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Karst water studies and environment in West Malaysia

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by

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KARST WATER STUDIES AND ENVIRONMENT IN WEST MALAYSIA

VOLUME I

Figures, tables and plates
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Summary of Thesis submitted for Ph.D. degree

by John Crowther

on

Karst water studies and environment in West Malaysia

The overall aim of the study is to investigate the environmental characteristics and karst water properties of selected limestone outcrops in Selangor, the Kinta Valley and the North-west region, West Malaysia. The meso-topography, soil and vegetation cover of 13 plots, which encompass the range of weathering environments observed in the limestone hills, are described. Reports are also given on regular measurements of throughfall, litterfall, throughflow, subcutaneous flow, soil moisture, soil carbon dioxide and karst water properties made over a one-year period.

Five main conclusions result from the investigation. First, the surface of the limestone hills comprises a mosaic of weathering environments, determined principally by topographic position and rock purity. Each is distinguished by the nature and depth of soil cover, by the characteristics of the vegetation canopy, and by its meso-relief. Secondly, rapid cycling of nutrients, especially potassium and calcium, takes place within the soil/vegetation system. Thirdly, marked systematic, spatial and temporal variations occur in karst water properties. The most important distinction is between groundwaters, which characteristically display only narrow hardness and temperature fluctuations, and surface waters, which exhibit much greater variability. Fourthly, a high proportion of solution occurs near the surface, and much secondary deposition takes place within the outcrops, especially where, as in the tower karst hills, groundwater flow is generally of low velocity. Variations in net denudation rates, which range from 38.4 to 81.0 mm/1000 yrs, are largely determined by differences in effective rainfall. Finally, detailed comparisons of the hydrological and denudational processes operative within the contrasted outcrops of the Setul Boundary Range and the tower karst hills suggest that

whilst karstification processes under humid tropical conditions seem particularly favourable for the formation and preservation of tower karst hills, dominant vertical and subvertical fractures are normally prerequisite for their development.

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CHAPTER I

A REVIEW OF KARST RESEARCH IN THE HUMID TROPICS

CHAPTER I. A REVIEW OF KARST RESEARCH IN THE HUMID TROPICS

> "There is doubtless no other discipline within geomorphology that is weighted down with so many disputable interpretations of forms, and controversial unsolved problems, as the field of karst morphology."
>
> (L. Jakucs, 1977, p. 9).

A. INTRODUCTION

General discussions on landform development and hydrology in regions of karst terrain and recent trends in karst research have been presented in several recent reviews (Ford and Cullingford, 1976; Herak and Stringfield, 1972; Jakucs, 1977; Jennings, 1971; Marker, 1978; Roglić, 1972; Shuster and White, 1971; Sweeting, 1972, 1976). This introductory chapter, therefore, focuses specifically upon humid tropical karst. It summarizes early observations and ideas, examines more closely the way in which the main hypotheses and methods of research have evolved over the last two decades, and evaluates the role of humid tropical studies as a component of the world-wide field of karst geomorphology. The need for further investigation of certain aspects of karst landscape in the humid tropics is demonstrated, and the purpose of the present study is established. The structure of the thesis is outlined in the final section.

B. EARLY OBSERVATIONS AND HYPOTHESES

There were few accounts of tropical karst before 1900, and these were mainly descriptive and non-analytical. One of the earliest, by Sawkins et al. (1869), concerned the dolines of Jamaica (cockpits, in later terminology). These, by analogy with the dolines of Europe, were regarded as solutional forms. In an early reference to the limestone outcrops of the Malay Peninsula, Daly (1879) recounts the discovery of Bt. Batu (Bt., bukit - hill). A

more detailed report on this hill and its cave systems was presented by Ridley (1898). Perhaps the first geomorphological observations were by the Rev. J.E. Tenison-Woods (1855, p. 1188), a geologist, who described the limestone towers of the Kinta Valley as forming "a series of detached almost conical hills seldom above 1500 feet high and presenting precipitous sides of greyish or blueish (sic) rock very beautifully variegated with stalactites and various infiltrations of iron oxides." However, he misinterpreted the origin of the granite detritus found in caves at heights of several hundred feet above the present level of the plain. Consequently, his suggestion that the limestone had been subject to "enormous denudation and of a very rapid kind" (ibid, p. 1189), was not properly substantiated by the evidence advanced.

At the turn of the century, Penck (1900) and Davis (1901) presented schemes for the cyclical evolution of the karst landscape. In these they postulated that tectonic uplift, followed by a period of fluvial erosion, subsequently gave way to karstification as bedrock joints became widened by solution and underground drainage systems developed. A peneplain was considered to be the ultimate product of karstic denudation. Shortly afterwards Daneš (1908, 1910), in the first substantial geomorphological studies of tropical karst, drew attention to the frequent occurrence of hills with convexo-concave slope profiles (cockpit karst; Ger. Kegelkarst) in Jamaica and in the G. Sewu region, Java (G., gunong - mountain). Grund (1914) interpreted the features as described by Daneš as having evolved from dolines, and regarded the cockpit terrain as representing a distinct stage in karst landform development, intermediate between doline karst and a corrosional plain. Thus, at a time when geomorphologists were becoming increasingly aware of the role of climate in landform development (eg. Davis, 1905; de Martonne, 1913; Passarge, 1926; Penck, 1910), the poss-

ibility of climatically-determined variations in denudation processes and landforms in karst terrain was largely overlooked. The azonal view of karstification, advocated by Grund (1914), remained largely unchallenged until 1936.

After the pioneering work by Daneš, the landforms of several tropical karst regions, notably Indonesia, were investigated in detail and painstakingly recorded (eg. Bothe, 1929; Escher, 1931; Hol, 1918). Over the same period, studies in the Malay Peninsula, chiefly by geologists, focused upon the origin of the tower karst hills (Ger. Turmkarst) of the Kinta Valley. Divergences of opinion on this matter arose mainly from different interpretations of the stratigraphy. Scrivenor (1913, 1923) proposed that the hills had resulted from block-faulting, Rastall (1927a, 1927b) that they were aligned along anticlinal ridges which had subsequently been faulted, and Cameron (1925a) that they were of younger, more massive and less disturbed limestones than those of the adjacent plains, which they unconformably overlay. However, recent surveys have shown that faults with the magnitude of throw necessary to support Scrivenor's thesis are absent, and that the limestones of the towers are continuous with and of the same age as those which form the adjacent karst plain. Clearly, a simple structural explanation is thus untenable. An erosional theory was first advocated by Jones (1917), who noted the close relationship between the drainage pattern of the valley and the distribution of the hills, and explained this by the uneven denudation of a strongly-jointed metamorphic limestone and the subsequent mechanical and chemical denudation of the surrounding limestone. The general principle of differential erosion is still held to be the main cause of tower karst in the Peninsula (Paton, 1964).

C. INVESTIGATIONS DURING THE PERIOD 1936 TO 1960

As more detailed information became available on the morphology of tropical karst regions, doubts were expressed concerning the validity of the universal cycle of karst erosion. Lehmann (1936) noted that depressions in the G. Sewu karst are star-like in plan, and therefore could not originate from dolines of the type characteristic of European karst, which are generally circular. He also postulated that the dome-like hills of the Sewu region were formed at a very early stage in denudation, and are not secondary features as is implicit in the cyclical model (Grund, 1914). From these observations first emerged the concept of a separate and highly distinctive tropical karstification. Between 1936 and 1960 cockpit karsts, with similar morphologies as the area studied by Lehmann, were reported from Jamaica (Lehmann, 1955; Sweeting, 1958), Indonesia (Sunartadirdja and Lehmann, 1960) and other parts of the humid tropics, and tower karsts were described from Cuba (Lehmann *et al.*, 1956) and S.W. China and Vietnam (Wissman, 1954). The fact that cockpit and tower karsts are confined mostly to tropical latitudes provided the basis for a climato-morphogenic approach to karst landform development. This school of thought gained much support, particularly in the early 1950s (eg. papers presented at the International Geographical Union, Symposium on Karst Morphogenetics, 1953). By 1960, the term 'cone karst' (Ger. *Kegelkarst*), which subsumes both cockpit and tower karst (Jennings, 1971, p. 187), became almost synonymous with the humid tropics, and attempts were made to define thresholds of rainfall and temperature which determined zones of development (Gerstenhauer, 1960; Sweeting, 1958; Verstappen, 1960a).

The proposition that the nature and effectiveness of geomorphic processes vary from one climatic zone to the next is central to the climato-morphogenic thesis. Lehmann (1936), for

instance, observed that infiltration in the G. Sewu region is impeded by a partial soil cover and that, in contrast with temperate karst regions, much surface runoff occurs during the frequent intense storms. Such runoff accounts for the development of linear erosion features, and these give depressions their distinctive stellate form.

Compared with karstlands in temperate environments, those in the humid tropics generally have a more accentuated mesorelief and appear to be mostly at a more advanced stage of erosion. These differences may be interpreted in two ways. First, they may reflect different rates of karstification in the two areas. In the humid tropics high temperatures, abundant rainfall, and high rates of release of biogenic carbon dioxide and of organic acids into the weathering environment may combine to produce rapid karstification, particularly of the surface and near-surface zone (Birot, 1954; Bögli, 1951; Lehmann et al., 1956). This was supported by the hardness of waters in Cueva de Indio, Cuba, which were saturated with calcium carbonate after passing through only 4 m of rock (Lehmann, 1955), and also by the high solute contents of runoff from solution runnels in Cuba (Lehmann et al., 1956). Secondly, the contrasts observed between mid-latitude and tropical karst may reflect differences in the duration of the present phase of karstification. Thus karst landform development in the humid tropical lowlands may be more advanced because it was not interrupted by the glaciations which affected cold and temperate regions during the Quaternary. This explanation was advocated by Corbel (1959a), who presented data from diverse climatic zones which indicated that chemical denudation was more effective at higher latitudes. According to Corbel, the inverse relationship between temperature and the amount of carbon dioxide absorbed by water (Henry's Law) is the dominant control in the dissolution of limestone. Much more data were required before either of these

opposed hypotheses could be tested adequately.

D. THEMES AND METHODS OF TROPICAL KARST RESEARCH OVER THE LAST TWO DECADES

L. Jakucs (1977, p. 134) describes the karst regions of the humid tropics as being "practically impenetrable owing to the strikingly-accidented relief, dense growth of vegetation, and the unhealthy oppressive humidity of the near-ground atmosphere". He goes on to suggest that this is why these areas are "among the least-known regions of the earth; and why the study of their forms and of the intensities of the forces forming them will be among the more arduous tasks of the future". Seen in this light, a noteworthy feature of tropical karst research over recent years has been the pioneering spirit with which it has been undertaken. Since 1960, the surface and underground features of previously unknown and uncharted limestone outcrops have been described, mapped and, in many cases, subjected to closer and more rigorous scientific investigation (eg. Aldabra Atoll - Trudgill, 1972b; Belize - McDonald, 1975; Madagascar - Rossi, 1975; New Guinea - Brook, 1976a; Williams, 1972a; Sarawak - Wilford and Wall, 1965; Sweeting, 1979; Sweeting et al., 1980; Tanzania - Cooke, 1971; Thailand - Pitman, 1978b; Venezuela - Gascoyne, 1974). These, and other studies, have brought a greater awareness of the spatial variability present in the form of tropical karst terrain and in the processes and controls operative within it. Even the most basic information on the geology, climate, soils, vegetation, and morphology of these remote regions augments considerably the data upon which geographical interpretations are founded. Furthermore, knowledge of the geography of these regions provides a rational basis for the selection of sites for closer scientific investigation.

Over the last two decades, three distinct trends may be

identified in tropical karst research. First, the form of tropical karst terrain has been described and analysed quantitatively. Secondly, the significance of geological control has been more widely recognized. Finally, hydrological and denudational processes, and some of the factors controlling them have been measured in a number of regions. As a result of these diverse studies, and reports on hitherto unknown karst regions in several parts of the world, the relative importance of climatic control has been closely examined. It is convenient to consider these various themes separately.

1. Description and analysis of karst forms

The sheer variety and intricacy of karst forms explain why a complex, unwieldy and genetically-based terminology has evolved in karst literature (Balázs, 1973b). Landforms may be defined more precisely and objectively by morphometric methods, and indices applicable to tropical forms have been utilized by Balázs (1973b), Day (1977a) and Williams (1971, 1972a, 1972b).

To date, morphometric procedures have been applied mainly to major landform elements. At a basic level, the frequency or areal extent of a particular phenomenon may indicate its significance and mode of formation. Thus, the high frequency of cliff-foot caves compared with other types of foot-slope around tower karst hills in Malaysia (Jennings, 1976) and Sulawesi (McDonald, 1976a) confirms the importance of basal undercutting in tower karst formation. Furthermore, the size distribution of cliff-foot caves around Bt. Batu, West Malaysia suggests that they develop initially from many small features, which merge by lateral extension into a few larger ones (Jennings, 1976). More complex morphometric studies in New Guinea have shown that tropical karst terrain, rather than being irregular and chaotic in form, may comprise a mosaic of contiguous, polygonally-shaped surface catchments (Williams, 1971, 1972a, 1972b). The distribution of stream sinks in this area

tends towards a uniform pattern. The size, shape, orientation and distribution of karst depressions in Jamaica and of limestone hills in northern Puerto Rico have been investigated by Day (1976, 1978).

At a micro-scale, indices devised to quantify solution flutes (Ger. Rillenkarren) have been applied in the Chillagoe karst region, North Queensland (Lundberg, 1977b). However, meso-scale features, such as the larger solution forms, rock pinnacles and hill slope profiles, await quantitative analysis.

2. Geological factors in karst landform development

Certain striking landforms reflect direct geological control. For instance, a number of the dome-like hills in the Sewu region are bioherms (Sartono, 1964 - cited by Verstappen, 1969), and not, as was previously believed, the products of a distinctive karstification. More generally, however, lithological and structural characteristics exert an indirect control by influencing the effectiveness and distribution of denudation processes within the landscape.

Some of the most marked contrasts have been observed between porous limestones and more massive ones of low porosity, with well-defined joint patterns. In North-Central Jamaica, for example, pure limestones of different porosity are juxtaposed. The Montpelier Beds of high porosity, but with few dominant joints, give rise to a smooth rounded topography with dry valley features, while the topography of the adjacent White Limestone is of the cockpit karst type, with dissolution being confined mainly to the major joints (Sweeting, 1958; Smith, 1972; Smith and Atkinson, 1976). However, in the Star Mountains, Indonesia, conical forms are better developed on more porous limestones than in more compact rocks, where sinkholes form the more dominant landscape elements (Verstappen, 1969). Rocks of particularly low primary porosity and with pronounced vertical structural lineations often give rise to tower karst forms, as for example in Sarawak (Wilford and Wall,

1965), Maros, Indonesia (Sunartadirdja and Lehmann, 1960), Chillagoe, North Queensland (Ford, 1978), and West Malaysia (Gobbett, 1971). In the G. Sewu area of Java, the spacing of horizontal planes of weakness may be an important determinant of the form of the residual hills (Tjia, 1969). Structural control has also been recognized in the form of asymmetric residuals in Cuba (Panoš and Štelcl, 1968).

The chemical and mineralogical composition of limestones also affect landform development. For instance, marked contrasts in surface erosion rates occur between different coral limestones on Aldabra Atoll (Trudgill, 1972a, 1972b, 1976b). Non-carbonate impurities are particularly important. In the Star Mountains, for example, they reduce infiltration in more porous limestones and thus favour sinkhole development (Verstappen, 1960b). Infiltration is similarly impeded where limestones in New Guinea are interbedded with cherts and mudstones (Williams, 1972a). Tower karst formation in the Malay Peninsula is restricted to the purer limestones (Paton, 1964).

The contrasted landscapes developed on the more porous lithologies in Jamaica and Indonesia, as outlined above, clearly demonstrate that the effects of a particular geological factor may differ according to local circumstances. Further research is required to elucidate the mechanisms of geological control in karst landform development. The relationships between the lithological and structural properties of limestones and the hydrological and chemical characteristics of karst groundwaters also remain to be established.

3. Studies of geomorphic processes and factors controlling them

3.1. Hydrology

Hydrological observations in tropical karst regions have been mostly confined to the surface and near-surface zone. In many

areas runoff to stream sinks forms an important component of karst drainage during intense storms. For example, Balázs (1971) estimates that in an area of relatively undissected karst in the Barung Islands, off the east coast of Java, as much as 40 per cent of the rainfall runs off as surface flow. Surface runoff is also a significant erosional agent (Sweeting, 1972, p. 270), as in the Star Mountains, Indonesia, where it has led to the formation of steep ravines which end in sinkholes (Verstappen, 1969). In contrast, Williams (1972a) finds that in New Guinea lateral movement of water towards the sinks, as diffuse surface flow, throughflow and seepage through the uppermost layers of weathered rock (termed 'subcutaneous flow' - Birot, 1966), is mainly responsible for the development of convexo-concave hillslope profiles. Also, the hardness of waters in semi-permanent pools in the cockpit region of Jamaica indicates that much of their water is derived from subcutaneous flow (Day, 1976). The hydrological characteristics of groundwaters are less well known.

3.2. Karst water quality

Groundwater chemistry has been investigated closely at a number of sites in the humid tropics. In Madagascar, for instance, distinct seasonal fluctuations in water chemistry have been recorded at two resurgences, with total hardness increasing during the wetter part of the year (Rossi, 1976). This pattern contrasts with one observed in caves along the lower gorge of the Mkulumuzi River, north-east Tanzania, where, despite a rapid increase in plant growth at the beginning of the wet season, there was little seasonal variation in the hardness of cave drips (Cooke, 1971). However, at sites of active underground deposition positive correlations were found between total hardness and the volume of flow. The latter relationship probably reflects the higher proportion of dissolved carbonate which may be deposited at lower discharges.

The average solute contents of waters in Madagascar and north-east Tanzania, and of those recorded for a wide range of sites in Indonesia (Balázs, 1971, 1973a) and Jamaica (Smith, 1969a), lie within the range of values recorded at sites in higher latitudes.

Data based upon less frequent or even single samples are available for many sites, and provide some measure of the magnitude and distribution of chemical denudation within the tropical karst landscape. In the Phangnga region of Thailand, for example, waters sampled on one occasion at each of 15 spring sites were mostly saturated or supersaturated with calcite, and had calcium carbonate concentrations in the range 74 to 191 parts per million (ppm) (Pitman, 1978b). Water sampled on one occasion at a point 200 m downstream from Cueva de Cuarto Vientos, Venezuela, was found to have a total hardness of 55 ppm lower than at the resurgence, the difference in hardness being indicative of rapid deposition (Gascoyne, 1974). Isolated 'spot' samples of this type have also been taken from stagnant waters in pools on rock surfaces (Balázs, 1968; Wilford and Wall, 1965) and from surface flow over bare and lichen-covered surfaces (Muxart et al., 1969).

Downpours of high intensity and magnitude are characteristic of the humid tropics. In Jamaica, total hardness maxima have been recorded at the beginning of flood pulses, as waters already present within joints are flushed out (Ashton, 1967). A different and more complex pattern was observed during a flood at the Hueque Resurgence, Venezuela (Gascoyne, 1974). Total hardness decreased rapidly with increased flow, but the minimum hardness occurred six hours after peak discharge. The time taken for the return of pre-flood base flow discharge and water hardness levels was 2½ and 6 weeks, respectively. Depletion of soil carbon dioxide during the storm, and the time taken subsequently to build up carbon dioxide concentrations in the soil are thought to account for the

sluggish response of water chemistry to changing flow conditions.

3.3. Distribution of chemical erosion within the landscape

Most published denudation rates for karst terrain are of two types. The first relate to net amounts of chemical erosion from entire outcrops, estimated from the solute contents of effluent waters and the total volumes of annual runoff (eg. Balázs, 1971, 1973a). The second concern denudation at single locations within the limestone system, based upon either the chemical quality of surface, soil and groundwaters, or upon direct measurement of surface lowering by means of a micro-erosion meter (High and Hanna, 1969) or erosional weight-loss tablets (Trudgill, 1975). Studies on Aldabra Atoll have revealed marked variations in surface erosion rates between different types of limestone and also between different degrees and types of soil cover (Trudgill, 1976b), while in Jamaica, higher rates of surface erosion have been recorded in the wet season as compared with the dry season (Trudgill, 1977a).

Visible evidence of secondary deposition is provided by case-hardening phenomena (Ireland, 1979; Monroe, 1966; Panoš and Štelcl, 1968; Thorp, 1934; Trudgill, 1976a), by speleothems in cave systems (Wilford, 1964), and by tufa formations along the beds of surface rivers (Hossfield, 1951; Marker, 1973a). Because of the massive scale of depositional forms in the humid tropics, net erosion rates may poorly reflect the gross movement and redistribution of limestone within the karst system (Bögli, 1960, p. 15; Jakucs, 1977, p. 107; Jennings, 1971, p. 193; Jennings and Bik, 1962). To date, however, studies which have monitored erosion at several locations within single limestone outcrops (eg. Corbel and Muxart, 1970; Day, 1976; Smith et al., 1972) provide no estimate of the depositional component of groundwater flow.

3.4. Mechanical erosion

Until recently, mechanical erosion had only occasionally

been considered as a significant denudational agent in karst terrain. Although detailed measurements such as those from Britain (Newson, 1970; Smith and Newson, 1974) are as yet unavailable for the humid tropics, existing data indicate higher rates of mechanical erosion in the latter regions than in temperate environments. Balázs (1973a), for example, estimates the rates to be between five and ten times higher in Indonesia than in Hungary. This difference is probably caused by the greater frequency of intense storms in the former area. Furthermore, in the mostly undissected karst of the Barung Islands, Indonesia, it has been estimated that 93.3 per cent of the surface denudation is due to mechanical action, and only 6.7 per cent to chemical processes (Balázs, 1971).

More usually, assessments of the effectiveness of mechanical erosion in the humid tropics have been based upon morphological evidence. Versey (1959, 1972), for example, asserts that the cockpit karst in Jamaica is caused mainly by surface collapse, which is accelerated by hydraulic and abrasive action in the underground drainage systems during floods. In the same area large amounts of poorly-sorted limestone debris are removed during storms (Day, 1976). Also, cliff-foot erosion in the tower karst of Belize and Sulawesi has been attributed, in part, to the mechanical action of ephemeral streams (McDonald, 1975, 1976a, 1976b). Since limestone dissolution is generally more rapid beneath deeper mineral soils where amounts of moisture and concentrations of carbon dioxide are maintained at high levels, the removal and redistribution of soils by mechanical erosion significantly affects the spatial pattern of chemical weathering. A self-perpetuating process may thus be envisaged whereby the floors of depressions are lowered more rapidly than adjacent slopes, which progressively lose their residual soils to the depressions. According to Jakucs (1977, p. 131), the development of pronounced micro- and macro-

relief forms in the humid tropics reflects the higher rates of soil erosion in these regions than elsewhere.

