the overall patterns of sediment distribution when measured by weight. In periods 4 and 5 the distribution of sediment shows that areas of greater concentration of sediment transported occurs in the middle section of the slope. The aptness of the BG x S model may reflect influence of lower slopes near the watershed (at trough 6) and the increase of vegetation near the base of the slope (at trough 1) to limiting sediment entrainment.

5.3 Long Term Variations

Already it has become evident that there will be temporal variations in sediment transport and discharge. Part of this variation will reflect seasonal effects in the occurrence of rainfall affecting runoff, vegetation cover, antecedent soil moisture conditions, and quantity of weathered material available for entrainment. In the latter case debris flushing occurs on all time scales from the individual storm (Bryan, Yair, and Hodges 1978) to the annual level. Thus Thornes 1976 emphasises the importance of the first rainstorms after the summer drought. However superimposed on these seasonal variations are long term changes in vegetation cover, slope development, infiltration and the character of the soil surface.

So far soil loss has been considered more in terms of hillslope and discharge characteristics (eg slope, distance from divide, bare ground area) rather than in the physical and chemical properties of the soil (eg particle sizes, cohesion). This largely reflects the dominance of the former type of parameters in erosion models. However here it is necessary to consider changes in the soil as a response to continued processes of soil detachment, transport, and deposition, particularly with reference to armouring. Changes in the surface characteristics through armouring can feed back into the system and change quantities and patterns of discharge and sediment production.

The extent of armouring on the two slopes is assessed by comparing the size fractions of the soil transported (trough soil samples), the soil surface left behind (from surface samples), and the 'original' distribution of soil size fractions (from soil samples taken at depth). The analysis concentrates on the gravel soil fraction (greater than 2mm diameter) as this is most sensitive to critical discharge values.

The variation in particle size with depth (discussed in Chapter 2) on the conglomerate soil (site 2) varies from c39.6% to c61.4% (between 0 and 55cms) reflecting a significant increase in stoniness with depth. The gravel fraction of the surface samples is compared to those samples taken from the 0-1 cms depth, assuming that if armouring occurs the difference to examine is between the surface layer and the soil lying immediately below. Figure 5.5 plots the variation in the coarse fraction with distance downslope for the soil matrix, the surface samples, and transported material for periods 4 and 5 for site 2. The soil matrix value is taken as a constant value of 39.6% which is the mean of the gravel fraction in the top horizons for pits 2, 3, and 4 (table 2.6).

The surface samples show an increase in gravel content in the downslope direction from 35.2% to 50.1% (figure 5.5 and table 5.9). On the upper part of the slope there is less gravel on the surface than within the underlying soil matrix, but below c30m from the divide the situation is reversed. This suggests that on the upper part of the slope there is a greater proportion of fines, and on the lower half there is a lag material with a greater proportion of gravel than within the soil.

Figure 5·5 Downslope variations in the percent gravel content

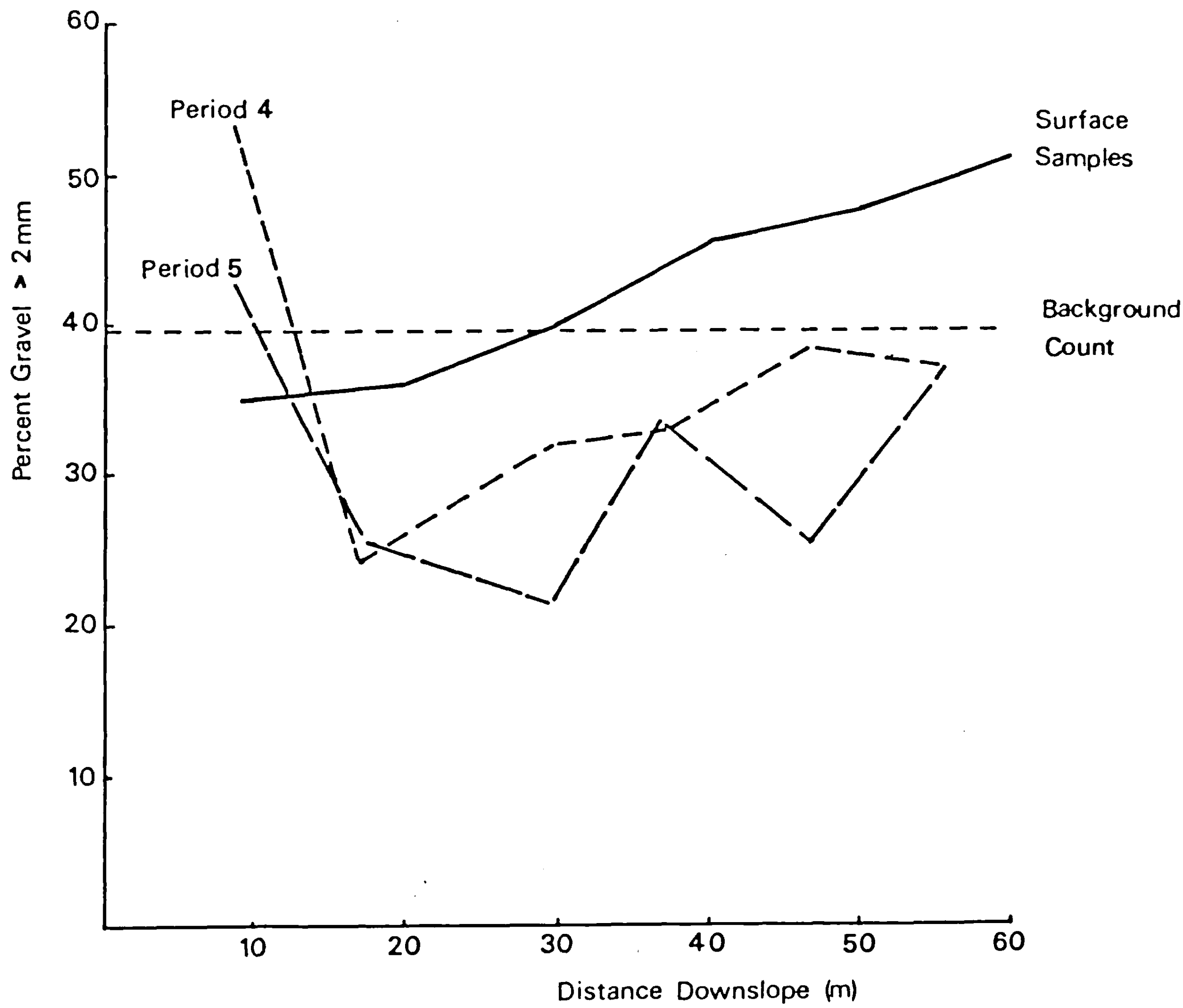


Table 5.9 Variations in surface content of gravel

Site 1

%>2mm	59.8	48.6	32.0	23.2	20.8	8.1	3.9	5.2	4.9
Distance Down (m)	63.0	83.0	104.0	122.0	136.6	155.5	182.5	201.0	205.5

Depth cms	Pit 5 %>2mm	Pit 4 %>2mm	Pit 3 %>2mm
0-13	25.8	23.9	7.5
14-27	30.0	24.7	3.6
			0.1

Site 2

%>2mm	35.2	35.8	40.0	45.4	47.4	50.1
Distance Down (m)	8.5	17.0	29.5	37.0	46.5	55.5

Depth cms	Pit 4 %>2mm	Pit 3 %>2mm	Pit 2 %>2mm
0-13	40.8	37.4	40.6
14-27	45.6	3.8	37.7
28-41	60.6	55.2	
42-55	61.5	61.3	

The trough samples show that (apart from trough 6) there is an overall increase downslope in the proportion of gravel entrained reflecting partly the increase in coarse material itself in the source area, and possibly other factors related to increases in the competance of the flow. Figure 5.6 examines changes in the ratio of trough samples to surface sample gravel fractions with distance from divide. This suggests that beyond c20m the proportion of gravels caught in the troughs increases at a similar rate to the proportion of gravels found on the surface for the mean case.

During periods 2 and 3 the proportion of gravel is generally low although the presence of one or two stones only can markedly affect the proportion of coarse material. However the data suggest that in both periods the actual discharge was not as effective in transporting the coarser fraction as in other periods. This is born out by figure 5.7 a to d, which shows the breakdown of the gravel fraction into the sizes 16mm+, 8-16mm, and 2-8mm, and then the base level of particles less than 2mm by percent weight. For periods 2 and 3 the modal particle size class is the less than 2mm fraction, and the largest size group represented in the frequency charts is mostly the 2-8mm fraction, although in period 3 troughs 2 and 4 have a large stone each. In period 4 there is a progressive decrease in each particle size group above 2mm from troughs 1 to 5 downslope, with an associated increase in the fraction less than 2mm. In trough 6 there were two large stones in the fraction greater than 16mm diameter which has upset the trend. Period 5 shows the gravel fraction is mainly greater at the top and bottom of the slope. The fine fraction tends to increase and then decrease downslope, and the gravel correspondingly decreases and increases.

On site 1 the problem is more complicated because there is a change in soil type from the conglomerate soil on the upper portion of the slope to marl. The marl contains few stones (Chapter 2) and because of this one would not expect armouring as such to the same degree as on the conglomerate soil. However, one may expect a graded condition covering the transitional area from the stony conglomerate to the relatively, stone-free marl where surface flows transport some of the stony material onto the marl.

Figure 5·6 Ratio of gravel in troughs and on thė surface on site 2

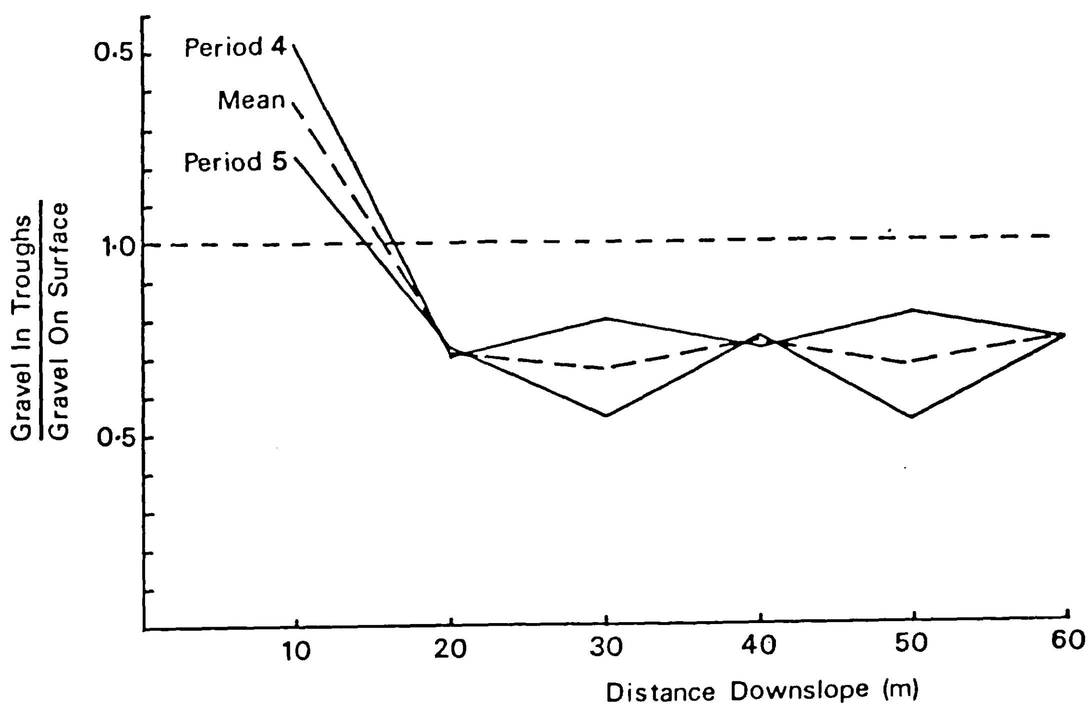
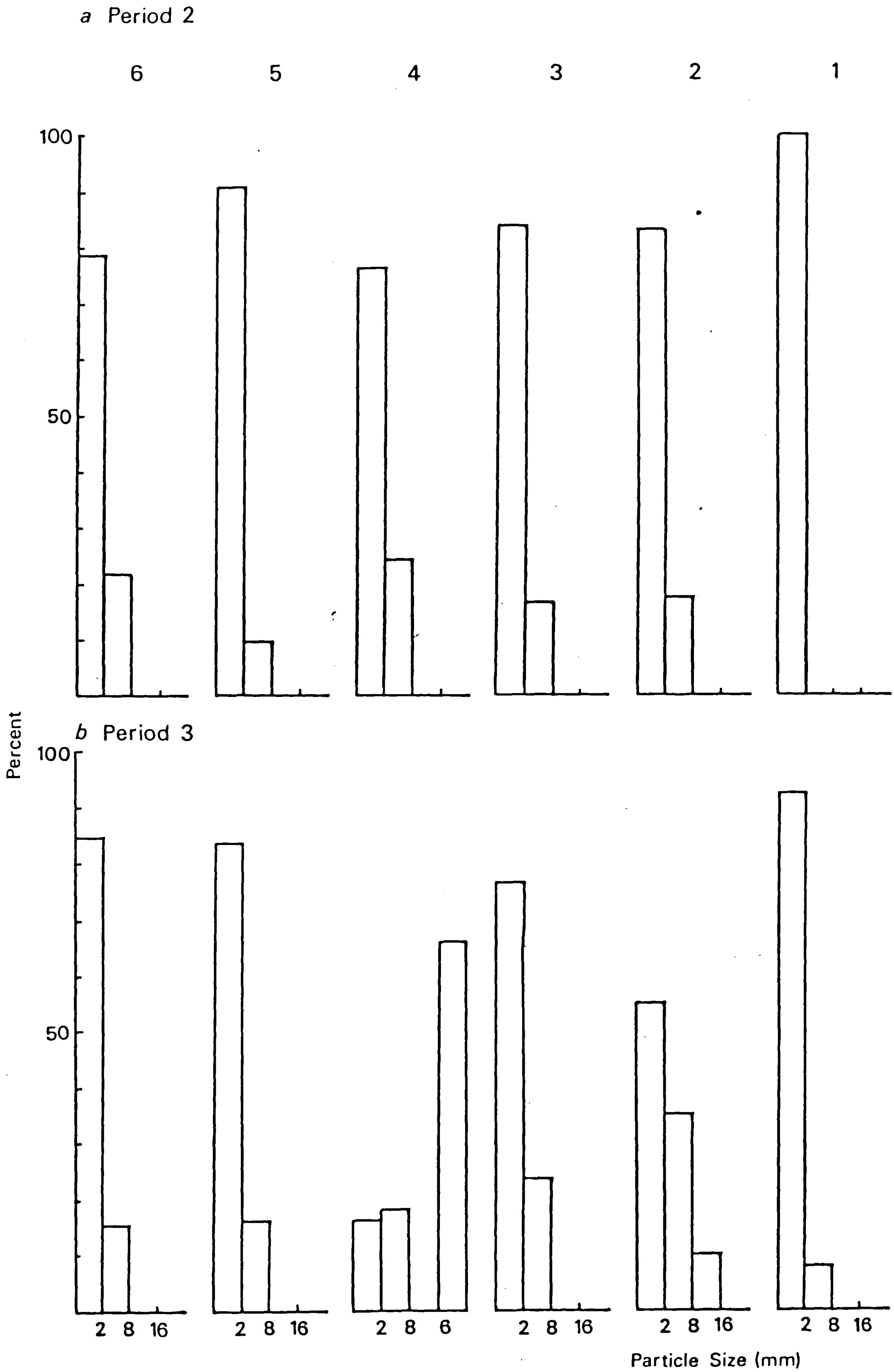
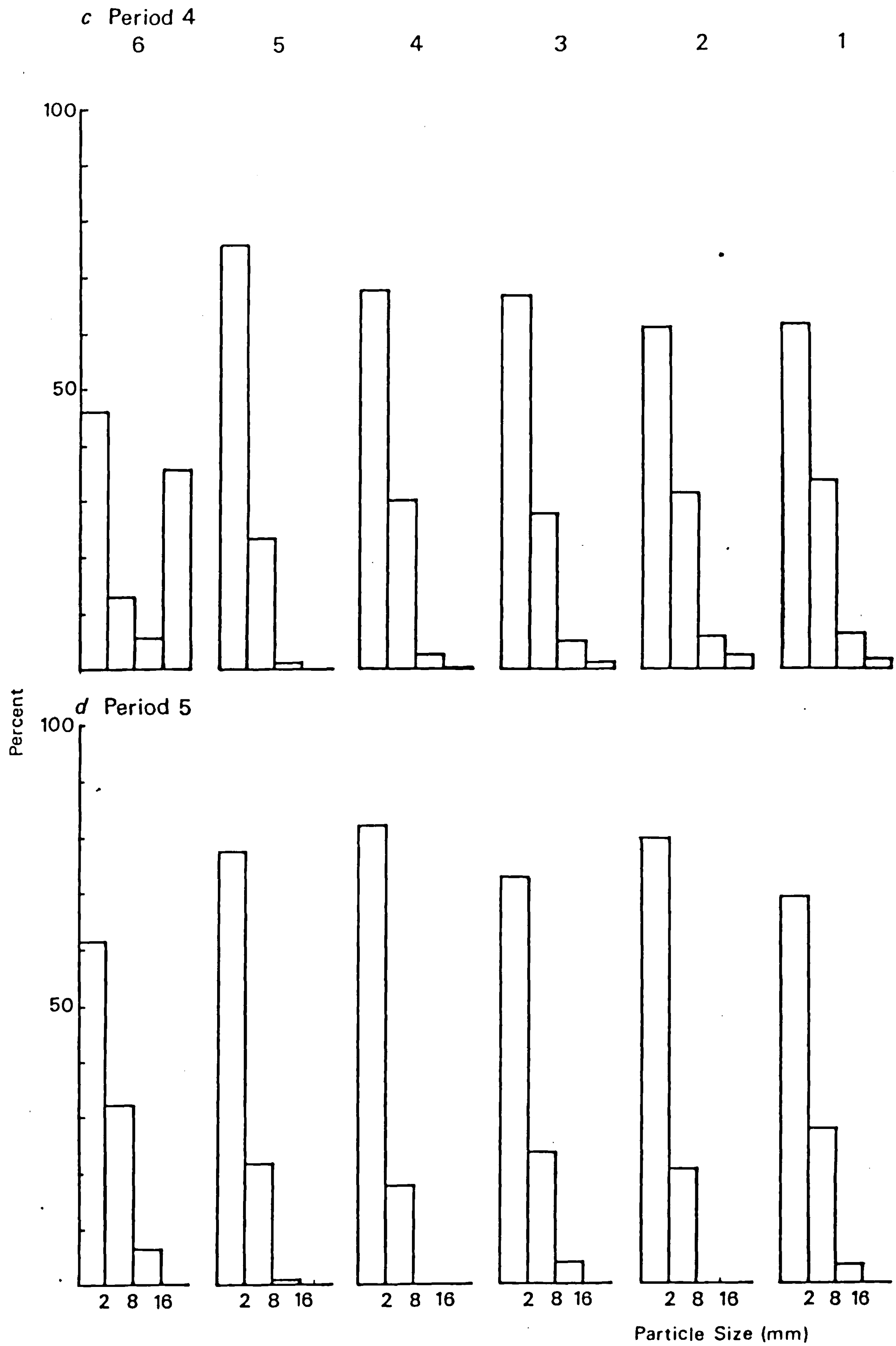


Figure 5·7 Coarse gravel size distribution on site 2 for periods 2 to 5




c Period 4
6
5
4
3
2
1
100
50
Percent
d Period 5
100
50
2 8 16
Particle Size (mm)

Such activity makes it difficult to delineate the junction between the two soil types.

The gravel content of the conglomerate on site 1 at 0-15cms depth for soil pits 4 and 5 is 23.9% and 25.8% (table 5.9). These pits, although within the conglomerate, are near the marl boundary and may be 'diluted' with marl fines in comparison to the deeper, more substantial conglomerate further over the slope on site 2. On the marl itself the gravel content falls from 7.5% between 0-14cms to only 0.1% at 30-45cms (table 5.9).

Figure 5.8 plots the changes in gravel content for the surface samples and trough samples for periods 1-5. On the lower part of the hillslope the surface and depth gravel fractions are fairly similar around 5% and 7.5% respectively. From 150m coming upslope the proportion of gravel lying on the surface increases rapidly to 59.8% c52m from the watershed. Troughs 1-7 included and 10 are categorised as on marl from visual observation mainly, but also from analysis of the sediment caught in the trough and changes in soil colour on ignition (the marl remains the same colour, but the conglomerate soils turns a bright orangey-red). On this basis troughs 1-4 lie on marl with a surface gravel component of 4-8%, and 5, 6, 7, and 10 lie on marl with an increasing cover of stones on the surface due to wash from upslope with surface gravel contents rising to c20-30%. Troughs 8 and 9 lie on the conglomerate soil with surface gravel contents rising from c30-60%.

This division between marl and conglomerate soils is substantiated by the proportion of coarse material deposited down the slope in the troughs, as generally troughs 1-7 and 10 have relatively higher proportions of fines than troughs 8 and 9.

There is also a division between periods 1, 4, and 5, where there are relatively large quantities of sediment caught in the troughs, and periods 2 and 3 when volumes were smaller. In period 2 the proportion of coarse material transported is less than in period 1 on site 1 (figure 5.9a-e). In period 3, despite lower recorded rainfall at Ugijar, the coarse fraction is more variable. However as the total weights are small, large proportions of coarse

Figure 5·8 Downslope variations in percent gravel content on the surface and in troughs on site 1