3.5. Soil properties and karst denudation

The properties of soils are recognised as being important factors in chemical weathering. In Madagascar, for example, soil carbon dioxide levels vary seasonally, being highest during the wetter part of the year, and this pattern is reflected in the solute content of karst groundwaters (Rossi, 1974, 1976). In Jamaica, marked contrasts have been observed in the carbon dioxide contents of different soils (Nicholson and Nicholson, 1969). Data from other published sources (eg. Day, 1976; McDonald, 1976a; Pitman, 1978b), while not encompassing completely the spatial, with-depth, and temporal variations present within the study areas, nevertheless provide some indication of the range of carbon dioxide levels extant in tropical karst soils. On present, though limited evidence, differences in the carbon dioxide contents of humid tropical and temperate soils are small. Other properties of tropical karst soils, though often documented in regional soil survey memoirs, are rarely reported in geomorphological literature.

3.6. Vegetation as a factor in karst denudation

Ramifying grooves on joint faces (Wilford and Wall, 1965, plate 18), pipe-like channels in bedrock (Jakucs, 1977, p. 137) and other micro-features provide visible evidence of the corrosional effect of plant roots in the humid tropics. The pronounced meso-relief of tropical karst terrain poses severe difficulties for investigations of vegetation dynamics, which require frequent sampling. Nonetheless, Aub (1964 - cited by Sweeting, 1972) was able to show that depressions in the Jamaican cockpit karst receive 14 per cent more rainfall than the hilltops. Further elucidation of the effects of vegetation upon hydrology in tropical karst terrain, and quantification of the dynamics of tropical karst ecosystems and their influence upon denudation processes, provide challenging fields.

4. The morpho-climatic debate

Recent research has provided much insight into the mechanisms and determinants of karst landform development. Nevertheless, the fundamental question of climatic control remains largely unresolved.

4.1. Climate and karst denudation processes

According to Jakucs (1977, p. 110-112), the intensity of karstification in the humid tropics, based on several, though unspecified, approaches, is eight times greater than in temperate climates, and twelve times greater than in periglacial and high-mountain regions. He attributes these differences mainly to the higher concentrations of carbon dioxide and organic acids in the soils of the former regions. In contrast, Corbel and Muxart (1970) maintain that karst waters in the humid tropics generally have lower hardness concentrations than elsewhere and that, additionally, the volumes of effective precipitation are often less, because rates of evapotranspirational loss are high. They note that a number of elementary life forms, such as mosses, which are important sources of carbon dioxide in other ecosystems are absent in the humid tropics. They also cite results from other studies which show that bacterial numbers in the soil are as high in Lapland as in much warmer regions, and further suggest that "les bactéries jouent un rôle plus faibles dans les zones tropicales modérement humides que dans les zones temperées humides et les zones arctiques humides". (ibid, p. 458, authors' italics).

Smith and Atkinson (1976), using published water hardness values show that the average calcium hardness concentration of river and spring waters increases from 83 ppm in arctic/alpine regions to 211 ppm in temperate environments, with an average of 174 ppm in the humid tropics. As they acknowledge, some of the data used are only approximations. Also, since widely different types

of site, with different proportions of allogenic catchment, are included within the three climatic groups, the effects of climate cannot be isolated completely. Nonetheless, solute levels recorded from arctic and alpine sites were notably lower than those from sites in temperate and humid tropical zones. In a more specific study in N. America and Mexico, limited to spring waters which were almost saturated with calcium carbonate, an overall positive correlation was found between mean temperature and mean hardness (Harmon et al., 1975). A related study by Drake and Wigley (1975) shows that this correlation is attributable to the greater concentrations of soil carbon dioxide in regions with higher mean annual temperatures.

An analysis of chemical erosion in different climatic zones shows that rates are highest in arctic and alpine regions, with a mean of 61.8 $m^3/km^2/yr$ (Smith and Atkinson, 1976). Values for the temperate and humid tropical regions are 56.9 and 45.5 $m^3/km^2/yr$, respectively. The high rates characteristic of the arctic and alpine regions are explained by the large volumes of flow, the latter being caused by the combined effects of heavy rainfall, low rates of evapotranspiration, and rapid runoff from bare and sparsely-covered rock surfaces. Douglas (1964) has demonstrated that discharge, rather than the concentration of dissolved solids, is the key factor which influences rates of chemical denudation. The above results lend support to Corbel and Muxart's hypothesis (1970). However the increase from the tropics to the arctic and alpine areas is relatively small, being only 36 per cent. In a more specific regional comparison, Balázs (1973a) has investigated chemical denudation rates of similar limestone formations in two different climatic zones. The study shows that 80 $m^3/km^2/yr$ of limestone is removed in solution from the Maros region, Celebes, whereas the corresponding figure for the Aggtelek Karst, Hungary

is 20 $m^3/km^2/yr$. This fourfold difference is caused entirely by differences in mean annual runoff, since dissolved limestone in spring and resurgence waters averages between 150 and 180 ppm in the former area, compared with a 250 to 350 ppm range in the latter.

Data on world-wide rates of mechanical denudation are few, and do not form a meaningful basis for inter-regional comparisons. However, Balázs (1973a) estimates mechanical erosion in Maros to be about 50 to 200 m^3/km^2, compared with 10 to 20 m^3/km^2 in Aggtelek, a five to tenfold difference.

4.2. Morphological considerations

The literature contains many general statements concerning morphological contrasts between different climatic zones. For example, limestone outcrops in the humid tropics are thought to become more massive as karstification proceeds, because of rapid secondary deposition, whereas those in colder climates become increasingly skeletal in form (Bögli, 1960; Jakucs, 1977, p. 107; Jennings, 1971). Also, the correlation of cockpit and tower karst with climatic factors has gained continued support, and further attempts have been made to determine climatic characteristics necessary for their development (eg. Balázs, 1962; Jakucs, 1977; Monroe, 1977; Sweeting, 1972).

Tropical karst is noted for its great diversity of forms (Jennings, 1971, p. 186; Verstappen, 1964a, 1964b). For instance, circular depressions resembling temperate dolines have been reported from New Guinea (Jennings and Bik, 1962), West Malaysia (Jennings, 1972), and several other regions. But equally, landforms similar to tower karst have been reported from extra-tropical regions (eg. Ford, 1975). For example, labyrinth and tower karsts from the Nahanni area, Northwest Territories, closely resemble their tropical equivalents (Brook and Ford, 1978). These observations cast further doubts into the morpho-climatic debate, and recently the role of

geological factors has been emphasized both in studies from the humid tropics and in world-wide comparisons (eg. Brook and Ford, 1977, 1978).

E. SIGNIFICANCE OF TROPICAL STUDIES IN KARST GEOMORPHOLOGY

Regions of lowland humid tropical karst are of considerable geomorphological importance for three reasons. First, whilst evidence suggests that climatic changes in the humid tropics during the Quaternary have been of greater magnitude than was previously thought (see review by Flenley, 1979), nonetheless, environmental changes over this period have had a less marked effect upon landform development in the humid tropical lowlands than in the present-day temperate and cold regions. In particular, fluvial processes have remained the dominant denudational agencies. The Malaysian landscape, for example, is thought to date back into the Tertiary period in terms of uninterrupted fluvial erosion, possibly under continuous humid tropical conditions (Jennings, 1971, p. 93; Morgan, 1976, p. 319). Humid tropical landscapes might therefore be regarded as 'normal', and those elsewhere as modifications of the 'normal' system (Douglas, 1969a, 1972). Also, because of their generally rugged terrain and thin soils, many limestone outcrops in the humid tropics remain almost undisturbed by human activity. Secondly, the erosional environment of the humid tropics is perhaps the most active and dynamic in the world, yet, because temperature variations are small, it is the least complex. Finally, humid tropical karst displays greater diversity of form than any other (Jennings, 1971, p. 186), and thus provides ideal field areas for examining a wide range of geomorphic controls.

F. OBJECTIVES OF THE PRESENT STUDY

The overall aims of the study are threefold: (1) to measure the

systematic, spatial and temporal variations of discharge, solute content and temperature of surface streams and groundwaters in the karst regions of Selangor, the Kinta Valley and the North-west region, West Malaysia; (2) to assess the rates and distribution of chemical erosion and solute movement within the karst outcrops; and (3) to interpret the measured variations in karst water quality and chemical budget in terms of the environmental characteristics of the limestone hills. The experimental design had three broad levels of enquiry. First, the meso-topography and soil and vegetation cover of 13 plots, which encompass the observed range of topographic units, were examined to provide an indication of the weathering environments and solute sources in the limestone hills (Chapter IV). Secondly, more specific process measurements were undertaken in selected plots to quantify the hydrology and dynamics of cation movement within the bedrock-soil-vegetation system. Thus, in 6 of the 13 plots, throughfall and litterfall were monitored over a one-year period to establish the rate of nutrient cycling and the scale of cation input at the surface of the outcrops (Chapter IV), and soil carbon dioxide was determined regularly to assess the potential degree of chemical aggressiveness derived from this source (Chapter VI). Additionally, the chemical properties of surface runoff from bare and thinly covered rock surfaces (Chapter V), of throughflow at the soil/rock interface (Chapter VI), and of subcutaneous flow (Chapter VI), were investigated in several plots to evaluate weathering rates and their controlling factors in the surface and near-surface zone. Thirdly, discharge, temperature and the solute contents of surface streams and groundwaters were determined at regular intervals at 156 sites, and single determinations were made at a further 168 locations. The sites investigated include deep groundwaters,

underground diffuse-flow seepages, springs, cave streams, surface streams and seepages in alluvium of the karst alluvial plains, these being the six principal types of karst water in the field areas. Net rates of chemical denudation are calculated from the total hardness data and estimated amounts of effective precipitation. Some indication of the distribution of limestone solution and deposition within the outcrops is gained by comparing the hardness contents of the various karst waters.

An outline of the physical geography of the field areas, based on a collation of existing information and basic data collected during the investigation, is presented in Chapter II, whilst the methods of sampling and techniques of laboratory and statistical analysis are discussed in Chapter III. Chapters IV to VII contain the main body of results, and Chapter VIII focuses upon several important general themes which may be evaluated in the broader context of an overall synthesis of the results.

CHAPTER II

PHYSICAL GEOGRAPHY OF THE STUDY AREAS

CHAPTER II. PHYSICAL GEOGRAPHY OF THE STUDY AREAS

A. LOCATION OF FIELD AREAS

Outcrops of limestone occur at scattered localities throughout the northern half of the Malay Peninsula (for details of distribution, see Hutchison, 1968, fig. 1, p. 46). Field areas were chosen so that the range of karst rock types, geological settings and climatic conditions were covered. Three karst regions, west of the Main Range (fig. 1A), were selected for detailed investigation:

1. Selangor, north of Kuala Lumpur (fig. 1B)
2. Kinta Valley, Perak (fig. 1C)
3. North-west region, ie. Perlis and northern Kedah (fig. 2).

B. CLIMATE

This section describes the climate of the field areas, and examines the weather conditions during the sampling period. First, the network of meteorological stations used in the study is discussed. These stations are identified by a four-figure code (tables 1 to 3), and their locations are shown in figures 3 to 5.

1. Sources of meteorological data

Meteorological stations are maintained principally by the Drainage and Irrigation Department (D.I.D.) and Malaysian Meteorological Service (M.M.S.). The density of rain gauges is fairly high to the west of the Main Range (Lockwood, 1967). For instance, the network maintained by D.I.D. in Selangor comprises one 20.3 cm gauge per 59.5 km^2 (Chia, 1968). In each field area the earliest rainfall records are pre-1910. By contrast, temperature, hours of bright sunshine, relative humidity and other climatic variables are recorded at only a few locations. The Kinta Valley, for example, has only one such station, and there are none in the North-west region, the nearest being 35 km to the south at Alor Star.

1.1. Reliability of rainfall records

Operator variance will almost certainly occur in the meteorological data, and several examples are cited by Lim (1969, p. 15 to 16, p. 69 to 70). Chia (1973) estimates that only 9.8 per cent of the stations in West Malaysia are 'completely reliable', and that only 39 to 45 per cent of the stations in the three field areas may be regarded as 'reliable or better'. Using different criteria (table 4) the records of 38 per cent of the 95 stations maintained by the D.I.D. in Selangor have been found to be of 'dubious reliability' (Leigh and Low, 1973).

Because much rainfall in the tropics occurs as isolated convectional storms, often less than 10 km in diameter (Watts, 1955), the problem of sampling is acute. Variability of rainfall may be considerable, even over small areas (Jackson, 1969; Nwa, 1977). For example, at three stations less than 10 km apart in Johore, each with similar average monthly and annual figures, correlations of rainfall for individual months are poor (Jackson, 1978). Clearly, differences in daily rainfall between adjacent stations will be even more marked.

1.2. Rainfall stations utilized in the study

Reliable daily rainfall records were essential, particularly in the following areas where temporal variations in karst water quality and soil and vegetation characteristics were investigated:

1. Bt. Batu, Selangor
2. Ak. Bt. Takun, Selangor (Ak. Bt., anak bukit - hillock)
3. G. Tempurong, Kinta Valley
4. Wang Tangga, North-west region.

When fieldwork began in July 1974, the D.I.D. operated gauges at Batu Caves Estate (1601), 0.5 km from Bt. Batu, and at Kaki Bukit Police Station (4601), within 1 km of Wang Tangga. The

first of these was closed on 31.10.74, and records from Kepong (1602), an adjacent station, were used for the remainder of the sampling period (see below). Weekly recording gauges ('Kent', siphon type) were installed at the other two sites. Security from disturbance was the prime constraint in siting these gauges. One was installed on the corrugated iron roof of a warden's hut in Templer Park (plate 1), approximately 1 km from Ak. Bt. Takun. The other was set up at Siput Tin Mines Ltd. (plate 2), Sg. Siput South (Sg., sungei - river). This is located less than 500 m from the sampling plots on G. Tempurong and approximately 1 km from Gua Tempurong (Chapter VII). These four stations formed the primary gauge network used in this study.

A number of D.I.D. and M.M.S. stations (table 1) were used to supplement the four key gauges when gaps occurred in their records, and to estimate rainfall over broader areas. As it seemed likely that storm patterns would differ between the north-east and south-west monsoons (Dale, 1960), these two periods were treated separately in the following analysis. Pearson correlation coefficients ($\underline{r}$) for daily rainfall at pairs of stations in each region are shown in tables 5 to 7. As anticipated above, the coefficients are generally low, and differences occur between the two monsoons. For example, during the approximate period of the north-east monsoon (table 5A), rainfall at Templer Park (2601) correlates most strongly with that at Kepong (1602), $\underline{r}$=0.551, whilst during the south-west monsoon (table 5B) it correlates most closely with that at Kg Sg. Tua (1604), $\underline{r}$=0.714.

Spatial extrapolation of rainfall data is commonly made by Theissen polygon and isohyetal methods. However, because of the highly localized nature of most storms, and the fact that during both monsoons single stations stood out as having rainfall most closely approximating that of the key stations, gaps in rainfall

amounts occur between pairs of closely-correlated stations, daily data has been adjusted in the same proportion. For example, rainfall at Sg. Siput South (3601) correlates most strongly with that of Kampar (3606), but total amounts average only 89 per cent of the latter. Rainfall records from Kepong (1602), 4 km west of Bt. Batu, have been used after closure of the gauge at Batu Caves Estate. Also, data from Kepong was substituted at Templer Park (2601) between 12.11.74 and 11.12.74 when the autographic gauge was moved temporarily to the North-west region.

2. Average climatic characteristics

West Malaysia lies within 7° latitude of the equator and consequently temperatures are uniformly high throughout the year. The peninsula comes under the influence of two major wind systems, the south-west monsoon from the Indian Ocean, and the north-east monsoon, from the South China Sea. Both air masses are warm, moist and unstable (Dale, 1959).

According to the Köppen classification system (Köppen, 1936) the climate is of class 'A' type (ie. 'tropical rainy climate', with mean temperature of coldest month 18°C). However, rainfall regimes vary significantly within the country. To the west of the Main Range, in the vicinity of Alor Star (6°N), an important climatic boundary distinguishes the North-west region from the more southerly limestone outcrops (table 8). In the former area the climate is of type 'Am' (ie. 'monsoon type forest climate', rainfall 60 mm in driest month, but total annual rainfall is sufficient to compensate for seasonal drought) whilst in the latter areas it is of type 'Af' (ie. 'hot damp forest climate', mean annual rainfall of driest month 60 mm). This marked contrast between the northernmost field area and the Kinta Valley and Selangor regions, with their similar climates (table 8), is reflected in several climatic regionalizations (Dale, 1959; Morgan, 1971;

Whitmore, 1975, p. 45).

2.1. Average monthly and annual rainfall

2.1.i. Selangor. Annual rainfall for four stations adjacent to the main sampling sites, Bt. Batu and Ak. Bt. Takun, averages 2495 mm (table 9). Considerable variability may be observed between the stations with average values ranging from 2352 mm (General Hospital, Kuala Lumpur, 1612) to 2738 mm (Rawang Reservoir, 1605).

Selangor lies within the western rainfall region recognized by Dale (1959). Rain is fairly evenly distributed throughout the year, with peaks in April and October/November (fig. 6; table 9), the inter-monsoonal periods (Chia, 1968). July is the driest month, and soil moisture deficit may occur at this time (Nieuwolt, 1965).

2.1.ii. Kinta Valley. Rainfall totals in the Kinta Valley (table 9) are slightly higher than in the Selangor karst region, averaging 2847 mm. The south-eastern part of the valley is wettest (Kampar, 3663 mm; Gopeng, 2923 mm). Rainfall at Sg. Siput South (3601) was 11 per cent less than at Kampar over the study period, giving an estimated mean annual total of 3260 mm. Amounts decrease westwards across the valley (Batu Gajah, 2528 mm) and also to the north (Ipoh, 2607 mm; Tg. Rambutan, 2514 mm). Like Selangor, the Kinta Valley lies within the western rainfall region and has a similar monthly pattern (fig. 6). Dry spells are less frequent in these areas than further north, but nonetheless, periods of 14 consecutive dry days, recur, on average, every two years.

2.1.iii. North-west region. Rainfall in the North-west region averages 1960 mm per annum (table 9). In general, amounts are lowest in the north-eastern part of the region (Padang Besar, 1704 mm). Rainfall at the two stations along the front of the Boundary Range for which long-term records are available (4601 and 4604)

averages 2089 mm. The latter figure, which is notably less than that of the other two karst regions, is assumed to represent the mean annual rainfall of the Boundary Range.

The North-west region has a more marked seasonal incidance of precipitation than the other field areas. Rainfall maxima, corresponding with the inter-monsoonal phases, occur in May and October, but the presence of a very pronounced rainfall minimum from December to March is of greater significance (fig. 6). At Kangar, for instance, during the period from 1909 to 1958, no rain was recorded in January on five occasions, and in February on four. Also, spells of 21 consecutive dry days occurred on 16 occasions, and of 14 days on 46 occasions.

2.2. Characteristics of single storms

In all three areas much of the rainfall occurs as high intensity convectional storms of short duration. Storms greater than 25.0 mm are common (table 10), whilst those exceeding 100.0 mm recur on average every two to three years. Over periods of up to an hour's duration, intensities may be exceptionally high, but these levels are not sustained over longer periods. At Alor Star, for instance, as much as 90 mm of rain has been recorded in an hour, but over 24 and 96 hours the maximum known precipitation is 152 and 243 mm respectively, or 6.33 and 2.53 mm per hour. The return periods for precipitation events of given magnitude have been determined for selected stations in West Malaysia (Lim, 1969; Lockwood, 1967). The probable maximum 24-hour precipitation (table 11) ranges from 422 mm (Selangor) to 681 mm (North-west region).

2.3. Relative humidity and potential evapotranspiration

The annual mean relative humidity (table 12) ranges from 82.0 (North-west region) to84.6 per cent (Selangor). In all three regions monthly variations in relative humidity correlate directly with rainfall (table 13) and inversely with sunshine hours

(fig. 7). Everywhere, diurnal contrasts are marked, with humidity levels rising from between 55 and 70 per cent by day to more than 95 per cent at night. Monthly evaporation data (table 14) provide the best measure of potential evapotranspiration in the field areas. Although the original data (after Nieuwolt, 1965) are based on evaporation pan measurements and are therefore suspect because of the 'oasis effect' (Lockwood, 1971, p. 7), alternative published sources are lacking. Annual evaporation increases northwards from 1430 mm in Selangor to 1646 mm in North-west region. The seasonal patterns correspond closely with the number of hours bright sunshine (fig. 7).

2.4. Temperature and insolation characteristics

2.4.i. Mean annual and monthly temperatures. Mean annual temperatures (table 15) range from 26.3 (Selangor) to 26.9°C (North-west region). The monthly temperatures are highest from April to June, and lowest from November to January.

2.4.ii. Sunshine hours. The number of sunshine hours per day (table 16) increases northwards from 6.2 in Selangor to 7.0 in the North-west region. Seasonal variations are more pronounced in the latter area, where values range from 5.6 hours per day in November to 8.5 hours in February and March. Sunshine in all three regions varies inversely with rainfall (table 13).

2.4.iii. Diurnal temperature fluctuations. Daily temperature fluctuations (table 17), though of greater amplitude than seasonal ones, are moderated by the insulating effect of the perpetually moist atmosphere. The annual mean diurnal range varies between 8.8 (North-west region) and 10.4°C (Kinta Valley).

2.4.iv. Soil temperature. Soil temperature was monitored regularly at between 13.00 and 14.00 hrs. on Ak. Bt. Takun (table 18). Mean temperatures vary little with depth, ranging from 24.2 (15.2 cm) to 24.4°C (7.6 cm). These values are approximately 2.0°C less than

the mean annual air temperature of the region. Temporal variations are small and decrease with depth. At 7.6 cm, for example, the observed range is 1.4°C (s. dev., 0.54°C), while the corresponding figure for 30.1 cm is 1.0°C (s. dev., 0.35°C). Although no data were obtained from the Kinta Valley and North-west region, there is no reason to believe that soil temperatures in these areas will differ substantially from those reported above.

3. Rainfall during the sampling period (1.8.74 - 31.7.75)

Rainfall was the only climatic variable for which meaningful and reliable daily records were available. However, since sunshine hours and relative humidities are strongly correlated with rainfall (Section 2, above), anomalies in the latter will affect the temppral pattern of other climatic characteristics. To establish how the weather during the sampling period compared with that of previous years, data for the sampling period were tested against mean annual figures calculated from pre-sampling period data. In each field area, the station closest to the main group(s) of study sites has been selected for detailed examination. In all three regions the annual totals for 1974/5 lie well within the 95% confidence limits calculated from the standard deviation of the long-term records (tables 19 to 21).

During the sampling period a major rainfall anomaly occurred on the north-east coast of the Malay Peninsula. October and November are usually months of heavy and prolonged rain in this area, but in 1974 only short spells of rain were recorded. According to a Meteorological Office spokesman, the full impact of the north-east monsoon was intercepted by cyclones moving across the South China Sea and the Philippines. This resulted in heavy rain over the latter areas, but left the north-east coast comparatively rain-free (Straits Times, December, 1974). These extraordinary circumstances may also have affected rainfall in the

study areas. In each region statistically significant differences (chi-square test, $p<0.001$) were found between the monthly rainfall pattern in 1974/75 and the long-term average figures.

3.1. Selangor - Kepong (1602)

Rainfall during the sampling period totalled 2318 mm (table 19), which is 5.0 per cent below average. In five months of the year, monthly amounts differed by more than 100 mm from the calculated average.

Rainfall in August 1974 was exceptionally low, and a period of 23 successive dry days was recorded. The monthly total of 71 mm was 80 mm below the mean. October was the driest month with a total of 51 mm, this being 53 mm lower than the minimum October figure over the 1913-58 period and 230 mm below the monthly average. There were 25 dry days in the month as compared with 15 in the pre-sampling period, and no storms of more than 25.0 mm were recorded (table 19). The highest monthly total, 340 mm (December, 1974), included a 24-hour maximum of 72 mm on the 10th. The highest 96-hour rainfall of the study period (154 mm), fell between the 7th and 10th December.

January 1975 was drier than average. In contrast, rainfall in February exceeded the monthly average by 104 mm. During this month there were only 10 dry days as compared with 20 in the pre-sampling period, and 130 mm of precipitation fell between the 10th and 13th. March was 68 mm drier than average, and April 47 mm wetter. The highest 24-hour maximum of the sampling period (81 mm) fell on the 3rd April. Rainfall in May was 101 mm below average, and no storms of greater than 25.0 mm were recorded. June was also drier than usual, whereas July, which is usually the driest month, was exceptionally wet. The monthly total of 282 mm was only 14 mm less than the maximum recorded over the 1913-58 period. Rain fell on 21 days, as compared with 8 in the 1970-73 period.

3.2. Kinta Valley - Kampar (3606)

The sampling period was 12.7 per cent drier than average, with precipitation totalling 3198 mm (table 20). August, with 117 mm, was considerably drier than average. In contrast, September was exceptionally wet. The monthly total (557 mm) exceeded the maximum recorded over the 1897-1958 period by 56 mm, and was 301 mm above average. 91 mm fell on the 2nd and over the 96-hour period between the 2nd and 4th, rainfall totalled 216 mm. October (103 mm) and November (46 mm) were exceptionally dry, their combined total being 657 mm below average. Rainfall in both months was less than the 1897-1958 minima. During the remainder of the sampling period monthly totals were mostly within 100 mm of the average figures. However, rainfall in July (251 mm) was exceptionally high, and included 118 mm on the 29th.

3.3. North-west region - Kaki Bukit (4601)

Rainfall during the study period totalled 2082 mm (table 21), 11.9 per cent above average. Anomalies were again observed in the monthly pattern. During the August-December period, which is typically the wettest part of the year, rainfall was 201 mm below average, and only 8 storms exceeded 25.0 mm compared with 16 in the pre-sampling period. In contrast, January, which is usually the driest month of the year, was the wettest. The monthly total of 314 mm was 259 mm above the average, and greater than the maximum of the 1946-58 records. Five storms exceeded 25.0 mm in depth, and two were of 76 mm. Between the 14th and 17th rainfall totalled 138 mm. The heavy rainfall of January 1975 was confined mostly to the northern part of the field area. 303 mm occurred at Padang Besar and 241 mm at Tasoh, but at Abi Head Works and Kangar, less than 12 km south of Kaki Bukit, rainfall was 92 and 79 mm, respectively.

February was the driest month with only 34 mm precipit-

ation. No rain was recorded between 23rd January and 26th February, giving a total of 34 successive dry days. The remaining months were wetter than average by between 17 and 55 mm.

3.4. Discussion

In each area, total rainfall over the sampling period differed little from the long-term mean, and therefore the results of the investigation may be taken as being broadly representative of average conditions. Possible effects of the rather anomalous monthly rainfall patterns upon ecological, pedological and denudational processes are considered in the appropriate chapters. Exceptional wet and dry periods occurred at each sampling site. Measurements taken under such extreme conditions are often valuable in elucidating ecological and geomorphological controls, and are examined closely in the following analysis.

4. Chemical characteristics of rainwater

Samples of rainwater were collected on several occasions during the study period. The results of this work are summarized in table 22. Individual properties display considerable variability within each region. In Selangor, for example, the observed pH range is 4.80 to 6.65, and calcium varies from 0.40 to 3.40 ppm. Average levels of specific conductance, sodium and potassium are very similar in all three field areas. In contrast, pH, calcium and magnesium differ quite markedly. pH is lowest in the Kinta Valley (mean, 5.31) and highest in the North-west region (mean, 6.02), whilst the average concentrations of calcium and magnesium in Selangor (1.48 and 0.14 ppm, respectively), are approximately three times higher than corresponding figures from the Kinta Valley. Overall, the calcium content of rainwater is higher in the karst regions than elsewhere in the Peninsula, whereas the sodium content is appreciably lower (cf. table 23). The higher calcium concentrations in the karst areas may be attributable to air-borne dust, origin-

ating from the quarrying of limestone. Explanation of the differences in sodium concentrations is more problematic, since distance from the sea is similar in each case. In view of the small numbers of samples on which these data are based, the above results should be regarded only as approximations. They do, nevertheless, provide an indication of the rate of cation input from the atmosphere (table 24), against which other components of solute movement within the karst regions may be compared. Inputs of calcium are greatest, ranging from 11.38 to 33.06 kg/ha/yr during the study period. Values of less than 8.0 kg/ha/yr were recorded for magnesium, potassium and sodium.

C. GEOLOGY

This section examines firstly those aspects of the geological evolution of the karst regions which are important to an understanding of the present landscape, and secondly the physical and chemical characteristics of the limestones. A summary of the chronology is given in figure 8.

1. Geological evolution of the karst regions

The Malay Peninsula forms part of a much larger structural unit, a cratonised block known as 'Chersonesia' (Gobbett, 1973a), which has evolved since Precambrian times through a series of stages related to the shifting margins of several tectonic plates (Hutchison, 1973a; see also earlier accounts by van Bemmelen, 1949, and Holmes, 1965, pp. 1138-44). Presently, it is bounded to the west and south by the Indian Ocean/Australian plate, subducting east and north, and to the east by the Pacific Ocean plate, subducting to the west.

1.1. Lower Palaeozoic

During the Lower Palaeozoic, the western half of the Peninsula was occupied by an eastwards migrating geosynclinal seq-

uence, subducting beneath a Precambrian landmass to the west (Jones, 1973). At this time, deeper basin environments alternated with stable shelf conditions, In the Cambrian period, the latter were associated with deltaic deposits, such as those of the Machinchang Formation, but deposition of limestone became dominant as sediment supply diminished during the Ordovician and Silurian. In the North-west region limestones of the Setul Formation lie conformably over rocks of the Machinchang Formation. In the Kinta Valley and Selangor, limestones of late Ordovician and early Silurian age, respectively, form the oldest outcrops. In the former region, limestone deposition was continuous from Silurian to mid-Permian times.

1.2. Upper Palaeozoic

The westwards subducting trench migrated eastwards during the Upper Palaeozoic and by the end of the Carboniferous period it lay well clear of the Malay Peninsula (Hutchison, 1973a). At some stage, possibly as early as the Carboniferous period, a second trench system developed to the west, migrating westwards and subducting eastwards beneath the Peninsula. In the North-west region and Selangor, major breaks in sedimentation occurred towards the end of the Devonian period, resulting from uplift, folding and possible minor granitic intrusion. In these regions the limestones are unconformably overlain by clastic sediments of the Singa and Kenny Hill formations respectively. In the North-west region, the Singa formation was replaced in mid-Permian times by a shallow-water organic limestone facies, the Chuping Formation. Deposition of limestone in the North-west and the Kinta Valley was terminated by tectonic activity in the Upper Permian/early Triassic (Gobbett, 1973b).

1.3. Mesozoic

Locally, in Kedah, shelf conditions returned in the

Middle and early Upper Triassic, leading to deposition of the Kodiang limestone. More importantly, the late Triassic was a period of massive uplift and intrusive activity. At this time, the granites of the Main Range were emplaced, and the country rocks were subjected to varying degrees of metamorphism. This marked the end of the depositional stage in the Malay Peninsula, and the region has been above sea level ever since. Palaeomagnetic evidence suggests that the peninsula already lay at a latitude of about 15°N in the period between the late Palaeozoic and Cretaceous (Haile et al., 1977; McElhinny et al., 1974). Consequently, the karst regions under consideration have been at low latitudes ever since their initial uplift. This finding is consistent with the long period of uninterrupted landform development postulated by Jennings (1971) and Morgan (1976).

Considerable structural deformations, which occurred in the Jurassic and Cretaceous periods, have considerably affected the geomorphic evolution of the karst regions. East-west compressive stresses produced a series of broad north-south trending corrugations in the Setul and Chuping limestones of the North-west region, and faults with north-westerly strike in the Kinta Valley (Gobbett, 1971). The latter structures may be traced from the Kinta Valley into the adjacent granites of the Kledang and Main Ranges. The north-easterly trending junction between the Kledang granite and Kinta limestone is thought to be a major fault line, which has been subsequently cut and offset by the minor north-westerly wrench faults. According to Gobbett (1971), the Kledang Range may thus be regarded as a granite horst, and, if suggestions of a similar fault line at the junction with the Main Range are substantiated, then the Kinta Valley itself may be interpreted as a graben feature.

1.4. Cenozoic

In the mid-Tertiary period major wrench faulting affected much of the Malay Peninsula (for detailed discussion, see Holcombe, 1977), and produced a complex set of second or third order joint and fault patterns in the Kinta Valley (Gobbett and Tjia, 1973). In general, however, the Peninsula is thought to have remained essentially tectonically stable throughout the Cenozoic, with activity being confined to epeirogenic uplift and tilting of limited magnitude and areal extent (Stauffer, 1973). Sea level changes during this period have affected landform development in the lowland karst regions. The origin of these changes, whether eustatic or of more local tectonic origin, and their absolute chronology, is uncertain.

1.4.1. Morphological evidence of former sea levels. The morphology of that part of Chersonesia, known as the Sunda Shelf, which is presently below sea level (Molengraaf, 1921), and, in particular, the sediments of the Malacca Strait (Keller and Richards, 1967), which forms part of the Shelf, suggest the existence of a former peneplain which was drowned by the eustatic rise in sea level at the end of the last glacial episode. Confirmation of this is seen in the fact that the sub-alluvial karst floor of the Kinta Valley is graded to a height of -122 m at the present estuary. All altitudes given relate to heights above mean sea level. Cliff-foot notches may be observed at depths of several metres below the present alluvial plain at the junction of the sub-alluvial karst surface and the limestone hills (plate 21; Crowther, 1978b). Also, in the archipelago of Pulau Langkawi, where contemporary wave-cut platforms are absent or, at most, poorly developed, limestone cliffs pass precipitously below sea level (Hodgkin, 1970). In a review of the available evidence, Haile (1971) concludes that sea levels of 122 m, or as low as 168 m, below the present level occurred during the Pleistocene.

Evidence of former high sea levels is less substantive. Several workers have postulated high sea levels at 76 m. Benches in the limestones and adjacent granites of the Kinta Valley (Walker, 1956), pronounced breaks of slope in Wang Tangga, Perlis, and cave levels in the Chuping Limestone (Jones, in press) have been reported at this height. However, in a more recent investigation of the heights of cave floors and cliff notches in the limestone hills of central Peninsular Malaysia, Wycherley (1975) finds that whilst certain levels, notably between 27 and 61 m are more frequent, every outcrop displays some unique features (plate 22). He concludes that "there is no convincing evidence of any widespread body of water such as the sea being responsible for their simultaneous country-wide formation" (ibid., p. 44). Haile (1971) also found no unequivocal evidence of sea levels at more than 6 m above the present level.

1.4.ii. The 'Old Alluvium'. The Old Alluvium is an extensive deposit of "grey to brown sandy clay, usually with frequent horizontal and gravel laminae" (Walker, 1956, p. 25; plates 3 and 41). It occurs widely to the west of the Main Range, and, locally, reaches a height of 70 m (Burton, 1964; Walker, 1956). More generally, however, it has been eroded to heights of between 30 and 60 m. The formation of the Old Alluvium has been the subject of much debate (Scrivenor, 1912; Stauffer, 1973; Walker, 1956). The results of recent stratigraphical and sedimentological investigations (Newell, 1971; Stauffer, 1973, p. 162; Sivam, 1969), and the total absence of marine fossils (Mohammed Ayob, 1970), clearly indicate a fluvial origin. Carbon-14 determinations for Selangor (Haile and Mohammed Ayob, 1968) and the Kinta Valley (Sivam, 1969) have yielded dates beyond the range of the method (ie. older than 39,900 years B.P.). Available stratigraphic and fossil evidence suggests that the morphology and structure of the Old Alluvium is

related to early- or mid-Pleistocene sea level fluctuations (Stauffer, 1973).

1.4.iii. Raised beach deposits. Raised beaches occur along most of the west coast of the peninsula, and are usually interpreted as products of Holocene sea level fluctuations (Tjia, 1970, 1973). On the coast adjacent to the Kinta Valley, for example, beach ridges have been identified at heights of 3 and 6 m (Courtier, 1962; Koopmans, 1964). Also, along the coast of the North-west region, Alexander (reported by Tjia, 1973) identifies a marine terrace at 3 m, dated as 5,200 200 B.P.

1.4.iv. Interpretation of sea level changes. Of published data, those from the Huon Peninsula, New Guinea (Chappell, 1974; Bloom et al., 1974) perhaps afford the best approximation of late Quaternary sea level changes in the Malay Peninsula. Here, there have been as many as 20 sea level maxima over the past 400,000 years, the highest probably being 6 m above the present level (approx. 125,000 years B.P.). The lowest sea level over the last 150,000 years is estimated at between -120 and -135 m (approximately 17,000 years B.P.). These levels appear to correlate closely with those recently postulated for West Malaysia. Clearly, if the level in the last interglacial (stage 5e) was higher than at any other time in the Quaternary (Shackleton and Opdyke, 1973), then the high sea level previously invoked (Walker, 1956), if substantiated, must be of local tectonic, rather than eustatic, origin.

2. Geological disposition and physical and chemical characteristics of the major limestones

A total of 91 rock samples were analysed in the present investigation. These analyses, and those previously undertaken by Hutchison (1966) and other workers are summarized in tables 25 to 28 and figure 9. The limestones are notably pure, with more than 90 per cent of the rocks analysed having acid insoluble residue

contents of less than 5.0 per cent (fig. 9A). Secondary deposits in caves are even purer (table 27), the mean acid insoluble residue content being 0.89 per cent (cf. 3.50 per cent in bedrock), and contain much smaller amounts of magnesium, the molar Mg:Ca ratio averaging 0.008 (cf. 0.230 in bedrock).

2.1. Selangor

The Kuala Lumpur limestone series, approximately 1830 m in stratigraphic thickness, is composed almost entirely of carbonates, and lies conformably over Hawthornden and Dindings Schists. The limestones were recrystallized by regional metamorphism and are finely-crystalline marbles of low porosity and high mechanical strength. Locally, adjacent to the granite, contact metamorphism has also occurred (Hutchison, 1973b; Gobbett, 1964). The original sedimentary structures of the limestones have been mostly obliterated by marmorisation, and the resultant marble commonly exhibits flow folding (plate 6). Generally, the limestone is grey/cream in colour and very pure, the amount of acid insoluble residue averaging 1.84 per cent (table 28A; fig. 9B). Whilst calcite forms the major mineral, small amounts of magnesium are frequently present and, locally, dolomite is dominant. The molar Mg:Ca ratio ranges from 0.002 to 1.023 (mean, 0.220). Samples with values greater than 0.548 are of dolomitic limestone, calcite dolomite or dolomite (ratio calculated from Chilingar et al., 1967). Of the samples analysed, 21 per cent fall in this category. In common with limestones of the other formations, potassium is barely detectable in the Kuala Lumpur Limestone. The sodium content averages 0.033 per cent by weight.

2.2. Kinta Valley

The stratigraphic thickness of the Kinta Limestone exceeds 3000 m (for detailed stratigraphy, see Suntharalingham, 1968). The limestones are interbedded with several argillaceous facies which,

locally, exceed 150 m in thickness. Both rocks have been recrystallized. The grain size of the marbles decreases regularly away from the granites of the Main and Kledang Ranges (Ingham and Bradford, 1960). Generally, the Kinta Limestones are mechanically strong and very pure, averaging 1.55 per cent acid insoluble residue (fig. 9C). They are composed mainly of calcite, the average Mg:Ca ratio being 0.325. However, one-third of the samples analysed were dolomitic. The sodium content is similar to that of the Kuala Lumpur Limestone, averaging 0.021 per cent by weight.

2.3. North-west region

The karst outcrops comprise limestones of three different ages (table 25). In contrast with those described above, the limestones of the North-west region were not buried at sufficient depth nor heated to the high temperatures necessary to induce regional metamorphism. On the mainland, all three limestones remained in the epizone. They were, however, subjected to thermal metamorphism and exhibit varying degrees of recrystallization. Offshore, in the archipelago of Pulau Langkawi, the limestones of the Setul and Chuping Formations were intensively deformed and recrystallized in the mesozone (Hutchison, 1973b).

2.3.1. Setul Formation. The limestones of the Setul Formation are mainly disposed on the eastwards dipping limb of a north-south trending anticline (fig. 10, from Gobbett, 1973b). Locally, where metamorphism has been slight, the original bedding and occasional bands of argillaceous and siliceous impurities are well preserved. The limestones are thickly bedded and generally dark-grey in colour due to the presence of considerable detrital and carbonaceous impurity (Jones, 1965). The amount of acid insoluble residue is usually high, averaging 9.89 per cent (fig. 9D) and this distinguishes these limestones from others in the Peninsula (Hutchison, 1968). Towards the anticlinal core the limestone is replaced by a

calcareous phyllite. The latter, which comprises lens-shaped nodules of carbonate enclosed by laminae of brown argillaceous material (Jones, in press), is exposed in the Gelang Kerbau Valley, near Abi Head water works. In this vicinity the limestones are less pure than in other parts of the Boundary Range. Of the 37 samples analysed, only two were dolomitic. The sodium content of the Setul limestones averages 0.063 per cent (range, 0.01 to 0.25 per cent), which is notably higher than that of the other limestones.

2.3.ii. Chuping Formation. Limestones of the Chuping Formation outcrop along the line of a north-south trending syncline. In the north, as at Mata Ayer (fig. 10) the syncline is simple, whilst further south it is split along its major axis by an anticlinal flexure. The latter serves to divide the outcrop into two distinct parallel bands (fig. 2), separated by shale and quartzite (Jones, in press). The Chuping is thinner than the Setul Formation, comprising around 600 m of strata, as compared with 1547 m. It is a light-grey, finely-crystalline rock with generally low levels of impurity (fig. 9E).

2.3.iii. Kodiang limestone. The Kodiang limestone is of very limited occurrence. It is fairly pure, with an average of 2.81 per cent acid insoluble residue. The molar Mg:Ca ratio averages 0.208.

D. THE KARST LANDSCAPE

Two aspects of the karst landscape are considered: surface topography and caves.

1. Surface topography

1.1. Morphological features of the limestone hills

In a pilot survey of the field areas four main morphological units were identified: hilltops, rock cliffs, soil-mantled slopes and alluvial flats. The spur at the southern end of G. Tempurong (plate 23) displays three of these units.

1.1.i. Hilltops. The tops of the karst outcrops take two main forms:

1. Rocky surfaces which are deeply dissected along solution widened joint planes. Intervening blocks of more massive limestone are highly-pitted, sharply-etched and often pinnacle-like in form. Locally, the pinnacles exceed 20 m in height (plates 13, 14 and 30).

2. Small enclosed depressions. These occur mainly at joint intersections and are filled with residual soils.

1.1.ii. Rock cliffs. Limestone cliffs occur frequently in all three regions, They are generally straight in plan, often vertical or overhanging, and locally exceed 100 m in height (plates 4, 11 and 23 to 26). The water they receive is derived from condensation, runoff from exposed cliff tops, and seepages from joints which intersect the cliff face. Such surfaces are usually coated with a thin patina of powdery calcite and adorned by stalactite forms which hang from projecting ledges (Crowther, 1978b; plates 7 and 24 to 26).

1.1.iii. Soil-mantled slopes. Two broad groups may be recognized:

1. Steep slopes (45 to 60°). These occur either on the flanks of the hilltops and ridges or immediately above cliffs. Typically, they have very thin soil covers and frequent rock outcrops (plate 23).

2. Moderately-inclined slopes (30 to 45°). In places these are continuous with the steep slopes which fringe the hilltops, and extend over several hundred metres down to the surrounding plains. Generally the cover of mineral soil is more complete than on the steeper slopes and soil depth increases downslope. The junction with the plains is often abrupt, though "short footslopes at 5-15°" may be present (Jennings, 1976, p. 94).

1.1.iv. Alluvial flats. These are widespread and take three main forms:

1. Contemporary flood plains (plates 19 and 39)

2. Residual surfaces of the Old Alluvium

3. Alluvial floors of large enclosed depressions, known locally as 'wangs', within the limestone outcrops (plate 48).

With few exceptions, such areas have been greatly disturbed by tin-mining activities.

1.2. Regional descriptions

1.2.i. Selangor. Natural exposures of the Kuala Lumpur Limestone are restricted to three localities: Bt. Batu, Bt. Takun and Ak. Bt. Takun (fig. 1B). Of these, Bt. Batu is the best known and forms a prominent landmark 10 km north of Kuala Lumpur (plates 3 to 7). It is an isolated tower, covering 1.3 km^2, which rises abruptly to an altitude of 304 m from the broad alluvial plains of the Sg. Batu and Sg. Gombak, at 53 to 61 m. In common with other large towers, the surface of Bt. Batu is broken by enclosed depressions and narrow, steep-walled corridors. In places, open shafts at the base of surface depressions connect directly with underlying caverns (plate 9). More than two-thirds of the undisturbed margin of Bt. Batu comprises tower walls (Jennings, 1976). Of these sections, approximately one-sixth have cliff-foot caves, which penetrate in some instances to a depth of 15 m. The latter features may be formed by sub-soil solution processes (Jennings, 1976), but elsewhere in the Peninsula, swamp or stream water seems a more likely agent (plates 22 and 34).

Bt. Takun and Ak. Bt. Takun (plates 11 to 14) lie 8 km further north in the headwaters of the Sg. Kanching. The former, the highest of the two outcrops, is bounded on three sides by towering cliffs. The latter appears to be at a more advanced stage of erosion. Its sides, which mostly comprise accumulations of angular limestone blocks derived from the free faces of higher cliffs, have an almost continuous cover of mineral soil.