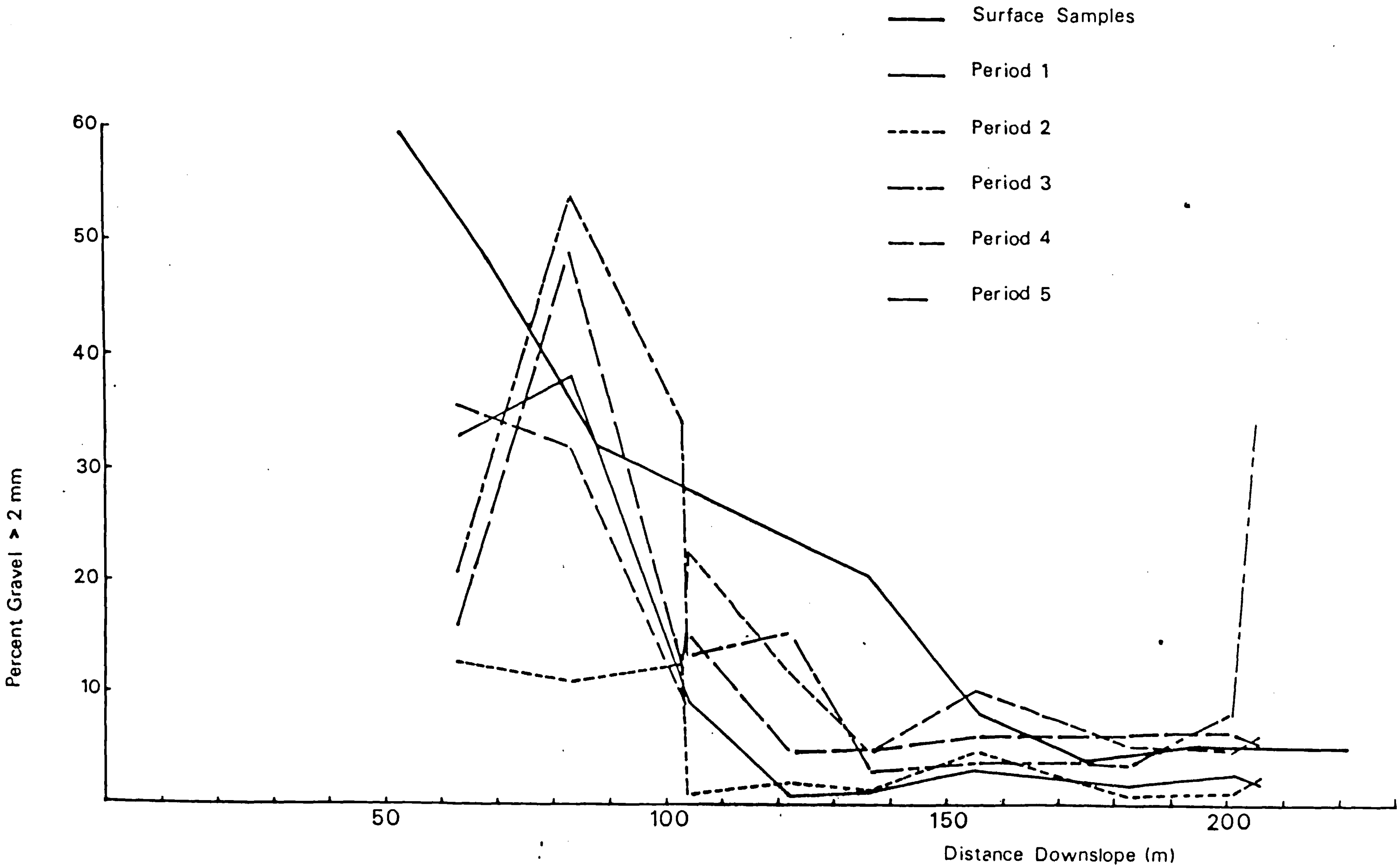
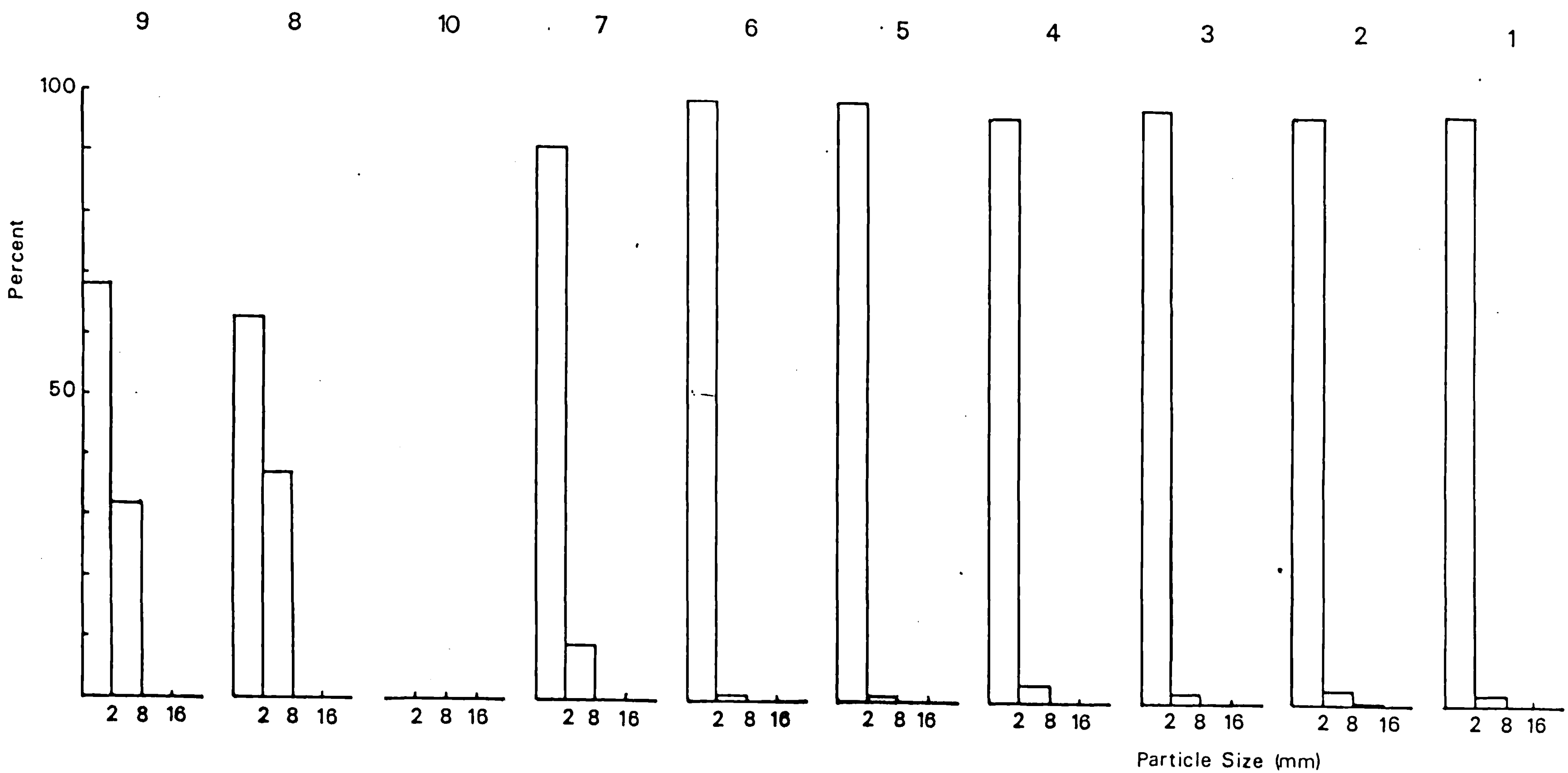
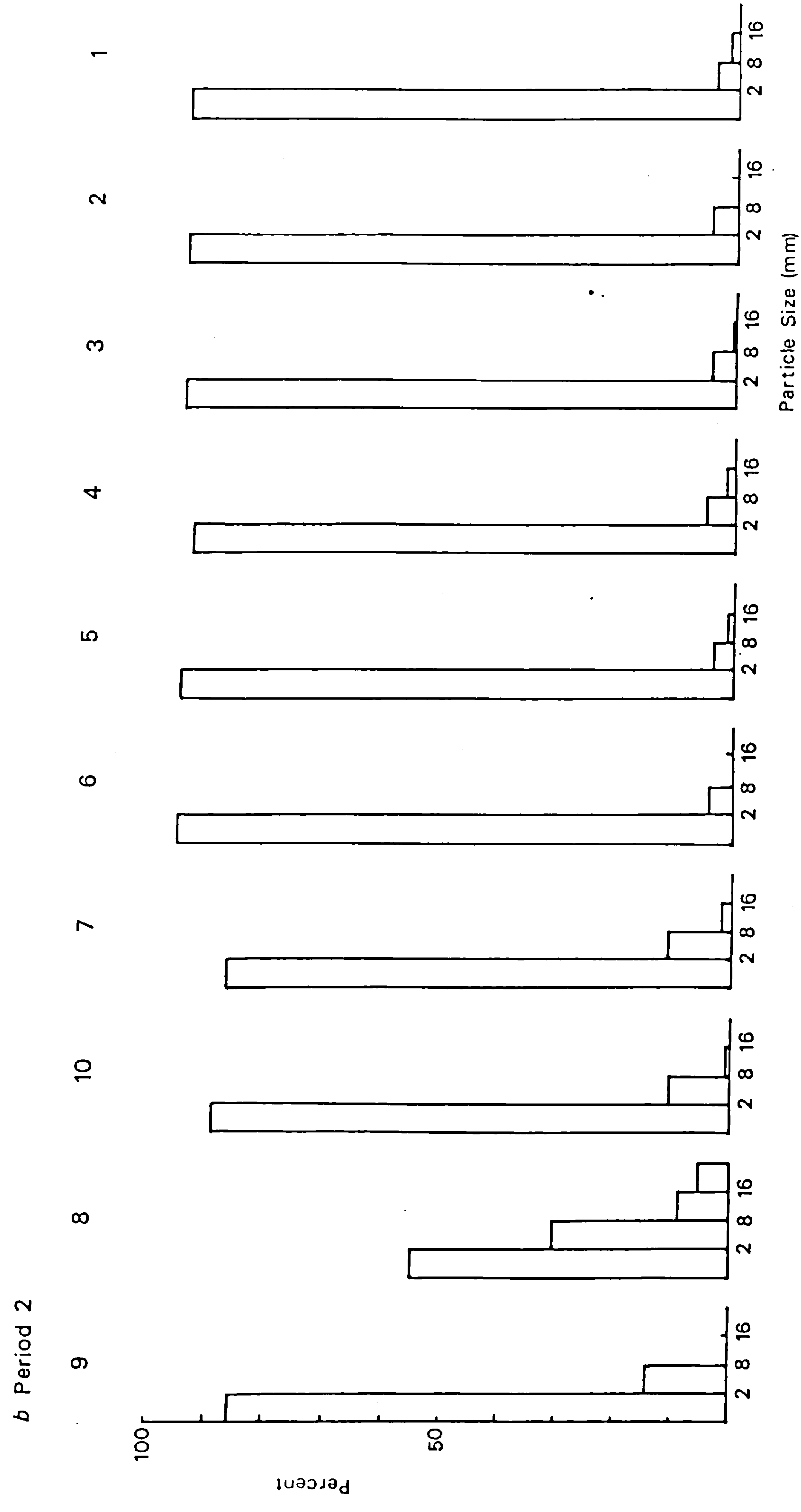
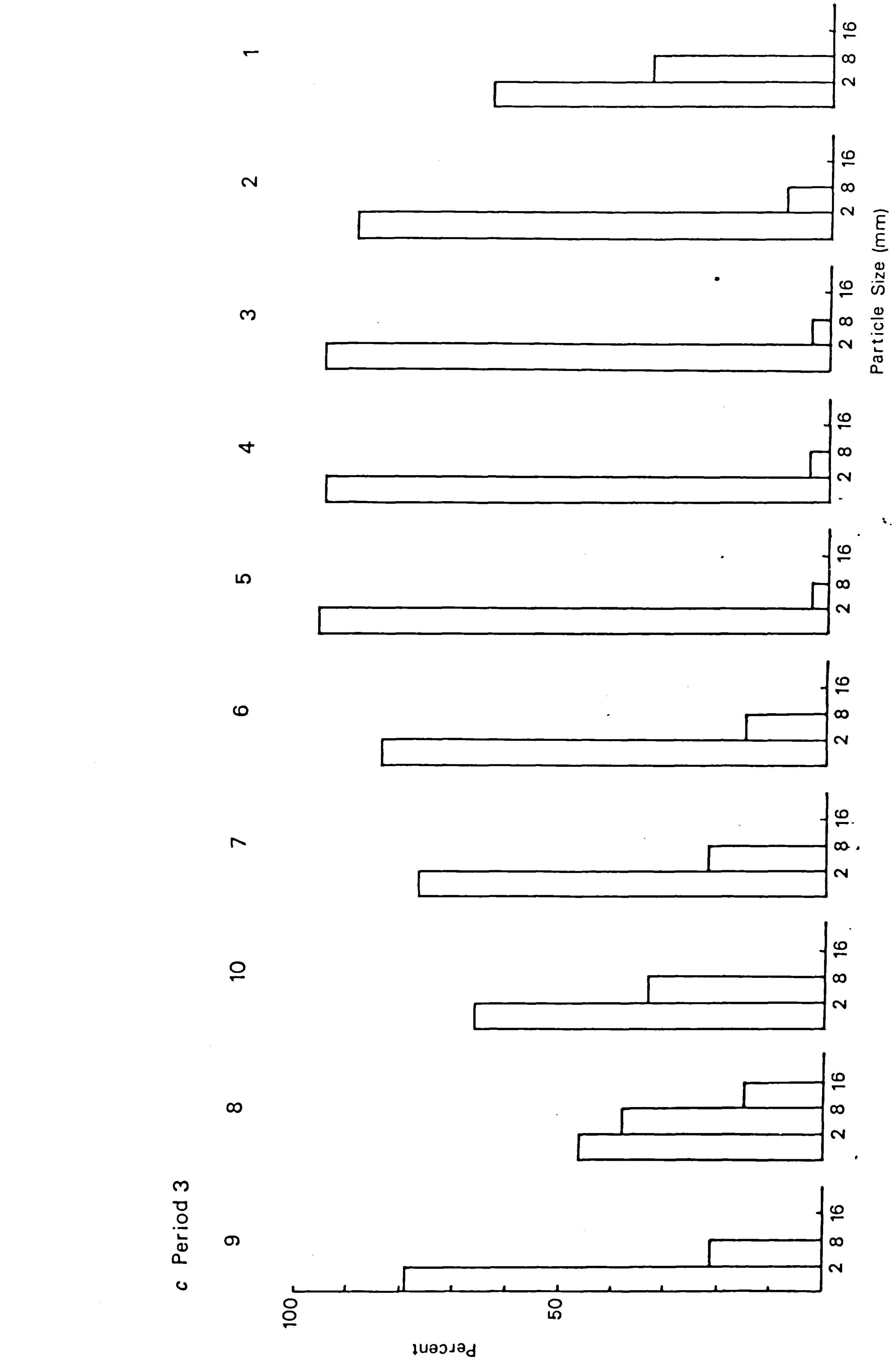


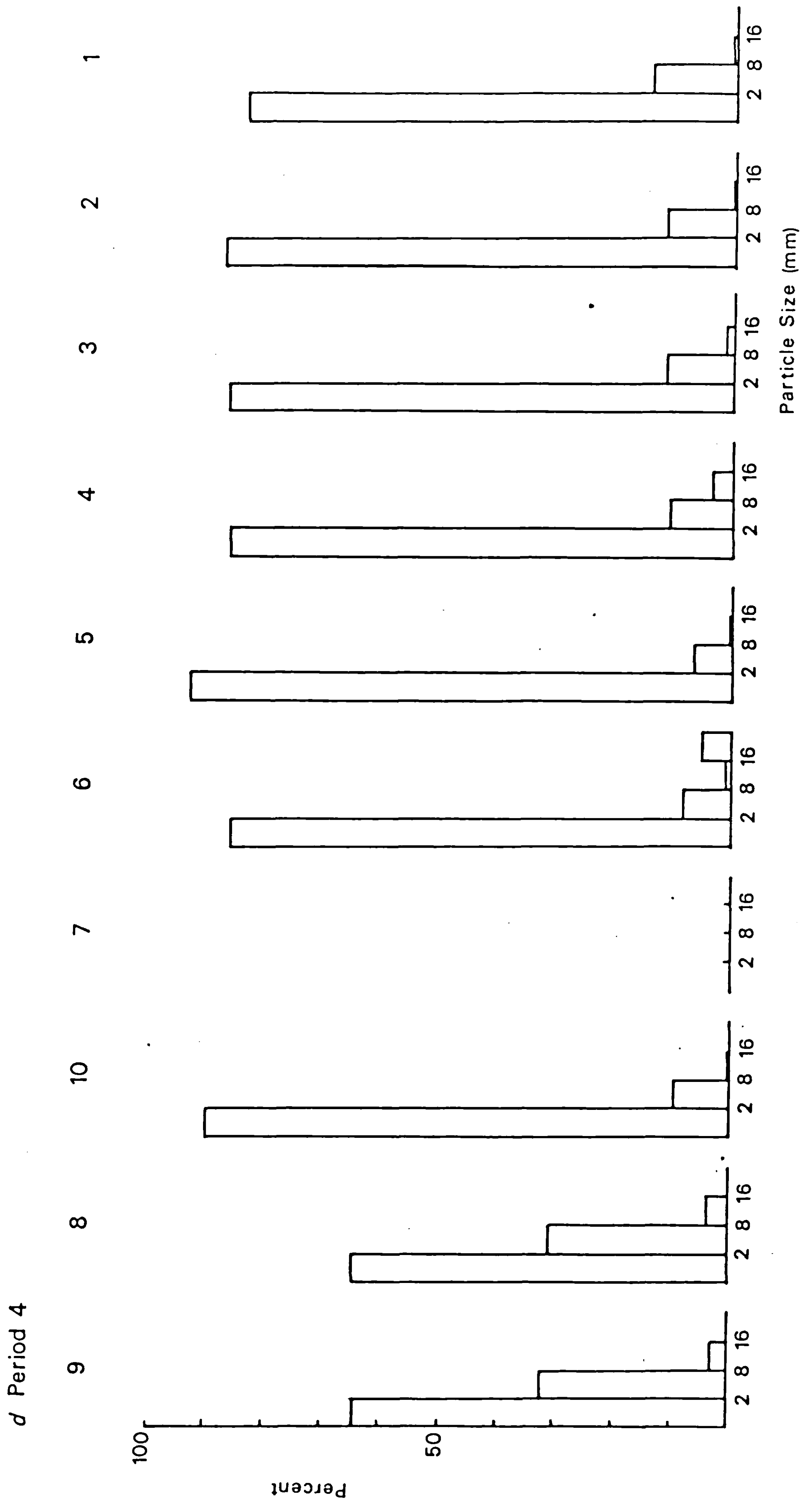
Figure 5·9 Coarse gravel size distribution on site 1 for periods 1 to 5

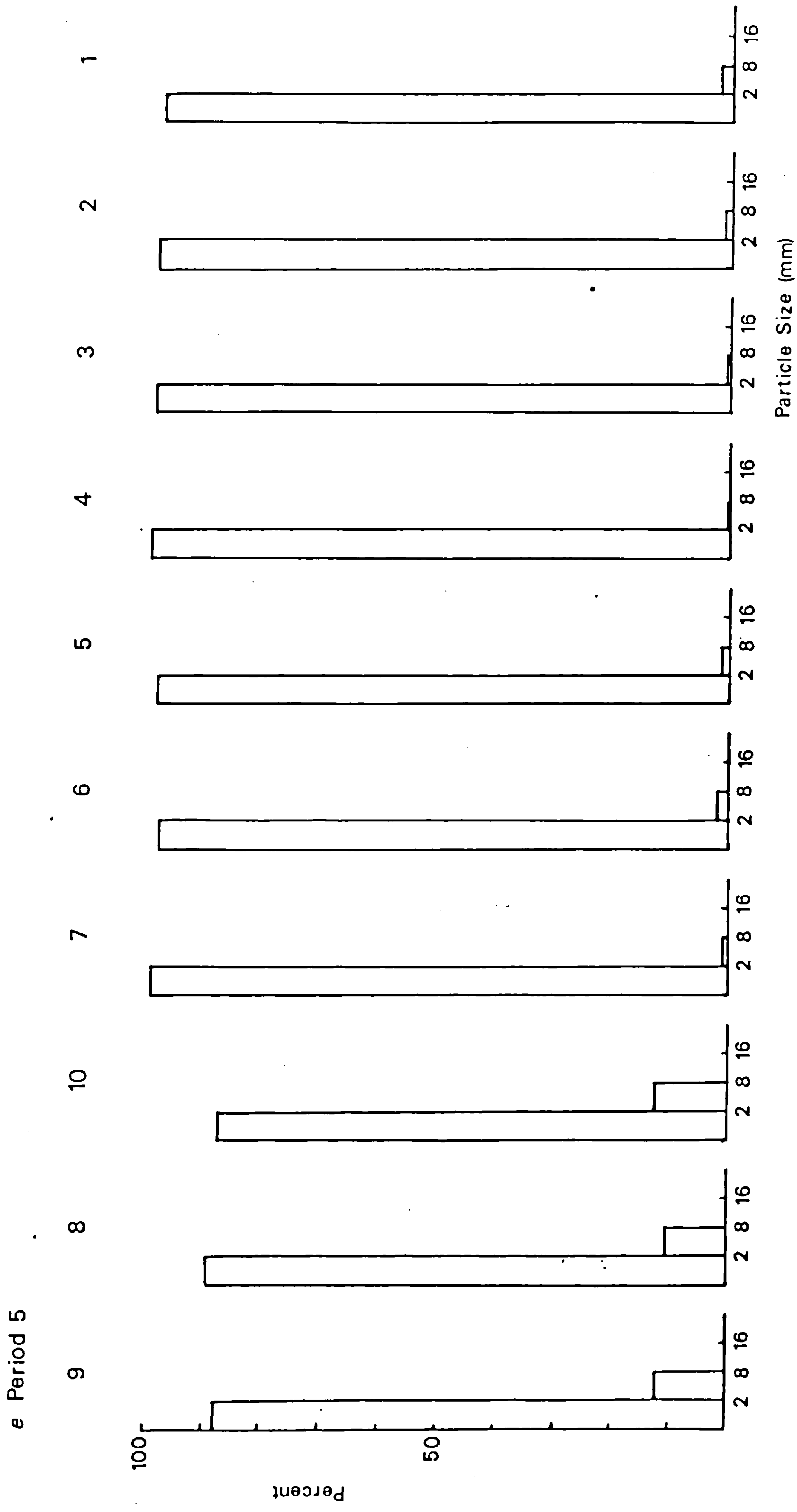
a Period 1




b Period 2
9
8
10
7
6
5
4
3
2
1
100
50
Percent
2 8 16
Particle Size (mm)


c Period 3
9
8
10
7
6
5
4
3
2
1
Percent
100
50
2 8 16
Particle Size (mm)


d Period 4
9
8
10
7
6
5
4
3
2
1
100
50
Percent
2 8 16
Particle Size (mm)


e Period 5
9
8
10
7
6
5
4
3
2
1
100
50
Percent
2 8 16
Particle Size (mm)

material, for example 34.5% at trough 1, may be due to the influence of one or two small stones. Figure 5.9d for period 4 differentiates between troughs 1-6 and 10 on marl and 8 and 9 on the conglomerate soil with the gravel fraction at 32.0% and 35.8% respectively for troughs 8 and 9, and between 5.1% and 22.7% on the marl. This also shows the emergence of stones 8-16mm in diameter lying on the marl for the length of the slope. This fits the general observation that scattered pebbles cover this lower slope and may represent some critical threshold velocity for entraining pebbles of this size in marl. Period 5 again brings out the distinction between troughs 8 and 9 with higher coarse fractions. On the marl coarse gravel is found in the troughs at several locations, but even so that gravel fraction is small, and for troughs 1 - 6 the percent of fines is above 90%.

The proportion of coarse material caught in the troughs is most variable for the periods of low sediment transported as the presence of one or two stones may make a considerable difference to the proportions. Thus on site 2 the percent of the coarse fraction varies from 0 to 78.6% for periods 2 and 3, but only between 21.5 to 51.3% during periods 4 and 5. On the marl the variability in the coarse fraction is related to a change of soil cover, but still there is some variation associated with the size of the sample. Thus in period 3 at trough 1 (figure 5.9c) 34.5% of the soil was greater than 2mm. In period 4 a larger proportion of coarse material was caught in the troughs so that for large storms it is possible that a greater proportion of gravel is transported further from the conglomerate over the marl.

The results suggest that there is evidence of armouring on the conglomerate soil with a notable increase of coarse sediment downslope compared to the amount of gravel material transported on the slope or lying within the soil matrix. On the marl there is no evidence of armouring as such, but there is a downwashing of coarser material on top of the marl from the conglomerate upslope.

5.4 Summary

Both the marl and conglomerate soils are suffering from rapid and substantial soil loss from surface wash. Estimates from small plots tend to be higher than denudation rates estimated for larger areas, but even so rates of soil loss ranging between 5 to 30 kg ha^{-1} yr^{-1} are substantial. In wetter years the figure may be expected to rise. The data presented here show that there is considerable variability in sediment yield caught in troughs not only for a given time between troughs, but also for the same trough at different times. Bryan and Campbell (1980) suggested that although there is great variability in runoff and erosion in the short term (due to for example, geology, relief, infiltration, and precipitation variability) the effect over the long term is to smooth out the variability so that semi-arid areas have relatively homogeneous high erosion rates. The evidence from this project does not support this contention. Rather, for small runoff events the loci of maximum erosion may be quite variable, but for large events and in the long term patterns of erosion are determined by characteristics of the slope and the soil. In this case maximum erosion will tend to occur in the middle slope sections where high slopes and low vegetation cover encourage erosion. Campbell and Honsaker (1982) found that there is considerable variability of erosion in time and space for badlands, and suggested that the variability with time was greater. This project confirms that there is considerable spatial and temporal variations in surface wash erosion in semi-arid areas.

Some of the factors affecting the variability of sediment yield and runoff in semi-arid areas are summarised, with respect to the study site and the published literature. These include observations on precipitation, slope angle and length, and vegetation cover.

Firstly with only five observation periods for the study site it is not sensible to come to any rigid conclusions about the relationship between rainfall and erosion. It was noted that for four occasions there was a positive relationship between accumulated rainfall for the period, and the sediment data. Other authors have examined this further using a more extensive data base. Pearce (1976) found that there was a high degree of correlation between total sediment loss with total rainfall kinetic energy in the runoff producing storms. Bryan and Campbell (1980) similarly suggest that there is a close relationship between precipitation and sediment yield, with a precipitation threshold necessary before surface runoff occurs. Bryan, Yair and Hodges (1978) suggest a precipitation intensity threshold of 0.5 mm/min for sandstones and 3 to 5 mm/min for shale in Alberta, Canada. Le Roux and Roos (1982) found a poor correlation between soil loss from Gerlach troughs on the one hand, with rainfall quantity and intensity on the other. However the product of rainfall quantity and intensity correlated highly with sediment yield.

Secondly the study suggests that on the marl there is a strong association between slope angle and sediment entrainment, but not on the conglomerate. Or at least on the conglomerate the relationship with slope is masked by other factors. It has been generally assumed that the runoff coefficient increases with increasing slope angle from 0° to 45°. Zingg (1940) suggested that total soil loss is a function of slope angle in the generalised form:

$$X = C.S^m$$

where X is the total soil loss in weight units, C is a constant of variation, S is the land slope in percent and m is an exponent of the land slope. He also suggested that increasing the length of slope decreased the total runoff. Smith and Bretherton (1972) also point out that a simple sediment transport law

$$Qs = f(Q,S)$$

is both theoretically and empirically sound. This suggests that sediment transport is related to slope.

Bryan (1979) undertook a laboratory experiment to examine the influence of slope on soil erodibility. His conclusion is that although there is no one relationship applicable for all the samples he used, the relationship between slope angle and rainsplash and wash is a polynomial function. This is more marked for the rainsplash with maximum transport rates occurring around 15° to 18°. However he notes that the relationship is blurred by other factors.

Field reports from semi-arid environments tend to point to poor correlations between slope and sediment entrainment. Kirkby and Kirkby (1974) found that rates of sediment transport by unchannelled processes were independent of slope gradient despite a strong dependence of particle distance travelled on grain size and slope. Yair and Klein (1973) studying slope and channel processes in arid conditions in Israel (annual precipitation 28mm) found no clear relationship between slope angle and slope runoff, and an inverse relationship between slope angle and slope erosion. They were dealing with debris covered slopes and suggested that with increasing slope angle there is a greater opportunity for infiltration associated with greater roughness and particle sizes. They quote from Evenari et al (1968) that there is a decrease in runoff coefficients with increasing slope from 27.1% runoff coefficient at a 10% slope to 14.1% runoff coefficient at a 10% slope for the Avdat watershed. Likewise Gerson (1977) found no correlation between suspended sediment concentration in runoff and environmental characteristics including slope angle, rainfall intensity, catchment area and lithology in the Mt Sdom area in the Dead sea region. Again his sites were very stony.

One explanation for these opposing observations is that on debris covered slopes or very stony soils the relationship between slope and sediment movement is affected by such factors as greater infiltration on steeper slopes reflecting larger sediment sizes, and higher inputs of energy required to entrain the sediment. On fine particle soils like the marls, there is no significant increase in roughness or infiltration with slope, and the effect of greater tractive force of the water on steeper slopes is to entrain more sediment.

Thirdly the data set from the conglomerate soil suggests that a decrease in the amount of bare ground (or inversely an increase in vegetation cover) can significantly affect erosion patterns. This effect is well known and documented by numerous authors, particularly with respect to crop cover and erosion (Adams et al 1959, Temple 1972). Some species of plant are particularly good at reducing soil loss around them so that the plant sits on a raised hummock of soil.

The evidence from the Gerlach troughs suggests that there is rapid and substantial surface wash erosion which would affect the gully head in two ways, firstly the sediment will fill up the gully if input exceeds output, and secondly the sediment may cause abrasion of the gully head itself. The abrasive effect of the sediment was not studied but is probably related to the sediment quantity and particle size. It is infered that the abrasive action will be greater on the conglomerate because despite lower quantities, the particles are larger.

An attempt was made to examine the effect of sediment transport into the gullies themselves. No account was taken of sediment transport rates out of the gully because no flow or sediment transport was observed or measured in the gully. The mean sediment loss per year is 0.12 cms m^{-2} yr^{-1} on marl and 0.068 cms m^{-2} yr^{-1}, representing volumes of 0.0012 m^{3} and 0.00068 m^{3}. Assuming a contributing area 3m wide and of varying lengths, the annual volume of sediment transported to the gullies are estimated. Table 5.10 shows that on both marl and conglomerate the annual sediment input to the gully for 1982 was small, even if the contributing areas extend to the watershed. It should be remembered that the erosion rate is less around the gully head than on the middle section of the hillslopes, so that the mean rate is probably an overestimate of sediment movement to the gully head. As an extreme example the maximum volume of input is calculated using the highest rates of sediment loss measured on the slopes for the maximum contributing area. On the conglomerate this still leads to slow sedimentation rates, although rates are faster on the marl.

In conclusion overland flow production appears to be greater on the marl than the conglomerate soil so that for a given storm a greater proportion of the water will occur as overland flow on the marl. The marl also has a greater propensity towards surface wash erosion, partly due to greater discharges, but also to a greater sensitivity to erosion on steep slopes and smaller particle sizes. On both sites 1 and 2 the location of greatest sediment movement is in the middle slope section so that the quantity of sediment being produced around the gully heads is relatively small. Whilst rates of wash erosion are substantial, they will

Table 5.10 Estimates of sediment yield to the gully head

Assume:

1 The contributing area is 3m wide
2 The length of the contributing area is variable
3 Volumetric soil loss m^{-2} yr^{-1} on marl is $0.0012m^{3}$
4 Volumetric soil loss m^{-2} yr^{-1} on conglomerate is $0.00068m^{3}$
5 Volume of marl gully is $180m^{3}$
6 Volume of conglomerate gully is $208m^{3}$

Length of Contributing Area (m)	Area (m^2)	Sediment Yield on marl (m^3)	Years to fill in gully	Sediment Yield on conglomerate (m^3)	Years to fill in gully
1	3	0.0036		0.0020	
5	15	0.0180		0.0102	
10	30	0.0360		0.0204	
20	60	0.0720		0.0408	
60	180	0.2160	2500	0.1224	1699
220	660	0.7920	227		
220*	660	17.0	10		
60*	180			0.2.196	94

*Maximum rate measured on the marl = 0.0258 m^{3} m^{-2} yr^{-1}
*Maximum rate measured on the conglomerate = 0.0122 m^{3} m^{-2} yr^{-1}

not fill in the gullies in the near future. It is thought that most of the debris in the gullies is likely to come from the gully sides rather than from the catchment areas. In the long term gully wall collapse may lead to shallower angles on the gully sides and gully stabilisation if there is not sufficient flow in the gully to evacuate the sediment. The effect of sediment movement on the conglomerate site is to produce a lag of coarse debris on the lower slopes due to armouring, whereas on the marl wash redistributes gravel material on the lower marl slopes from the conglomerate.

CHAPTER 6 SUBSURFACE HYDROLOGY

There are two major areas for discussion with respect to the subsurface hydrology of gully head catchments, soil moisture content and movement. The soil moisture content will vary with time, depth and hillslope location as a response to climatic and edaphic controls. Two aspects of the soil moisture content are of interest. Firstly, the development of near saturated soils and the conditions for that development are important with respect to flow rates, raising pore water pressures on the gully sides and inducing saturated overlandflow. Secondly, on a more general theme the accurate monitoring of soil moisture conditions provides a rare source of data for the effects of drought on soil conditions in south east Spain and provides a measure of the soil moisture deficit which is relevant to botanical and agricultural problems. The first part of the chapter is concerned with the measurement and analysis of the soil moisture status, and the second part considers the likelihood of soil moisture movement, in particular to the gully heads.

6.1 Soil Moisture Content

Techniques And Experimental Design

The soil moisture conditions of the site were monitored with time, depth and hillslope location using a Wallingford neutron probe and a network of access tubes. This system was used because once the access tubes are inserted there is no further disturbance to the soil, and the soil moisture content of the same sample positions can be measured accurately and repeatedly. The operational procedures for installing the access tubes and using the probe were followed as outlined in Eeles (1969), Bell (1973), and the Users Handbook, Institute of Hydrology (1981).

Altogether nine permanent access tubes were used with five on site 1 and four on site 2 (figure 2.5). On site 1 tubes numbered 1, 2, and 3 were placed in a row across the top of the gully head in order to ascertain differences in soil moisture just above the head cut itself. Tubes numbered 4 and 5 were placed upslope from tube 2 to examine the variation in soil moisture along a flow line into the gully. On site 2 all four tubes were placed on a flow line into the gully head.

A tenth "temporary" access tube was used for the calibration procedure required to convert the count rate measured by the probe to soil moisture figures (Moisture Volume Fraction). .The count rate is a measure of the number of neutrons reflected back to the probe, which is affected primarily by hydrogen atoms, and to a lesser extent boron, chlorine, and iron atoms. The most abundant source of hydrogen is in soil moisture, but also includes bound water and organic matter. The count rate is also affected by the bulk density of the soil. For a given soil, assuming a constant chemical background and bulk density, changes in the count reading are ascribed to changes in soil moisture content. The calibration should be undertaken for each soil type for accurate results, although standard curves are now available for some common (British) soils. The calibration procedure is a time consuming process, involving the extraction of a large number of samples covering a range of soil moisture values, and is subject to sampling errors especially compression of the sample which changes the bulk density.

To perform the calibration a temporary access tube was inserted near one of the permanent tubes and readings were taken at known depths. Undisturbed soil samples of known volumes were taken from the corresponding depth, and the soil moisture was determined thermogravimetrically (Reynolds 1970). The field calibration was undertaken during the April field session. The soils were quite dry, and in order to obtain the higher range of moisture values one tube on sites 1 and 2 were irrigated more than 24 hours prior to sampling. Altogether the temporary access tube was used five times, with two runs on site 2 (one wet, one dry), and three on site 1 (two wet and one dry). The calibration of each tube took several days to complete so that each morning probe readings were taken from the remaining profile to check that the soil moisture contents had not changed appreciably.

The Users Guide recommends that six samples are collected for each horizon to counteract sampling errors. However as it was difficult to take good samples from the stony soils, and since time was short, between 1 to 6 samples were collected for each level at 20cms intervals. On the marl and clay horizons the samples were taken with an auger designed to extract undisturbed cores with a volume of 473.8cc. The auger bit was damaged on the stony soils so the samples on the conglomerate were taken using a bulk density ring with a volume of 1000cc. Each sample was weighed on site to an accuracy of 0.1 grm, and then oven dried at 105°C and reweighed to measure the soil moisture content.

The calibration equation is a regression of the count rate (R) divided by a standard (Rs) on the moisture volume fraction where

$$MVF = \frac{\text{volume of water in soil by drying}}{\text{volume of sample}}$$

and

$$MVF = m \cdot R/Rs + c.$$

The Rs value is usually taken as the count rate measured when the probe is inserted into an access tube in a large body of water. This has two advantages, firstly it identifies trends in count readings should the probe malfunction, and secondly it allows comparison between probes. It is recommended that the Rs is measured every day the probe is used before going into the field. Over a long period if there is no trend the values can be averaged. The rig required to measure the water standard was quite awkward to set up in the field, and the Rs value was finally taken as the shield count determined by the manufacturers of 625. Providing there was no drift in the probe readings this will not affect the final results, although the calibration equation obtained is not directly comparable to others determined using a water standard.

The calibration was obtained firstly for each soil type, the marl, clay and conglomerate, assuming differences in chemical composition and bulk density would affect the regression, and secondly for all the data points. All the data points are plotted in figure 6.1 which shows a positive relationship between count rate and MVF. Table 6.1 summarises the regression results, and

Figure 6.1 Calibration of probe values

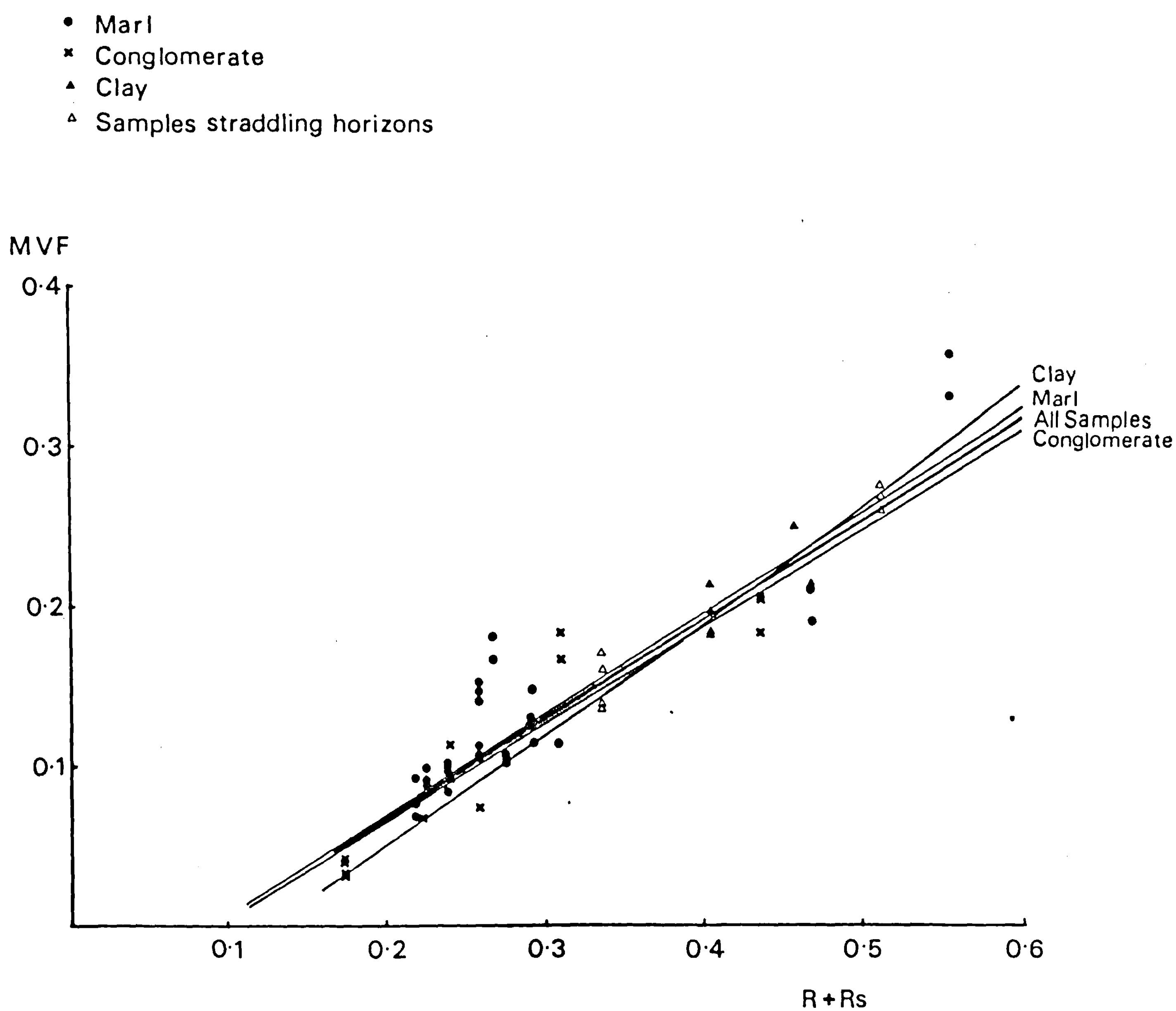


Table 6.1 Regression results for calibration of neutron probe

1 All samples

$n = 59$
$A = -0.0622$
$B = 0.6464$
$r^2 = 0.9257$

2 Samples for marl only

$n = 31$
$A = -0.0633$
$B = 0.6579$
$r^2 = 0.9105$

3 Samples for conglomerate only

$n = 13$
$A = -0.0611$
$B = 0.6311$
$r^2 = 0.8865$

4 Samples for clay only

$n = 7$
$A = -0.0880$
$B = 0.6965$
$r^2 = 0.5245$

shows the curves for the marl and conglomerate soils have very similar gradients and intercepts.

The Users Handbook shows that the calibration curves for finer grained sediments plot above the coarser grained soils, with a slightly steeper gradient. The regression lines for the marl and conglomerate fits in with these expectations as the marl plots above the conglomerate with a steeper slope. The regression for the clay horizon does not follow the accepted pattern which may be a result of a small number of samples. When all the samples are taken together the coefficient of correlation rises slighly to 0.962. Several tests were undertaken to examine the robustness of the calibration.

Firstly the regressions are tested to determine whether they are significant, as although the coefficients of correlation are high, the sample numbers tend to be low. The significance of the correlations was tested using the F value which shows that the regression equations for the marl, conglomerate, and all sample cases are significant at the 0.001 level, but the clay correlation is only significant at the 0.05 level.

However this test does not indicate whether the regressions are significantly different from each other. To compare the marl and conglomerate the gradients of the two regressions are compared using the test

$$t = b_1 - B_0 / \hat{s}b_1$$

b_1 = gradient of calibration with the conglomeratic soil
B_0 = gradient of the calibration on marl

$$\hat{s}b_1 = \hat{s}_{yx} / \sqrt{\textstyle\sum (x_i - \bar{x})^2}$$

$$\hat{s}_{yx} = \sqrt{\frac{\sum (y_i - \hat{y}_i)^2}{N-2}}$$

$$t = 0.6311 - 0.6579 / 0.1899 = -0.141$$

With N-2 degrees of freedom (11) $t_{0.001} = 4.437$ so the null hypothesis that there is no significant difference between the gradients of the calibrations on marl and conglomerate soils is accepted.

A partial correlation test was employed to determine the effect of bulk density on the full data set (n=59). The correlation between MVF and count rate is examined whilst holding for the effect of bulk density using the formula

$$r_{ij.k} = \frac{r_{ij} - (r_{ik})(r_{jk})}{1-r_{ik}^2 \;.\; 1-r_{jk}^2}$$

$$= \frac{0.9621 - (-0.280 \;.\; -0.243)}{0.922 \;.\; 0.941}$$

$$= 0.960$$

This shows that the bulk density has little affect on the calibration for the sample given here as the r value whilst holding for bulk density is 0.960 whereas the r value for the single regression is 0.9621. Thus considering the small number of samples for the separate horizons, and the small differences between the calibration regressions for marl and conglomerate, it was decided to use the regression involving all the data points for the calibration.

Table 6.2 lists the data needed to plot the confidence limits at the 99% level for the regression using all the samples (n = 59). This shows that for a R/Rs value between 0.3 to 0.35, 99 out of 100 observations will lie within a range of 0.014 MVF. Towards the outer limits of the data set the range increases to 0.03 MVF. These data are used in the following section to describe the significance of changes in soil moisture values with respect to the standard errors of the regression.

Soil Moisture Variability

Values for soil moisture content are described for each field campaign to show the values and variations to be found with time, depth, and location.

July 1982

For the summer period soil moisture data is only available for the marl on site 1. Figure 6.2 shows for each tube the changes in soil moisture between 16 and 30 July 1982 for various horizons. The first impression obtained from this graph is the constancy of the soil moisture values for all horizons. At tube 2, for

Table 6.2 Calculations for the 99% confidence limits

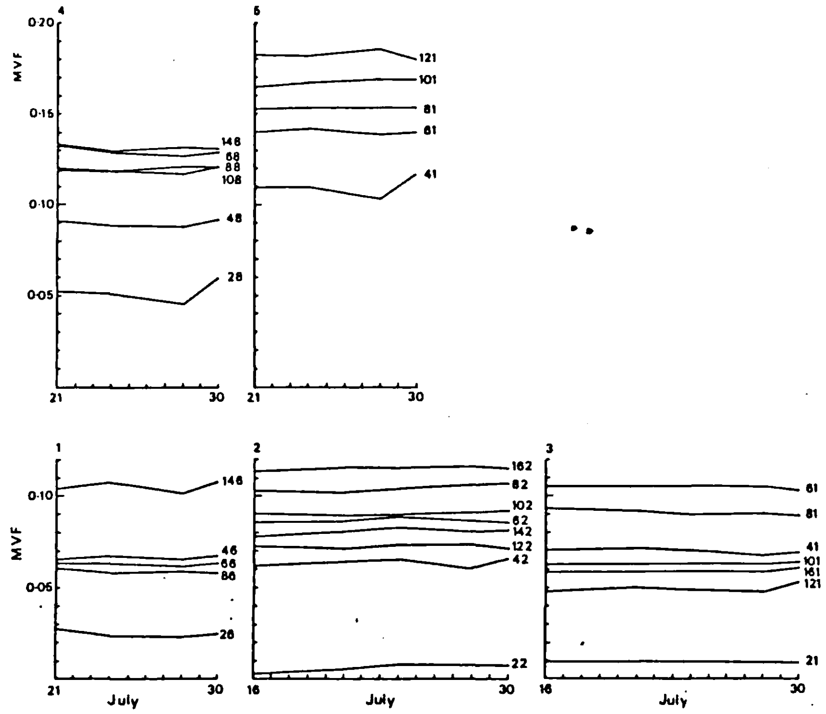
R/Rs	Upper Limit MVF	Calculated MVF	Lower Limit MVF	Range MVF
0.10	0.0184	0.0024	-0.0136	0.032
0.15	0.0481	0.0348	0.0215	0.027
0.20	0.0778	0.0671	0.0564	0.021
0.25	0.1079	0.0994	0.0909	0.017
0.30	0.1388	0.1317	0.1246	0.014
0.35	0.1710	0.1640	0.1570	0.014
0.40	0.2048	0.1964	0.1880	0.017
0.45	0.2394	0.2287	0.2180	0.021
0.50	0.2742	0.2610	0.2478	0.026
0.55	0.3093	0.2933	0.2773	0.032
0.60	0.3446	0.3256	0.3066	0.038

Range of soil moisture in the field:

Marl - 0.05 to 0.32 MVF

Conglomerate - 0.03 to 0.22

Figure 6·2 Soil moisture values in July 1982 on marl for different depths (cms)

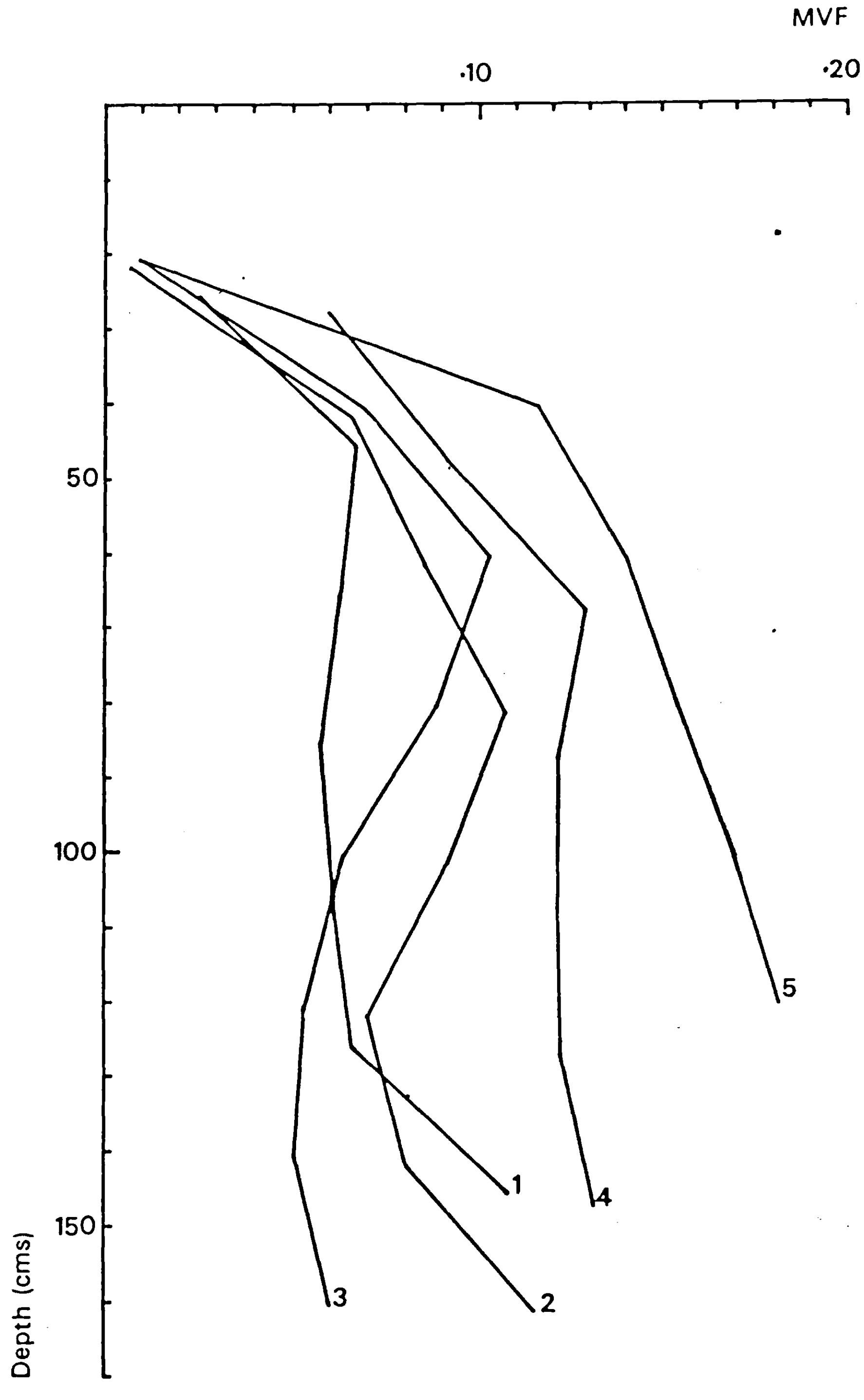


example, between 16 to 30 July the MVF varied by 0.000 to 0.004 at different depths. Such changes are not significant as at these low values of soil moisture, for a given count rate the MVF may vary by between 0.02 and 0.03 at the 99% confidence level. The rainfall noted on 18/19 July 1982 (Chapter 3) has made no impact on the soil moisture conditions.

The highest soil moisture content is 0.186 MVF at tube 5, 121 cms depth, which is approximately 37% of the saturated value, however elsewhere values tend to be lower. As expected, the lowest moisture values recorded are for the shallowest horizons with less than 0.02 MVF at tube 2, 22 cms and tube 3 at 21 cms. These low values are a result not only of drainage and evapotranspiration, but may be affected by the boundary problems associated with using the neutron probe. The probe cannot be used to take readings near the surface due to the loss of neutrons to the atmosphere. In wet soils the radius of the sphere of influence of the probe is about 20 cms, but in dry soils it may reach 30cms (Bell 1973). Excluding the 20-30cms layer, the difference between the wettest and driest horizon is about 0.06 MVF within approximately 1m depth.

At tube 5 the soil moisture increases with depth from 0.110 to 0.183 MVF between 41 to 121cms. At the other locations the variations with depth are different. Figure 6.3 gives a better idea of variation with depth, and shows that at tubes 1, 2, 3, and 4 the soil moisture increases, decreases, and then starts to increase again. The variations with location are also set out clearly in this figure. Tubes 1, 2 and 3 show soil moisture variations across the gully head, tube 2 being directly above. Despite a distance of 2 to 3m between the tubes their profiles are not too similar. Tube 2 is not the driest as one may expect if the gully head cut was draining the soil immediately upslope, indeed between 70-130cms the highest soil moisture values are recorded here. A comparison of tubes 2, 4, and 5 shows that soil moisture content increases upslope. This may reflect several factors. Firstly the lower soil moisture values around tubes 1, 2, and 3 may indicate the draining capacity of the gully head on a relatively large area upslope. Secondly there may be variations in the soil profiles upslope which alter infiltration rates or

Figure 6·3 Soil moisture variation with depth on 30 July 1982



moisture storage capacities. Thirdly the increase in soil moisture upslope may reflect the influence of the consumptive use of water by plants. Photographs 2.4 and 2.5 show that whilst tubes 1, 2, and 3 are sited in relatively dense vegetation, tubes 4 and 5 are in more open areas. It may be over long periods that soil moisture losses by evaporation alone may be less than those by evapotranspiration despite the greater influence of shading reducing the evaporation component. Finally the change in soil moisture downslope may represent divergence of subsurface water down the spur.