1.2.ii. Kinta Valley. Outcrops of Kinta Limestone number more

than fifty in Perak and cover almost 50 km^2 (plates 19 to 31). Most occur as groups of towers along the eastern flank of the Kinta Valley, where the G. Gajah-Tempurong Massif, 612 m in height, forms the largest single outcrop. The latter has been described more fully elsewhere (Crowther, 1978b). Aerial photographs reveal clear, north-westerly tectonic lineations in the form of the hills (Gobbett, 1971) and individual towers show marked structural control (plate 28). Enclosed amphitheatre-like depressions, which in some cases exceed 1 km in length along their major axis, occur frequently within the towers. The limestones forming the outcrops are continuous beneath the adjacent plains and are frequently exposed at the bottom of tin-mine workings (plates 42 and 43). The sub-alluvial karst surface is deeply dissected along major joints. Locally, in isolated pockets, the alluvial fill is more than 20 m deep (Ingham and Bradford, 1960, fig. 8, p. 79).

1.2.iii. North-west region

1. Tower karst. The Kodiang limestone and limestones of the Chuping Formation give rise to tower karst forms which are similar to those described above. Most of the hill margins are steeply cliffed. The Kodiang limestone is exposed as five towers along a narrow belt in northern Kedah (fig. 2) and reaches a maximum height of 178 m. Outcrops of Chuping limestone are more numerous (fig. 2; plate 44). In central Perlis they occur principally along the lines of two parallel synclinal troughs. Mata Ayer (511 m) forms the highest summit. Many of the hills are bounded by low coastal plains, mostly less than 5 m above sea level. The base of the seaward side of Bt. Kodiang (plate 45), and of several other towers, has notches identical in form to those of contemporary wave-cut forms in the adjacent islands. Offshore, in the archipelago of Pulau Langkawi, limestones of the Chuping and Setul Formations occur as isolated sea stacks with precipitous cliffed walls (see Jennings,

1971, plate 39, p. 220).

2. Setul Boundary Range. The Setul Boundary Range which comprises limestones of the Setul Formation, forms an extensive line of hills along the western end of the Malaysian-Thailand border (fig. 2; plate 46). Though less than 5 km wide in Perlis, the outcrop is continuous over the frontier. The area is drained principally by the Sg. Pelarit and Sg. Gelang Kerbau. Relief decreases from north to south. In the north, Bt. Pelarit and Bt. Wang Mu reach heights of 554 and 540 m respectively, whilst near the coast, summits rarely exceed 300 m. Jennings (1972a, p. 340) described the Boundary Range as an "exaggerated Yugoslavian holokarst". When viewed from central Perlis the Boundary Range appears outwardly solid and imposing, with a plateau-like form. Is is, however, deeply dissected by steeply-cliffed ravines, dolines and many large enclosed depressions. Of the latter, Wang Tangga, drained by the Sg. Pelarit, is the most impressive (plate 48). It is totally enclosed and access may be gained only by a 300 m walk-way, suspended 5 to 10 m above the subterranean river course (plate 49). The flat alluvial floor of the depression, at an elevation of approximately 60 m, covers 0.7 km^2. Basal cliffs are less frequent in the Boundary Range than in the adjacent tower karst, and slopes are usually less steep, being mainly in the range 30 to 55°.

2. Caves

The caves of the three karst regions are described elsewhere (Crowther, 1978a, 1978b; Chapter VII, Section C1.1) and only their general characteristics are summarized here.

With the exception of the less pure members of the Setul Formation, the limestones in all three areas are massive, mechanically strong and are able to maintain large underground passages and caverns. For example, the caverns of Gua Batu, Bt. Batu (Selangor), are 40 m wide and 80 m high in places (plate 8), whilst

in Alam Cavern, Gua Tempurong (Kinta Valley), one cross-section is more than 130 m wide (fig. 45).

On hydrological grounds, a fundamental distinction may be made between the limestones of the Setul Boundary Range and those of the tower karst outcrops. Inthe former, autogenic vadose streams are present in many of the lower cave passages (plates 53 and 54). Permanent underground flow is favoured in this cave by the extensive catchment areas afforded by the Boundary Range and by the presence of less permeable, impure beds in the eastwards dipping strata. In contrast, the tower karst hills in all three regions are of insufficient size to maintain autogenic vadose flow. Also, the regional watertable lies well below the level of the adjacent plains and, consequently, there are no major springs around the base of the towers. Locally, allogenic streams pass through the towers (eg. G. Kroh, plate 38). Most noteworthy is the Sg. Tempurong, in the Kinta Valley, which rises on G. Chante in the Main Range and follows a 1.5 km east-west course through the G. Tempurong Gajah Massif (fig. 2; plate 35; Crowther, 1978a). Such cases are, however, exceptional, and caves in the tower karst are mostly infilling with speleothems, guano and other deposits (plates 9 and 10).

E. SOILS

Shallow soils and rugged topographies render the limestone hills mostly unsuitable for agriculture and forestry. Consequently, there have been few pedological studies in the karst regions and those have concerned mainly the deeper soils formed on lower, gentler slopes, often at the margins of the karst outcrops (eg. Joseph, 1965, 1968; Plateborze and Ee, 1972).

In a preliminary survey of the field areas, two broad groups of soils were identified: hill and foothill. A detailed

analysis and discussion is presented in Chapter IV.

1. Hill soils

Since the limestones are mostly very pure, hill soils are generally very shallow. Often, particularly on hilltops, the soils are restricted to isolated pockets between sharply-etched rock outcrops. On the summit of Ak. Bt. Takun (plate 14), for example, only 7.5 per cent of the surface is soil-covered. Here the soils are organic-rich and occur in the bottom of rock flutes and in crevices on cliffs. Typically they are slightly acid to slightly alkaline in reaction. Five samples from Ak. Bt. Takun had an average pH of 6.7 (range, 6.1 to 7.2).

Deeper and more continuous mineral soils occur on gentler slopes, and where local circumstances have favoured the accumulation of weathered residual material in depressions in the limestone surface. Such soils are characteristically dark reddish-brown in colour (2.5 YR 3/4), fine textured, very friable and of slightly acid reaction. One profile investigated in Wang Tangga had an average pH of 5.96. Also, relatively deep soils, exceeding 70 cm in depth, have formed in areas of less pure limestone, such as the Gelang Kerbau Valley. Here the soils are typically strong brown in colour (7.5 YR 5/6 to 5/8), clay-rich, and contain a mottled horizon with lateritic concretions at 45 to 65 cm. They are acidic in reaction, with a mean pH of 5.50.

2. Foothill soils

Joseph (1965, 1968) has established three soil series developed from limestone parent material in Kedah: the Langkawi Series, the Kaki Bukit Series, and the Kodiang Series. Mapping has not been extended to other karst regions and so the full extent of these soils is not known. The last two series are not considered further here since they generally occur on the plains at some

distance from the limestone hills.

Soils of the Langkawi Series generally occur at the foot of the limestone hills at elevations not exceeding 100 m (Malayan Soil Survey, 1974). They are deeper than the hill soils formed on pure limestones, the typifying pedon being 130 cm deep. However, in other respects they are very similar, being dark reddish-brown, friable clay soils, with a weak crumb structure in the A horizon. Clay skins are moderately developed in the upper part of the B horizon and become well developed lower down the profile. The pH (water, 1:2.5) ranges between 5.9 and 6.1 in the A and upper B horizons (0 to 68 cm) but is alkaline in the lowest B horizon (86 to 130 cm), the pH being 8.0 (Malayan Soil Survey, 1974).

A further soil type was identified on the lower slopes in Selangor and the Kinta Valley. A profile investigated on G. Tempurong (124 m) is representative of this group. It is dark yellowish brown (10 YR 4/4), weakly structured, and of slightly acid pH (range, 6.1 to 6.85). In contrast with those described above the soil contains a high proportion of sand, being as much as 40 per cent.

F. VEGETATION

The limestone hills of the Malay Peninsula support an extremely rich variety of flora (Chin, 1973; Henderson, 1939; Molesworth Allen, 1961; Wycherley, 1967, 1971). Hitherto, botanical work has been concerned primarily with the identification and recording of rare plant species. The results of a quantitative investigation of vegetation characteristics, litterfall and throughfall in the three field areas are discussed in Chapter IV.

Lowland Dipterocarp rain forest forms the climax vegetation at lower altitudes throughout the Malay Peninsula (Richards, 1952; Whitmore, 1975; Wyatt-Smith, 1964). However, the limestone outcrops are exceptional in that frequent rock outcrops and shallow

soils, with their correspondingly low water holding capacities, severely inhibit tree growth. Hopea dryobalanoides (family, Dipterocarpaceae) has been recorded on Bt. Batu (Wycherley, 1971), but with few exceptions, dipterocarps are absent from the limestone hills in Selangor and the Kinta Valley. In these areas the tree flora are of types characteristic of the understorey of tall lowland or disturbed secondary forest (Wycherley, 1971). On bare rock summits, scattered, stunted and misshapen trees with small tough leaves form an irregular and discontinuous canopy (plate 14). Vitex siamica, Memecyclon spp., and Garcinia spp. are typical of such environments (Henderson, 1939). On summits with deeper soils, such as G. Layang Layang in the Kinta Valley (Chapter IV, Section C2.1), the cover is characteristically more dense, and the trees spindly.

The North-west region is closer to the semi-deciduous and deciduous dipterocarp forests of Thailand and Burma, and consequently Dipterocarp species are more common on the limestone hills (Richards, 1952). In Perlis and Langkawi Hopea ferrea, Pentacme siamensis and Shorea sericeiflora occur where soils are deep (Wyatt-Smith and Panton, 1963). Unlike the limestone flora of Selangor and the Kinta Valley, many trees in the North-west region are semi-deciduous and lose their leaves in January and February (cf. plates 46 and 48).

In all three regions shrubs and herbs commonly form the dominant plant cover on rocky hilltops and cliffs (plate 30). Locally, differences in topography, aspect, drainage and soil characteristics provide a great diversity of habitat, and this is reflected in the floral composition (Wycherley, 1971). The bare rock surfaces support a wide variety of cremnophyte (rock crevice plants) and calicole species. Certain herbs, such as those of the Gesneriaceae family (eg. Boea spp., Chirita spp., Monophyllaea spp.,

and Paraboea spp.) occur frequently on exposed cliffs and are able to resist prolonged dry spells by reversible desiccation of their tissues (Whitmore, 1975). Others, such as those of the Balsaminaceae and Begoniaceae families, have tubes or bulbs which similarly allow them to survive the drier periods. Where deeper mineral or organic soils are present, the upper horizons are often acidic and support a calcifuge flora. However, conditions are not so extreme as those on limestone summits up to 610 m altitude in Sarawak where the floral assemblage has close affinities to the low altitude heath forests of Borneo (Anderson, 1965).

Certain plant species are restricted to particular limestone outcrops. For instance, Stenothymus ridleyi, of the Acanthus family, occurs only on the hills near Tambun in the Kinta Valley (Molesworth Allen, 1961). Similarly, certain species of mollusca have been identified as being particular to individual hills (Tweedie, 1942).

CHAPTER III

METHODS OF STUDY

CHAPTER III. METHODS OF STUDY

This chapter examines briefly the basic structure of the investigation and outlines the various field and laboratory techniques used.

A. OVERALL EXPERIMENTAL DESIGN

Being the first investigation of its kind in West Malaysia, the present study is necessarily of an exploratory nature. The main criterion used in the choice of sampling sites was that the full range of denudational environments was encompassed. The initial field and literature surveys of the karst regions highlighted the main soil and vegetation types (Chapter II, Sections E and F), and provided a rational basis for selecting study plots. Slope profiles, soil properties and tree densities were measured in 13 plots, and in six, soil moisture, soil carbon dioxide, litterfall and throughfall were monitored regularly over a period of one year. The data obtained from these investigations (Chapters IV and VI) are broadly representative of soil-covered karst in the three field areas. Also, the weathering processes measured on the rocky summit of Ak. Bt. Takun (Chapter V) are very similar to those on bare limestone outcrops elsewhere in the Peninsula.

The establishment of a network of water sampling sites was more problematical for two reasons. First, few springs, resurgences and caves are mapped on the 1" Malaysian Topographic Survey, and comparatively little information is available on individual caves, apart from brief descriptions and sketch maps of Gua Batu (Heynes-Wood and Dover, 1929; Soepadmo and Ho, 1971) and Gua Anak Takun (Dunn, 1965) in Selangor, and Goa Boa (Jones, 1965) in the Setul Boundary Range. Consequently, the exercise of locating potential sites relied heavily on local assistance, and was time consuming. Secondly, the characteristics of a water sampling site, unlike soil and vegetation, cannot be assessed on the basis of a single obser-

vation or measurement. Therefore, to reduce the chance of significant sites being omitted, as many as possible were included in the experimental framework. Thus, virtually all the major rivers, cave streams and springs in the field areas, and 113 underground diffuse-flow seepages in caves, were sampled regularly over the study period. In some cases, potentially interesting sites were necessarily excluded by their sheer inaccessibility.

B. METHODS OF FIELD SAMPLING AND MEASUREMENT

1. Slope transects

Soil and vegetation studies were carried out in 13 plots (ideally, 30 x 30 m square). The plots were established orthogonal with the local slope. In each, slope profiles were measured along 1 m segments using a pantometer (Pitty, 1968a). Because of the discontinuous form of the micro-relief, with its frequent enclosed hollows, slope profiles were measured along straight lines which were aligned parallel with the two steeper sides of the plots. By adopting this procedure, the steepness of relief in areas with deeply-dissected meso-forms was somewhat underestimated. Soil depth was determined at the end of each 1 m segment.

2. Method for collection and storage of soil air samples

2.1. General background

Bunting and Campbell (1975), Grable (1966) and Macfadyen (1970) present discussions on the problems of sampling soil air. The type of technique adopted depends largely upon the means available for sample extraction and the sample volume needed.

Samples may be extracted by pump (eg. Draeger method - Miotke, 1974), by syringe (Cooper, 1975; Dowdell et al., 1972) or by means of a carbon dioxide permeable 'container' (eg. Harley and Brierley, 1953; Macfadyen, 1970; Martin and Pigott, 1965;

Nicholson and Nicholson, 1969). The author had no access to a suitable pump, and even had one been available, then its use would have proved impracticable in the field areas. Initially, tests were undertaken on the device used by Nicholson and Nicholson (1969) in the limestone soils of Jamaica. The method proved unsatisfactory in three respects. First, soil was disturbed considerably when the plastic pipe was removed for sampling. Secondly, glass sampling tubes were found unsuitable for field use. Finally, problems were encountered in drawing air from the glass sampling tube into the analyser as it was difficult to maintain the pressure "by blowing up the balloon in the sampling tube" (Nicholson and Nicholson, 1969, p. 137). The other methods which use permeable 'containers' were considered inappropriate for regular sampling at depth. Consequently, the method of syringe extraction was adopted. Samples of at least 30 ml were required to determine carbon dioxide concentrations by means of a Lloyd gas analyser. Since volumes as large as this can rarely be extracted directly from soil pores without a pump, chambers or air reservoirs were utilized.

2.2. The soil chambers

The chambers (fig. 11) comprise thick-walled steel pipe, 34 mm internal diameter, and of a length equal to the sampling depth. The pipe is open at the bottom to allow free entry of soil air, and an access tube, 6 mm outside diameter, passes through a metal plate, which is welded at the top. Throughout, the sampling system air-flow is controlled by Hoffman clips attached to flexible PVC tubing of 6 mm bore. One such valve is pushed firmly onto the access tube forming an air-tight seal. Because the area of PVC tubing exposed below the clip is small, and because carbon dioxide diffusion through such tubing is likely to be slow, errors arising from this source are considered to be negligible. The PVC tubing did, however, prove to be a weakness in a further unexpected way.

It fell victim to the insatiable appetite of the ubiquitous termite! After only one week in the field, holes of up to 5 mm in diameter were eaten in the tubing. Of the range of remedial measures tested, plasticine, moulded round the tubing, proved to be the most satisfactory deterrent - and it also reduced the rate of carbon dioxide loss.

2.3. The syringe

Even though less than 100 ml of air were removed at one sampling, a large and robust syringe (fig. 11) was required to extract this volume from wet and clay-rich soils. The outer casing, constructed of clear perspex, allows immediate detection of defects in the piston. The latter is made from aluminium with a rubber seal, and the rod is of light-weight aluminium tube. The sample intake assembly comprises a glass tube connected to the base of a polypropylene y-shaped adaptor (6 mm outside diameter) via a short length of PVC tubing. One arm of the adaptor leads to the soil chamber, the second to the sample containers. In order to make the sample intake assembly more resilient to knocks, it is mounted in a rubber bung - the latter acting as a shock absorber. The joint between the rubber bung and the perspex case, and also the internal walls of the syringe are lightly coated with silicone grease to ensure that air-tight seals are maintained.

2.4. Sample storage

Transporting field samples to the laboratory without change of composition is a major problem in soil air studies (Bunting and Campbell, 1975; de Jong and Schappert, 1972; Macfadyen, 1970). A wide range of containers, including glass sampling tubes (similar to those used by Nicholson and Nicholson, 1969) and heavy duty polythene bags (as described by Macfadyen, 1970), were tested. Football bladders were found most suitable. Several types were tested and one, of Czechoslovakian manufacture, proved to be

considerably less permeable than the others.

Throughout the study, soil air samples were taken between 13.00 and 14.00 hrs and were transported to the laboratory in a well-ventilated position in the rear of the field vehicle. The rate of carbon dioxide loss under these conditions was determined experimentally. When expressed as:

$$\frac{\text{amount of carbon dioxide lost in 1 hour}}{\text{initial concentration of carbon dioxide}} \times 100$$

the values for the 15 sampling bladders used in the study ranged between 1.6 and 2.3. A typical curve, for bladder no. 1 (2.0 per cent), is shown in figure 12. Such curves were used to estimate carbon dioxide loss during storage - the storage time being kept to a minimum, and in all cases less than 4 hours. The experiment was repeated on 5 of the bladders towards the end of the study period in order to test whether the permeability of the rubber had changed through field use. No significant difference was found.

2.5. Sampling procedure

In an attempt to minimize dilution of the soil air sample by foreign air present within the sampling equipment, the following standard procedure was adopted (letters refer to valve clips shown in fig. 13):

1. The bladder is fitted to the syringe. With B closed, C and D open, and the piston fully depressed, the piston is pulled out as far as possible to remove air from the bladder. The piston is held until C is closed. Air is expelled from the syringe via B, and the bladder detached.

2. The syringe is fitted to the soil chamber. To check for leakages: with B open and A and D closed, the piston is withdrawn as far as possible, held for 10 seconds, then released. If the piston returns to the base of the syringe, the system is air-tight.

3. To flush the intake assembly of foreign air: with D closed, A and B open, 15 ml of soil air is drawn into the syringe to flush

out foreign air in the intake assembly. With B closed and A open, this air is expelled through D. This is repeated and on the second occasion the expelled air is used to flush the neck of the bladder above C.

4. The bladder is attached immediately.

5. With C and D closed, A and B open, 40 ml of air is drawn into the syringe. In wet or clay-rich soils it was necessary to draw as much as 100 ml into the syringe under greatly reduced pressure, close B, and then release the piston - giving a final volume of approximately 40 ml.

6. With A and B closed, C and D open, the sample is expelled into the bladder, C closed, and the bladder removed.

2.6. Discussion of carbon concentrations measured by the present method

In an early study of soil aeration using a sampling device of the gas chamber type, Boynton and Compton (1944) observed that greater tensions were required to extract samples of air from smaller pore spaces. Vine et al. (1942) also noted that on many occasions no gas could be withdrawn, even under greatly reduced pressure, from the lowest soil layers - presumably (they suggest) because of the higher moisture content at depth. These observations highlight a major difficulty in soil aeration studies. Under dry conditions air will diffuse readily between the soil macro-pores and the gas reservoir and, during sample collection, air will be drawn freely by mass flow from the interconnected macro-pores. In contrast, under wetter conditions, gaseous diffusion and mass flow will be inhibited. Thus, at the greater tensions required to extract soil air, air will be drawn not only from the macro-pores which connect with the open end of the chamber, but also from smaller pores. Under contrasted moisture conditions, therefore, soil pores of different sizes contribute differentially to the air sampled. Since gaseous exchange is likely to occur more readily in

larger, interconnected pore spaces, the latter will generally have lower carbon dioxide concentrations. Consequently, temporal and spatial variations in measured carbon dioxide levels may be partially caused by the sampling of different pores. It should be noted that this problem is likely to be less marked when the gas chamber method is used, than in the case of direct extraction from capillary tubes (eg. the Draeger method), since the bulk of the sample is generally obtained from the reservoir. Further aspects of the sampling of soil air are discussed in Chapter VI, Section D.6.

3. Other soil sampling methods

Samples for soil moisture determination, 15 to 20 g in weight, were taken by means of a small screw auger and stored in small screw-capped aluminium containers. Usually, samples were weighed within several hours of collection. When storage exceeded 24 hours, as with samples from the North-west region, corrections were made for moisture loss. Over the soil moisture range normally present this averaged 0.05 g per day. Cylindrical tubes (35 mm diameter) were used to extract cores of known volume for bulk density determination.

4. Throughfall and litterfall measurements

Throughfall refers to rainfall which reaches the ground after passing through a vegetation canopy. As in previous studies (eg. Carlisle et al., 1966; Nye, 1961; Voigt, 1960) it was collected using rain gauges. These comprised a polythene funnel, 185 mm in diameter (with upper rim removed) and an 18 l plastic collecting bottle (fig. 14; plate 32). In order to reduce evaporation loss, a length of polythene tube was connected to the base of the funnel. Particulate matter was retained on a plug of fine plastic netting in the neck of the funnel. When installed,

the rim of the gauge was at least 55 cm above the ground surface. The volume of throughfall was measured and a sub-sample taken for chemical analysis. These devices were also used as temporary rain gauges, and to collect rainwater samples.

The term 'litter' is taken to include "dead leaves, twigs, scales and bracts, fruits and flowers and other small unidentifiable material including some insect frass which may come down with the leaves that fall on to the forest floor" (Lim, 1978, p. 375 - after paper presented at International Biological Programme Synthesis Meeting, Kuala Lumpur, Aug. 1974), and 'litterfall' as the production of this litter. According to Klinge (1978), this should be termed 'fine litter'. A variety of devices have been used to collect litterfall (eg. belts of plastic netting - Kira et al., 1967; square or rectangular litter trays - Kenworthy, 1971; Lim, 1978; circular funnels - Ogawa, 1978). In the present study 1 m square litter trays were used. These comprised a fine-grade plastic netting, mounted at the base of a wooden frame (15 cm deep). The trays were supported on wooden legs at a height of 50 cm above the ground surface (plate 32). Disposable polythene gloves were worn when handling the litter in order to prevent contamination with salts from the hands.

5. Water sampling methods

Water samples were taken using narrow-mouthed, screw-capped polythene bottles (mostly 250, 500 or 650 ml). Each bottle was soaked in dilute nitric acid and then washed thoroughly in distilled water before use. At each site the bottle and cap were flushed with the water to be sampled before filling. In the case of springs and cave streams, samples were taken from the point of emergence. Where possible, bottles were capped beneath the water, ensuring that no air was present. Polythene funnels were used at drip sites in caves. When the drip rate was low and sampling

bottles were only partly filled during the time available the bottles were squeezed to expel the air before capping.