November/December 1982

Figure 6.4 shows that soil moisture contents in the marl were more variable in this period as a result of the storms occurring on 5 to 6, and 26 November 1982 (see section 3.2). The temporal patterns illustrated in the graphs can be classified into three groups:

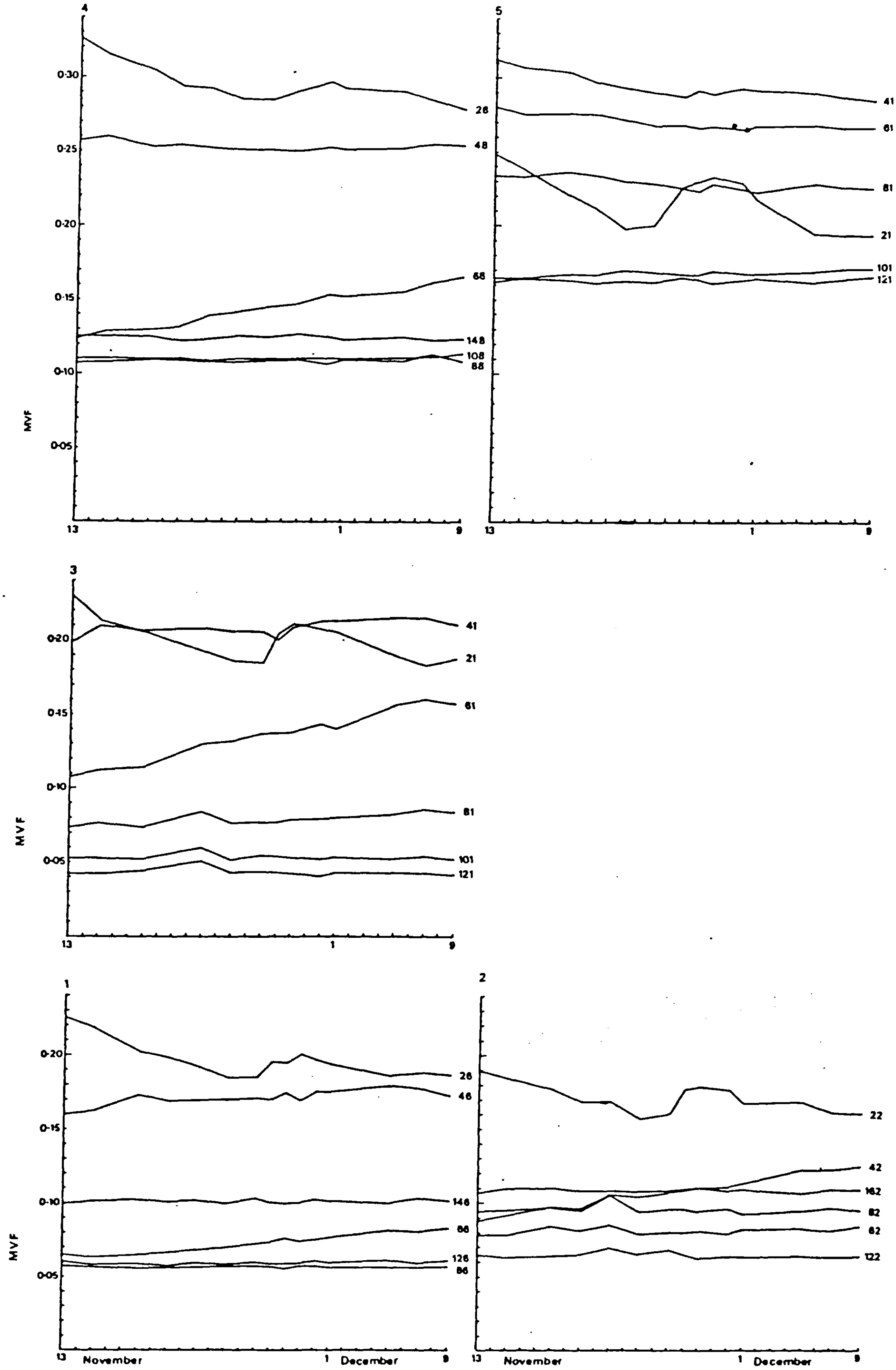
1 Patterns of soil moisture responding to individual storms (20-30cms)
2 An overall trend of increasing or decreasing soil moisture where storms may or may not form a small peak
3 No significant change with time (80cms +).

These divisions represent the decreasing effect of the storm water with depth, as the time taken for the soil moisture to reach the lower horizons diffuses individual events.

Between 13-25 November the shallowest horizons are draining as the storm water from 6 November either evaporates or percolates downwards. The MVF values decrease by 0.032 to 0.048 in the depth range 21-28cms for the five tubes which are greater than the expected variation in MVF for a given R/Rs value, and assuming a porosity of 0.5 in this horizon represents a loss of 6 - 10% of the total available storage. The readings on 26 November were taken between 1.15pm at tube 5 and 4.30pm at tube 4, approximately 10-13 hours after the start of the storm, and already increases in soil moisture are recorded.

Between 40 and 68cms a different picture emerges, although there is some variation between sites for the period 13 to 25 November. Soil moisture tends to be increasing at several locations, for

Figure 6·4 Soil moisture values in November/December 1982 on marl for different depths (cms)



example at tube 2 there are increases of 0.017 MVF at 42cms and 0.001 at 62cms, and at tube 3 soil moisture increases by 0.008 MVF at 41cms and 0.024 MVF at 61 cms. Not all these increments are significant, however where there are successive changes in soil moisture (either increasing or decreasing) over several days it suggests that there is a genuine trend in the data.

The effect of the storm is brought out in table 6.3 which shows that the deeper the horizon the later the peak of soil moisture and the smaller the change in soil moisture content. At 21/22 cms the recorded peak is on 28 November, at 26cms on 29 November, and at 28cms on 30 November. The change in the MVF between 24 November and the day of peak soil moisture ranges between 0.012 to 0.033 which are probably significant (table 6.2). In the 40-50cms layer the response is more variable with the day of maximum soil moisture ranging from 30 November at 41 cms and still increasing by 9 December at tube 2 at 42 cms. The increases in MVF are smaller too with a range of 0.003 to 0.020 which are too small to be significant. Further down the profile the storm has made no impact. There is only a difference of 0.1% of the saturated MVF (0.5) at tube 1, 86 cms, and tube 4, 88 cms between 13 November and 9 December.

Figure 6.5 illustrates the soil moisture distribution on 28 November on the marl and shows a classic wetting front. The soil moisture values in the 20-30cms layer are high (around 0.200 MVF) but decrease rapidly between 40/50cms and 60/70cms to around 0.100 MVF. Not only is there a large variation in the MVF, but this is accomplished in a short distance, so that hydraulic gradients will be steep.

The variations with location are similar to the July period. Soil moisture increases upslope, but across the gully head tube 2 is now the driest to 60cms, but still wetter below 80 cms. This may indicate rapid drainage under wetter conditions. Tube 1 (which is downslope from a rill) does not have the greatest soil moisture content so that the position of the rill has not affected soil moisture conditions nearby.

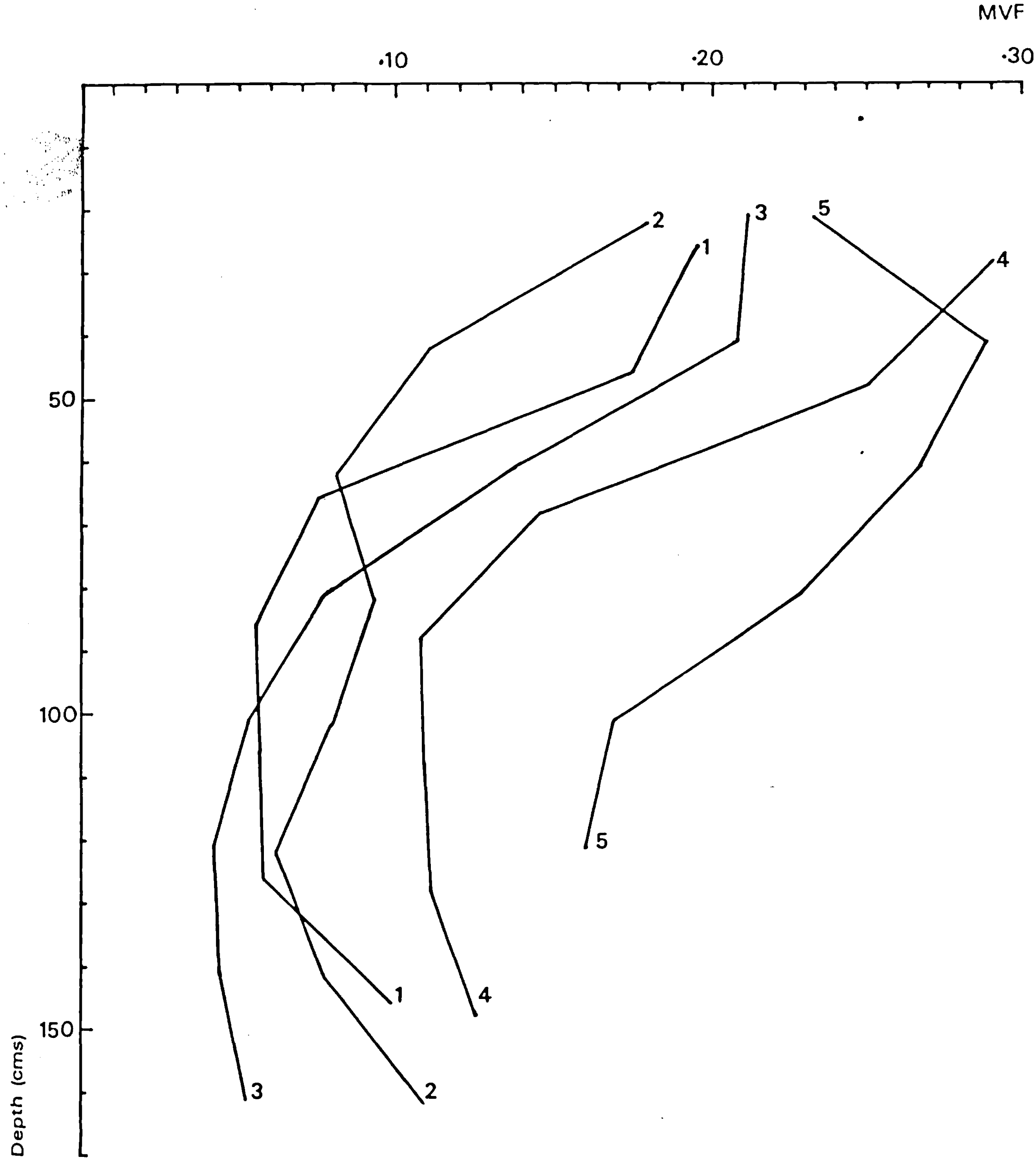
Table 6.3 The effect of the storm on marl

Tube	Depth cms	Day of Peak	Peak MVF	Size of Peak	% of Porosity (Porosity = 0.5)
1	26	29 Nov	0.201	0.016	3
2	22	28 Nov	0.179	0.022	4
3	21	28 Nov	0.211	0.025	5
4	28	30 Nov	0.297	0.012	2
5	21	28 Nov	0.233	0.033	7
1	46	5 Dec	0.180	0.010	2
2	42	9 Dec	0.155	0.020	4
3	41	5 Dec	0.215	0.009	2
4	48	7 Dec	0.255	0.004	1
5	41	30 Nov	0.293	0.003	1

Table 6.4 The effect of the storm on conglomerate

Tube	Depth cms	Day of Peak	Peak MVF	Size of Peak	% of Porosity (Porosity = 0.35)
1	30	27 Nov	0.185	0.048	13
2	24	26 Nov	0.170	0.050	14
3	24	29 Nov	0.223	-	-
1	50	28 Nov	0.190	0.016	4
2	44	28 Nov	0.166	0.018	5
3	44	29 Nov	0.179	-	-

Figure 6·5 Soil moisture variation with depth on 28 November 1982 on marl



On site 2 tubes 1 and 2 were monitored between 24 November and 8 December, and tube 3 between 26 November and 8 December. The profile of tube 1 consists of conglomerate from 0 to c117 cms, the clay horizon from 117-152 cms, and marl from then onwards. The other tubes lie completely within the conglomerate horizon. Figure 6.6 plots the soil moisture variations with time and depth for site 2. This diagram shows two interesting points, firstly the influence of the storm on the soil moisture contents, and secondly the difference with location.

The graph again shows the division of response patterns into three groups, those which respond quickly to inputs, those affected by prolonged rises and falls in moisture, and finally those unaffected by the rainfall. Table 6.4 lists the day of peak soil moisture and the increse in the MVF for different depths and tubes. At tube 2 the peak occurred on 26 November, the same day as the storm, but occurred a day later at tube 1 by 30cms. At tube 3 the soil moisture continued to increase until 29 November. Increments were about 14% of the total saturated volume (assuming porosity equals 0.354) and the size of the peaks are greater than the variation of MVF for a given R/Rs value (table 6.2). By 44/50cms the peak days were 28 and 29 November with only 5% increments in soil moisture content but even at this depth the change in soil moisture is probably significant. At tube 1, soil moisture increased from 28 November at 90 and 110cms. This may partly reflect the boundary effect with the clay horizon with a change in permeability allowing a build up of moisture in the soil immediately above. Below 110cms there is little variation in soil moisture with depth.

The soil moisture distributions look very different between tubes 1, 2, and 3. The highest values are recorded at tube 1 with 0.224 MVF (64% assuming porosity equals 0.35) and the range of soil moisture at tube 1 is 0.090 MVF over 170cms. At tubes 2 and 3 the range in soil moisture values increases to 0.110 MVF and 0.150 MVF respectively. Also at tubes 2 and 3 there appears to be a division of soil moisture at depth where values are low (around 0.070 to 0.080 MVF) and stable between 124 to 184cms, and the overlying soil where soil moisture changes are more variable.

Figure 6·6 Soil moisture values for November/December 1982 on the conglomerate for different depths (cms)

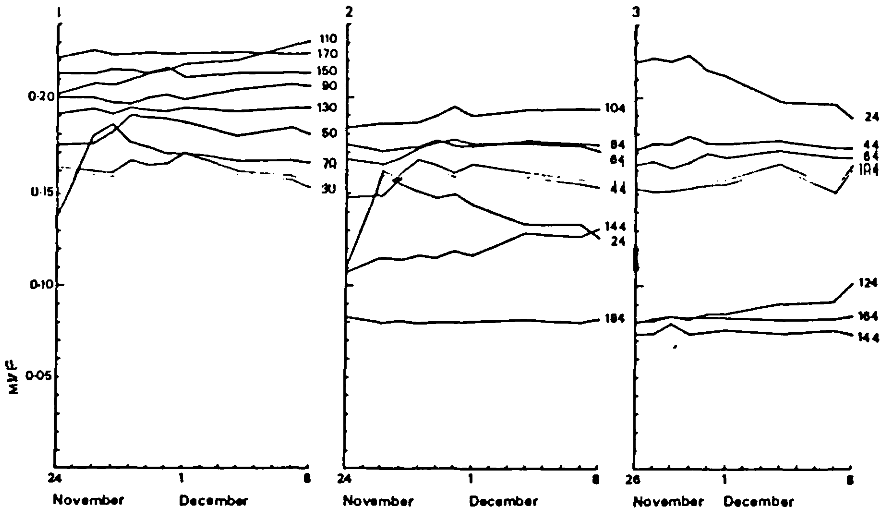


Figure 6.7 plots the soil moisture distribution with depth for 28 November. This shows that tube 1 is the wettest with the moisture generally increasing with depth. At tube 2 the soil moisture increases from 0.151 to 0.187 MVF at 104 cms, and then decreases rapidly to 0.080 MVF at 184 cms. At tube 3 the soil moisture decreases from 0.220 at 24 cms to 0.151 at 104cms and then falls rapidly to 0.084 MVF at 124cms.

April/May 1983

The overall impression of soil moisture variability on the marl is of an uninterrupted decline of soil moisture values (figure 6.8). The rate of fall in soil moisture is greatest for the shallowest horizon and declines as shown in table 6.5a as the influence of evaporation penetrates downwards. Some horizons below c80 cms have slight increases of soil moisture for example at 86cms tube 1 the MVF rises by 0.005, although this increase is not significant, the run of successive increases may suggest a genuine upward (albeit very small) trend in the data. On the whole without any further rain these values will fall towards the July figures. Indeed figure 6.9 (showing the situation on 3 May) already resembles figure 6.3 for the previous July with a profile showing an increase in soil moisture to about 50 cms and then decreasing.

A similar drawdown pattern occurs on site 2. Figure 6.10 shows that the rate of drawdown is different between locations and that the variation in soil moisture decreases upslope. Firstly the soil moisture decreases with time during the spring. However table 6.5b shows that the greatest decrease is in the 60 to 90 cms depth, unlike the marl where the largest decrease is for the shallowest horizon. This suggests that by early April the upper horizon has already drained considerably.

Figure 6.10 clearly shows that the range of soil moisture values with depth is greatest at tube 1 and least at tube 4. By 3 May at tube 4 there is only a difference of 0.021 MVF between the wettest and driest horizon, which is about 6% of the saturated value. Figure 6.11 shows the soil moisture profile with hillslope location. The range of soil moisture at tube 1 is due to low moisture values near the surface and large values with depth.

Figure 6·7 Soil moisture variation with depth on 28 November 1982 on conglomerate

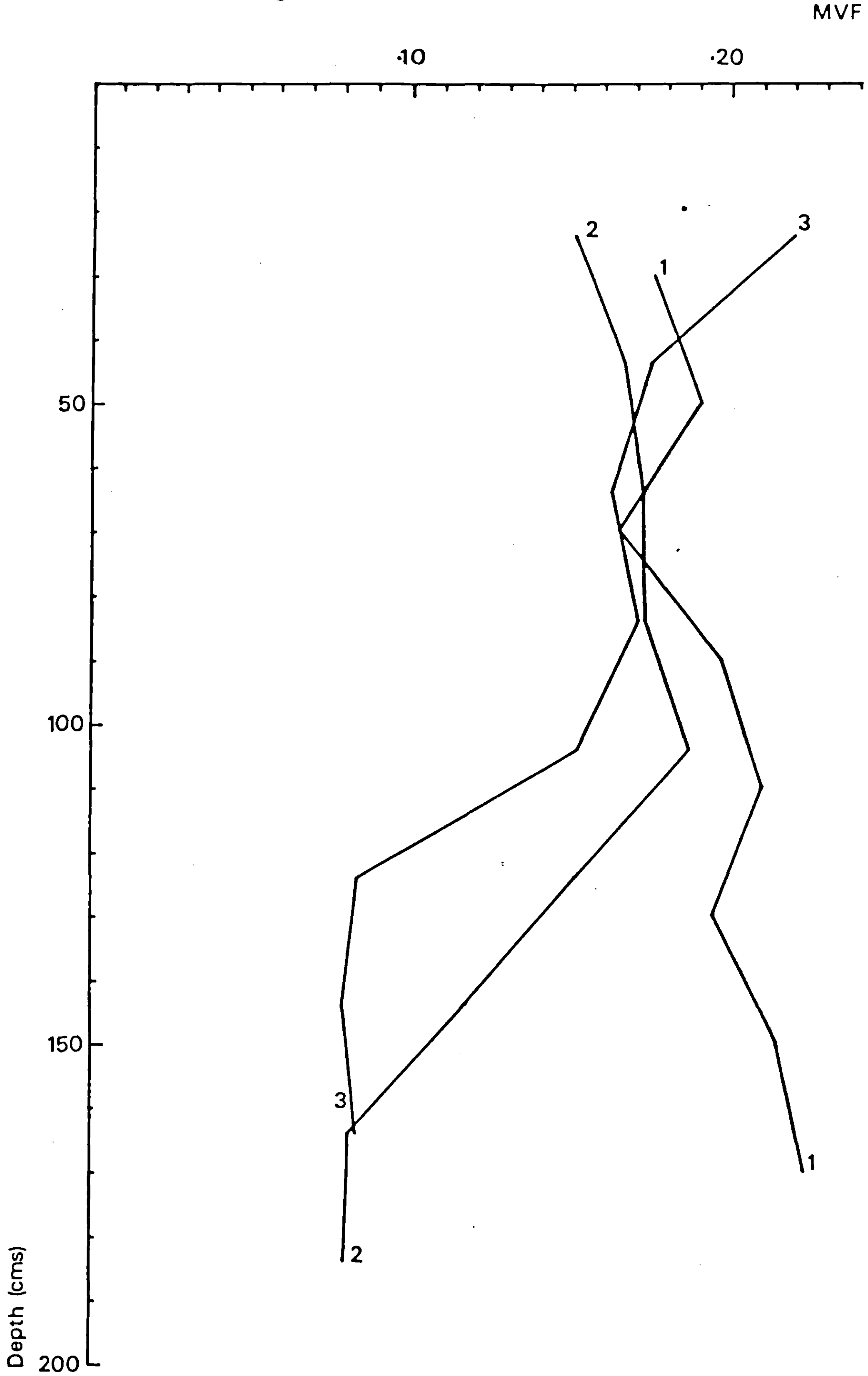


Figure 6·6 Soil moisture values for April/May 1983 on marl for different depths (cms)

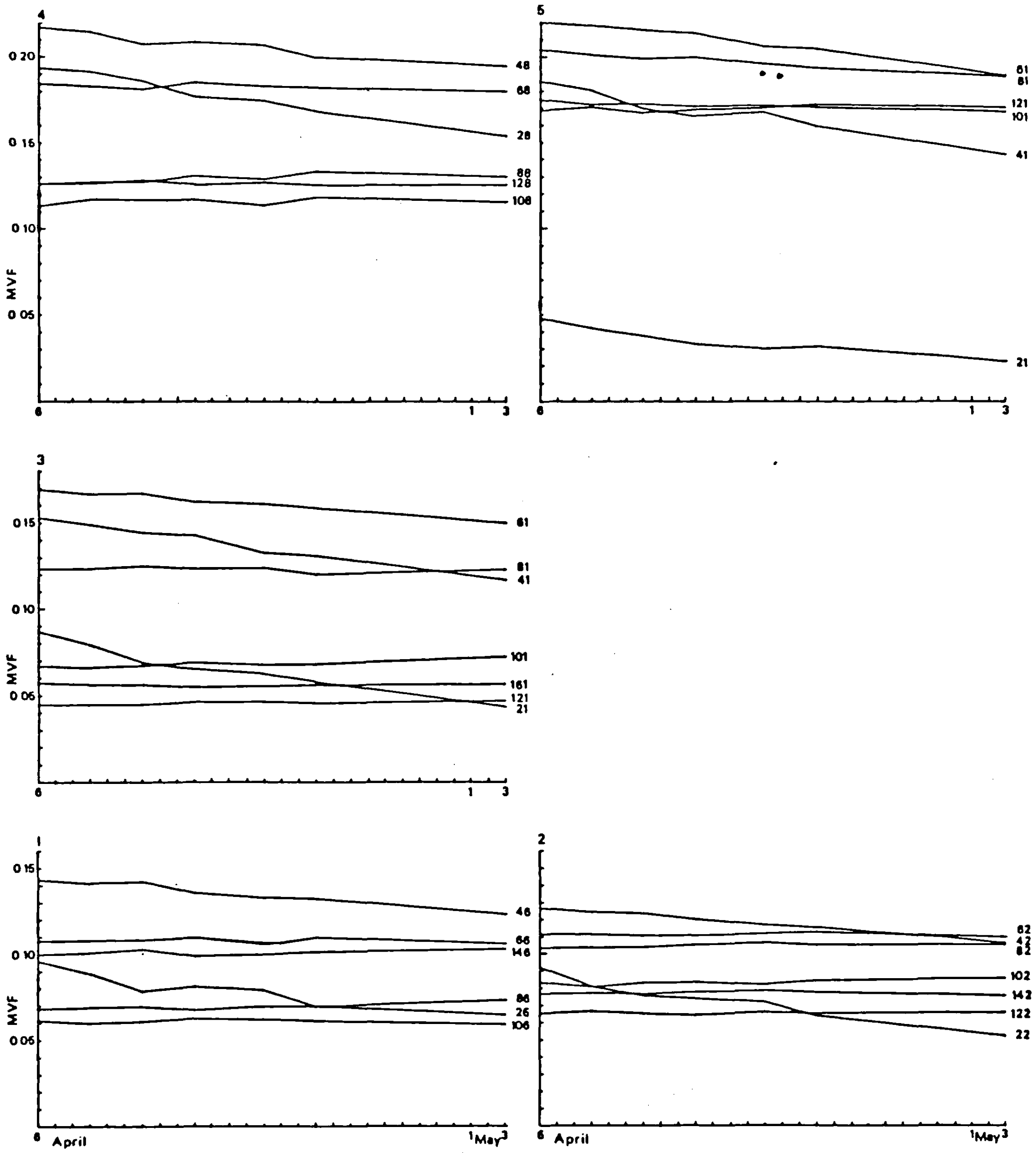


Table 6.5 Changes in soil moisture between 6 April and 3 May 1983

Marl

Tubes	Horizons			
	1	2	3	4
	20-30	40-50	60-70	80-90 cms
1	-0.031	-0.020	-0.002	+0.005
2	-0.040	-0.021	-0.002	+0.000
3	-0.043	-0.036	-0.019	-0.000
4	-0.040	-0.023	-0.005	+0.004
5	-0.024	-0.042	-0.032	-0.016

Conglomerate

	1	2	3	4
	20-30	40-50	60-70-	80-90 cms
1	-0.010	-0.032	-0.050	-0.061
2	-0.030	-0.024	-0.027	-0.020
3	-0.055	-0.037	-0.028	-0.020
4	-0.038	-0.031	-0.017	-0.010

'-' = a decrease in soil moisture
'+' = an increase in soil moisture

Figure 6·9 Soil moisture variation with depth on 3 May on marl

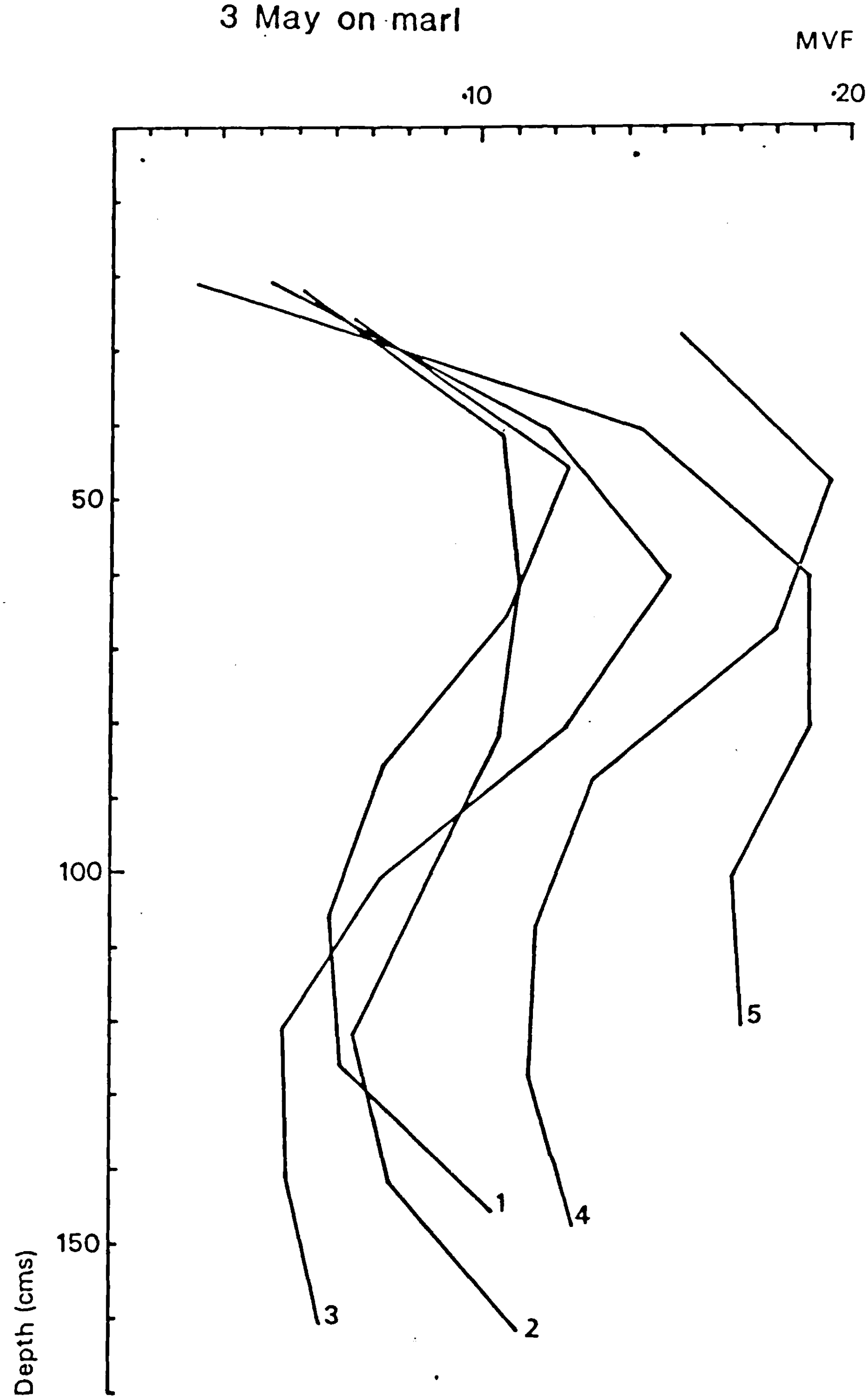


Figure 6·10 Soil moisture values for April/May 1983 on the conglomerate for different depths (cms)

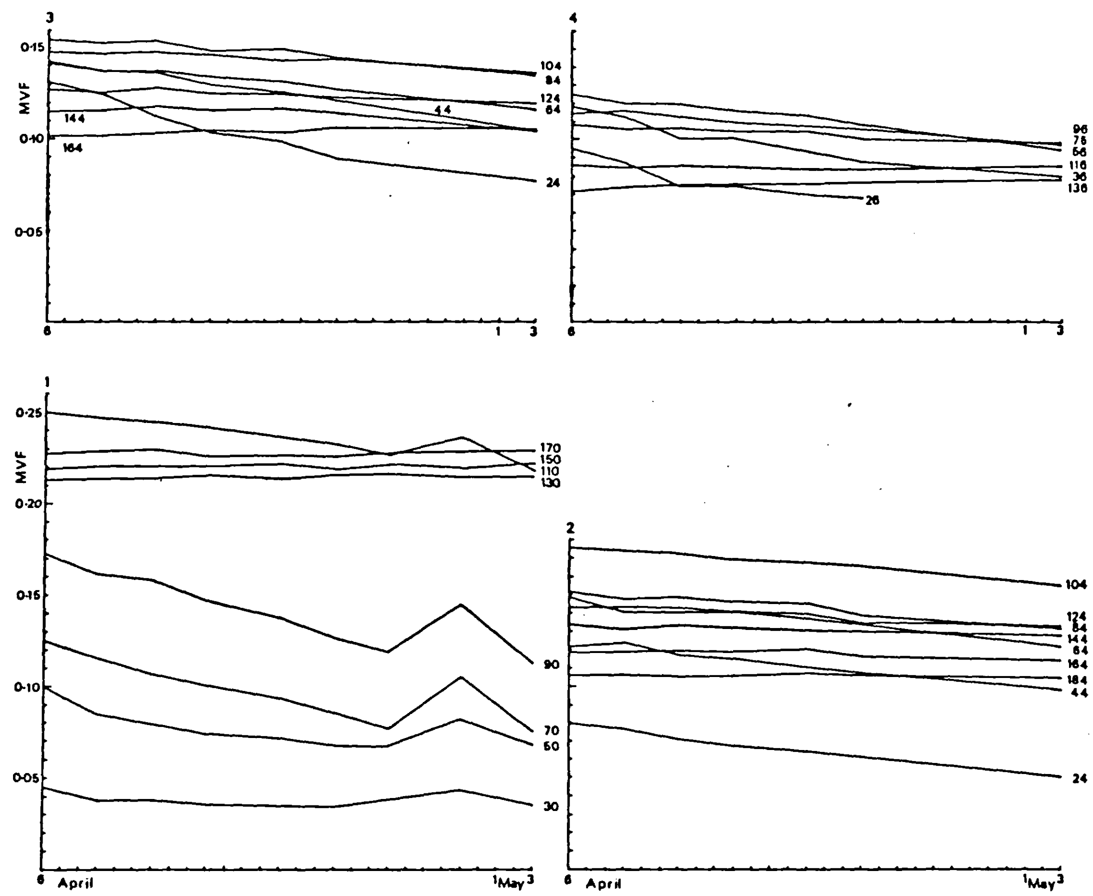
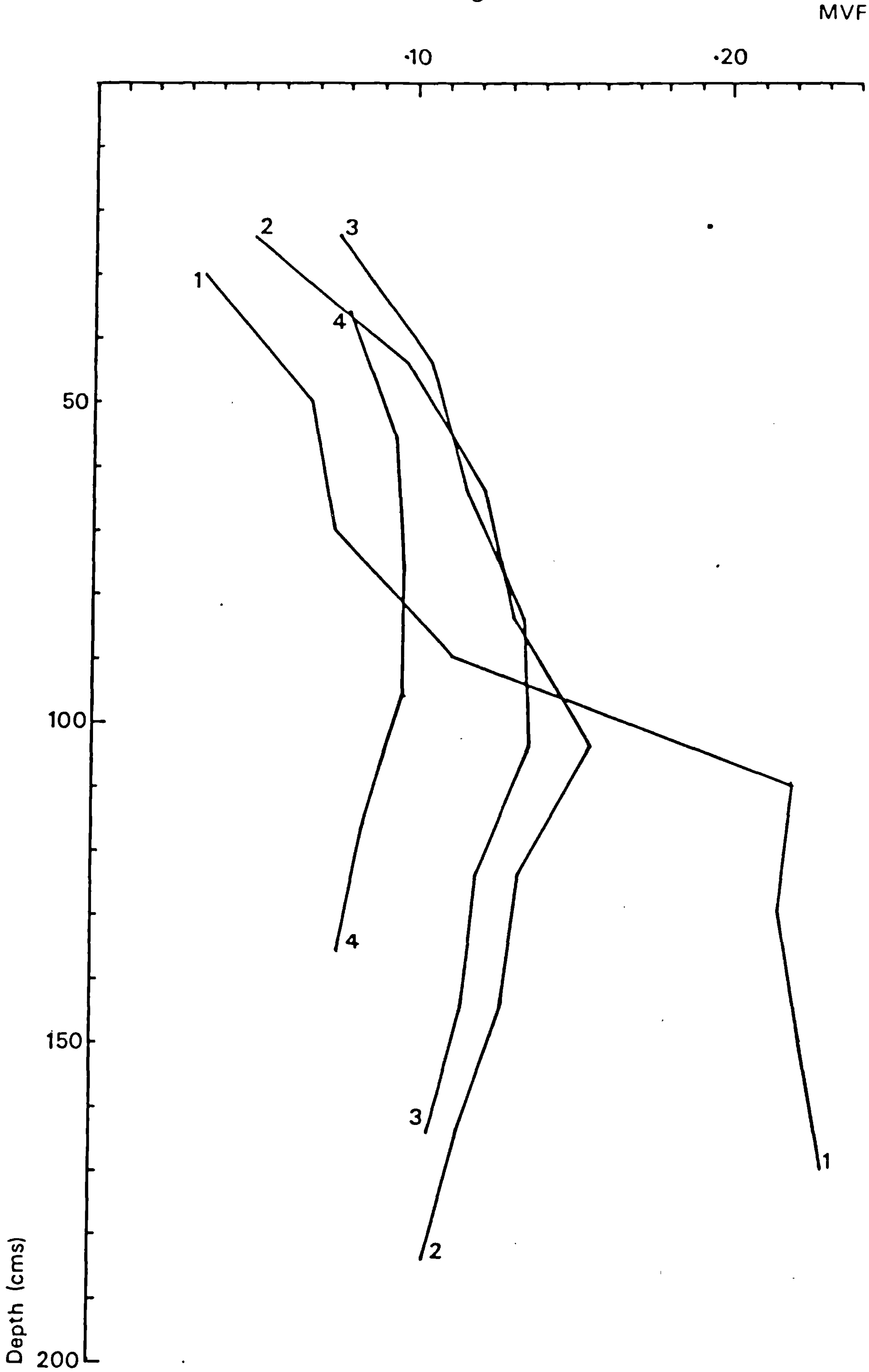


Figure 6·11 Soil moisture variation on 6 April on the conglomerate



Tube 1 is the driest to 80cms but between 80 and 110 cms the soil moisture increases rapidly and then levels out between 0.21 to 0.22 MVF. The profiles of 2 and 3 are similar to each other although tube 2 tends to be drier above 50cms and wetter below 90 cms. Both profiles increase in soil moisture to c104cms and then decrease. Tube 4 is relatively dry with little variation with depth. The draining effect of the gully head may be responsible for the low soil moisture values to about 90 cms at tube 1, but below this the higher soil moisture values are probably due to the effect of the conglomerate/clay boundary. Where this boundary is much deeper, the conglomerate drains towards an equilibrium soil moisture profile, which is almost attained at tube 4.

On Thursday 28 April at 9.30am, 5 litres of water were poured around tube 1. By 10.05am the next day the effect of this irrigation had reached 110cms depth. The greatest increase recorded was for 90cms, reflecting perhaps that above this the majority of the water had already seeped through, and below this the peak either had not arrived through the less permeable clay horizon or had attenuated with depth. By 3 May the soil moisture had returned to about the value recorded on 25 April 1983.

Summary

Saturated soil moisture conditions were not observed on the hillslope. On the marl soil moisture values were recorded in excess of 0.3 MVF which may represent 60% of the saturated capacity a week after the storm on 6 November. It seems quite probable that the soils were saturated during the storm but only to a limited depth of 30-40cms (see figure 6.4 tubes 4 and 5). Recalling the soil profiles, there is a horizon of loosely packed aggregates some 15 cms deep resting upon a marl bedrock which decreases in porosity with depth. Following heavy or prolonged rainstorms the top layer may well become saturated, especially as the drop in permeability of the bedrock will retard further downward percolation. This has implications for saturated overland flow development, shallow throughflow, and slaking of the marl. Unfortunately soil moisture values are not available for this shallow horizon.

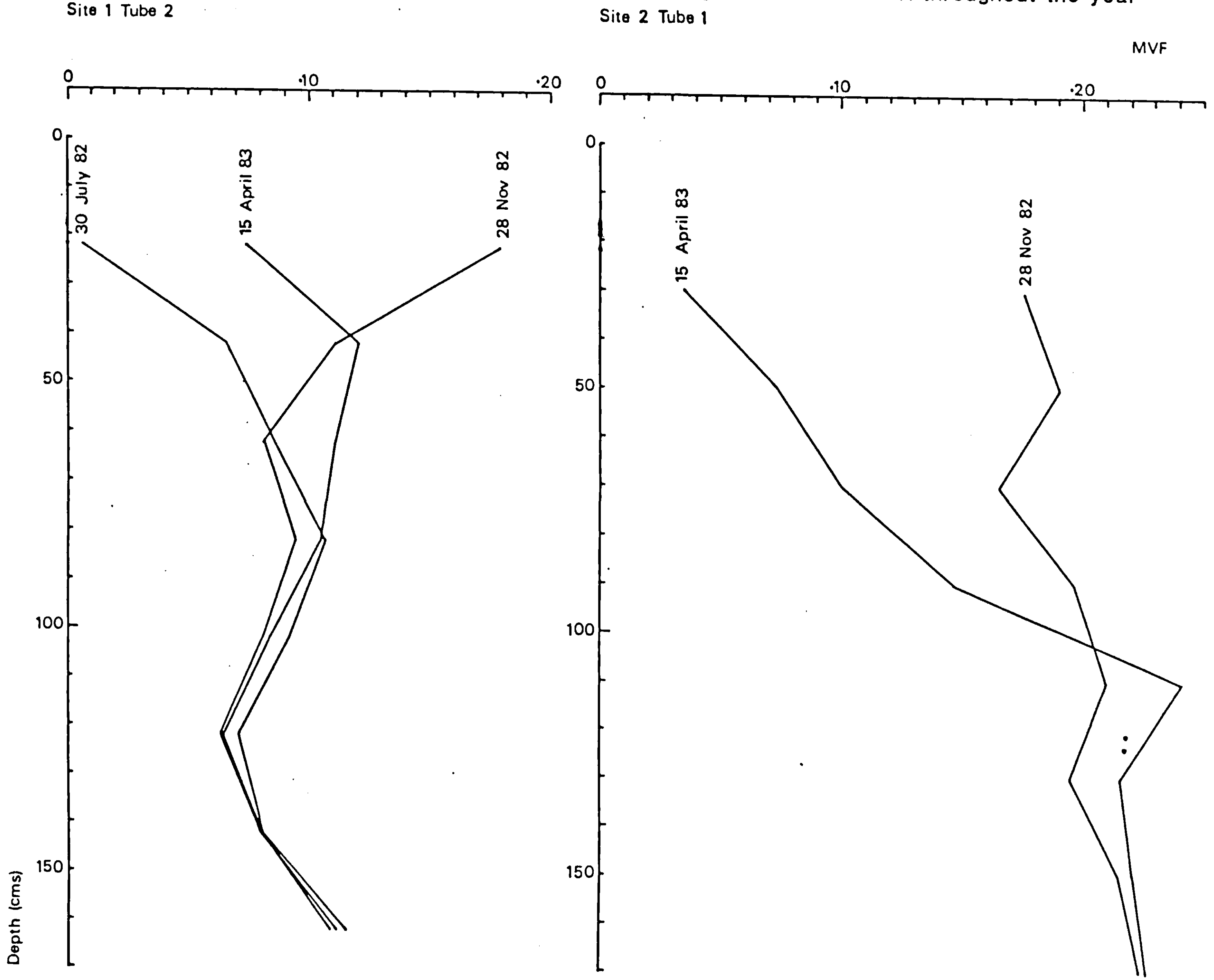
On the conglomerate soil the soil moisture values recorded near the surface reach 0.185 MVF at tube 1 (which is about 50% porosity) and 0.223 MVF at tube 3 (about 60% of the porosity). The lack of observed higher soil moisture readings may reflect the problem of taking the readings too infrequently or the drainage characteristics of the soil are too rapid to allow greater build up of soil moisture.

The data for marl shows that most temporal variation occurs above c80cms increasing towards the surface as a response to evaporation and precipitation. Below 80cms there is little change, although it is interesting to see in fig 6.12 that the soil was wetter in July 1982 than April 1983 despite 202.2mm of rainfall in the meantime. This may be testament to the severity of the continuing drought. However the difference between the readings may not be significant as at 102 cms the difference in moisture between 30 July and 15 April is only 0.008 MVF. Soil moisture variation in the conglomerate extends to a greater depth, and is particularly marked for tube 1 above the gully where soil moisture builds up at the conglomerate/clay interface.

There are a number of differences in hydrological properties between the marl and conglomerate. The data for November 1982 shows that water moves more quickly through the conglomerate as peak soil moisture levels occur sooner. The values recorded for the 20-30cms horizon (figures 6.4 and 6.5) show that soil moisture varies between 0.160 to 0.223 on site 2 and 0.179 to 0.298 MVF on the marl. However despite higher values on the marl, the variation between readings recorded before the storm and peak values after are 13 to 14% on the conglomerate and 3 to 7% on the marl suggesting that relatively more water is moving through the conglomerate. These values are misleading because of the time lag between the events and the recorded data so that the peak soil moisture values may not have been recorded.

During the autumn period the marl profiles show a marked wetting downwards particularly between 40-70cms. Below this depth soil moisture values are higher on the conglomerate than the marl where soil moisture values stay between 0.15 to 0.2 MVF to c110cms. On site 2 at tube 1 the change to clay at c110cms probably causes the

Figure 6·12 Comparison of soil moisture values on conglomerate and marl throughout the year



soil moisture build up of water between 24 November and 8 December, and may induce some lateral throughflow here. The affect of this change is more pronounced in the spring where figure 6.12 shows a marked increase in moisture with depth. The marl here is maintained in a relatively wet condition by the flow of water from above compared to site 1 where there is no build up of moisture, so that on 6 April there is 0.227 MVF in the marl on site 2 tube 1 at 150cms, compared to 0.076 and 0.109 MVF at tube 2 site 1 at 142cms and 162 cms respectively. Further upslope where the conglomerate/clay interface is much deeper there is no build up of soil moisture with depth, in fact the soils tend to drain towards an equilibrium soil moisture distribution. Figure 6.12 shows also an increase in soil moisture with position downslope suggesting possibly throughflow.

6.2 Soil Moisture Movement

The movement of moisture through the soil is very tortuous resulting from the complex structure of the soil particles. Thus estimates of flow through soil deal with gross velocity in a volume of soil rather than the velocity of flow in particular channels. The fundamental basis of water movement in soils is Darcy's Law which states for saturated soils that

$$q = -K. \Delta \phi$$

where q is the flux (volume of water flowing through a unit cross-sectional area per unit time, t), K is the hydraulic conductivity and $\Delta\phi$ is the hydraulic gradient. When the soil is unsaturated the hydraulic conductivity and gradient are related to soil moisture content so the equation is extended to

$$q = -K(\theta) \frac{\partial \phi}{L} \qquad \phi = z \mp \Psi(\theta)$$

The solution of Darcy's Law requires the knowledge of the soil moisture content within the soil, and the characteristic curves between soil moisture against pressure and conductivity. The theoretical background and development of Darcy's Law has been described in many papers and standard texts (Hubbert 1940, Freeze and Cherry 1979, Hillel 1982).

There are limitations to the use of Darcy's Law, in particular the equation should not be used to estimate flows in soils where 'by passing' occurs (Bouma and Dekker 1978). This short circuiting can occur on the scale from macropores to pipes in such a way that flow rates in the soil are dominated by a proportionately small number of channels. In extreme cases flow in pipes can be considerable with 'flashy' discharges ranging up to 10 litres sec^{-1}, and some techniques for tracing, collecting and measuring flow in pipes are described by Atkinson (1978). This problem was avoided by choosing non-piped sites, despite the knowledge that some marl gullies in south east Spain are piped (Harvey 1982).

A more difficult problem is by passing in macropores including interaggregate pore spaces, shrink-swell cracks, root channels and faunal tunnels (Skopp 1981). Even relatively small continous pores can transport large quantities of water; Bouma (1981) reported saturated conductivities of 60cms/day in clay as a result of this action. A number of field and laboratory experiments using dye staining techniques have shown that by passing occurs (Anderson and Bouma 1973, Bouma et al 1977, Bouma and Dekker 1978, and Bouma and Wösten 1979). Theoretical or empirical solutions to calculating flow are hindered by this division of flow into two units, micro and macro pore transport. Furthermore an understanding of flow in macropores is complicated by observations suggesting that macropores are not necessarily full as water flows down the sides of pores, and that this flow is governed by 'pore necks' which are very difficult to observe directly. Very little is known about the hydraulics of flow in macropores, but Beven (1981) has suggested that should the hydraulics of flow be different in macro than micro pores then substantial by passing already points to the break down of traditional Darcian concepts of water flow in soils.

Despite these criticisms Darcy's Law is still widely used to calculate flow rates in soils although its use should be guarded when dealing with soils with numerous shrinkage cracks or well defined ped structures. It was assumed on this project that Darcy's Law is appropriate for describing flow rates in the soil and no experiments were undertaken to verify in situ flow patterns or rates.

Techniques

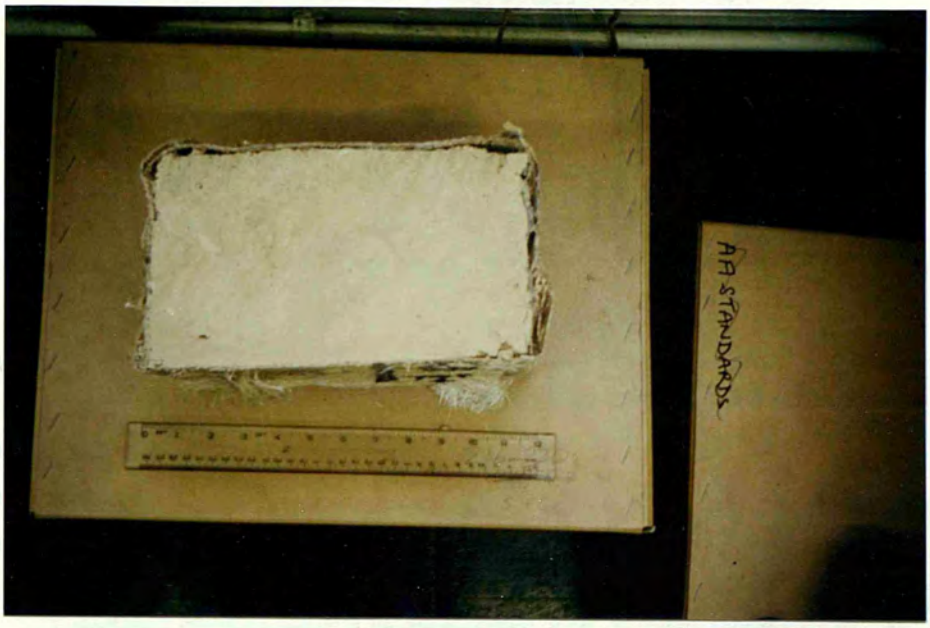
Values for the characteristic curves of soil moisture against soil tension and hydraulic conductivity were obtained using a selection of field, laboratory, and numerical techniques. A field based correlation of soil moisture against tension was attempted by siting two pairs of tensiometer nests by neutron probe access tubes. However tensiometers can not be used in soils for tensions greater than 0.8 bars although soil tensions may reach 20 bars. For much of the field monitoring the soil was too dry as 0.8 bars represents 0.127 MVF in the conglomerate and 0.311 MVF in the marl (using the final characteristic curve).

The characteristic curves presented here are based on undisturbed soil samples analysed in the laboratory. The soil cores were obtained by cutting soil monoliths out of the soil, and coating the free sides with resin and fibre glass (photograph 6.1). A rigid laminating B Polyester Resin was used with a liquid hardener (supplied by Alec Tirante Ltd, Theale, Reading, Berks). When dry the marl can maintain vertical walls so the monoliths were relatively easy to excavate. The conglomerate soil was more friable and difficult to cut. Two monoliths from the marl and one from the conglomerate were finally taken from pit 3 on site 1 and pit 5 on site 2. These large monoliths were cut up into blocks using a diamond saw, and photographs 6.2 and 6.3 show cross sections in both soil types. When the subsamples were a suitable size they were orientated correctly with respect to their position in the slope, and the resin coats on the top and bottom were cut off.

This method of taking samples was thought to cause least structural damage to the soil sample. All methods of augering involve some degree of compaction especially as insertion is difficult in both stony soil or the more cemented marl. Two problems arise, the conglomerate soil tended to crack, and secondly the samples obtained are an awkward shape. As the conglomerate was less cohesive or cemented the monoliths were harder to obtain and tended to crack. Subsequently all samples

Photograph 6·1 Taking a soil monolith on the marl

Photograph 6·2 Cross section of a marl monolith



Photograph 6·3 Cross section of a conglomerate monolith

from this were taken from the uncracked parts. Using the diamond saw did not appear to markedly affect the soil structure, and did not cause any further cracking. The second problem of sample shape is considered below with reference to permeameters.