5.1. Runoff from exposed and partially-covered rock outcrops

Runoff from bare limestone surfaces and from rock flutes with soil and litter covers (Ak. Bt. Takun, Chapter V) was conducted to collecting bottles via lengths of PVC tubing (fig. 15). Over the 2 weeks' duration of the experiment, plasticine was used to hold the tubing in place since it was found not to alter the chemical quality of runoff during its brief contact with the water, and could be moulded on to the rock surface to form a seal and divert flow. Ideally, the total runoff from the rock catchment would have been collected, its volume measured, and a sub-sample then taken. However, in the case of an average-sized rock flute, 1 m in length and 20 cm wide, the potential runoff from a 30 mm storm would be approximately 4.5 l. From the larger catchments 50 litres or more may be expected. The containers available ranged between 65 ml and 18 litre capacity, these being mostly 650 ml. It was found experimentally, using a flow of 25 ml per minute (typical of field rates), and a 650 ml sampling bottle, that when sampling bottles overflow, then the most representative samples are obtained when the outlet of the PVC tubing is inserted to approximately one-third the depth of the bottle. Clearly, the results would differ according to the size and shape of the bottle, and according to the volumes and rates of flow. However, since runoff characteristics could not be predicted in advance, and varied significantly from site to site, the tubing was inserted consistently to one-third the depth of the bottles. By adopting this standard procedure sampling errors were reduced, though not eliminated completely.

5.2. Throughflow collection at the soil/rock interface

Waters seeping through mineral soils at the soil/rock

interface (termed 'throughflow' in the present study; Chapter VI), were intercepted by means of zinc plates, 30 cm long, 15 cm wide, and 2 mm thick. The front and sides of the device were bent upwards to form raised rims, 2 cm high, which were welded at the front corners (fig. 16A). A zinc outlet pipe was connected to plastic collecting bottles via PVC tubing. Pits were excavated to bedrock, which in the cases studied comprised irregular blocks of limestone, and the plates were installed so that they received runoff from the rock surface (plate 33). On occasions, plasticine was used to seal the plate in place. The water was piped to a collection point several metres downslope, thus allowing the original pit to be infilled (fig. 16B) and thereby preventing artificially accelerated gaseous diffusion from the deeper soil layers. The method proved successful at first. However, after about 4 months the collection devices failed, for two reasons. First, the PVC tubing which carried water to the collecting bottle became badly eaten by termites. Secondly, the plasticine which was used to seal the zinc plate came adrift from the rock at one site (ABT-I - see Chapter VI, Section E1.1).

6. Field measurements of karst water characteristics

6.1. Water temperatures

Total immersion thermometers (graduated to 0.1°C) were used, and temperatures were estimated to 0.01°C. Thermometers were calibrated against each other to ensure consistent measurement. In the case of cave drips, the temperature of the actual sample was determined, always ensuring that at least the bottom 10 cm of the thermometer was covered. Experiments showed that the errors arising from incomplete immersion of the thermometer were negligible under these circumstances. Where the rate of flow was slow, samples were transferred to a narrow-bore, polypropylene measuring cylinder to provide the necessary depth of water. If insufficient water was

collected over a 20-minute period, then no reading was taken.

6.2. Discharge measurement

Stage-discharge curves were available for many of the rivers (unpublished D.I.D. records), and wherever possible surface sampling sites were established close to gauging poles. Streams issuing from karst risings, and also underground streams and seepages, were mostly too shallow to permit the use of a flow meter. In these cases, stream velocity was estimated by means of a velocity head rod (Wilm and Storey, 1944). Flow rates at drip sites in caves were readily determined using polypropylene measuring cylinders.

7. Methods of assessing erosion rates on limestone surfaces

7.1. General background

Surface erosion rates may be estimated directly by measuring the depth of surface lowering over a given time increment. Many studies of this type have relied upon either natural datum levels, such as those provided by glacial erratic boulders (Bögli, 1961; Sweeting, 1966; Williams, 1968) or upon artificial surfaces of known date, such as gravestones (Geikie, 1880; Mathias, 1967; Rahn, 1971; Tolley, 1978) and buildings (Raistrick and Gilbert, 1963). Although this approach provides useful estimates of long-term erosion rates at particular sites, it is insufficiently sensitive to detect short-term micro-scale changes in solutional activity. The micro-erosion meter (High and Hanna, 1969; Trudgill, 1972a) and the erosional weight-loss tablet technique (Trudgill, 1975) have been developed in an attempt to meet this need, and recent studies have clearly demonstrated their value in assessing spatial variations in surface and subsoil erosion rates (High and Hanna, 1969; Jennings, 1977; Trudgill, 1972a, 1972b, 1975, 1976b, 1977a, 1977b). However, both techniques share three significant limitations. Firstly, where a soil cover is present, this is

inevitably disturbed during installation of the datum studs and tablets and on subsequent occasions when measurements are made. A period of 'unnatural' erosion, diminished or accelerated, necessarily follows, and may persist for several months after disturbance (Trudgill, 1977a). Secondly, the very act of direct erosion measurement may erode the rock surface. Impact of the probe tip of the micro-erosion meter (Trudgill, 1977a) and handling of the rock tablets during transport, installation, removal, cleaning and weighing may thus effect erosion of a magnitude which might be highly significant over short time intervals. Finally, amounts of surface lowering and weight-loss are so small as to limit observations to those of inter-seasonal variations (Trudgill, 1977a).

In the present study, rock tablets were used to estimate sub-aerial (Chapter V) and sub-soil erosion rates (Chapter VI). Additionally, surface erosion rates on Ak. Bt. Takun were estimated indirectly from the total hardness content of superficial runoff.

7.2. Erosional weight-loss tablet technique

The method used is that described by Trudgill (1975), but with two modifications. First, the tablets were square rather than circular, and measured 5.08 x 5.08 x 0.64 cm - a shape and size determined by the available rock cutting facilities. Secondly, they were not enclosed in nylon mesh as this would have served no useful purpose in the present study. The main objective was to assess the relative solutional potential of a range of different weathering environments, rather than to estimate the absolute rates of erosion in the field areas. Consequently, it was essential that the rock tablets were identical, and not of the local rock. These were cut from a single homogeneous block of Great Scar Limestone, density 2.65 g/cm^3, from Langcliffe, Yorkshire.

7.3. Solute method

Surface erosion rates may be assessed by monitoring the

volume and chemical quality of precipitation and runoff on bare limestone outcrops. The method has been used frequently in estimating regional denudation rates (eg. Corbel, 1959b; Jennings, 1972b, 1972c; Hellden, 1973), but, with the exception of recent work by Jennings (1978), has not been used in micro-scale studies. The technique may be used to measure the erosional effects of single storms, or parts of storms. Furthermore, if measurements are made over the range of weather conditions which obtain in an area, and if long-term climatic records are available, then mean annual erosion rates may be estimated with some confidence. Runoff from steeply-inclined rock outcrops on Ak. Bt. Takun is rapid and the relative humidity is high, particularly during storms. Consequently, evaporation loss is likely to be small, and erosion rates on such surfaces may be estimated from the solute content of runoff collected by the sampling procedure adopted (Section 5.1, above). Evaporational and evapotranspirational losses might be significant on rock surfaces with partial soil and litter covers and ideally, under these circumstances, total runoff volumes would have been collected.

7.4. Solute and direct methods of assessment compared

Where circumstances permit its use, the indirect method of assessment compares very favourably with the micro-erosion meter and rock tablet techniques. The equipment required is cheap, readily obtainable, and easily installed. It permits measurement of subsoil erosion without site disturbance. Absolute erosion estimates are obtained and these relate to clearly defined areas within which a range of possible independent controls may be evaluated. Most importantly, erosional losses over short time intervals are detectable with a degree of precision superior to that of the alternative methods, and without the inaccuracies of the latter methods.

The degree of precision attainable by the three techniques

may be compared under conditions which approximate to those which obtain in the karst regions of West Malaysia (fig. 17). Calculations are based on a standard rock tablet, 1.5 cm in diameter and 0.5 cm in thickness. The density of the bedrock and of the tablets is assumed constant at 2.5 g/cm^3. The likely error at the 95 per cent confidence limits has been calculated for a bare rock surface and, to facilitate comparisons, it is also assumed that the total hardness of runoff over each time period is constant at 40 ppm. At such a site, erosion of the rock tablet will be confined to the top surface and sides. The curve for the micro-erosion meter is based on reported instrument reliability (Trudgill, 1977a) and for the rock tablet technique is calculated from the weighting reliability of microbalances (Trudgill, 1977a, from Newson, 1970). The third curve is derived from 20 replicatory titrations of total hardness, using 50 ml samples, which gave a standard deviation of 0.48 ppm. The confidence limits of measurement are presented as percentages of the actual erosional losses associated with different amounts of precipitation. An approximate time-scale appropriate to the rainfall of West Malaysia is shown. This is based on average rainfall data for stations in Selangor and upon storm characteristics recorded at Templer Park (2601) during the study period.

The direct techniques have a low level of precision over short time intervals, the errors being 36.4 and 27.5 per cent for 0.2 mm rainfall when surface lowering equals 0.0000032 mm, but this increases considerably over longer time periods. In the case of 2000 mm precipitation, typical of annual rainfall in West Malaysia, erosional loss amounting to a surface lowering of 0.032 mm may be estimated to 0.6 and 0.08 per cent by these methods. In contrast, the level of precision attainable by titration remains constant (4.0 per cent) over all time scales, provided a 50 ml sample is obtained (equivalent to 0.2 mm rainfall over 0.25 m^2). Errors

increase proportionately with reduced sample volume. On the grounds of instrument precision, and under the above conditions, solute loss is the most reliable technique for all erosional events, single storm or monthly, of less than 64 mm precipitation (surface lowering, 0.001 mm). Beyond this, the micro-erosion meter is superior. Instrument and operator variance may be reduced considerably by replicate measurement. For instance, the standard procedure of triplicate titration reduces the error of water hardness determination to 1.36 per cent, and similar improvements may be achieved by direct methods.

C. METHODS OF LABORATORY ANALYSIS

Standard laboratory methods were used throughout the study and only reference manuals or brief descriptions are given here. Most laboratory work was carried out at the Universiti Malaya, Kuala Lumpur. However, when working in the Kinta Valley and North-west region, a mobile laboratory was set up in the field areas in order that soil air composition and the pH, specific conductance and alkaline hardness of water samples could be determined within 5 hours of collection.

1. Methods of soil analysis

Soil moisture content was measured gravimetrically by oven-drying at 105°C for 24 hours (Bascomb, 1974). It is expressed as a percentage by volume of soil, this being calculated from bulk density measurements. Particle density was determined by means of a 100 ml specific gravity bottle, the air being removed from the soil sample by boiling (Blake, 1965). Particle-size analysis was undertaken by the pipette method (British Standards Institution, 1967). Soil reaction, loss on ignition, cation exchange capacity (C.E.C.) and exchangeable calcium, magnesium, sodium and potassium were determined by the methods described by Bascomb (1974). Organic

carbon was determined by the Walkley and Black method, using ferrous sulphate, and orthophenanthroline-ferrous complex as indicator (U.S. Dept. Agriculture, 1972). The latter indicator was used in preference to barium diphenylamine sulphonate since it gave a clearer end-point in the soils analysed (for discussion, see Hesse, 1971, p. 211-16). Soil air samples were analysed using a Gallenkamp-Lloyd gas analyser, model GC-370 (Lloyd, 1958). The carbon dioxide content of each sample was determined twice. When concentrations differed by more than 0.04 per cent, a further analysis was carried out, otherwise an average value was taken. Calcium carbonate was estimated by observing the reaction when a few drops of 10 per cent hydrochloric acid were applied to the soil (Hodgson, 1974, pp. 56-7).

2. Analysis of rock samples

Rock samples were dissolved in 150 ml of 3N hydrochloric acid at 50 to 60°C (Bisque, 1961), and then filtered through a Whatman No. 42 ash-free filter paper. The weight of acid insoluble residue was determined. The filter paper and residue were then ashed, and a sample of 50 mg was digested in a teflon decomposition vessel using the procedure described by Bernas (1968). The cationic composition of the original filtrate, and of the digested residue were then determined by atomic absorption and flame emission methods (Section 4, below), using appropriate blanks and standard solutions. The same method was used in the analysis of snail shells (Chapter V, Section D2.2).

3. Analysis of plant material

For most samples, the dry ashing method described by Isaac and Kerber (1972, p. 28) was used. Plant tissue was dried at 80°C for 12 hours and then finely ground. A sample of 1.00 g was then ashed at 500°C for 4 hours, and the ash dissolved in 5 ml 2N hydrochloric acid. With some samples, however, the residue did not dissolve completely, even when the solution was warmed. According to Leigh (personal communication, 1975), this was probably due to

silica. In these cases, dissolution was effected using hydrofluoric acid.

4. Analysis of water samples

Because of the difficulties of obtaining reliable measurements of pH in the field using portable meters (Bray, 1971; Stenner, 1969), pH was determined in the laboratory by means of a bench pH meter. Provided that samples are kept tightly sealed, then fairly consistent results may be obtained up to several hours after collection (Harmon et al., 1972). Experiments showed that pH, and alkaline hardness, the properties of water which are most susceptible to change (Brown et al., 1970), altered markedly during storage. For instance, the alkaline hardness of one sample decreased from 232 ppm to 174 ppm over a four-day period. These properties, and specific conductance were determined within 3 to 5 hours of sample collection. Samples were then filtered, if turbid, and preserved by acidifying with sulphuric acid (Golterman and Clymo, 1969, p. 19). Tests were carried out on a range of samples to ensure that the cationic composition of the waters was not affected by acidification.

Alkaline hardness was measured by titration using 0.02N sulphuric acid and British Drug Houses (B.D.H.) 4.5 mixed indicator solution. Calcium hardness (expressed as ppm, calcium carbonate - see Douglas, 1968b) was determined by E.D.T.A. titration using B.D.H. indicator tablets (Stenner, 1969). The calcium content of rainfall and throughfall was determined using a Perkin Elmer Atomic Absorption Spectrophotometer. A range of samples were tested using first an air-acetylene flame, with lanthanum chloride solution added (Brown et al., 1970; Isaac and Kerber, 1972), and second, a nitrous oxide-acetylene flame. No significant difference was found in the results. Since the addition of lanthanum was both expensive and time-consuming, a nitrous oxide-acetylene flame was used through-

out the study. Tests also showed that the calcium contents of karst waters determined by titration and atomic absorption were the same. Consequently, the latter method was used when the sample volume was insufficient for titration.

In karst water studies, magnesium is often calculated by subtracting calcium hardness from total hardness, both being determined by titration methods (eg. Halliwell, 1977). The method has been criticized (Bray, 1969). An alternative technique of direct titration for magnesium has been described by Bray (1969) but as Halliwell (1977) notes, this method is both more complex and expensive. In the present study, magnesium was determined by atomic absorption, using a nitrous oxide-acetylene flame. Total hardness was then calculated as the sum of magnesium hardness (expressed as ppm calcium carbonate) and calcium hardness. Non-alkaline hardness (expressed as ppm calcium carbonate) was determined by subtracting alkaline from total hardness. The aggressiveness of water towards calcium carbonate was assessed by measuring the total hardness of samples before and after saturation with AnalaR calcium carbonate, the latter being added in the field to a duplicate sample at the time of collection (Stenner, 1969). Sodium and potassium were determined by flame emission methods, and iron, aluminium and silicon by atomic absorption, using an air-acetylene flame (Brown et al., 1970).

In order to facilitate comparisons of the cationic compositions of rainwater, throughfall, litterfall, exchangeable bases in soil, limestone bedrock and karst waters, the molar ratios Ca+Mg:K+Na, Mg:Ca and K:Na have been determined.

D. METHODS OF STATISTICAL ANALYSIS

Standard methods of statistical analysis have been used throughout the study. Except where stated otherwise, the Pearson correlation coefficient (referred to as '*r*') has only been determined for data which does not differ significantly from the normal distribution.

Certain data have been normalized by logarithmic transformations. The term 'r_s' denotes the Spearman rank correlation coefficient.

CHAPTER IV

SPATIAL VARIATIONS IN SOIL AND VEGETATION CHARACTERISTICS

CHAPTER IV. SPATIAL VARIATIONS IN SOIL AND VEGETATION CHARACTERISTICS

A. INTRODUCTION

Preliminary field observations revealed marked variations in the soils and vegetation of the karst outcrops. The aim of this chapter is to describe those features of the soil and vegetation cover which are of geomorphological significance and to examine the factors which influence their spatial distribution. Soil moisture and carbon dioxide are discussed later (Chapter VI).

1. Literature review

1.1. Variations in soil type in tropical karst terrain

According to Owen (1951, p. 27), soils derived from limestone in the Malay Peninsula are "fairly uniform, varying slightly in intensity of red colour and reaction". At the time they were grouped together provisionally as one series - the Perlis Series. In a later and more detailed survey of the lower slopes and hill-foot sections of limestone outcrops in N. Kedah, Joseph (1968) has identified three soil series. These form a toposequence which ranges from comparatively base-rich soils (Langkawi Series) on the lower slopes, through soils of the Kaki Bukit Series developed at the foot of the limestone hills, to base-deficient soils (Kodiang Series) on the adjacent plains. Typically, base saturation values in the top 15 cm of the soil are 40, 15 and 7 per cent, respectively. There are no comparable data for the higher slopes and hilltops in the North-west region, or for karst soils in the Kinta Valley and Selangor.

Topographically-controlled variations in soil properties have been observed in Jamaica (Price, 1959; Smith, 1969a), Sarawak (Anderson, 1965; Wilford and Wall, 1965), and Belize (Furley and Newey, 1979). In the last region soil depth, moisture, exchangeable magnesium and percentage fines were found to increase from the hill-

tops to the base of cone karst depressions, while pH, calcium carbonate, other exchangeable bases, and the coarser soil fractions decreased downslope. Differences in the purity and chemical, mineralogical and structural characteristics of the limestone may also affect soil properties. In Jamaica, for example, soils of the White Limestone outcrop contrast with those of the less pure Yellow Limestone. In the former area, slightly alkaline, loam and clay loam soils form a shallow and discontinuous cover, while in the latter, soils are clay-rich, neutral in reaction and typically between 30 and 90 cm in depth (Smith, 1969a). Marked local differences in soil type also occur in Puerto Rico (Monroe, 1976) and on Aldabra Atoll (Trudgill, 1976b).

1.2. Variations of vegetation in tropical karst terrain

Regional and more local distribution patterns are now well-established for many of the plant species of the karst outcrops in West Malaysia (Chapter II, Section F). Detailed investigations of floristic composition have also been undertaken in Belize (Furley and Newey, 1979) and Sarawak (Anderson, 1965). However, comparatively few measurements of fundamental ecological variables, such as vegetation density, basal area of trees at chest height and plant biomass, have been made in tropical karst regions.

1.3. Hydrological significance of vegetation in the humid tropics

A considerable proportion of precipitation in rain-forested catchments is lost by evaporation and transpiration. For example, in areas of relatively undisturbed rain forest in the Main Range, Malaysia, 18 to 42 per cent of rainfall is intercepted by the forest canopy (Kenworthy, 1971; Low, 1972, 1974) and the combined evaporational and evapotranspirational losses account for 50 to 70 per cent of the total precipitation input (Kenworthy, 1969; Low and Goh, 1972). Interception rates within the above limits have been reported from Sarawak (Brunig, 1970), El Verde, Puerto Rico (Kline et al., 1968), Uganda (Hopkins, 1960), and Mauritius (Vaughan and Viehe,

1957). The vegetation canopy may also be important in redistributing throughfall, as in the Jamaican cockpit karst (Aub, 1964 - cited by Sweeting, 1972).

1.4. Organic matter production and nutrient cycling in the humid tropics

It is well-established that rates of organic matter production in tropical rain forests are greater than in other regions (Bazilevich et al., 1971; Jones, 1979, pp. 26-35; Lieth, 1973; Whitmore, 1975, pp. 94-98; Whittaker and Likens, 1973). Consequently inputs of organic matter at the soil surface are high. For instance, in the Pasoh Forest Reserve (F.R.), Negri Sembilan, West Malaysia, gross productivity ranges from 70 to 100 t/ha/yr (Kira, 1978), and fine litter production varies between 7.5 and 10.6 t/ha/yr (Gong, 1972; Lim, 1978; Ogawa, 1978). Similar rates of litterfall have been reported from rain forests in E. Amazonia, Brazil (Klinge, 1978), Ghana (John, 1973; Nye, 1961), Trinidad (Cornforth, 1970) and Columbia (Jenny et al., 1949). In the Malay Peninsula many rain forest trees exhibit a seasonal pattern of litterfall, with peaks occurring usually during the drier months (Koriba, 1958; Medway, 1972; Ogawa, 1978; see also review by Whitmore, 1975, pp. 52-58).

Rapid cycling of nutrient elements is a characteristic feature of rain forest ecosystems (Douglas, 1969; Kenworthy, 1971; Whitmore, 1975), the nutrient elements being returned to the soil principally in litterfall, throughfall and stemflow. At Pasoh (1972/73), for example, the amounts of calcium, magnesium and potassium returned to the ground in fine litter averaged 69.5, 18.2 and 31.5 kg/ha/yr, respectively (Lim, 1978). Data on the solute contents of throughfall and stemflow have been reported from many rain forest regions (eg. Pasoh F.R. - Manokaran, 1978; Ulu Gombak F.R., Selangor - Kenworthy, 1971; Kade, Ghana - Nye, 1961; Trinidad - Cornforth, 1970; El Verde, Puerto Rico - Jordan, 1970).

Since data of this type are lacking from karst regions

in the humid tropics, the results of studies in the Pasoh and Ulu Gombak forest reserves form the basis for comparative discussion in this chapter. The Pasoh F.R. is an area of primary lowland Dipterocarp forest, located on acid soils derived from deeply-weathered granites, sandstones and shales, at an altitude of between 75 and 150 m (Soepadmo, 1978). The Ulu Gombak Reserve also comprises Dipterocarp forest, of which 50 per cent is primary (Kenworthy, 1971). It is located on deeply-weathered granites at an altitude of between approximately 150 and 600 m.

2. Outline of objectives

The principal objectives of this chapter are:

1. To describe and quantify the meso-relief, soil properties, and vegetation characteristics of the karst regions, and interpret the spatial variations observed.