Macrochannels between the soil and resin coating were sealed by pouring melted parafin wax around the edge of the sample and a final resin coat around the sample ensured that the samples were watertight. The samples were placed in distilled water and allowed to saturate under a small vacuum for at least 24 hours.

The characteristic curves of soil moisture against tension were determined using a pressure chamber apparatus. A standard procedure was carried out whereby the samples were reweighed at each pressure interval until the weights stabilised. Altogether four marl, three conglomerate and four clay samples were used.

The relationship between soil moisture and conductivity was determined by measuring the saturated hydraulic conductivity and using this value in a standard equation involving parameters from the relationship between soil moisture and tension.

In the first place the saturated hydraulic conductivities were measured using both constant head and falling head permeameters. The constant head permeameter is best suited for measuring the conductivity of sands and fine gravel whereas the falling head permeameter is most suited to clays, silts and fine sands, a reflection of the level of accuracy of both tests. The clay samples were taken from pit 2 site 2 using the auger for undisturbed cores and were already shielded in a plastic container. This was easily fixed into a falling head permeameter. The marl and conglomerate monolith samples however were not uniformerly shaped so that the permeameter had to be built around the sample. This is the problem alluded to above. The most successful design was a constant head type obtained by building the resin case up above the sample, waterproofing it and providing an outlet. Water was drip fed into the top of the sample and the outfall was collected at known intervals. In this way the less suitable technique was used on the sample as it was thought preferable to be less accurate on an undisturbed sample, than more accurate on a disturbed sample.

In the second place several methods have been proposed to calculate hydraulic conductivities for different soil moisture conditions from moisture retention curves (Childs and Collis George 1950, Marshall 1958, Millington and Quirk 1959, Brooks and Corey 1966, and Campbell 1974). These methods are based on a conductivity function which is matched to one measured value at a known water content, normally at saturation. The equation from Brooks and Corey (1966), for example, takes the form

$$K(P_c) = Ks\ (P_b/P_c)^n$$

where $K(P_c)$ is the unsaturated conductivity, Ks is the saturated conductivity, P_b is the bubbling pressure of the soil, P_c is the capillary pressure and $n = 2+3\lambda$. Both P_b and λ are derived from the soil desorption data. Another similar equation comes from Campbell (1974) with the form

$$K = Ks\ (\theta/\theta s)^{2b+3}$$

where θ is the soil moisture value, θs is the saturated soil moisture content, and b is the gradient of the line of the retention curve when both soil moisture and tension are logged. Campbell tests this equation on five soils by comparing actual values measured and those calculated using Millington and Quirk's method and his own method. Campbell's equation provides quite good agreement with measured values of conductivity, and in several cases improves upon calculated values by Millington and Quirk. Campbell's method was adopted because it is easily parameterised, and gives accurate results for a range of soils to relatively dry water contents of 0.1 cm^3/cm^3.

Problems

There are four main suites of problems associated with the use of the curves of soil moisture against tension and pressure which are namely

1 The representiveness of the samples for a variable medium
2 Hysteresis
3 Soil structure and stability with wetting and drying
4 The range of moisture values.

On the first problem reference has already been made to the variability of soil parameters including $CaCO_3$ content, infiltration rates, and soil structure, such that soil properties have distributions which can be quite wide. The possible diversity is such that Beckett and Webster (1971) suggest that up to half the variance within a field may already be present within any metre square, although the range of variability will depend on the variable. With specific reference to hydraulic properties, several authors point to the spatial variability of parameters. Keisling et al (1977) tried to estimate the number of samples needed to obtain results within 10% of the mean for a given soil. They found, for example, only 1 to 2 samples were needed to estimate water content at 0.1 bar, but 7 to 24 samples for water content at 15 bars and 2 to 16 samples for the mean log conductivity. Carvallo et al (1976) tried to measure and evaluate the spatial variability of in situ saturated hydraulic conductivity. They found significant spatial variability in conductivity by the 1% level without any significant difference in soil texture. Both Freeze (1980) and Nielsen et al (1973) estimate that the hydraulic conductivity for a given soil water content is log normally distributed for each soil depth, and Freeze states further that the value of Ksat can vary over 12-14 orders of magnitude, although he puts the normal range of near surface Ksat at 0.0001 to 0.1 cms/sec. Freeze has attempted to incorporate this diversity into mathematical modelling by using distributions of some soil parameters instead of constants. One of his findings was that the hydraulic conductivity influenced overland flow generation directly in 'Hortonian' type flow, and indirectly through the control of the water table in 'Dunne' type flow. Finally Stockton and Warrick (1971) found that one standard deviation either side of average soil water characteristic curve resulted in 20-30% variation in unsaturated hydraulic conductivity.

Given the well known distribution of hydraulic parameters sample sizes of 2 to 4 to establish characteristic curves are small. The sample sizes are even too small to estimate the required sample size needed to attain a prespecified accuracy. The problem of getting values which would reproduce the field conditions could be resolved by either taking more samples, or by taking very large samples. The former response was restricted because of the logistical problems, although the samples used were quite large with volumes ranging from 125.5cc to 826.4cc. Despite the small number of samples, the relatively close agreement between the values for Ksat (particularly on the marl) and the characteristic curve of soil moisture against tension do allow at least a first approximation of the hydraulic conditions on these soils.

Secondly the problem of hysteresis is well known for both curves of soil moisture against tension and conductivity (for example Poulovassiliss 1962, 1969), although different authors have found conflicting evidence as to whether the wetting or drying cycle gives higher results (Poulovassilis 1969, Youngs 1964). According to some authors (Nielsen and Biggar 1961, and Topp and Miller 1966) its effect on conductivities is almost negligible. In this work the characteristic curves were determined for the drying curve, because the pressure chamber apparatus is better suited to this approach and the soil moisture data shows the drying state is more commonly encountered in the field.

Thirdly changes in volume of soil with wetting and drying complicate the relationships between soil moisture, tension and permeability as water retention cannot be related to initial conditions (Gumbs and Warkentin 1975). Sometimes structural changes on drying are irreversible (Croney and Coleman 1954). When the soil samples were saturated the clay samples showed significant increases in volume as a response to swelling, but the marl and conglomerate samples did not, although more sensitive expansion tests were not carried out. In the subsequent analysis the relationship between soil moisture and conductivity and flux rates were not calculated for the clay horizon.

Finally the data used for the retention curves are well within the wet range of the soil moisture. The maximum pressure used in the pressure membrane apparatus was 1.68 bars, by which pressure soil moistures are about 0.3 cm^{-3}/cm^{-3} on the marl and 0.1 $cm^{3}\,cm^{-3}$ on the conglomerate. Field measurements of soil moisture on the marl vary between 0.05 to 0.32 MVF and on the conglomerate between 0.03 to 0.22 MVF. Thus the characteristic curves determined on the wet range are being used to estimate soil moisture pressures in the field for the dry range. This is more of a problem on the marl as most of the field data are below the range of the retention curves. This is important because the retention curve is a logged function, and small soil moisture values using this calibration give very high tension values. It is not known whether the log relationship holds for the entire range of soil moisture so that these high tension values at the lower range may be incorrect.

Campbell's method for estimating hydraulic conductivity from retention curve data was not tested independently on the marl by comparing actual and calculated values. However some idea of the accuracy of Campbell's method is obtained by comparing the marl values and those for the soils tested by Campbell. The 'b' coefficient and porosity values for the marl of 7 and 0.55 $cm^{3}.cm^{-3}$ resemble most closely the values for Guelph loam of 4.5 and 0.52 $cm^{3}\ cm^{-3}$ used by Campbell. The 'b' coefficient was taken from the retention curve for a range of only 0.05 to 1 bar only (similar to the marl). A comparison of calculated and measured hydraulic conductivities by Campbell shows that the calculated values provided a good match to a rate of 1×10^{-5} cms/min by which time the water content was approximately 0.25 $cm^{3}.cm^{-3}$. Campbell did not plot the data beyond this point so it is impossible to tell whether the correlation continued to be sound. It is felt that the correlation produced for water content against conductivity on the marl would similarly be accurate to at least the same range. Although fluxes are estimated for moisture contents below 0.2 $cm^{3}.cm^{-3}$ further on, it is considered that these will be open to error. However as the main contention is estimating fluxes which will be significantly large to erode or raise pore water pressures, it must be remembered that fluxes are at a maximum for saturated conditions and then fall quickly to relatively small rates. Thus in terms of the problem of flow rates to gully heads it would be expected that flow rates for about 50% of the saturated case would already be too small to be significant.

Results

The relationship between soil moisture and pressure is plotted in figure 6.13 and noted in table 6.6 for the marl and conglomerate samples. The relationship plots as a straight line on log:log paper for all cases with coefficients of determination (r^2) greater than 0.94 and all the correlations significant by the 0.01% level using the F distribution. A number of points on the characteristics of porosity and retention curves can be drawn from the data.

Porosity in the samples has been measured in two ways, firstly

Porosity = 1 - (Bulk Density/Particle Density)

where the bulk density is the dry weight of the sample divided by its volume and the specific densities were measured using a pyncometer (Chapter 2), and secondly, porosity equals,

$$\frac{\text{Saturated weight of monolith - dry weight of monolith}}{\text{Volume of the monolith}}$$

which is the same as

$$\frac{\text{Volume of water}}{\text{Volume of soil}}$$

Table 6.7 shows that for the conglomerate samples the first method gives slightly higher values with porosities ranging between 0.389 to 0.412 compared to values of 0.353 to 0.386 for the second method. The difference between the two methods is about 2 to 3% of the total volume. However on the marl the difference is larger with higher values measured by the second methods of 0.547 to 0.587, whilst the first method gives porosities of 0.448 to 0.474.

Average soil porosities vary around 50%, with sandy soils having a lower porosity and clayey soils a higher porosity. It is not exceptional to have porosities of 0.55 for the marl, however it is disturbing that the two methods yield results which differ by 7 to 13% of the total volume of the sample. Both methods were repeated and gave the same result so that the difference is consistent. The porosity measured by the saturation method was finally taken as this forms an integral part of the retention curve analysis.

The curves for the marl and conglomerate plot out as two separate families showing that they behave differently. The conglomerate

Figure 6·13 Characteristic curve of soil moisture against suction for the marl and conglomerate

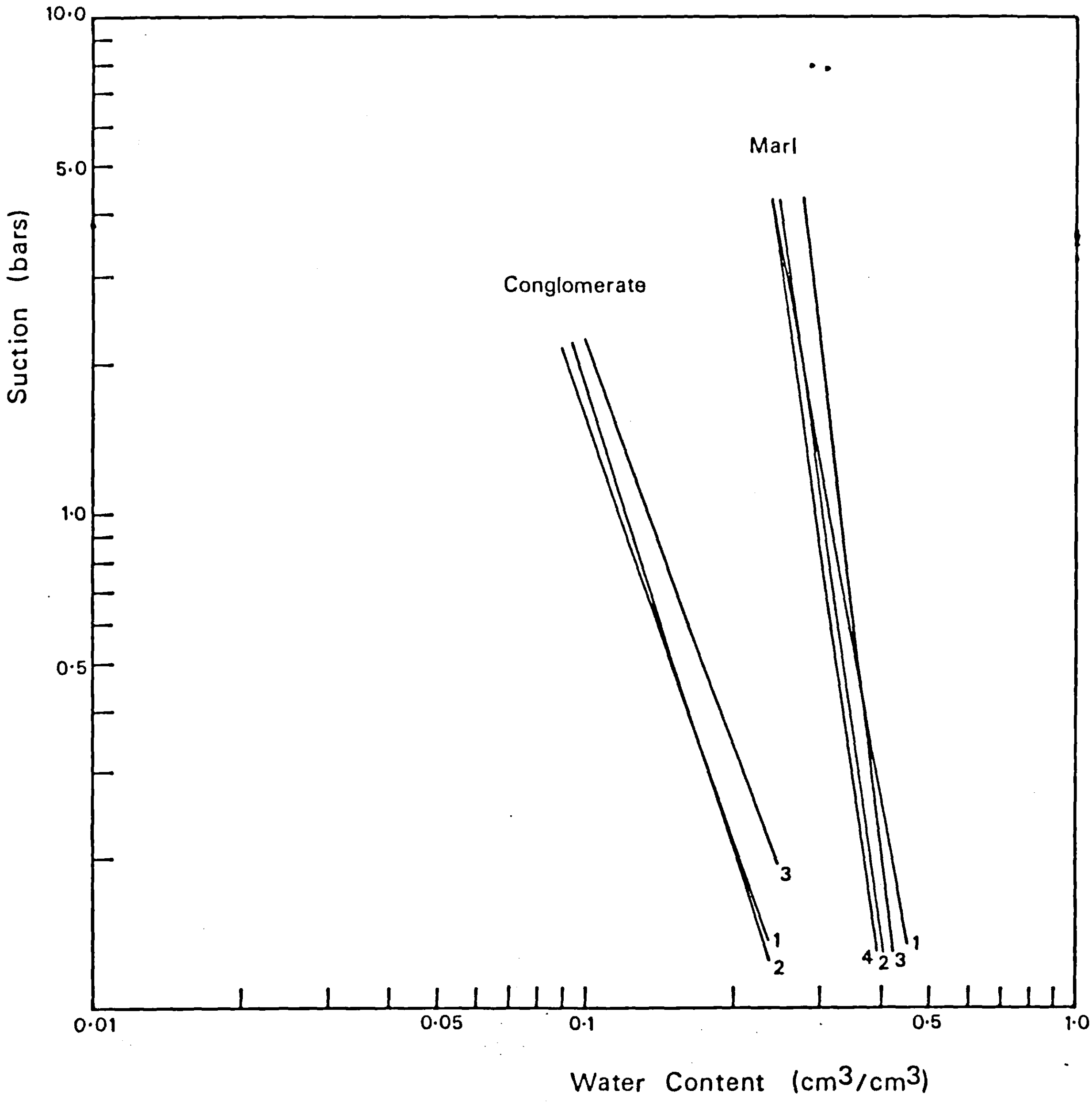


Table 6.6 Retention Curve Data: the values for MVF and Ψ have to be logged for the regression analysis

Marl

1		2		3		4	
MVF	Ψ bars	MVF	Ψ bars	MVF	Ψ bars	MVF	Ψ bars
0.5861	0.00	0.5468	0.00	0.5542	0.00	0.5868	0.00
0.5273	0.07	0.3443	0.34	0.3702	0.34	0.3316	0.34
0.4320	0.14	0.3323	0.52	0.3590	0.52	0.3240	0.52
0.4049	0.21	0.3216	0.69	0.3480	0.69	0.3155	0.69
0.3712	0.34	0.3037	1.03	0.3305	1.03	0.2996	1.03
0.3528	0.48	0.2960	1.24	0.3261	1.24	0.2879	1.24
0.3361	0.69	0.2852	1.44	0.3120	1.44	0.2746	1.44
0.3246	0.90	0.2768	1.68	0.3050	1.68	0.2699	1.68
0.3070	1.10						
0.2898	1.65						
A = -2.772		A = -3.7130		A = -3.856		A = -3.726	
B = -5.477		B = -7.1354		B = -7.994		B = -7.002	
r = 0.9831		r = 0.9757		r = 0.970		r = 0.941	
F = 407		F = 200		F = 166		F = 112	

Conglomerate

1		2		3	
MVF	Ψ bars	MVF	Ψ bars	MVF	Ψ bars
0.354	0.000	0.386	0.000	0.375	0.00
0.214	0.207	0.207	0.207	0.228	0.207
0.153	0.414	0.156	0.414	0.184	0.414
0.132	0.621	0.144	0.621	0.173	0.621
0.125	0.896	0.124	0.896	0.142	0.896
0.115	1.034	0.121	1.034	0.131	1.034
0.110	1.241	0.115	1.241	0.124	1.241
0.103	1.448	0.106	1.448	0.112	1.448
0.101	1.655	0.103	1.655		
A = -2.608		A = -2.763		A = -2.356	
B = -2.802		B = -3.009		B = -2.696	
r = 0.981		r = 0.993		r = 0.971	
F = 310		F = 851		F = 167	

Table 6.7 Porosity values measured using two methods

		$1 - \frac{\text{Bulk density}}{\text{Particle density}}$	$\frac{\text{Volume of Water}}{\text{Volume of Soil}}$
Conglomerate	1	0.389	0.353
	2	0.412	0.386
	3	0.400	0.375
Marl	1	0.455	0.586
	2	0.474	0.547
	3	0.448	0.554
	4	0.474	0.587

Table 6.8 Values for saturated hydraulic conductivity (cms/min)

From field work

Marl	Conglomerate	Clay
0.0039	0.0548	0.0239
0.0129	0.0768	0.0424
0.0147	0.2832	
0.0168	0.5880	

From Molapo (1981)

Values for saturated hydraulic conductivity in a gully wall, Lesotho.

Horizon	Vertical Conductivity (cms/min)	Horizontal Conductivity (cms/min)
1 Clay loam	0.00557	0.00784
2 Clay loam	0.00291	0.00119
3 Clay	0.00024	0.00101
4 Clay	0.000008	0.00001

From Gilmour and Bonell (1979)

Horizon		Conductivity (cms/min)
1 0 - 10 cms	Oxisols	2.25
2 11 - 20 cms	Oxisols	0.11
3 below 20 cms	Oxisols	0.02

curves have a shallower gradient with a more rapid decline in soil moisture with pressure. Thus 50% of the soil moisture was lost by a pressure between 0.23 and 0.39 bars. On the marl 50% of the soil moisture was not removed until an estimated pressure between 1.00 and 3.99 bars, according to the regression equations.

Values for the saturated hydraulic conductivity are shown in table 6.8. Conductivities are highest and most variable for the conglomerate soil with values ranging between 0.0548 to 0.588 cms/min. Rates on the clay and marl are 0.0239 to 0.0420 cms/min and 0.0039 to 0.0129 cms/min respectively. These values represent the vertical conductivities of the samples. The small number of samples does not allow spatial variation due to anisotropy or heterogeneity to be examined, although other authors have shown that conductivity decreases with depth (Gilmour and Bonnell 1979) or is greater in one plane than another (Basak 1972 and Molapo 1981).

Using the saturated hydraulic conductivities and the retention curve data, the relationship between soil moisture and conductivity is calculated. Figure 6.14 and figure 6.15 show for each soil block the relationship between soil moisture and conductivity for all four estimates of saturated hydraulic conductivity for both conglomerate and marl soils. For each soil block the gradient of the four lines is the same, but the position of the line is set by the total porosity of the soil and varies between soils according to the 'b' parameter taken from the retention curves. In both lithologies the saturated hydraulic conductivity varies by about an order of magnitude, and this is reflected in calculated values for the unsaturated values. This is confirmed in table 6.9 which shows that a 1% or 10% change in the saturated hydraulic conductivity will lead to approximately 1% or 10% change in the conductivity value at 50% soil moisture content. Changes in the 'b' value cause non-linear changes in the calculated value of conductivity (table 6.9). Relatively small changes in 'b' will therefore have a large influence on the conductivity rates.

Table 6.10 shows the values for conductivity at the 50% soil moisture content for each soil block for the four estimates of the

Figure 6·14 Relationship between soil moisture and hydraulic conductivity on the conglomerate

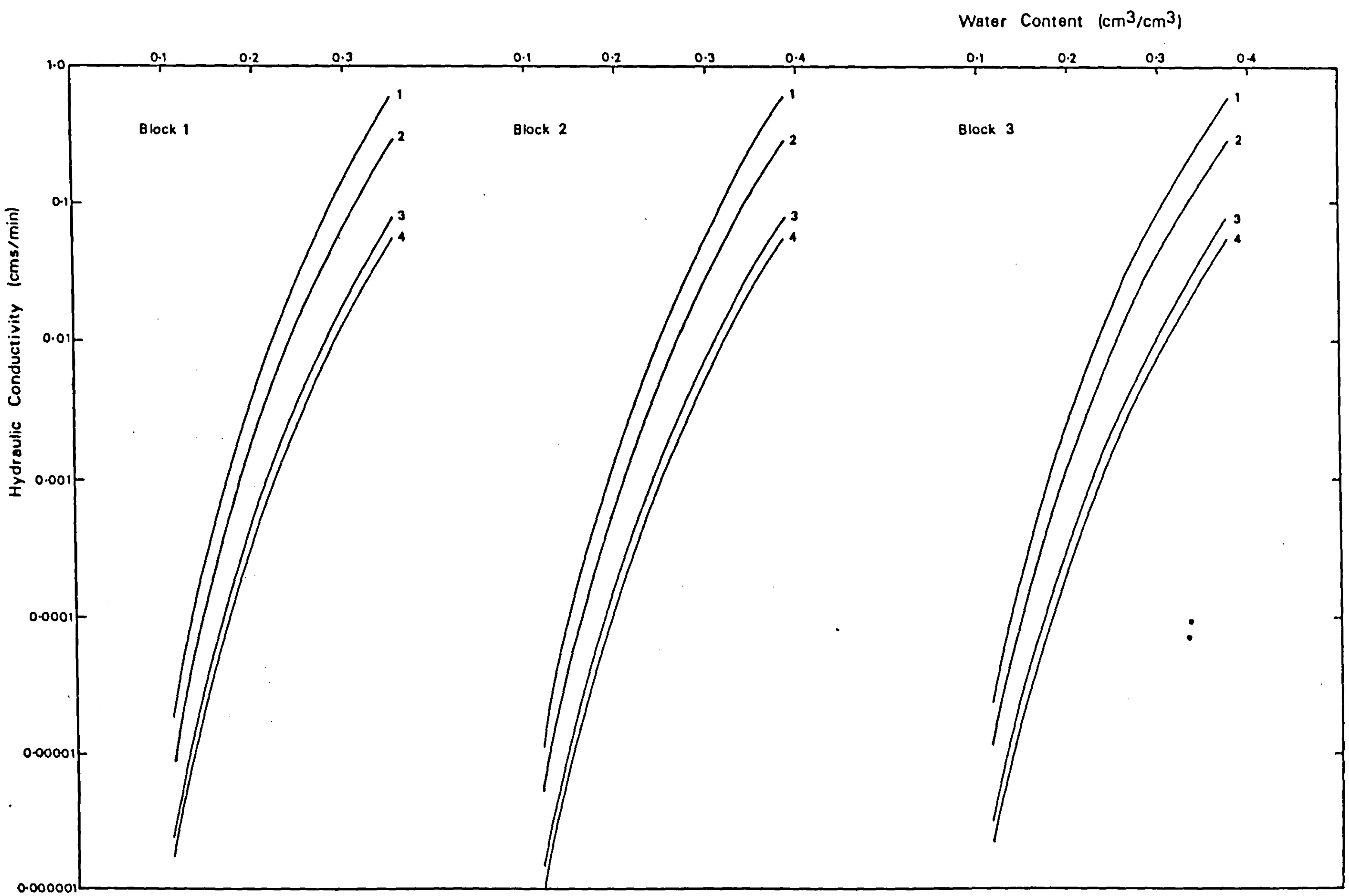
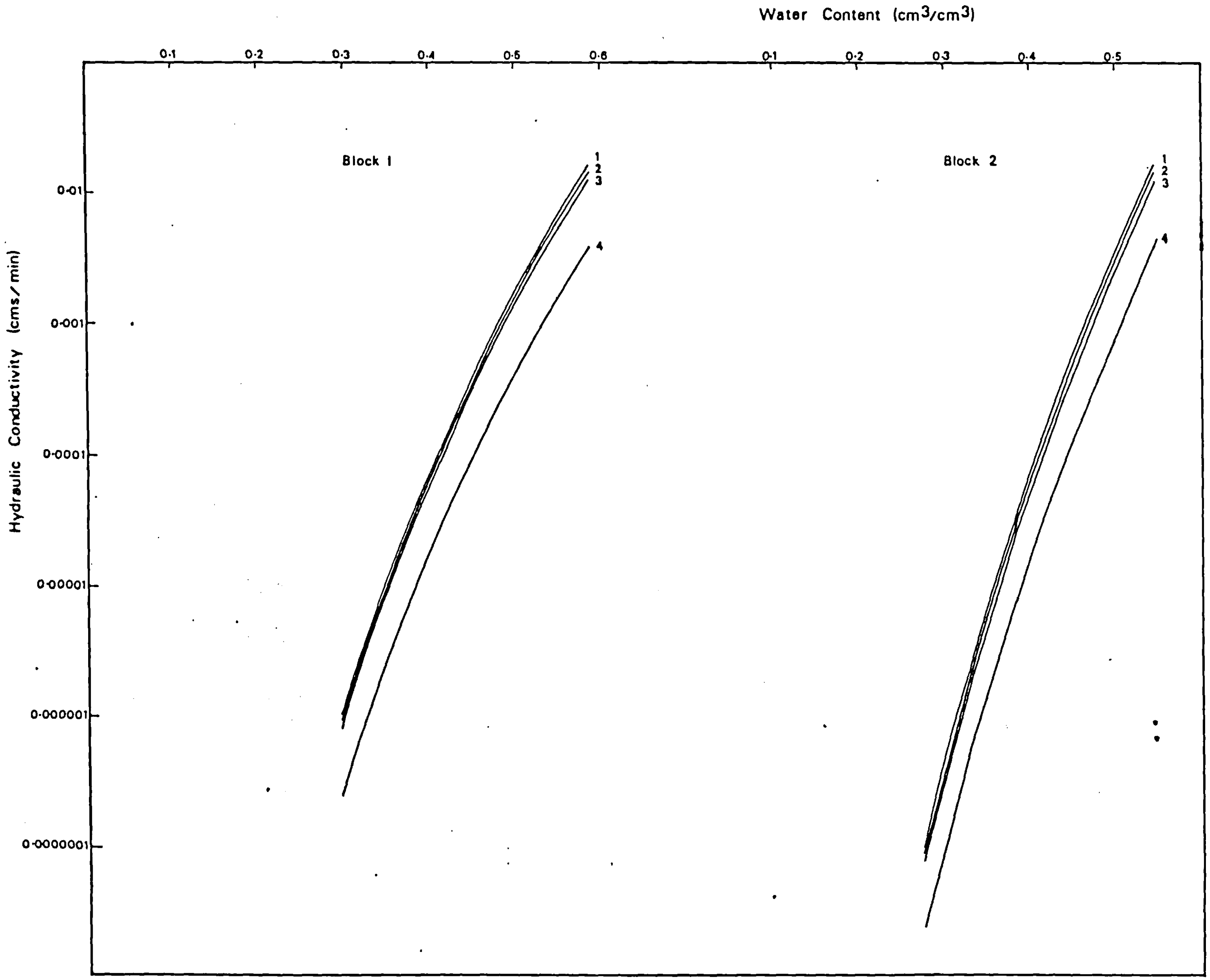


Figure 6·15 Relationship between soil moisture and hydraulic conductivity on the marl



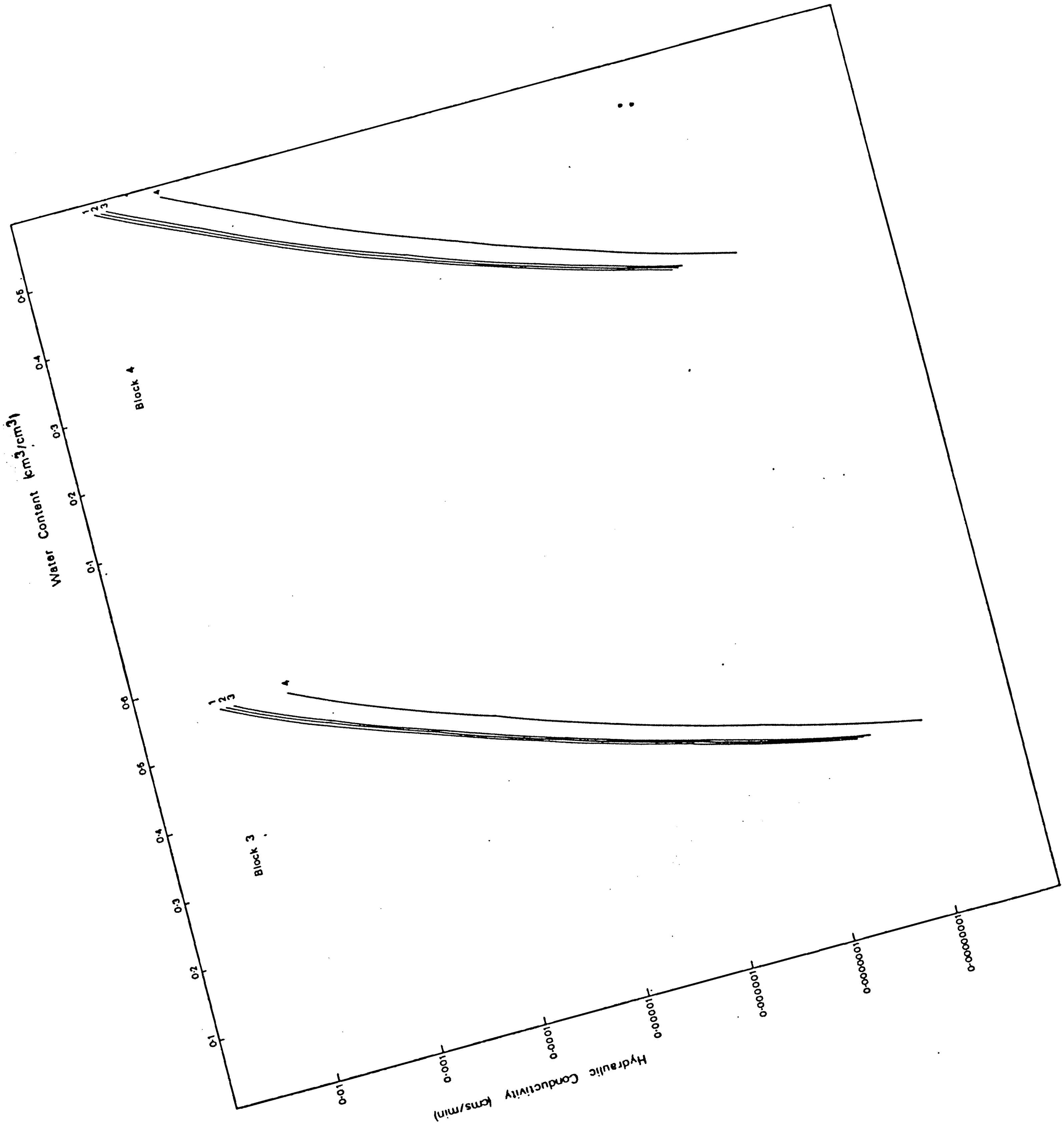

Water Content (cm3/cm3)
Hydraulic Conductivity (cms/min)
Block 4
Block 3
0·1
0·2
0·3
0·4
0·5
0·6
0·01
0·001
0·0001
0·00001
0·000001
0·0000001
0·00000001
1
2
3
4

Table 6.9 Effect of variations in Ksat and b on calculated hydraulic conductivity rates in cms/min.

(a) To show the effect of a 1% and 10% change in the Ksat to the calculated value of the conductivity at 50% soil moisture.

Change in Ksat	-10%	-1%	0	+1%	+10%
Conglomerate	4.38×10^{-4}	4.81×10^{-4}	4.86×10^{-4}	4.91×10^{-4}	5.35×10^{-4}
% change in K	10%	1%		1%	10%
Marl	8.32×10^{-8}	9.16×10^{-8}	9.23×10^{-8}	9.31×10^{-8}	1.01×10^{-7}
% Change in K	10%	1%	0	1%	9%

(b) Effect of a 1% and 10% change in the 'b' coefficient to the calculated value of the conductivity at 50% soil moisture.

Change in Ksat	-10%	-1%	0	+1%	+10%
Conglomerate	2.29×10^{-4}	1.58×10^{-4}	1.51×10^{-4}	1.45×10^{-4}	9.97×10^{-5}
% Change in K	52%	5%	0	4%	34%
Marl	2.50×10^{-7}	1.08×10^{-7}	9.84×10^{-8}	8.93×10^{-8}	3.73×10^{-8}
% change in K	164%	10%	0	10%	62%

Table 6.10 Calculated values for the hydraulic conductivity (cms/min) at 50% soil moisture on the marl and conglomerate for each sample using four values of the saturated hydraulic conductivity (cms/min).

Marl

Ksat	Block 1	Block 2	Block 3	Block 4
0.0039	2.46×10^{-7}	2.47×10^{-8}	7.50×10^{-9}	2.97×10^{-8}
0.0129	8.13×10^{-7}	8.16×10^{-8}	2.48×10^{-8}	9.81×10^{-8}
0.0147	9.33×10^{-7}	9.39×10^{-8}	2.85×10^{-8}	1.11×10^{-7}
0.0168	1.05×10^{-6}	1.06×10^{-7}	3.23×10^{-8}	1.28×10^{-7}

Conglomerate

	Block 1	Block 2	Block 3
0.0548	1.99×10^{-4}	1.49×10^{-4}	2.30×10^{-4}
0.0768	1.41×10^{-4}	1.11×10^{-4}	1.63×10^{-4}
0.2832	7.28×10^{-4}	5.46×10^{-4}	8.43×10^{-4}
0.5880	1.49×10^{-3}	1.12×10^{-3}	1.73×10^{-3}

saturated hydraulic conductivity. This shows that conductivities are about twice as high for the conglomerate than the marl. The variation of estimates within each block represents the effect of a different Ksat, whilst differences between blocks show the effect of the 'b' parameter. In the conglomerate examples the calculated unsaturated hydraulic conductivities at 50% soil moisture varies between 1.73×10^{-3} to 1.11×10^{-4}, and on the marl between 1.05×10^{-6} to 7.50×10^{-9} (cms/min).

Steady state fluxes can now be estimated for the soil moisture values measured during the study period. The vertical flux (from Darcy's Law) is:

$$Qv = \frac{(z_1 - \Psi_1) - (z_2 - \Psi_2)}{L} \times K(\theta)$$

where z_1 and z_2 are depth values above a datum, Ψ_1 and Ψ_2 are the soil tension values determined from the retention curve, L is the distance between points z_1 and z_2, and $K(\theta)$ is the mean conductivity between positions 1 and 2. Similarly the downslope flux is:

$$Qd = \frac{(z_1 - \Psi_1) - (z_3 - \Psi_3)}{dist} \times K(\theta)$$

where this time the location subscripted '3' is downslope rather than vertically downwards.

Harr (1977) shows that the resultant flux is:

$$Qr = ((Qd + Qv.\sin\alpha)^2 + (Qv.\cos\alpha)^2)^{0.5}$$

where α is the slope angle, and the flux angle γ is:

$$\gamma = \sin^{-1}(Qd.\cos\alpha/Qr).$$

The direction and angle of fluxes in two dimensions is estimated for a number of days during the study period for the soil moisture conditions immediately above the gully head, that is tube 2 on site 1 and tube 1 on site 2. The tubes were too far apart to determine the changes in soil moisture just above the gully head, so to simulate the downslope flux a number of 'reasonable' soil moisture values were used for the downslope position.

On the marl soil the flux rates and angles were calculated using three sets of constants to portray the range of possible values obtainable from the different values of saturated hydraulic conductivity and retention curve data. These were as follows:

1 a = -2.772, b = -5.477, ksat = 0.0168 (from block 1)

2 a = -2.772, b = -5.477, ksat = 0.0036 (from block 1)

3 a = -3.713, b = -7.135, ksat = 0.0168 (from block 2).

The first model gives the highest values whilst models 2 and 3 indicate the lower range of values that may be expected. For 30 July 1982 the flux rate from 22cms depth is about 5.1 x 10^{-5} mm/24hours and lower down where the soil moisture increases rates are slightly higher at 1.4 to 2.5 x 10^{-4} mm/24hours. In the short term these rates are so small they can be considered negligible. However in the long term, especially during a drought period, very small fluxes may lead to a significant depletion of soil moisture. Table 6.11 should be examined in conjunction with figure 6.16 which shows the flux angle for three situations:

1 Where the soil moisture for a given horizon is the same, that is where there is no change in soil moisture downslope.

2 Where the soil moisture at 0.5m downslope is 1% (of the porosity) greater (ie plus 0.0035 MVF on conglomerate and 0.0055 MVF on the marl).

3 Where the soil moisture at 0.5m downslope is 1% (of the porosity) smaller (ie minus 0.0035 MVF on the conglomerate and 0.0055 MVF on the marl).

The flux lengths are not scaled to the flux itself, and the angles are similar regardless of which model is used. On 30 July the flux direction is vertically upwards and a 1% increase in soil moisture downslope makes no significant difference to the angle of flow. By 62cms depth the flux is downwards and a soil moisture increase or decrease of soil mosture downslope wil induce some change in the flux angle.

In the November/December period the soil moisture is fairly high in the upper horizon (about 62cms) but decreases rapidly to values similar to the summer (see figure 6.16). The flux rates calculated for 13, 27, and 28 November bear this out as the highest flux is from 22 cms between 9.0 to 1.4 x 10^{-1} mm/24 hours depending on the Ksat value, and decreases by three orders of magnitude to the next horizon. Figure 6.16 shows that the flux direction has changed from the previous period to downward fluxes

Table 6.11 Fluxes

(a) Fluxes (mm/24hours) estimated for different days and depths at tube 2 on the marl.

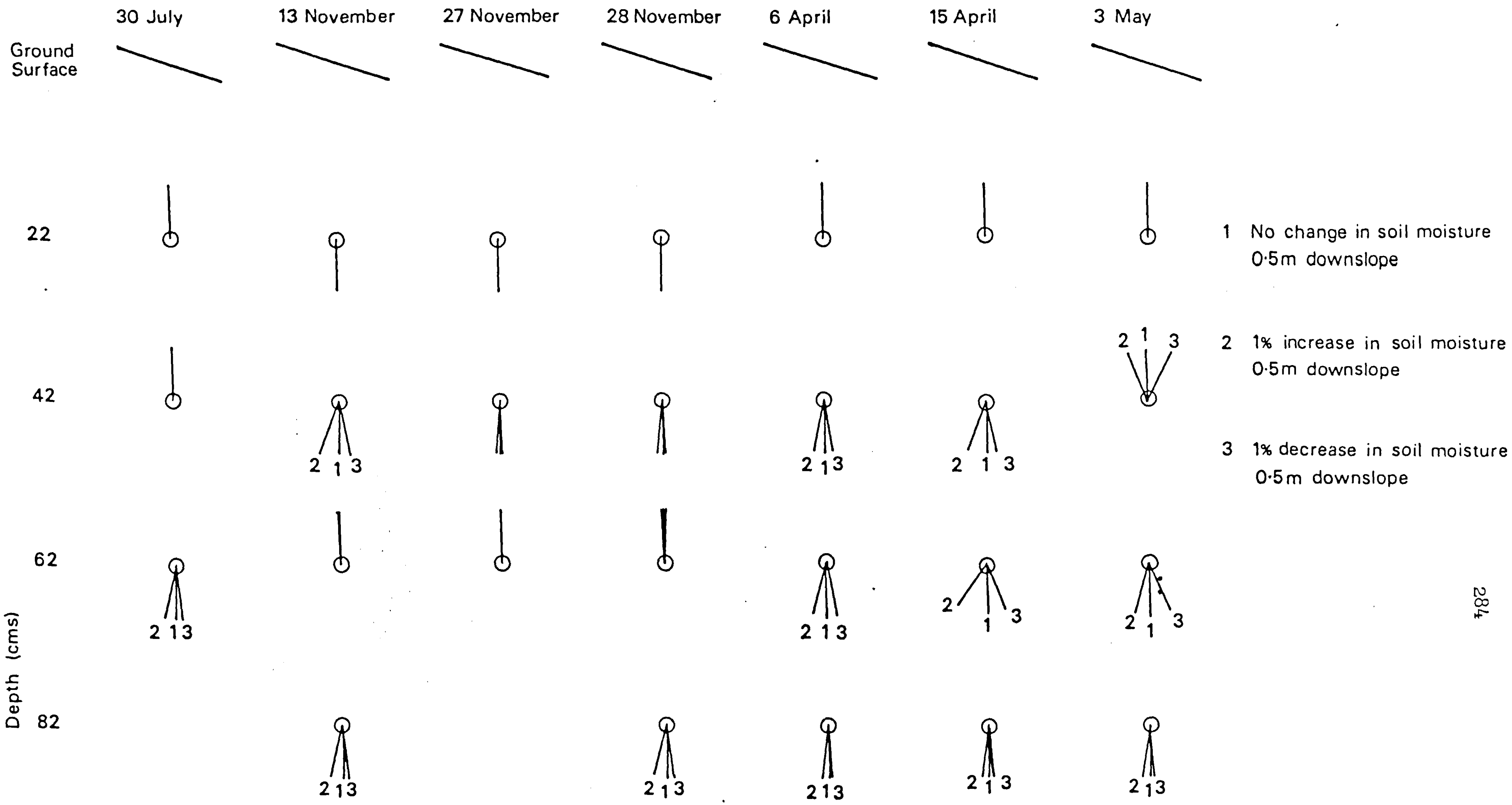
	Horizon 1	Horizon 2	Horizon 3	Horizon 4
July 30 1982				
1	$5.1x10^{-5}$	$2.5x10^{-4}$	$1.4x10^{-4}$	
2	$1.1x10^{-5}$	$5.4x10^{-5}$	$3.0x10^{-5}$	
3	$9.8x10^{-7}$	$6.6x10^{-6}$	$3.3x10^{-6}$	
November 13 1982				
1	$9.0x10^{-1}$	$2.1x10^{-5}$	$6.6x10^{-5}$	$5.6x10^{-5}$
2	$1.9x10^{-1}$	$4.6x10^{-6}$	$1.4x10^{-5}$	$1.2x10^{-5}$
3	$1.4x10^{-1}$	$3.5x10^{-7}$	$1.4x10^{-6}$	$1.2x10^{-6}$
November 27 1982				
1	$1.0x10^{-1}$	$5.2x10^{-4}$	$7.9x10^{-5}$	
2	$2.2x10^{-2}$	$1.1x10^{-4}$	$1.7x10^{-5}$	
3	$9.0x10^{-3}$	$1.6x10^{-5}$	$1.6x10^{-6}$	
November 28 1982				
1	$1.1x10^{-1}$	$6.4x10^{-4}$	$5.5x10^{-5}$	$5.6x10^{-5}$
2	$2.3x10^{-2}$	$1.4x10^{-4}$	$1.2x10^{-5}$	$1.2x10^{-5}$
3	$9.6x10^{-3}$	$2.6x10^{-5}$	$1.1x10^{-6}$	$1.1x10^{-6}$
April 6 1983				
1	$2.1x10^{-3}$	$5.0x10^{-4}$	$2.4x10^{-4}$	$1.9x10^{-4}$
2	$4.5x10^{-4}$	$1.1x10^{-4}$	$5.1x10^{-5}$	$4.0x10^{-5}$
3	$8.4x10^{-5}$	$1.5x10^{-5}$	$6.1x10^{-6}$	$4.6x10^{-6}$
April 15 1983				
1	$3.9x10^{-3}$	$2.2x10^{-4}$	$6.4x10^{-5}$	$2.5x10^{-4}$
2	$8.4x10^{-4}$	$4.8x10^{-5}$	$1.4x10^{-5}$	$5.3x10^{-5}$
3	$1.8x10^{-4}$	$5.9x10^{-6}$	$1.4x19^{-6}$	$6.4x10^{-6}$
May 3 1983				
1	$4.9x10^{-3}$	$4.2x10^{-5}$	$4.9x10^{-5}$	$2.1x10^{-3}$
2	$1.1x10^{-3}$	$9.1x10^{-6}$	$1.0x10^{-5}$	$4.6x10^{-5}$
3	$2.7x10^{-4}$	$9.4x10^{-7}$	$1.1x10^{-6}$	$5.4x10^{-6}$

Table 6.11 (Cont.)

(b) Fluxes (mm/24 hours) estimated for different days and depths at tube 1 on the conglomerate.

	Horizon 1	Horizon 2	Horizon 3	Horizon 4
November 24 1982				
1	15.0	6.8	28.7	5.7
2	159.2	71.8	303.4	59.9
November 27 1982				
1	5.0	13.0	26.1	3.2
2	52.8	140.3	276.7	33.7
November 28 1982				
1	6.3	18.7	20.0	6.7
2	65.3	198.0	211.5	70.9
April 6 1983				
1	5.3	2.1	21.5	208.3
2	56.3	22.0	227.7	2204.6
April 15 1983				
1	0.7	0.9	10.5	288.2
2	7.7	9.6	110.7	3059.9
May 3 1983				
1	0.4	0.1	2.4	300.1
2	3.9	0.6	25.5	3176.1

Figure 6.16 Flux angles on marl



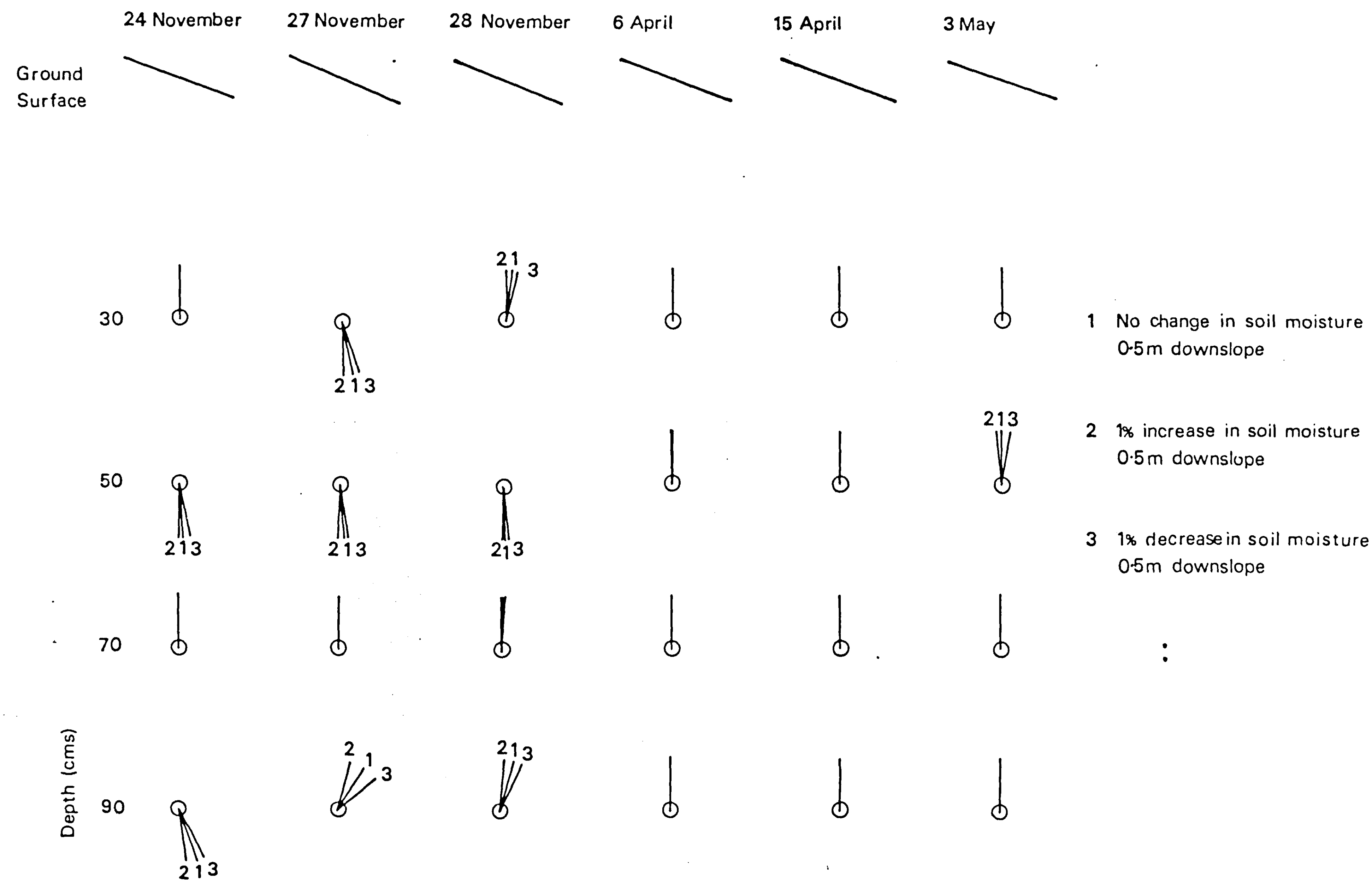
from 22 cms and 42 cms. On 13 November at 42 cms figure 6.16 shows that a 1% decrease in soil moisture over a 0.5 m distance downslope will induce a flux angle of 11° indicating a possibility of lateral movement although the volume of water involved will be small. The wettest recording of soil moisture at 22 cms was recorded on 13 November with 0.189 MVF, a week after the storm on 6/7 November with 132 mm of rain and a return period of 100 years. Assuming the falling soil moisture is linear,.we may project a soil moisture content of 0.2 MVF at 22cms and 0.088 MVF at 42 cms, which would give us a vertical flux of 4.66 mm/24hours.

In the spring period, drier soils mean lower fluxes, and already the direction of water movement for 22cms depth is upwards. Between 42cms to 82cms the fluxes are dominantly downwards, although slight decreases in soil moisture downslope can draw out lateral flow. Again the rate of flow is very small in the order of 1 x 10^{-3} to 10^{-6} mm/24 hours.

On the conglomerate at S2T1 table 6.11b shows that the magnitude of flux rates is much higher than the marl. The fluxes are calculated for the highest and lowest values of the saturated conductivity (0.0548 and 0.588 cms/min). The 'a' and 'b' parameters are taken for one block only as the three soil monoliths give very similar values. Before the storm on 24 November the fluxes are largest at 30 and 70cms with 15.0 and 28.7 mm/24 hours (or 159.2 to 303.9) respectively. Figure 6.17 shows that these fluxes are vertically upwards. At 50 and 90 cms rates are lower at 6.8 and 5.7 mm/24 hours, but the direction is downslope especially at 90 cms. On 27 November figure 6.17 shows that throughflow is occurring at 30 and 50 cms with flux rates about 5.0 to 13.3 (or 52.8 to 140.3) mm/24 hours depending on the values for ksat, whilst at 70 and 90cms the flux is upwards, but at 90 cms the direction has a downslope component. By 28 November the main difference is an increase in the flux at 50 cms, and the initiation of upward flux direction at 30 cms.

In April and May the fluxes at 30 and 50 cms are an order of magnitude lower than at 70 cms. For a given horizon the flux decreases with time and is 5.3 mm/24 hours at 30 cms on 6 April and 0.4mm/24 hours on 3 May. At 70 cms the flux is 21.5 mm/24

Figure 6.17 Flux angles on conglomerate



hours on 6 April and declines to 2.4 mm/24 hours on 3 May. However at 90 cms the flux is not only high at 208.3 mm/24 hours, but increases to 300.1 mm/24 hours by 3 May, and is substantially higher than the flux for the same location in November and December. This has been achieved by high suctions at 70 cms where there is a large difference in soil moisture between 70 cms and 90 cms. Figure 6.17 shows that during this period fluxes are dominantly vertical with downslope changes in soil moisture of 1% having little impact on the flow direction.