2. To report on interception, litterfall and the cycling of nutrients in litterfall and throughfall.

3. To examine the interrelationships between nutrient cycling and the chemical properties of bedrock and soils.

B. EXPERIMENTAL DESIGN

1. Selection of study areas

The main criterion in selecting the study areas was that they should be representative of the range of relief, soil and vegetation types present in the three regions. As noted previously (Chapter II, Section D), the relief of the karst regions comprises four main morphological units. In the present study, attention focuses on two of these: hilltops and soil-mantled hill slopes. Additionally, because of their distinctive soil covers, two further groups of sites were included: first, regions of impure limestone in the Setul Boundary Range; second, areas of coarser-textured soils on

the foothill slopes of Selangor and the Kinta Valley. Rock cliffs were not considered since they are practically devoid of soil and vegetation cover. Alluvial plains in all three field areas, and the gently-inclined hill margin zones of the North-west region (as studied by Joseph, 1965), were also excluded on the grounds that they have been disturbed so much by mining or agricultural practices that their soil and vegetation covers are too diverse to be encompassed meaningfully by the present study.

The choice of field areas was limited by the inaccessibility of many hills. Ak. Bt. Takun (Selangor), G. Layang Layang and the south-eastern ridge of G. Tempurong (both in the Kinta Valley) may be scaled without ropes. Access to the higher slopes of other tower karst hills is virtually impossible. The Chuping and Kodiang outcrops were thus excluded and studies in the North-west region were confined to the Setul Boundary Range.

Thirteen study areas were selected and a sampling plot (table 29) was located randomly within each. Throughfall and litterfall were monitored throughout the year at six of the more accessible plots. Of these, the two on Ak. Bt. Takun (ABT-I and ABT-II) and one on G. Tempurong (GT-I) are on the moderately-inclined foothill slopes. The second site on G. Tempurong (GT-II) and two sites in the Boundary Range (WT-I and GKV-I) are typical of moderately-inclined hill slopes, the first two being on relatively pure limestones and the last on impure limestone. The remaining seven plots include steep hill slopes (GT-III, WT-II and GKV-II) and hilltop sites (GT-IV and GLL) in the Kinta Valley and North-west region which could not be included in the regular sampling programme. The remaining plots (ABT-III and ABT-IV) are located near the summit of Ak. Bt. Takun, the field area used in investigating weathering processes on bare and thinly-covered outcrops (Chapter V).

2. Sampling within each plot

Ideally all of the plots would have been of the same size (30 x 30 m square) and slope profiles, soil depths and the proportions of bare rock, rock covered by organic litter, and mineral soil would have been measured along three 30 m transect lines. However, in more than half the cases (table 30) local circumstances and practical considerations limited the size of the plot. For example, the soils on the summit of G. Layang Layang were so deep that only one 20 m transect was completed, and a 30 x 20 m plot established, in a full day's fieldwork (the time allocated to that particular site). Each plot was divided into sub-plots, 10 x 10 m square. The number of trees and their girths at chest height were recorded for each sub-plot. Only trees exceeding 15 cm girth were included (cf. Kira et al., 1967, who use a limit of 4.5 cm diameter at chest height). Usually one soil profile from the central sub-plot was investigated. Since many of the soils lack clearly differentiated horizons, samples were taken at 10, 20 and 30 cm, and at three or four further depths below 30 cm. Where only shallow soils were present, samples were taken at between 2 and 7 cm from at least two different locations.

Throughfall and litterfall measurements were confined to the central 10 x 10 m sub-plot within each of the six principal sampling plots. Previous workers have used different numbers of throughfall collection devices (eg. 3 - Kenworthy, 1971; 4 - Hopkins, 1960; 6 - Low, 1974, and Nye, 1961; 13 - Manokaran, 1978). In the present study, four collecting bottles were installed. Volumes of throughfall were measured at two- to three-week intervals. On each occasion the collecting bottles were relocated, thereby preventing bias in the sample (Kenworthy, 1971). Four litter trays were used in each plot. This is fewer than the number recommended by Bullock (1978), but consistent with other studies (eg. 4 - Kenworthy, 1971;

6 - Nye, 1961). Litter was collected every two to three weeks. Over this period of time the litter loses only small amounts of nutrients (Nye, 1961).

C. DESCRIPTION OF FIELD AREAS

The thirteen plots studied fall into five topographic groups: rocky hilltops, hilltop depressions, steep hill slopes, moderately-inclined hill slopes and moderately-inclined foothill slopes (table 29). The main characteristics of the plots are summarized in tables 30 to 33 and figures 18 to 25.

1. Rocky hilltops

Rocky hilltops in the karst regions take two main forms: first, broad areas of pinnacle-like terrain (eg. ABT-IV); second, long narrow ridge-crests (eg. GT-IV). A third form, comprised mostly of accumulations of rockfall debris, is also recognized (eg. ABT-III).

1.1. GT-IV

1.1.i. Relief and surface cover. GT-IV spans a 30 m long section of the southernmost spur of G. Tempurong, at a height of 400 m (plate 23). Three transect lines, 20 m in length, were established across the ridge, each centred on the ridge-crest. The part of the ridge studied is asymmetrical, with the degree of asymmetry increasing from the northern to the southern transect (fig. 18). The gradients of the two eastern slopes average 53.8 and 63.0^{o}, respectively, while the corresponding figures for the western slopes are 47.2 and 31.7^{o}. Rock exposures are common (fig. 22, plate 30), especially on the steeper, eastern slopes (80 per cent bare rock). The outcrops mainly comprise flat facets, with pronounced angular junctions. Everywhere, except beneath overhangs, the rock surfaces are smooth. Rounded solution forms occur, but are limited both in their number and their degree of development. The forms of the outcrops suggest that they are joint controlled, and that rockfall

rather than chemical weathering is the main denudational process operative. Although very little rockfall debris was found within the plot, in several places joints were being forced open by the roots of trees and shrubs (plate 30). 36.5 per cent of the plot is soil covered (fig. 22). The deepest and most continuous cover occurs on the gentler slopes of the southwestern part. For instance, the western half of the southern transect has a complete soil cover, with a maximum depth of 60 cm. Elsewhere, mineral soils are shallow and are mostly restricted to isolated crevices and the occasional gently-graded rock facet.

1.1.ii. Description of soils. One 28 cm profile and three shallow soils, less than 10 cm deep, were investigated (table 34). The deepest soil varies in colour from very dark grey (5 YR 3/1) at 10 cm, to dark reddish brown (5 YR 3/2) at 20 cm. It is a clay soil, with a moderately-developed granular structure. Soil reaction is slightly acid near the surface (pH, 6.35 at 10 cm), but is neutral at depth (pH, 6.90 at 20 cm). At 10 cm, the organic carbon content (4.73 per cent) and C.E.C. (34.76 m.e./100 g) are high, but decrease to 2.20 per cent and 19.61 m.e./100 g, respectively, at 20 cm. Calcium and magnesium are the dominant exchangeable cations, the molar ratio Mg:Ca averaging 0.217.

The shallow soils vary from clay to clay loam in texture and from neutral to slightly acid in reaction (pH range, 6.15 to 7.05). They have high organic carbon contents (mean, 6.17 per cent) and C.E.C. (mean, 36.02 m.e./100 g). In shallow soil 3, which has the highest pH, exchangeable calcium is 25.73 m.e./100 g. The carbonate content of all four soils is low, reaching a maximum of 0.5 per cent.

1.1.iii. Vegetation. There are 12 trees in the plot (200 ha^{-1}), the largest being 74 cm girth. The cross-sectional area of trunks at chest height (i.e. 'basal area' index - Whitmore, 1975, p. 90) is

3.3 m^2/ha. Boea spp., Chirita spp., Diospyros, Neolitsia, Vitex (identification by staff of the School of Biological Sciences, Universiti Malaya) and a range of other shrubs occur in the crevices of rock outcrops. For the most part, however, exposed rock surfaces are devoid of plant cover (plate 30).

1.2. ABT-IV

1.2.i. Relief and surface cover. ABT-IV is located in an area of deeply-dissected terrain near the summit of Ak. Bt. Takun. Its relief is dominated by pinnacle-like residuals. The transect line through the plot (fig. 18) poorly depicts the magnitude of the meso-topography since, at the time of the study, the chasm between the two pinnacles shown was blocked by a boulder which had fallen from the overhanging section of the higher pinnacle. The rock walls of the chasm actually extend precipitously to a depth of 7 m or more below this boulder. Although there is clear evidence of rockfall within the plot, most of the rock surfaces are deeply grooved by solution forms. Superficial solution appears, therefore, to be the dominant denudational process. (A more detailed description of these forms, and discussion of the solution process, is presented in Chapter V). The rock pinnacles are mostly bare of soil (plates 14 and 15), with 83.3 per cent of the surface being of bare rock (table 30). The remaining 16.7 per cent comprises accumulations of leaf litter (7.4 per cent) and shallow mineral soils (9.3 per cent). Such accumulations are mostly confined to the lower parts of the deeper solution runnels.

1.2.ii. Description of soils. Two shallow soils, both less than 10 cm deep, were sampled (table 35). The first is a clay soil of neutral reaction (pH, 7.20), while the second is slightly acid (pH, 6.25), silty clay soil. The former contains fragments of snail shells which are composed almost entirely of calcium carbonate (Chapter V, Section D2.2.i). Visible fragments were removed before

analyses were undertaken. Both soils are black (10 YR 2.5/1) in colour and have high organic carbon contents (11.37 and 10.69 per cent, respectively) and C.E.C. (46.21 and 48.90 m.e./100 g, respectively). Calcium and magnesium form the major proportion of the exchangeable cations. For example, in the first soil, which has a base saturation of 80.61 per cent, exchangeable calcium and magnesium are 31.97 and 4.80 m.e./100 g respectively, and exchangeable potassium and sodium together account for only 0.48 m.e./100 g. The average molar ratio Mg:Ca of exchangeable cations of the two soils is 0.177, which compares with a value of 0.014 for the underlying bedrock (table 31). Carbonates are present in both soils (maximum, 5.0 per cent).

1.2.iii. Vegetation. Seven trees occur within the plot, the largest having a girth of 45 cm. The density of trees (233 ha^{-1}) is higher than in GT-IV, but the basal area (2.2 m^2/ha) is much lower. Cremnophyte species are present, but less frequent than in GT-IV.

1.3. ABT-III

1.3.i. Relief and surface cover. ABT-III (153 m) lies adjacent to ABT-IV and is at the foot of a 15 m high cliff. The plot is littered with large angular blocks of limestone, up to 3 m square, which have fallen from the cliff. One such block occurs at the lower end of the transect line (fig. 18). Rock surfaces are exposed over 60.6 per cent of the plot (table 30) and are invariably grooved by karren forms (Chapter V). The meso-relief of ABT-III is more subdued than that of ABT-IV. Almost one-third (31.3 per cent) of the surface is covered by mineral soil (fig. 22) and more than half of this is deeper than 5 cm (maximum, 27 cm). The fact that many of the larger blocks have soil covers suggests that they have been *in situ* for a long period of time.

1.3.ii. Description of soils. Three shallow soils were analysed (table 35). They are black (10 YR 2.5/1) to very dark brown

(10 YR 2.5/2), clay to silty clay soils, which range from slightly acid to neutral in reaction (pH range, 6.10 to 7.15). The third sample, the most alkaline soil, contained small pieces of calcareous snail shells. Overall, the organic carbon content and C.E.C. are high, averaging 6.38 per cent and 34.32 m.e./100 g, respectively. Calcium and magnesium are the dominant exchangeable cations, the molar ratio Mg:Ca for the three soils averaging 0.249 (cf. bedrock, 0.014). Carbonates occur in all three soils, reaching a maximum of 5.0 per cent.

1.3.iii. Vegetation. ABT-III has a much denser tree cover than the other rocky hilltop sites. There are 43 trees within the 30 x 30 m plot (478 ha^{-1}). Of these, eight (89 ha^{-1}) exceed 100 cm girth, the largest being 230 cm. The basal area of trunks is 26.0 m^2/ha.

2. Hilltop depressions

Deep accumulations of residual soils occur in isolated pockets on many hilltops in the field areas. One example, plot GLL, is described below.

2.1. GLL

2.1.i. Relief and surface cover. GLL extends across approximately one-half of a depression on the summit of G. Layang Layang (290 m), Kinta Valley. The depression is roughly circular in plan (diameter, 40 m) and rises towards the rim at an angle of approximately 15°. Bedrock is exposed in places around the outer edge of the depression (fig. 19), but such exposures make up only 0.3 per cent of the area of GLL (table 30). The depression itself has an unbroken, soil-mantled surface. The transect line investigated (fig. 22), which is aligned along one radius, illustrates the irregular form of the underlying limestone. Soil depth increases towards the centre, reaching a maximum depth of 240 cm. The pocket in the limestone surface is drained internally. Clearly, in order for such a depth of

soil to develop on a pure limestone (acid insoluble residue, 0.40 per cent) at a hilltop location, joints in the bedrock must either be blocked by sediment, or be too narrow to allow the removal of soil material.

2.1.ii. Soil profile description. One profile, 145 cm in depth, was investigated (analytical data, table 36). It is a dark reddish brown (2.5 YR 3/4) clay soil, with a marked textural profile. The clay content increases with depth from 10 cm (60.7 per cent) to 70 cm (78.2 per cent), and then decreases (140 cm, 69.3 per cent). Gravel, which is composed entirely of small rounded lateritic concretions, also reaches a peak at 70 cm (8.7 per cent). The soil has a friable crumb structure. This is reflected in the low bulk density of the soil, which ranges from 0.71 g/cm^3 at 10 cm to 1.14 g/cm^3 at 140 cm, and in the large volume of pore space which ranges from 61.0 to 74.6 per cent.

Organic carbon decreases regularly down the profile from 1.34 per cent at 10 cm to 0.08 per cent at 140 cm. Base saturation and pH increase with depth. For instance, at 10 cm the soil is moderately acid in reaction (pH, 5.20) while at 140 cm it is only slightly acid (pH, 6.50). Calcium and magnesium are the main exchangeable cations. The molar ratio Mg:Ca is high throughout the profile (mean, 0.333), and decreases progressively with depth from 0.590 at 10 cm to 0.190 at 140 cm. These figures compare with 0.088 for the underlying rock. Levels of exchangeable potassium and sodium are extremely low. Potassium decreases from 0.05 m.e./100 g at 10 cm to 0.02 m.e./100 g at 50 cm. Below 50 cm, exchangeable potassium is less than 0.01 m.e./100 g. The highest sodium value is 0.01 m.e./100 g. Carbonates occur in small amounts at 20 and 30 cm (0.5 per cent), but are virtually absent in the remainder of the profile.

2.1.iii. Vegetation. There are 107 trees within the 30 x 20 m plot, the highest density (1783 ha^{-1}) of the thirteen plots. The trees

are all small (largest, 90 cm girth), with 93 being less than 40 cm girth. The absence of larger trees, which is reflected in the relatively small basal area of the trunks (17.5 m^2/ha), is not easily explained. Although rather spindly in form, the trees appear to be extremely healthy, and their size is certainly not limited by edaphic conditions. The trees of the summit appear, therefore, to be all of similar age - a fact which may be explained by the wholesale destruction of an earlier tree cover, and subsequent regrowth. Since the summit is virtually inaccessible, it is inconceivable that it has been logged. Fire, probably caused by lightning, provides the most likely explanation. While there is no direct evidence in GLL of such an occurrence, fires of this type are quite common on the limestone outcrops of the Kinta Valley (T.M. Lee, personal communication).

3. Steep hill slopes (45 to 60°)

Three plots of similar gradient (range, 45.3 to 46.9°) were investigated. The first two, GT-III and WT-II, are on relatively pure limestones, while the third, GKV-II, is located in an area of impure limestone.

3.1. GT-III

3.1.i. Relief and surface cover. GT-III (287 m) is located on the eastern flank of the southern spur of G. Tempurong, 113 m below GT-IV (plate 23). Massive outcrops of limestone occur regularly along the slope transect (fig. 19). They appear to be mostly of solid bedrock rather than rockfall debris, and their surfaces are almost invariably etched by solutional forms. The outcrops are broken by a series of near-horizontal ledges, spaced approximately 8 to 10 m apart, which probably reflect the joint pattern of the bedrock. 50.5 per cent of the surface is soil-covered, the maximum depth being 65 cm (fig. 22).

3.1.ii. Description of soils. One soil profile, 23 cm deep, and

three shallow soils were analysed (table 37). The deepest soil is a dark reddish brown (2.5 YR 3/4), clay to clay loam, with a strongly-developed crumb structure. The bulk density is low (mean, 0.68g/cm^3), while the porosity is correspondingly high (mean, 76.0 per cent). Both organic carbon and C.E.C. are highest near the surface, the values being 2.60 per cent and 25.86 m.e./100 g, respectively, at 10 cm. pH is 6.70 at both sampling depths.

The three shallow soils range from very dusky red (2.5 YR 2.5/2) to dark reddish brown (2.5 YR 3/2) in colour, and from clay to clay loam in texture. These soils have higher concentrations of organic carbon (mean, 7.01 per cent), and a higher C.E.C. (mean, 35.16 m.e./100 g) than the deeper soil. They are neutral to slightly alkaline in reaction (pH range, 7.35 to 7.55), with base saturation levels in excess of 79 per cent. Calcium and magnesium are the dominant cations in all four soils, averaging 18.29 and 3.97 m.e./100g, respectively. Carbonates are present, but nowhere exceed 2.0 per cent.

3.1.iii. Vegetation. GT-III has an almost continuous tree canopy. The trees are mostly small and stunted, although two trees, located in deeper pockets of soil, are greater than 100 cm girth (maximum, 120 cm). Overall, the density of trees is 800 ha^{-1} and the basal area is 16.4 m^2/ha.

3.2. WT-II

3.2.i. Relief and surface cover. WT-II (244 m) is situated beneath a rocky ridge at the south-eastern corner of Wang Tangga. In contrast with GT-III, the profile of the rock surface is very irregular (fig. 19). Deep pockets of mineral soil occur within the plot (maximum depth, 145 cm). These are probably formed by preferential weathering along bedding planes and joints, which are more common in the Setul Limestone than in the other karst rocks. More usually, however, the soils are shallow (fig. 22). Thus, while 50.6 per cent

of the surface is soil-covered, less than 3 per cent of the soil depths recorded exceeded 60 cm.

3.2.ii. Description of soils. One profile, 28 cm deep, and three shallow soils were investigated (table 38). The former is a dark reddish brown (5 YR 3/3) clay soil, with a moderately-developed crumb structure. Bulk density is low (mean, 0.90 g/cm^3), and the volume of pore space is correspondingly high (mean, 67.5 per cent). The soil is neutral to alkaline in reaction, with pH increasing from 7.15 at 10 cm to 7.60 at 20 cm. At the latter depth base saturation is 90.0 per cent. The carbonate content is 0.5 per cent at both sampling depths.

The three shallow soils range from black (5 YR 2.5/1) to very dark grey (5 YR 3/1) in colour. Like the deeper profile investigated, they are neutral to alkaline, clay soils with a moderately-developed crumb structure. However, they differ in having consistently-higher organic carbon contents (mean, 4.14 per cent).

3.2.iii. Vegetation. The total number of trees in WT-II (1033 ha^{-1}) and the number greater than 100 cm girth (100 ha^{-1}) are both considerably higher than in GT-II, even though both plots are of similar gradient and have similar amounts and types of soil cover. The density of undergrowth is also much greater, with *Strebilus illicoides*, a small thorny shrub, being common (identification by staff of School of Biological Sciences, Universiti Malaya).

3.3. GKV-II

3.3.i. Relief and surface cover. GKV-II (327 m) lies within 50 m of the rocky ridge-top which forms the western watershed of the Gelang Kerbau Valley. As in WT-II, the rock surface is irregular with sporadic pockets of deeper soil. The bedrock is less pure (acid insoluble residue, 6.40 per cent) than in the other steep hill slope plots (GT-III and WT-II, table 31). This difference is not reflected in the area of soil-covered surface or in the maximum soil

depth, which in GKV-II are 53.1 per cent and 60.0 cm, respectively. However, it may account for the relatively small proportion of GKV-II (3.1 per cent) which is covered by shallow soils of less than 5 cm (fig. 22). Corresponding figures for GT-III and WT-II are 11.8 and 13.9 per cent.

3.3.ii. Soil profile description. One profile, 32 cm deep, was investigated (table 39). It is a brown (7.5 YR 5/2) to strong brown (7.5 YR 5/6), clay soil, with weakly-developed granular peds in the surface horizons, and a moderately-developed blocky structure at depth. Bulk density is high (mean, 1.20g/cm^3) and the volume of pore space relatively small (mean, 55.0 per cent). The gravel fraction of the soil is composed of rounded lateritic concretions and fragments of partially-weathered bedrock impurity (metamorphosed argillaceous sediments). Gravels occur at all sampling depths but are most frequent near the base of the profile (2.1 per cent at 30 cm). The soil is neutral in reaction (pH range, 7.0 to 7.35), with calcium and magnesium forming the main exchangeable cations. The Mg:Ca and K:Na molar ratios both decrease with depth, the first from 0.342 (10 cm) to 0.251 (30 cm), the second from 3.80 (10 cm) to 2.667 (30 cm). Only small amounts of carbonate (0.5 per cent) are present at each sampling depth.

3.3.iii. Vegetation. The tree cover in GKV-II has a similar density to that in WT-II, the total number and number greater than 100 cm girth being 1167 ha^{-1} and 100 ha^{-1}, respectively. GKV-II differs, however, in having very little undergrowth.

4. Moderately-inclined hill slopes (30 to 45°)

Two of the plots studied, GT-II and WT-I, are on relatively pure limestones, which have acid insoluble residue contents of 0.38 and 2.35 per cent, respectively. The third, GKV-I is sited in an area of impure limestones (mean acid insoluble residue, 9.65 per cent).

4.1. GT-II

4.1.i. Relief and surface cover. GT-II lies immediately below GT-III at an altitude of 196 m. In contrast with the latter plot, parts of the surface of GT-II are littered with fallen blocks. One such block is shown in figure 20. Others may also be present along the upper part of this transect line, but this was not confirmed. The blocks may be of immediately local origin, being formed by undermining and surface collapse, or may be derived from further upslope, possibly from the ridge-crest (GT-IV). The rock profile in the lower part of the transect (fig. 20) contains ledges similar to those of GT-III. Soil forms a continuous cover in this section of the transect. Overall, 59.1 per cent of the surface is soil covered (maximum depth, 155 cm) with bare rock outcrops and litter-covered rock surfaces accounting for 34.0 and 6.9 per cent, respectively.

4.1.ii. Soil profile description. The profile studied is 72 cm deep (table 40). It is a dark reddish brown (2.5 YR 3/4) clay soil with a loose, friable crumb structure. The clay content increases with depth to 40 cm (60.3 per cent), and then decreases. Gravel-sized particles, which are composed entirely of lateritic concretions, occur at most sampling depths, but are most frequent at 10 and 50 cm. Crumb structures are strongly developed, especially in the surface horizons, and the soil has a low bulk density (range, 0.67 to 0.87 g/cm^3). Total pore space is high, ranging from 70.0 per cent at the base of the profile to 76.3 per cent near the surface.

The upper horizons are comparatively rich in organic matter, the organic carbon content being 2.16 per cent at 10 cm. This probably accounts for the slightly darker colour (2.5 YR 3/2) and the higher C.E.C. (17.50 m.e./100g) of the soil at this depth. Below 10 cm, the C.E.C. ranges from 5.84 to 8.32 m.e./100 g. The soil is neutral to slightly acid in reaction. Highest pH values occur near the surface (6.75 at 10 cm) and at the base of the profile

(6.55 at 7.0 cm). The carbonate content ranges from 0.1 to 1.0 per cent, being highest at 50 and 70 cm.

4.1.iii. Vegetation. The frequency of trees in GT-II, 656 ha^{-1}, is less than in the plots on the steeper slopes, and the number of trees greater than 100 cm girth is also relatively small (56 ha^{-1}). However, the 30 x 30 m plot supports several very large trees, three of which exceed 200 cm girth (maximum, 300 cm). Clearly, therefore, soil conditions in GT-II do not severely limit tree growth (cf. GT-III, above).

4.2. WT-I

4.2.i. Relief and surface cover. WT-I is located at a height of 180 m on the southern slopes of Wang Tangga. It lies above a precipitous cliff, 6 m in height, the top of which is shown in the transect line (fig. 20). The limestone surface, which is exposed sporadically between pockets of mineral soil, is highly irregular in form. Typically, soil pockets are between 4 and 10 m in diameter, and 1 m or more in depth (maximum depth, 135 cm). Compared with the adjacent plot, WT-II, which is steeper and at a higher elevation, a greater proportion of WT-I is soil-covered (58.6 compared with 51.6 per cent), and bare rock outcrops are correspondingly less frequent (29.4 compared with 40.7 per cent).

4.2.ii. Soil profile description. One profile, 79 cm deep, was investigated (table 41). It is a dark reddish brown (2.5 YR 3/4) clay soil, with a well-developed crumb structure. The clay content increases with depth from 10 cm (56.2 per cent) to 40 cm (75.5 per cent), and then falls to below 72 per cent at 60 and 70 cm. In contrast, the sand fraction is highest near the surface (18.1 per cent at 10 cm). Gravels, which are composed entirely of lateritic concretions, are confined mostly to the 30 and 40 cm sampling depths. The bulk density is low, ranging from 0.72 g/cm^3 at 10 cm to 0.94 g/cm^3 at 70 cm.

Organic matter is fairly evenly distributed down the profile, the carbon content decreasing from 1.31 per cent at 10 cm to 0.84 per cent at 70 cm. The C.E.C. also remains relatively constant down the profile (range, 14.44 to 16.87 m.e./100 g). In contrast, pH and exchangeable bases show a marked trend. Both decrease through the upper 50 cm of the soil, then increase towards the base of the profile. The soil is slightly acid in reaction (pH range, 5.70 to 6.20). As in the underlying limestone (table 31) the molar ratio Mg:Ca is high (range, 0.470 to 0.668 m.e./100g). The carbonate content is 0.5 per cent from 10 to 60 cm depth, but increases to 1.0 per cent at 70 cm.

4.2.iii. Vegetation. The density of trees in WT-I averages 1050 ha^{-1} and of these the equivalent of 100 ha^{-1} are greater than 100 cm girth (maximum, 240 cm). As in WT-II, the undergrowth is quite dense, with Strebilus illicoides again being common.

4.3. GKV-I

4.3.i. Relief and surface cover. GKV-I is situated on a small spur on the western side of the Gelang Kerbau Valley at an altitude of 189 m. In contrast with the two previous plots, the slopes of GKV-I have an almost complete soil mantle (93.9 per cent). The maximum soil depth recorded was 138 cm. Comparisons may be made between GKV-I and WT-I under fairly precisely controlled conditions since they lie within the same climatic region, are of similar gradient (32.1 and 35.2^{o}) and altitude, and are located in similar topographic positions. The average soil depth in the former plot is 43.6 cm (modal group, 60 to 80 cm; fig. 22), which is more than double that of the latter plot (mean of 21.1 cm; modal group, 0 to 20 cm). This difference may be attributed largely to the contrasted purity of the limestone in the two areas. A further consequence of the high proportion of impurity in the limestones of the Gelang Kerbau Valley appears to be that the bedrock becomes

weakened mechanically during chemical weathering. As a result, cave development is inhibited (no penetrable cave passages occur within the vicinity of the plot) and it seems likely that solution-widened joints are subject to collapse or become blocked with clay-rich fill.

4.3.ii. Soil profile description. The profile investigated in GKV-I is a strong brown (7.5 YR 5/6 to 5/8) clay soil, 102 cm in depth (table 42). Clay increases from 59.3 per cent at 10 cm to a maximum of 73.1 per cent at 30 cm. Highest concentrations of sand and gravel occur at 70 cm (12.2 and 11.7 per cent, respectively), with a lesser peak at 10 cm. These fractions are composed of small lateritic nodules. Peds are only weakly-developed and consequently, the bulk density is relatively high (range, 1.14 to 1.33 g/cm^3).

Organic matter occurs throughout the profile, with organic carbon ranging from 0.80 per cent at 10 cm to 0.07 per cent at 100 cm. The soil is slightly acid in reaction. pH increases slightly down the profile from an average value of 5.45 in the upper horizons (10 to 50 cm) to 5.60 at 70 and 100 cm. While calcium (mean, 1.83 m.e./100 g) and magnesium (mean, 0.80 m.e./100 g) are the main exchangeable cations, the amount of exchangeable potassium (mean, 0.208 m.e./100 g) is also relatively high. The carbonate content is 0.5 per cent at all depths.

4.3.iii. Vegetation. GKV-I is located in the Bukit Bintang Forest Reserve, an area which is exploited commercially for timber. There was no evidence of felling within the plot studied. The following tree species are abundant in the vicinity of GKV-I: Annoneae spp., Calophylum incrassatum, Cynometra spp., Hopea ferrea, Intsia palembanica, Morinda citrifolia, Randia scortechini, Salmalia valetonii and Swintonia schwenki (identification by R.Husin, Department of Forestry, Perlis). The overall density of trees in the plot (1167 ha^{-1}) is quite high, but the most significant feature

is the number of trees (233.0 ha^{-1}) greater than 100 cm girth, the largest tree being 305 cm. This is reflected in the high basal area (70.2 m^2/ha).

5. Moderately-inclined foothill slopes (Selangor and Kinta Valley)

Three plots (ABT-I, ABT-II and GT-I) were investigated. Each is of similar altitude (range, 113 to 124 m) and gradient (range, 32.8 to 36.6^{o}), but they differ considerably in their topographic location.

5.1. ABT-I

5.1.i. Relief and surface cover. ABT-I lies on the southern side of Ak. Bt. Takun, east of the towering cliff which surmounts the entrance of Gua Anak Takun (plate 12). It is approximately 30 m above the surrounding plain. The soil mantle is almost complete, covering 84 per cent of the surface, and locally exceeding 100 cm in depth (fig. 21). The highly irregular profile of the rock surface and the form of the few rock exposures suggest that the slope is predominantly one of rockfall accumulations. Further evidence of this is seen on steeper slopes below ABT-I, where angular blocks of limestone are exposed over much of the surface.

5.1.ii. Soil profile description. The soil profile studied is 75 cm deep (table 43). It is dark brown (10 YR 3/3) in the upper horizons (10 and 20 cm), becoming dark yellowish brown (10 YR 3/4) at depth. Though essentially a clay soil, it contains relatively high proportions of silt and sand. The silt fraction, which averages 32.3 per cent, exhibits no clear distribution pattern. In contrast, the clay fraction increases with depth, while the sand fraction decreases. The latter is composed mostly of quartz grains. Moderately-developed granular peds occur near the surface, but the lower horizons contain only weakly-developed blocky structures. Bulk density ranges from 0.97 g/cm^3 at 10 cm to 1.26 g/cm^3 at 70 cm.

Organic carbon (mean, 0.44 per cent) decreases from 0.85

per cent at 10 cm to 0.42 per cent at 30 cm, before levelling off at an average of 0.31 per cent (40 to 70 cm). The soil is neutral to slightly acid in reaction. pH falls from 6.65 at 10 cm to 6.10 at 50 cm and then rises to 6.50 at 70 cm. Calcium (mean, 8.96 m.e./100 g) and magnesium (mean, 0.53 m.e./100 g) form the bulk of the exchangeable cations. The molar ratio Mg:Ca (mean, 0.059) is lower than in the other soils sampled, and decreases with depth from 0.131 at 10 cm to 0.026 at 70 cm. Levels of exchangeable potassium (mean, 0.04 m.e./100 g) and sodium (mean, 0.013 m.e./100 g) are both low. Carbonate concentrations are very low through most of the profile (0.1 to 0.5 per cent) but increase to 1.0 per cent at 70 cm.

5.1.iii. Vegetation. The density of trees in ABT-I (556 ha^{-1}) is less than in any of the hill slope plots described above. Trees exceeding 100 cm girth are also few in number (56 ha^{-1}) and the basal area of trunks (22.2 m^2/ha) is also small. Although there is no visible evidence of logging, it is likely that some trees will have been cut by local tin miners. Since the soil is relatively deep and base content of the soil is high, human activity, rather than edaphic factors, is probably the main cause of the relatively sparse tree cover.

5.2. ABT-II

5.2.i. Relief and surface cover. ABT-II has the same orientation as ABT-I, but lies to the west of the cave entrance, at about 20 m above the plain (plate 12). Whilst the extent of the soil cover (81.0 per cent) is approximately the same as in the latter plot, it is generally shallower, and nowhere exceeds 100 cm in depth. Detached blocks of limestone occur in the lower half of the plot (fig. 21), but rock exposures in the upper part may be outcrops of solid bedrock. The plot is located on the floor of a corridor, which extends almost to the top of the hill, and therefore receives runoff from the surrounding slopes. Its position also corresponds roughly with that

of the band of metamorphosed shale observed in the underlying cave system (Chapter VII, Section B1.2.i).

5.2.ii. Soil profile description. The soil profile studied (table 44) is 68 cm deep. It is very dark grey (10 YR 3/1) in colour at all depths, and is the darkest of the profiles investigated. In texture it varies from a clay to a clay loam. Generally, the proportion of clay increases with depth, while that of sand decreases. The latter fraction comprises mainly quartz, with no lateritic concretions. Near the surface the soil has moderately-developed granular peds, but contains only weak, blocky structures at depth. Bulk density ranges from 0.93 (10 cm) to 1.28 g/cm^3 (50 cm).

The organic carbon content (mean, 0.94 per cent) is more than double that of the profile in ABT-I and remains high down the profile, ranging from 1.40 per cent (10 cm) to 0.75 per cent (60 cm). This may, in part, explain the dark colour of the soil. The C.E.C. is also relatively high, being 27.82 m.e./100 g at 10 cm (cf. 14.76 m.e./100 g in ABT-I), and averaging 22.53 m.e./100 g through the rest of the profile. The soil is neutral in reaction (mean pH, 6.77) with an average base saturation of 74.5 per cent. The carbonate content is low throughout the profile (range, 0.5 to 1.0 per cent).

5.2.iii. Vegetation. Tree density (600 ha^{-1}) is about the same as in ABT-I, but the number of trees greater than 100 cm girth (17 ha^{-1}), and the basal area of trunks (9.5 m^2/ha), is much lower. Selective felling of the larger trees is the most likely explanation of the relatively poor tree cover, but edaphic factors may also be limiting at this site (Chapter VI, Section C4).

5.3. GT-I

5.3.i. Relief and surface cover. GT-I (124 m) is situated on a small spur, 52 m below GT-II and approximately 60 m above the alluvial plain. Outcrops of limestone, which account for 38.7 per cent of the surface (table 30), appear to be mostly of local bedrock since the rock profile in the lower half of the transect (fig. 21)

is fairly regular in form (cf. GT-III, fig. 20). Large fissures, in some instances a metre or more in width, were observed in certain bedrock exposures. These occur mainly at the edges of the spur, and are suggestive of surface collapse. The openness of the karst surface and the topographic location of the plot may explain the less complete soil cover (61.3 per cent).

5.3.ii. Description of soil profile. The profile investigated is 77 cm deep (table 45). Its colour varies from dark brown (10 YR 3/3) at 10 cm to dark yellowish brown (10 YR 4/4) at depth. The texture ranges from sandy clay (10 cm), through clay loam (20 cm), to clay (30 to 70 cm). The sand content is generally high, especially in the surface horizons (49.2 per cent at 10 cm). Gravel also occurs at this depth, but is relatively unimportant lower down the profile. These coarser fractions appear to be composed mainly of quartz. Clays increase progressively down the profile (maximum, 49.6 per cent at 60 cm). The soil has a moderately-developed, fine, sub-angular, blocky structure near the surface, but is more massive at depth. Bulk density increases from 1.06 g/cm^3 at 10 cm to 1.65g/cm^3 at 70 cm.

Organic matter is mostly confined to the upper 20 cm of the soil, with the organic carbon content ranging from 0.84 per cent at 10 cm to 0.16 per cent at 70 cm. C.E.C. varies between 15.53 and 19.38 m.e./100 g, but displays no clear distribution pattern. The soil is neutral to slightly acid in reaction. pH decreases regularly from 6.85 at 10 cm to 6.10 at 50 cm, and then increases to 6.50 at 70 cm. Carbonates increase with depth from 0.1 per cent at 10 and 20 cm, to 1.0 per cent at below 30 cm.

5.3.iii. Vegetation. Although the number of trees in GT-II (456 ha^{-1}) is less than in the foothill plots on Ak. Bt. Takun, the frequency of large trees (111 ha^{-1}) greater than 100 cm girth is higher. Within the 30 x 30 m plot, eight have girths exceeding 200 cm, with two of 420 cm. According to local tin miners, tree

felling in the area has been confined mostly to the lower foothill slopes below GT-I (L.T. Sin, Sg. Siput Mines Ltd., pers. comm.). Thus GT-I may be taken as representative of relatively-undisturbed foothill slopes. The tree density and basal area of trunks (63.2 m^2/ha) in the plot therefore provide some indication of the tree covers which formerly may have been present in ABT-I and ABT-II.

D. DISCUSSION OF SOIL ANALYTICAL DATA

1. Particle-size distribution

1.1. Gravel

In two of the profiles investigated, GLL and GKV-I, the maximum gravel content exceeds 5 per cent by weight (fig. 23). The highest concentration (12.2 per cent) occurs at 70 cm in GKV-I. In the latter soil, gravels are confined mostly to the lower half of the profile and are composed entirely of lateritic concretions. While the limestones of GKV-I contain a high proportion of acid insoluble residue, 9.65 per cent, iron (expressed as Fe_2O_3) accounts for only 4.6 per cent of this impurity (table 31). The total iron content of the soil is also small, averaging 6.4 per cent through the profile (table 33). Since the iron-rich nodules occur within a relatively iron-poor clay matrix, conditions in the soil clearly favour iron translocation. As noted previously, the subsoil is comparatively dense (bulk density, 1.33 g/cm^3 at 100 cm) and joints in the bedrock are plugged by clay residues. Consequently, drainage in this plot is relatively impeded. During storms the lower part of the profile may become saturated, thus creating reducing conditions under which iron, in the ferrous form, may be mobilized.
At such times, iron, in the ferric form, will be precipitated at the surface of the saturated zone and along free-draining root cavities. Similar accumulations of lateritic gravels occur near the

top of the watertable elsewhere in the Malay Peninsula (eg. Pasoh F.R., C.H.Leigh, pers. comm.). In GKV-II, the second plot on the impure limestones of the Gelang Kerbau Valley, lateritic gravels occur at all depths and reach a maximum of 2.1 per cent at 30 cm, the deepest sampling depth. The clay matrix contains only 4.8 per cent iron and, as in GKV-I, translocation most probably occurs under saturated conditions.

In GLL, as in GKV-I, lateritic gravels are concentrated between 50 and 100 cm (maximum, 8.7 per cent). However, the circumstances under which they have formed differ fundamentally in four respects. First, the bedrock in GLL is very pure, comprising 0.40 per cent acid insoluble residue, of which 30.0 per cent is iron (Fe, expressed as Fe_2O_3). Secondly, the iron content of the fine-earth fraction is high, averaging 14.8 per cent down the profile (table 33). Thirdly, because of its well-developed granular structure the soil drains freely and shows little evidence of waterlogging. Finally, the soil lies near the centre of an enclosed depression. Soil development is therefore unlikely to have been affected substantially by erosion or deposition, and the gravel accumulations within the profile are most probably the product of a long period of undisturbed pedogenesis, rather than a particularly rapid pedogenic process. In the absence of a permanent watertable, the vertical distribution of the gravels is not easily explained. In theory, reducing conditions may occur temporarily in the less well-aerated lower horizons during exceptionally prolonged or intense storms. Iron, mobilized at such times, will be reprecipitated in the most well-aerated zones of the deeper horizons. Plant roots are common between 50 and 100 cm. By taking up water from the soil, plants may be instrumental in creating localized pockets of unsaturated soil in an otherwise saturated matrix. Clearly, further field observations are required to substantiate this hypothesis. However,

some indication of the volume of water removed by vegetation under humid tropical conditions is seen in the Sg. Gombak catchment, Selangor, where diurnal variations in transpiration cause measurable fluctuations in the level of the watertable (Kenworthy, 1971).

Soils in five other plots (GT-IV, GT-III, WT-II, GT-II and WT-I) contain lateritic gravels, though nowhere do amounts exceed 3 per cent. Generally, the distribution of gravels down the profile is bi-modal, with peaks occurring near the surface and towards the base. For example, in GT-II the gravel content at 10 and 50 cm is 1.4 and 2.7 per cent, respectively, whereas no gravels are present at 20 cm. The soils of these plots are similar to those of GLL. They are derived from relatively pure limestones which contain iron-rich impurities (table 31), have high iron contents in the fine-earth fraction (range, 9.8 to 13.4 per cent), and are freely drained. However, in contrast with GLL, the soils of these plots are located on steep slopes and therefore are subject to erosion and deposition. This has two consequences. First, conditions are less favourable for marked profile development, and this is reflected in the lower concentrations of gravel at depth in these soils. Second, preferential erosion of finer soil fractions may result in the progressive accumulation of gravels at the surface, as a lag deposit, and this may explain the peak concentration at 10 cm in these soils. Similar surface accumulations occur in other soils of the Peninsula (eg. soils of Pasoh F.R. - C.H. Leigh, personal communication).

The gravel fractions of the three foothill soils (ABT-I, ABT-II and GT-I) are composed mostly of quartz and are lacking in laterite. In the first plot, gravel occurs only at 50 cm (0.6 per cent). Concentrations are higher in ABT-II and GT-I. In both cases, as in the hillslope soils discussed above, the highest values (1.5 and 1.9 per cent, respectively) occur at 10 cm. Since the limestones of the foothill slopes are very pure (table 31), it seems

unlikely that the quartz gravels are derived from the underlying bedrock. Two hypotheses may be put forward to explain the gravels. First, they may have originated from veins of quartz, outcropping up-slope of the plots. Such veins have been reported in the Kinta Limestones (Ingham and Bradford, 1960, plate X, p. 34). Also, since the granites adjacent to the Kuala Lumpur Limestones are frequently dissected by extensive quartz dykes (Gobbett and Tjia, 1973), it seems likely that similar veins are present in the limestones of Ak. Bt. Takun. Secondly, alluvium, possibly part of the Old Alluvium, may form the parent material of these foothill soils. These alternative hypotheses are considered further in later discussions.

1.2. Sand

The sand content of the soils varies greatly between the soils studied (fig. 23). Lowest concentrations occur in the North-west region, the four plots (WT-II, GKV-II, WT-I and GKV-I) having average values in the range 5.4 to 7.6 per cent. In contrast, the hilltop soils (GT-IV, ABT-IV, ABT-III and GLL) and soils of the steep, and moderately-inclined, hill slopes in the Kinta Valley (GT-III and GT-II) have average sand contents in the range 12.5 to 21.8 per cent. In all ten plots, the sand fraction is composed almost entirely of lateritic concretions, the lack of quartz grains reflecting the fineness of the impurities within the limestone. The mean sand content of these soils correlates directly with the percentage of iron in the insoluble residue of the underlying bedrock (r_s=0.852, $p<0.001$). This reflects the importance of iron in the formation of the soil fraction. The higher rainfall in Selangor and the Kinta Valley, by permitting more frequent mobilization of iron, may explain why the proportions of lateritic sands are higher in the soils of these areas than in the North-west region. However, there is no significant correlation (r_s) between the gravel and sand contents of the ten soils with lateritic gravels, and no

significant difference (randomization test - Siegel, 1956, pp. 152-56) between the gravel contents of soils in the North-west and the more southerly karst regions. It would appear, therefore, that the relationship is complex, and more soils must be investigated before firm conclusions may be reached.

The foothill soils in Selangor and the Kinta Valley contain high proportions of sand (range, 19.8 to 37.8 per cent), of which most is quartz. As with the gravel fraction of these soils, the sands are probably of colluvial or alluvial origin.

Generally, the sand content of the profiles investigated decreases with depth (fig. 26A). In six of the seven soils from which six or more samples were taken, the sand content correlates inversely with depth. In two cases, WT-I (r_s=-0.786) and ABT-I (r_s=-0.857) the correlation is statistically significant at the 95 per cent confidence level. This pattern is most likely caused indirectly by the eluviation of clays from the upper part of the profile, and their deposition in an alluvial B horizon. Support for this hypothesis is provided by the consistent inverse correlations found between the sand and clay contents of the various sampling depths. In five of the seven deep profiles the correlation coefficients (r_s) were statistically significant ($p<0.05$). The soil in GKV-I is exceptional in that below 20 cm the sand content generally increases with depth. The highest concentration (11.7 per cent) occurs at 70 cm, the depth of maximum gravel accumulation.

1.3. Silt

The shallow, organic-rich soils of the rocky hilltops have the highest mean silt contents (range, 38.9 to 43.2 per cent; fig. 23). Soils of GT-III, which also mainly comprise organic-rich soils, fall within the above range. Those of the remaining steep hill slopes (WT-II and GKV-II) and of the moderately-inclined hill (GT-II, WT-I and GKV-I) and foothill slopes (ABT-I, ABT-II and GT-I) have lower mean silt contents, varying from 23.4 to 33.3 per cent. The

lowest proportion of silt, 16.4 per cent, occurs in GLL. The general trend, therefore, is for the silt content to decrease with soil depth, the correlation between profile depth and mean silt content (r_s=-0.872, $p<0.001$) being statistically significant. The silt fractions of the shallow soils probably originate directly from the insoluble limestone residues. Over time the silts may be weathered chemically to form clays. Additionally, in the hill soils, lateriz-ation results in the formation of the coarser sand and gravel fractions.