Conclusions

There are several problems with the modelling of the soil moisture fluxes. On the marl the analysis does not apply to horizon 1 where, in Chapter 4 on infiltration, flows (as indicated by final infiltrabilities) may be substantially higher, with a downslope component induced by high porosities and a sharp change in permeability at the bedrock boundary. The simulation procedure employed here uses high estimates of porosity on the marl of about 0.5 and does not take into account the decrease in porosity with depth observed (table 2.2). Furthermore the samples used to obtain the soil moisture characteristic curves do not necessarily portray the distribution of possible parameters. Despite the experimental and sampling problems the work on fluxes shows that soil moisture movement is different between the two lithologies.

In the marl rates of soil moisture movement are very low and would only lead to significant soil moisture changes over a long time period, for example during the summer drought. The highest rates simulated are only 0.9 mm/24 hours for 22cms depth on 13 November for model 1. In July the fluxes are dominantly upwards and can be interpreted as rates of actual evaporation assuming that the rate of evaporation is determined by the hydraulic characteristics of the soil. As such the actual evaporation rates caluclated for horizon 1 in table 6.11 are far lower than the potential evapotranspiration rates presented in table 6.12 which are estimated from monthly potential evapotranspiration data calculated using the Thornthwaite method (table 2.10). In the November period, following rain, fluxes are downward to 42 cms, and although small changes in soil moisture downslope may cause

Table 6.12 Potential evapotranspiration (mm/24 hours)

Month	Potential Evapotranspiration (mm/24 hours)
July 1982	4.7
November 1982	0.9
December 1982	0.4
April 1983	1.6
May 1983	2.1

some lateral flow, for example on 13 November at 22 cms, the quantities involved are very small, and probably within the limits of the actual evapotranspiration, thus having little influence on the stability of the gully head. By April the flux rates are approaching the July values and at 22 cms the direction of the flux is upwards as the effect of the drying begins to penetrate the soil. Already the actual evaporation rates are very small, for example the maximum value recorded at horizon 1 is 2.1×10^{-3} mm/24 hours on 6 April 1983. Thus in the marl flow is negligible, is within potential evapotranspiration rates, and is dominantly in the vertical plane, so throughflow does not occur in sufficient quantity to change the pore water pressures as a response to single storms or a prolonged rainy period, nor lead to the development of saturated conditions.

On the conglomerate during the November period there does appear to be some evidence for throughflow, which can be quite large with only a 1% decrease in soil moisture 0.5m downslope. For example on 27 November (depending on the value for Ksat) fluxes may vary between 5 to 13.0 mm/24hours (or 52.8 to 140.3 mm/24 hours depending on Ksat) at 30 and 50 cms depth at an angle of 19° at 30cms and 6° at 50cms (from 0° vertically down). This is to be compared with values of potential evapotranspiration (table 6.12). In November and December the potential evapotranspiration may be 0.9 to 0.4 mm/24 hours which is considerably less than the simulated seepage rates, providing excess moisture in the gully. In the spring fluxes are vertical and soil moisture changes downslope do not affect the direction of the flux. In the spring the potential evapotranspiration rises towards 2 mm/24 hours with increasing mean monthly temperature. Table 6.10b shows estimates of actual evapotranspiration from horizon 1 of 5.3 mm/24 hours on 6 April, falling to 0.7 mm/24 hours on 15 April, and 0.4 mm/24 hours on 3 May (or 56.3, 7.7 and 3.9 mm/24 hours depending on the value for Ksat) which indicate the fall of actual evapotranspiration as the soil becomes progressively drier.

It is interesting to note the very large flux from the 90cm level of 208.3 to 300.1 (or 2204.6 to 3176.1 mm/24 hours depending on the value for Ksat) which would be very important for maintaining vegetation growth at the lower part of the slope. This section is towards the conglomerate/clay interface where soil moisture values are much higher. Further up the hillslope at tubes 2, 3, and 4, comparable depths of 80 to 100 cms are much drier and the change in soil type was not encountered so the soils here drain to a greater depth. Although saturated conditions were not encountered on the conglomerate, higher flux rates and the greater incidence of throughflow would cause wetter conditions around and in the gully head than for the marl. This has several implications:

1 Seepage lines may develop along the conglomerate/clay interface and cause some localised erosion. It is noted for example that the depth of this boundary varied between pits 1, tube 1 and pit 2 across the gully head by 65cms, 117cms, and 105cms respectively.

2 The wetter soil conditions will favour vegetation growth in the gully floor and where the depth to the clay horizon is within the rooting zones of the plants.

3 The soil conditions will also favour the possibility of obtaining saturated soils on the gully floor with more rapid surface flow and erosion as a result of throughflow rates in excess of evapotranspiration, especially during the winter when soil moisture is high and the evaporative power of the atmosphere is reduced compared to the summer.

CHAPTER 7 CONCLUSION

This thesis has attempted to answer a specific question, namely what is the role and significance of surface and subsurface hydrology on gully head growth. Gully growth is often seen as a surface wash phenomenon although there is a lot of evidence from bank collapse as a result of high pore water pressure (Leopold 1964, Bradford and Piest 1977), seepage (Ireland et al 1939), and piping that the subsurface hydrology can have a significant impact on gully growth. Field work on gully growth processes have been limited to observations, mapping changes in morphology, measuring the effect of storms on headward recession, and measuring associated variables such as infiltration, and cohesion. The work by Bradford is an exception with quantification of the influences of pore water pressure on gully bank stability. Similar work has not been undertaken in semi-arid environments where, as Leopold et al (1964) suggest subsurface hydrology stills plays a significant part in gully erosion. Instead many reports tend to ignore the subsurface component assuming surface wash processes are dominant (Bryan and Yair 1982). Even in heavily piped gullies where there is a lot of subsurface water movement during storms the hydrology has not been studied. This thesis is therefore an original piece of work on the significance of surface and subsurface hydrology contributing to gully erosion for a semi-arid environment. There are two other important aspects of this research. Firstly it provides a good source for data on the soil properties of marl and conglomerate soils, both of which are widespread throughout south east Spain, for example data on $CaCO_3$ variation, porosity, and conductivity and retention curves. Secondly the soil moisture data are a rare record of detail and accurate soil moisture changes over three seasons in this environment. They give a good indication of the severity and effect of the drought, which has implications for research into vegetation growth and conservation, agriculture, soil erosion, and watershed management.

The main findings of the project cover four main aspects, climatic parameters, infiltration, and the effects of surface and subsurface hydrology.

Firstly precipitation has three main impacts on erosion through the seasonal distributions of available moisture for erosion and plant growth, interannual variability, and the return periods for different sized storms. The precipitation data clearly show that there is a marked seasonal distribution of rainfall with most of the rain falling between October and April. Temperatures also show a seasonal trend with high values from June to August. This creates an annual moisture stress during the summer which prevents the growth of a protective cover of vegetation. Storms may have a substantial erosional impact on the poorly vegetated ground, especially during the summer or at the beginning of the winter when there is also plenty of loose debris on the surface.

Interannual variability is considerable with annual rainfall varying by 30% on either side of the mean. Surprisingly there is no autocorrelation between the apparent irregular runs of years with above and below average rainfall. However the area is prone to long periods of below average rainfall with a spell of five consecutive years with below annual rainfall in 1962-68 and 1978-83, the latter period being a particularly severe drought. During the drought periods, vegetation cover can be reduced substantially, thereby increasing the amount of erosion per rainstorm. The records show that rainfalls of up to 172mm/24 hours have occurred, although this storm has a return period of about 1000 years. These high magnitude low frequency storms cause considerable erosion and landform change when they occur. At a smaller scale the apparent trend towards more storms greater than 10mm/24 hours in the 40 years record has significant implications for soil erosion.

Secondly the results from the infiltration experiments suggest that there is no significant difference between the marls and conglomerates for this site. This finding reflects on the changes caused to the soil structure by ploughing, albeit in the recent past. In cultivated areas the effects of ploughing on infiltration may last for 20 years or more. However evidence from the surface discharge suggests that runoff is actually greater on marl, which poses problems of measuring infiltration rates on a crusted surface using a cylinder infiltrometer. The infiltration would be expected to yield higher results, which would not be comparable to measured runoff. However despite this, tests on crusted marls by Scoging have still differentiated between soils and patterns of infiltration within soils. This still leaves the suggestion that ploughing increases the small scale variability and reduces systematic hillslope variations.

Thirdly the runoff and sediment transport data, although limited, does provide some insights to the role and significance of surface wash on gully head migration. On the 26 November 1982 the discharge on the conglomerate was about half the volume on the marl. On the conglomerate discharges were lower at the top and base of the slope and greater in the middle. On the marl the pattern is less clear, but again the least discharge occurred at troughs 1, 2, and 3 above the gully head. Near the gullies some factors such as increases in infiltration through armouring or vegetation cover or lower slopes, have caused a decrease in surface runoff. For neither lithology does the discharge correlate significantly with the amount of sediment transported, however if the quantities of sediment transport are assumed to represent the competence of the flow for erosion we see for large events that most erosion is caused on the steeper slopes, and least just above the gully heads or near the watershed. Thus the maximum impact of discharge and sediment transport is not near the gully heads themselves.

The marl is more susceptible to both overland flow production and sediment transport than the conglomerate. Mean values tend to be higher on the marl, furthermore the maximum values are considerably higher. For example maximum erosion on the marl for 1982 is 3.61 kg $m^{-2} \cdot yr^{-1}$ but only 1.95kg $m^{-2} \cdot yr^{-1}$ on the conglomerate. For period 4 there was no significant difference between the marl and conglomerate. This may reflect two things, firstly the impact of the wide range of sediment transport on the marl as indicated by the standard deviations, or the greater quantities of sediment transport on the conglomerate due to some erosion threshold. Overall the denudation rate for 1982 is high despite the low precipitation value, and in wetter years one may expect even higher rates of denudation.

It is difficult to judge the impact of sediment and water moving into the gully heads. Overland flow was not observed, however, photograph 7.1, taken about a week after the storm on 5-6 November 1982, shows that overland flow had occurred in the last storm. The wet part of the gully face was cut away to reveal a dry soil underneath proving this was a result of overland flow rather than throughflow. The photograph shows a lot of debris at the foot of the gully head which may have been deposited recently from bank collapse or sediment entrained from further upslope. Estimates of the effect of sediment transport suggest that even assuming no removal of sediment from the gullies the sediment transport by surface wash would not be sufficient to fill in the gullies in the middle term. It is expected that there will be a different threshold of sediment entrainment in the gully with respect to precipitation. An accumulation of debris in the gullies would provide high infiltration rates so that more water is required to produce overland flow in the marl gully than on the hillslopes. Gully clean out may occur less frequently than sediment deposition on the gully.

The data evidence suggests that long term changes would have an effect on gully growth on the conglomerate. Armouring would probably lead to greater lag deposits at the foot of the slope causing higher infiltration rates and less surface flow, thus reducing the surface component and increasing the subsurface one. On site 1 armouring does not appear to occur although coarse sediment is washed downslope. Another long term change is vegetation recovery. On many sections of the slope the vegetation cover is low and evidence from the sediment transport data suggests that increases in vegetation cover could lead to decreases in sediment transport.

Finally the analysis of the subsurface hydrology shows that on the marl seepage rates are negligible, but they may be appreciable on the conglomerate. Saturated conditions were not monitored anywhere, and maximum soil moisture values reached 55-75% saturation on the marl and 50-60% saturation on the conglomerate. However these were for a very dry year, and during a wetter period some locations either in the upper 40cms in the marl or in the conglomerate profile could reach saturation. The soil moisture

Photograph 7·1 Surface wash over the gully face on site 1

Photograph 7·2 The dry surface on the gully head face under wet conditions

data shows marked temporal variations. On a short term basis soil moisture in the upper horizon fluctuates showing the impact of individual storms, although the impact decreases with depth. Following rainfall the soil drains, for example in the spring session a drying profile is sustained for several weeks. Finally soil moistures are so low they change very slowly with time as shown for the July period.

The rates of soil moisture movement are negligible on the marl, the fastest rate being 0.9mm/24 hours. Rates of movement are greater nearer the surface. On the conglomerate the fluxes are much higher and evidence suggests that throughflow can develop in the soil. This has implications for seepage to the gully head but may not be sufficient to affect the stability of the gully head.

On both lithologies gully head migration will be dominated by processes resulting from surface wash rather than from throughflow. However on the conglomerate the seepage will be sufficient to affect vegetation growth and time to runoff in the gully floor.

Several problems have come to light in the course of the research project, including problems of variability and sample size, the use of different techniques to measure the same parameter, and the importance of accuracy and precision.

A number of parameters have been shown to be quite variable either spatially (as a result of soil variability), or temporally (due to seasonality and antecendent conditions). For example the values for $CaCO_3$, infiltration characteristics, and the saturated conductivities. In order to examine the distribution of these variables, large numbers (or large samples) should be taken. Logistically this is very difficult when undertaking field research based abroad. Whilst it may be simple to return 100 small samples to measure $CaCO_3$ values, it is impractical to return large numbers of soil monoliths for hydrological analysis. However whilst there may not be sufficient samples to study the distribution of, for example, saturated conductivity, a few samples can be used to get first estimates of the order of magnitude of the results providing care is taken to reduce sampling and instrumental errors.

Both accuracy and precision of measurements are desired, where accuracy means obtaining results about the right values, and precision refers to the replicability of these results. On occasions non-standard equipment had to be used which may have been less accurate than standard equipment, but more suitable to the problem in hand. For example with the saturated conductivities it is no use taking soil cores with an auger which will undoubtedly affect the soil structure in order to use a standard permeameter. Instead undisturbed cores were excavated and used in a less precise permeameter constructed around the sample.

A final problem has been that different techniques measuring the same parameter yield different results, for example, measuring infiltration or porosity.

The problems associated with measuring infiltration are well established. At one level infiltration rates obtained by sprinkler methods tend to be an order of magnitude less than those measured using a cylinder infiltrometer. The results from sprinklers tend to be considered nearer the true infiltration rates as the method resembles the infiltration process more closely and the final infiltrabilities approach saturated conductivities. However the difference between the two methods is a result of the measurement of two different suites of parameters associated with infiltration. Thus the cylinder infiltrometer measures water percolation from a standing head over a small area, whilst the sprinkler measures percolation from falling drops which may or may not develop into ponded conditions. The results have shown that a small change in the infiltrometer design itself is sufficient to cause differences in infiltration variables which emphasises the importance of maintaining consistency in the experimental design.

The two measures of porosity for the soil monoliths similarly show that different methods give different results. Both methods use very similar parameters, but in the first case porosity is determined largely by the relationship between bulk density and particle density, whereas in the second case porosity is the volume of water required to saturate the monolith. The first method is the standard way of measuring porosity, although there is no reason why the second method should be less efficient for non-swelling media.

There are four areas where future research may profitably be conducted following the results obtained in this thesis.

1 Modelling

A two dimensional hillslope hydrological model may be developed to examine changes in soil moisture around the gully head, in particular as a result of high magnitude low frequency storms, or following a sequence of moderate sized storms. This would help establish whether saturated conditions are feasible for wetter conditions which could not be established by field work during a drought. This could be combined with models of gully wall stability or headward progression.

A more detailed model would be to examine the influence of the gully head itself on soil moisture patterns upslope as a result of the draining ability of the head with lowered base level, throughflow, and evaporation from the gully wall. Atkinson (1978 p80) discusses the effect of soil pits measuring throughflow on future lines of flow in the soil. The pit actually interferes with the drainage lines within the hillslope. In semi-arid environments the effect of evaporation will also be an important factor as high temperatures encourage high potential evaporation rates. This may encourage rapid initial drainage, however evaporation rates for drier soils are governed by the moisture retention characteristics in the soil. In soils with low conductivities a relatively small volume of dry soil around the gully head may be sufficient to retard further evaporation. The existence of such a narrow strip of dry soil on gully walls is portrayed in photograph 7.2 which clearly shows a thin, dry layer of soil on the marl gully headcut backed by relatively wet soil.

Another model to test would be the effect of a draining gully head cut on vegetation growth upstream of the gully. This would involve more field work on the consumptive use of water and rooting characteristics of plants. However the effect of the gully head on the upslope soil moisture contents may be responsible for reducing soil moistures and killing off vegetation.

2 Vegetation

A second area for research has already been alluded to, namely the relationships between xerophytic vegetation, soil moisture and soil erosion. Very little information is available on such factors as the spatial and temporal consumptive use of water of these plants, nor upon their effectiveness in reducing rainsplash and surface wash erosion. In semi-arid areas, where there seems to be a delicate balance between vegetation cover and erosion, this would appear to be a very important area for research.

3 Soil Crusts

Some research is needed to examine the development, but more importantly the stability and break down of crusts under rainfall, and the effectiveness of crusts in inhibiting infiltration.

4 Subsurface Hydrology and Piping

A final avenue for research is to further the analysis on the subsurface hydrology of the marls, with particular reference to the influence of the chemistry of the marls on subsurface hydrology, and the development of piping. It seems paradoxical that very large pipes can develop in a lithology with such small conductivities. Although water entry is probably by cracks in the soil surface, initial movement of water to the outflow (before the pipe has developed) must be by throughflow. From first impressions, piping seems to occur under two conditions, firstly where the soil moisture and hydraulic gradients are artificially high such as on terraces, and secondly where the marl has high contents of sodium or gypsum, for example, around Tabernas.

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APPENDIX 1 Daily rainfall at Ugijar 1980 to 1984

1980/81

Day	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep
1	-	-	-	-	-	-	-	4.5	-	-	-	-
2	-	25.5	-	-	-	-	-	-	-	-	2.0	-
3	-	-	-	-	-	-	6.5	-	-	-	2.5	-
4	-	13.0	-	-	-	-	-	-	-	-	-	-
5	-	-	-	-	-	-	-	-	-	-	-	-
6	-	-	-	-	-	-	-	-	-	-	-	-
7	-	-	-	-	-	-	-	-	-	-	-	-
8	-	17.5	-	-	-	-	-	-	-	-	-	-
9	-	-	-	-	-	-	-	-	-	-	-	-
10	-	-	-	-	-	-	-	-	-	-	-	-
11	-	13.5	-	-	-	-	-	-	-	-	-	-
12	-	4.5	-	-	-	-	-	-	-	-	-	-
13	-	-	-	-	-	-	-	-	-	-	-	-
14	-	-	-	-	10.0	-	-	-	-	-	-	-
15	-	-	-	-	-	-	-	-	-	-	-	-
16	2.5	-	-	-	-	16.5	16.5	-	-	-	-	-
17	-	-	-	-	-	-	-	-	-	-	-	-
18	-	-	-	-	-	-	-	-	-	-	-	-
19	-	-	-	-	-	-	-	-	-	-	-	-
20	-	-	-	-	-	5.0	10.0	-	-	-	-	-
21	-	-	-	-	3.5	-	6.5	-	-	-	-	-
22	-	-	-	-	-	-	22.5	-	-	-	-	-
23	-	-	-	-	-	-	-	-	-	-	-	-
24	-	-	-	-	-	-	-	-	28.0	-	-	-
25	-	-	-	-	-	-	-	-	6.5	-	-	-
26	2.0	-	-	-	-	-	-	-	-	-	-	-
27	-	-	-	-	-	-	-	-	-	-	-	-
28	-	-	-	-	-	-	10.0	-	10.5	-	-	-
29	-	-	-	-	xxxxx	-	8.0	-	-	-	-	-
30	-	-	-	-	xxxxx	10.5	12.0	-	-	-	-	-
31	-	xxxxx	-	-	xxxxx	-	xxxxx	-	xxxxx	-	-	xxxxx
Total	4.5	74.0	0.0	0.0	13.5	32.0	92.0	4.5	45.0	0.0	4.5	0.0

1981/82

Day	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep
1	-	-	-	-	-	-	-	-	-	-	-	-
2	-	-	-	-	-	-	-	-	-	-	-	-
3	-	-	-	3.5	-	-	-	-	-	-	-	-
4	-	-	-	-	-	-	-	-	-	-	-	-
5	-	-	-	-	-	-	-	-	-	-	-	-
6	-	-	-	-	-	-	-	-	-	-	-	-
7	-	-	-	-	-	-	-	-	-	-	-	-
8	-	-	-	-	-	-	-	-	-	-	-	-
9	-	-	-	-	-	-	-	-	-	-	-	-
10	-	-	-	-	-	-	10.0	3.0	-	-	-	-
11	-	-	-	22.5	-	-	-	-	-	-	-	-
12	-	-	-	15.4	-	-	-	-	-	-	-	-
13	-	-	-	6.7	-	-	-	-	-	-	-	-
14	-	-	-	-	8.0	-	-	-	-	-	-	-
15	-	-	-	-	13.0	-	-	-	-	-	-	-
16	-	-	-	15.8	-	-	-	-	-	-	-	-
17	-	-	-	7.5	-	-	-	-	-	-	-	-
18	-	-	-	-	-	-	-	-	-	-	-	-
19	-	-	-	-	-	-	-	-	-	-	-	-
20	-	-	-	-	-	-	-	-	-	-	-	-
21	-	-	15.0	-	-	-	14.5	-	-	-	-	-
22	4.5	-	-	-	-	-	-	-	-	-	-	-
23	-	-	-	-	-	-	6.0	-	-	-	-	-
24	-	-	-	-	-	-	-	-	-	-	-	-
25	-	-	-	-	-	-	-	-	-	-	-	-
26	-	-	-	-	-	7.0	-	-	-	-	-	-
27	-	-	-	-	-	-	-	-	-	-	-	-
28	-	-	-	-	-	-	-	-	-	-	-	-
29	-	-	32.7	-	xxxxx	-	2.0	-	-	-	-	-
30	-	-	23.0	-	xxxxx	3.5	-	-	-	-	-	-
31	-	xxxxx	-	-	xxxxx	32.0	xxxxx	-	xxxxx	-	-	xxxxx
	4.5	0.0	70.7	71.4	21.0	42.5	32.5	3.0	0.0	0.0	0.0	0.0

1982/83

Day	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep
1	-	1.2	-	-	-	-	-	-	-	-	-	-
2	-	-	-	-	-	-	-	-	-	-	-	-
3	-	-	-	-	-	-	-	-	-	-	-	-
4	-	-	-	-	-	-	-	-	-	-	-	-
5	-	-	-	-	-	-	-	-	-	-	-	-
6	-	62.0	-	-	-	-	-	-	-	-	-	-
7	-	70.0	-	-	-	-	-	-	-	-	-	-
8	-	-	-	-	-	-	-	-	-	-	-	-
9	-	-	-	-	-	-	-	-	-	-	-	-
10	-	-	-	-	-	-	-	-	-	-	-	-
11	-	-	-	-	-	-	-	-	-	-	-	-
12	-	-	-	-	7.0	-	-	-	-	-	-	-
13	-	-	-	-	-	-	-	-	-	-	-	-
14	-	-	-	-	9.0	-	-	-	-	-	-	-
15	-	-	-	-	-	-	-	-	-	-	-	-
16	-	-	-	-	-	-	-	-	-	-	-	-
17	-	-	-	-	-	-	-	-	-	-	-	-
18	-	-	-	-	-	-	-	-	-	-	-	-
19	-	-	-	-	-	-	-	-	-	-	-	-
20	12.0	-	-	-	-	-	-	-	-	-	-	-
21	15.5	-	-	-	-	-	-	-	-	-	-	-
22	-	-	-	-	-	2.5	-	-	-	-	-	-
23	-	-	-	-	-	-	-	-	-	-	-	-
24	-	-	-	-	-	-	-	-	-	-	-	-
25	-	-	-	-	-	-	-	-	-	-	-	-
26	-	23.0	-	-	-	-	-	-	-	-	-	-
27	-	-	-	-	-	-	-	-	-	-	-	-
28	-	-	-	-	-	-	-	-	-	-	-	-
29	-	-	-	-	xxxxx	-	-	-	-	-	-	-
30	-	-	-	-	xxxxx	-	-	-	-	-	3.5	-
31	-	xxxxx	-	-	xxxxx	-	xxxxx	-	xxxxx	-	-	xxxxx
Total	27.5	156.2	0.0	0.0	16.0	2.5	0.0	0.0	0.0	0.0	3.5	0.0

1983/84

Day	Oct	Nov	Dec	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep
1	-	-	-	-	-							
2	-	-	-	-	-							
3	-	48.0	-	-	-							
4	-	15.0	-	-	-							
5	-	10.5	-	-	-							
6	-	5.0	-	-	-							
7	-	-	-	-	-							
8	-	-	-	-	-							
9	-	-	-	-	-							
10	-	12.0	-	-	-							
11	-	-	-	-	-							
12	-	-	-	-	-							
13	-	-	-	-	-							
14	-	11.5	-	-	-							
15	-	3.7	25.5	-	-							
16	-	35.0	-	-	-							
17	-	22.0	8.5	-	-							
18	-	-	15.5	-	-							
19	-	-	-	-	-							
20	-	33.5	19.0	-	-							
21	-	3.5	-	-	-							
22	-	-	-	-	-							
23	-	-	-	-	-							
24	-	-	-	-	3.0							
25	-	-	-	-	3.5							
26	-	-	-	-	0.5							
27	-	-	-	-	35.0							
28	-	-	-	-	10.5							
29	-	-	-	-	xxxxx							
30	-	-	-	-	xxxxx							
31	-	xxxxx	-	-	xxxxx							
Total	0.0	199.7	68.5	0.0	52.5							