1.4. Clay

The profile in GLL has the highest clay content (mean, 71.7 per cent). Values in the range 60 to 70 per cent were recorded in the four plots of the North-west region. Soils of the rocky hill-tops and hill slopes in Selangor and the Kinta Valley have lower mean clay contents (range, 42.3 to 47.7 per cent). The clay content of the ten hilltop and hill slope soils correlates directly with the percentage acid insoluble residue in the underlying limestone (r_s=0.600, $p<0.05$). The foothill soils studied have comparatively low clay contents (range, 33.1 to 46.9 per cent; fig. 23), this being consistent with a colluvial or alluvial origin.

In the seven deepest profiles, the clay content correlates positively with depth. The correlation coefficients are statistically significant ($p<0.05$) in the three foothill soils: ABT-I (r_s=0.893), ABT-II (r_s=0.886) and GT-I (r_s=0.786). In a number of soils the clay fraction increases with depth down most of the profile, and then decreases towards the base (fig. 26B). For example, in GLL, clays increase progressively from 60.7 per cent at 10 cm to 78.2 per cent at 70 cm, and then fall to 69.3 per cent at 140 cm. This distribution is indicative of clay translocation. Thus, the soil between 30 and 70 cm may be interpreted as a weakly-developed ill-uvial B horizon.

2. Other physical properties

2.1. Bulk density

In all seven deep profiles investigated, bulk density correlates directly with sampling depth. The correlation coefficient (r_s) is significant at the 95 per cent confidence level in each case, and in two soils, GLL and GT-II, bulk density increases progressively down the entire profile. In GLL, for example, bulk density increases from 0.71 g/cm^3 at 10 cm to 1.14 g/cm^3 at 140 cm. This general pattern reflects the more strongly-developed structure of the surface horizons of the soils, and also their higher organic matter content.

To facilitate comparisons between the various profiles, attention in the remainder of this section focuses upon the bulk density at 20 cm, a depth which is thought to be most representative of the upper mineral horizons of the soil. Data were not obtained for the rocky hilltop plots. Bulk densities measured ranged from 0.68 (GT-II) to 1.14 (GKV-I) g/cm^3 (fig. 24A). The soils fall broadly into two groups. The first, comprising the soils of the hilltop depression (GLL) and of the hill slopes on the purer limestones (GT-III, WT-II, GT-II and WT-I), has relatively low bulk densities (range 0.68 to 0.93 g/cm^3). The second, which comprises the hill slope soils developed on the impure limestones (GKV-I and GKV -II) and also the three foothill soils, has bulk densities at 20 cm in the range 1.02 to 1.14 g/cm^3. As noted in the profile descriptions, the soils of the first group have more strongly-developed crumb structures in the surface and near surface horizons than those of the second group. They also have higher proportions of iron and aluminium in the fine-earth fraction of the soil (table 33). The iron content of the fine earth, expressed as Fe, ranges from 9.8 to 14.8 per cent in the first group, and from 2.2 to 6.4 per cent in the second. Corresponding ranges for aluminium are 12.0 to 29.5 per

cent, and 4.3 to 12.6 per cent. Inverse correlations between bulk density at 20 cm and the iron (r_s=-0.741, $p<0.05$) and aluminium (r_s=-0.711, $p<0.05$) contents of the fine earth suggest that hydrous oxides of iron and aluminium may be instrumental in soil aggregation. However, detailed micro-morphological studies are required to substantiate this hypothesis.

2.2. Particle density

Positive correlations (r_s) between particle density and soil depth occur in six of the seven deep profiles investigated. In four cases the correlation coefficient is significant at the 95 per cent confidence level. For example, in GT-I (r_s=0.901) particle density increases progressively with depth from 2.62 g/cm^3 at 20 cm to 2.73g/cm^3 at 70 cm. Down the same six profiles, particle density is inversely correlated with the organic carbon content. The organic fraction of the soil,which has a low density, diminishes with depth and thus explains the pattern observed. Particle density in the remaining plot; GKV-I, varies from 2.77 to 2.82 g/cm^3. The lack of significant correlations in this soil probably reflects the small range of particle density.

On the assumption that the highest particle density recorded in a soil profile is most representative of the mineral fraction of the soil, comparisons may be made between the various soils (fig. 24B). Values above 2.85 g/cm^3 were recorded in three soils (GLL, GT-II and WT-I). These are the group of deeper hilltop and hill slope soils formed on pure limestones. Maximum particle densities of below 2.80 g/cm^3 occur in WT-II, where the profile investigated was only 28 cm in depth, and in the three foothill soils. The comparatively low particle densities of the latter soils may be accounted for by their low iron contents. Overall, a significant positive correlation (r_s=0.723, $p<0.05$) was found between maximum particle density and the iron content of the fine earth fraction.

2.3. Pore space

In all seven of the deeper profiles investigated, the percentage pore space correlates inversely with sampling depth. For example, in GT-I (r_s=-1.000, p<0.001) pore space decreases progressively with depth from 59.8 per cent at 10 cm to 39.6 per cent at 70 cm (table 45). Pore space varies inversely with bulk density, both down individual profiles (eg. GT-I, r_s=-0.991, p<0.001) and between different soils (eg. at 20 cm, r_s=-0.948, p<0.001). Consequently, differences in pore space between the various soils (fig.24C) may be explained on the same grounds as bulk density (Section 2.1, above). Soils of the hilltop depression (GLL) and of the hill slopes on purer limestones (GT-III, WT-II, GT-II and WT-I) have large volumes of pore space (eg. range at 20 cm, 66.5 to 76.6 per cent). In contrast, those formed on impure limestones in the Gelang Kerbau Valley and on the foothill slopes in Selangor and the Kinta Valley have relatively small volumes of pore space (eg. range at 20 cm, 53.5 to 62.9 per cent).

3. Organic matter

3.1. Carbon

In all seven profiles, statistically significant (p<0.05) inverse correlations were found between sample depth and the percentage of carbon. In two plots, ABT-II (table 44) and GT-I (table 45), the carbon content decreases progressively with depth through the entire profile (r_s=-1.000, p<0.001). Such distributions indicate that the major organic matter input is at the surface, and that translocation of colloidal organic particles down the profile, if it does occur, is insufficient to form organic-rich illuvial horizons.

The mean carbon content of shallow soils in the three rocky hilltop plots and the carbon content at 10 cm in the remaining plots are shown in figure 24D. Soils of the rocky hilltop plots have

the highest concentrations of carbon (range, 4.73 to 11.03 per cent). Those of the steep hill slopes (GT-III, WT-II and GKV-II), the shallowest of the profiles investigated (maximum depth, 32 cm), also have high carbon contents (range, 2.17 to 2.60 per cent). Lowest carbon levels (0.80 to 2.16 per cent) were recorded in the seven deep profiles.

Litterfall was measured in six of the latter plots. Organic carbon at 10 cm correlates inversely with total litterfall (r_s=-0.886, $p<0.05$). One possible explanation is that soil conditions which favour tree growth, and hence litter production, also favour rapid litter decomposition. Thus where, for example, a considerable proportion of the surface is occupied by bare rock, as in GT-II (34.0 per cent), not only will tree growth be restricted but so too will the population of micro-organisms within the plot. This may explain the relatively high organic carbon content of GT-II (2.16 per cent at 10 cm), and is consistent with the higher organic matter levels of the rocky hilltop and steep hill slope soils. However, the data suggest that two other factors may also affect the organic carbon content at this depth. First, organic particles are likely to move downwards more freely where the soil has a well-developed structure than where peds are only weakly developed. Of the six soils, those in GT-II and WT-I have the lowest bulk densities at 10 cm (0.67 and 0.72 g/cm^3 respectively) and correspondingly high carbon contents at 10 cm (2.16 and 1.31 per cent). In both soils the carbon content remains relatively high throughout the profile (eg. at base of profile, 0.55 and 0.84 per cent, respectively). The other four soils, in contrast, have higher bulk densities (at 10 cm, range, 0.93 to 1.22 g/cm^3). In three of these, GKV-I, GT-I and ABT-I, the carbon content ranges from 0.80 to 0.85 per cent at 10 cm and falls rapidly with depth to between 0.07 and 0.31 per cent at the base of the profile. ABT-II is exceptional in that

its carbon content is relatively high at 10 cm (1.40 per cent) and throughout the profile (0.75 per cent at base). A second factor, soil moisture, may be important here. As noted in Section D, ABT-II is located in a water receiving site, possibly with impeded drainage. The soils are generally wetter than those of the other plots studied and may become waterlogged during storms (Chapter VI). Conditions, therefore, are less favourable for microbial activity and this may explain the higher levels of organic matter observed (further discussion, Chapter VI, Section C4.2).

3.2. Loss on ignition

Generally, the materials lost during ignition at 850°C for 30 minutes are organic matter, water of crystallization from clays, and carbon dioxide from the breakdown of carbonates. The last effect is relatively unimportant in the soil profiles investigated, since their carbonate contents are mostly less than 2 per cent. In each of the deep profiles, loss on ignition varies inversely with depth, and in two plots, GLL (r_s=-0.750) and GT-II (r_s=0.786), the correlation is statistically significant ($p<0.05$). In five soils (GLL, GT-II, ABT-I, ABT-II and GT-I), loss on ignition varies directly with the carbon content. Also a significant positive correlation (r_s=0.857, $p<0.001$) occurs between loss on ignition at 10 cm (including mean value for shallow soils in three rocky hilltop plots) and the corresponding carbon concentrations. Spatial variations in loss on ignition (fig. 24E) may therefore be attributed largely to differences in carbon content (Section 3.1, above). Thus, loss on ignition at 10 cm is highest in the rocky hilltop soils (range, 32.9 to 77.3 per cent). The clay fraction appears to be of relatively minor importance, as there are no significant correlations between clay content and loss on ignition in any of the deep profiles. Also, over the thirteen plots, there is no significant correlation between these properties at the 10 cm sampling depth.

4. Cation exchange properties

4.1. C.E.C.

It is well-established that cation exchange sites within the soil occur principally at the surfaces of colloidal organic particles, clays and, to a lesser degree, iron and aluminium hydroxides and amorphous silica (Black, 1968, pp. 213-26). In five of the deep profiles investigated C.E.C. decreases with depth, and in two soils, GT-II (r_s=-0.750) and GT-I (r_s=0.893), the relationship is statistically significant at the 95 per cent confidence level. In all five soils, C.E.C. correlates directly with the carbon content. For example, in GT-II (r_s=0.679, $p<0.05$) the highest C.E.C. (17.5 m.e./100 g, at 10 cm) corresponds with the highest carbon content (2.16 per cent), while the lowest C.E.C. (5.84 m.e./100 g, at 70 cm) corresponds with the lowest carbon content (0.55 per cent). In the two remaining deep profiles, WT-I and GKV-I, the correlation between C.E.C. and depth is positive, while that between C.E.C. and carbon content is negative. However, none of these correlations is statistically significant. In WT-I, clay content correlates directly with C.E.C. (r_s=0.929, $p<0.001$) and appears, therefore, to be an important control. No significant correlation was found in GKV-I, and this probably reflects the extremely small range of C.E.C. measured in this soil (13.72 to 14.53 m.e./100 g).

Differences in C.E.C. between the various soils may be explained largely by variations in organic matter content. C.E.C. at 10 cm in the profiles studied, and the mean C.E.C. of the soils in plots where only shallow soils are present (fig. 25A), vary directly with carbon content (r_s=0.819, $p<0.001$). The rocky hilltop soils have the highest C.E.C. (range, 40.29 to 48.90 m.e./100 g), whilst in the shallow, organic-rich profiles of the steep hill slopes, the C.E.C. ranges from 21.34 to 33.40 m.e./100 g. Of the remaining soils, only ABT-II (C.E.C. at 10 cm, 27.82 m.e./100 g), which has

a relatively high carbon content, has values in excess of 20 m.e./100 g.

4.2. Molar ratios of exchangeable cations

The absolute concentrations of individual exchangeable cations depend, in part, upon the C.E.C. of the soil. Consequently, changes with depth and spatial variations are difficult to interpret. To simplify the discussion attention is focused, therefore, upon the molar proportions of the various cations.

4.2.i. Ca+Mg:K+Na ratio. Of the exchangeable cations determined, calcium, magnesium, potassium and sodium, the first two are dominant in all the soils studied. The mean molar ratio of calcium and magnesium to potassium and sodium, henceforth referred to as the Ca+Mg:K+Na ratio, has an overall mean value of 37.4. Lowest values were recorded in GKV-I (4.9) and GKV-II (16.1) the plots on the impure limestones of the Gelang Kerbau Valley. In these soils, exchangeable potassium (0.208 and 0.253 m.e./100g, respectively) and the potassium content of the fine earth fraction (3.8 and 3.7 per cent, respectively), are both high. Argillaceous impurites in the rock are the most likely source of potassium. Elsewhere, the Ca+Mg:K+Na ratio of exchangeable cations is higher, ranging from 21.3 (ABT-III) to 89.6 (ABT-I).

In five of the seven deep profiles (GLL, GT-II, WT-I, GKV-I and GT-I) statistically significant correlations (r_s; $p<0.05$) were found between the above ratio and sample depth. Potassium, which in each soil varies directly with depth (Section 4.2.iii, below), is the key cation in this relationship.

4.2.ii. Mg:Ca ratio. Generally the Mg:Ca ratio decreases down the soil profile, and in five of the deep profiles (GLL, GT-II, WT-I, GKV-I and GT-I) the correlation with sample depth is significant at the 95 per cent confidence level. For example, in GLL (r_s=0.929) the ratio falls from 0.590 at 10 cm, to 0.190 at

140 cm. Three processes may account for this pattern. First, calcium may be removed disproportionately from the upper horizons of the soil by leaching or by selective absorption by plant roots. Secondly, magnesium may be removed disproportionately from lower horizons by either of the above mechanisms. Finally, inputs of nutrients at the soil surface may have a high Mg:Ca ratio. Although the exact mechanism is uncertain, observations suggest that the first process is operative in the plots with magnesium-rich soils. Studies of nutrient cycling in the remaining plots provide direct evidence of the second and third processes.

In the six moderately-inclined hill and foothill plots, the molar Mg:Ca ratio of total nutrient input at the surface ranges from 0.163 to 0.256 (table 54; Section E, below). Thus, where, as in GT-II, ABT-I and ABT-II, the average Mg:Ca ratio of exchangeable cations is low (range, 0.059 to 0.127), the relatively high magnesium content of the surface input may explain the high Mg:Ca ratio of upper soil horizons. In each case, the Mg:Ca ratio of the surface input exceeds that of exchangeable cations at 10 cm depth (table 54). In GT-II, for example, the values are 0.211 and 0.143, respectively. Field evidence suggests that the amounts of magnesium taken up by plants in these plots is disproportionately high compared with that of calcium (Section E6.2, below). This may explain the decrease in the Mg:Ca ratio of exchangeable cations with depth.

The pattern observed in the remaining three plots, WT-I, GKV-I and GT-I, is more problematical, since in each case the Mg:Ca ratio of exchangeable cations at 10 cm is higher than that of the surface input. In WT-I and GKV-I, the average ratio of exchangeable cations is also greater (0.837 and 0.565, respectively; cf. surface inputs, 0.256 and 0.229). In this group of plots, preferential leaching of calcium affords the most likely explanation of the relatively low calcium content of the upper part of the soil.

The disproportionately high uptake of calcium by plants which occurs in these plots (Section E6.2, below) seems inadequate to prevent the net depletion of calcium from the soil. Clearly, further observations of nutrient movement in magnesium-rich soils are required to substantiate this. If selective leaching of calcium is confirmed, then it may also occur in those plots with a low Mg:Ca ratio of exchangeable cations. Here this process is reinforced by the incongruous cycling of magnesium.

The mean Mg:Ca ratios of exchangeable cations in each of the plots are shown in figure 25. Values range from 0.059 (ABT-I) to 0.581 (WT-I). Over the thirteen plots, significant correlations were found between this ratio (mean for each plot) and the molar Mg:Ca ratio of the fine-earth fraction (r_s=0.559, $p<0.05$) and of the limestone bedrock (r_s=0.683, $p<0.05$). When the three foothill slopes are excluded the strengths of the correlations with the fine earth fraction (r_s=0.867, $p<0.001$) and the bedrock (r_s=0.830, $p<0.001$) are increased considerably. The latter correlations indicate a strong relationship between the chemistry of the soil and the parent limestone in the hilltop and hill slope plots. The reduced significance of the correlation coefficients when the foothill soils are included, suggests that these soils are less strongly influenced by the underlying rock. This observation supports the hypothesis that the foothill soils are of colluvial or alluvial origin.

4.2.iii. <u>K:Na ratio</u>. The principal inputs of potassium and sodium into the soil occur at the surface, in the form of throughfall and litterfall. The molar K:Na ratio of the total surface input in the six plots ranges from 3.46 to 9.78 (table 55). Such inputs account for the relatively high mean K:Na ratios of exchangeable cations in the soil (fig. 25B), which range from 1.79 (ABT-II) to 5.82 (ABT-III).

Of the two cations, potassium and sodium, the former is usually the least mobile in the soil because it becomes incorporated

readily within the lattices of clay particles (Black, 1968). On theoretical grounds, therefore, it may be hypothesized that exchangeable sodium will decrease with depth more rapidly than potassium, since it is more readily leached from the soil. In reality, such a relationship was not found. The K:Na ratio decreases with depth in all seven deep profiles studied. In six of the seven soils (GLL, GT-II, GKV-I, ABT-I, ABT-II and GT-I) the inverse correlation between this ratio and sample depth is statistically significant at the 95 per cent confidence level. In GKV-I (r_s=-0.942), for example, K:Na decreases from 5.29 at 10 cm to 2.00 at 140 cm. This general trend may be explained by the disproportionately large amounts of potassium which are absorbed by plant roots. Evidence of this mechanism is discussed below (Section E6.3)

4.3. Base saturation and pH (water, 1:2.5)

Base saturation and soil reaction provide independent measures of the base content of the soil. Direct correlations (r_s), with probabilities of 0.080 or less, were found between these properties in six of the seven profiles studied (GLL, GT-II, WT-I, ABT-I, ABT-II and GT-I). Also, the mean pH of soils in each of the thirteen plots correlates significantly with the mean base saturation (r_s=0.879, $p<0.001$). The lack of a significant correlation in profile GKV-I probably reflects the small range of base saturation (15.37 to 24.27 per cent) and especially of pH (5.4 to 5.6) in this soil.

4.3.1. Variations with depth. Variations in base content with depth were observed in all seven deep profiles. In two soils, GLL (r_s=0.991, $p<0.001$) and GKV-I (r_s=0.837, $p<0.05$). pH varies directly with sample depth. In the first plot, for example, pH decreases progressively from 5.2 at 10 cm to 6.5 at 100 and 140 cm. Here, the upper soil horizons are leached of bases (base saturation at 10 cm, 22.8 per cent), while the lower horizons are comparatively

base-rich (base saturation at 100 cm, 62.1 per cent). This clearly reflects the importance of the weathering front (ie. soil/rock interface) as the principal source of bases, chiefly calcium and magnesium, within the soil. In the remaining hill slope plots with deep profiles, GT-II and WT-I, the relationship between base content and sample depth is more complex. As in the previous plots, pH increases towards the bottom of the profile. For example, in WT-I, pH increases from 5.7 (base saturation, 47.2 per cent) at 50 cm to 6.0 (base saturation, 51.6 per cent) at 70 cm. However, in both profiles the maximum pH occurs near the surface. Thus, at 10 cm in WT-I, pH equals 6.2 and the base saturation is 75.1 per cent. The high base content of the surface horizons may be explained by the high rate of nutrient input at the soil surface (table 52).

In the three foothill soils pH increases at the lowest sample depth(s). However, the dominant trend in ABT-I (r_s=-0.618), ABT-II (r_s=-0.353) and GT-I (r_s=-0.613) is for pH to decrease with depth. The comparative weakness of the effect of the basal weathering front is consistent with the soil parent material being of alluvial or colluvial origin.

4.3.ii. Spatial variations. In order to facilitate discussion of the spatial variations in base content, attention is now focused on the mean pH (fig. 25D) and the base saturation of the profile and/or shallow soils sampled in each plot. A corollary of the hypothesis proposed above to explain the increases in bases in the lowest parts of individual profiles, is that a shallow soil will have a higher mean base content than a deeper one, since a higher proportion of the soil is closer to the weathering front. Generally, among the hilltop and hill slope plots, there is an overall inverse relationship between pH and soil depth (r_s=-0.670, $p<0.05$). Soils of the three steep hill slope plots have the highest mean base contents (pH range, 7.14 to 7.23; base sat-

uration range, 73.9 to 80.8 per cent), and also include the single sample of highest pH, 7.6 (WT-II). In contrast, soils of the moderately-inclined hill slopes and hilltop depression have consistently lower base contents (pH range, 5.50 to 6.46; base saturation range, 20.2 to 61.3 per cent), with the two lowest mean pH values being recorded in the deepest profiles, GKV-I (pH, 5.5) and GLL (pH, 5.94). In the latter profile the minimum pH is 5.2, the lowest recorded in the field areas.

Soils of the rocky hilltops, though mostly shallower than those of the steep hill slopes, generally have lower base contents (pH range, 6.61 to 6.73; base saturation range, 64.0 to 67.2 per cent). In contrast with the latter soils, the former frequently have a continuous surface mat of decaying organic litter. It may be hypothesized that organic acids, released into the soil solution during decomposition, accelerate the leaching of bases from the soil. Support for such a mechanism is provided by the levels of non-alkaline hardness (mean, 8.97 ppm) in runoff from solution runnels with shallow, organic-rich, soil covers on Ak. Bt. Takun (Chapter V, Section D2.1.1).

The mean pH values of the foothill soils range from 6.41 to 6.77 (base saturation range, 64.8 to 74.5 per cent). These soils have a generally higher base content than the moderate hill slopes, which are of comparable depth. Although no blocks of limestone were observed in the profiles investigated, rounded blocks were found in other pits excavated in ABT-I and ABT-II. In places they occur within several centimetres of the surface. One possible explanation of the high base content of these soils is that the release of bases from the limestone, rather than being confined to the basal weathering front, may occur throughout the profile. Under these circumstances, solution of limestone may occur close to the surface. Consequently, the resulting soil waters will be base-rich

and have a reduced capacity to leach exchangeable bases from the soil.

5. Carbonates

In four of the seven deep profiles investigated (GT-II, WT-I, ABT-I and GT-I) the highest carbonate content occurs closest to the basal weathering front, at the lowest sampling depth. Statistical analysis of the data for individual profiles has not been undertaken, since the method of carbonate determination used provides only an approximate figure and yields many tied values. For example, the estimated carbonate content in GKV-I is 0.5 per cent down the entire profile.

A mean carbonate content has been calculated for the profile and/or shallow soils in each plot (fig. 25). Highest carbonate contents occur in ABT-IV (3.50 per cent) and ABT-III (3.00 per cent), the two plots with the shallowest soils. In GT-III the mean carbonate content exceeds 1.0 per cent. Elsewhere the highest value is 0.7 per cent, with GLL (0.21 per cent), the deepest profile, having the lowest figure. The mean carbonate content correlates directly with pH over the thirteen sites (r_s=0.593, $p<0.05$).

E. RESULTS OF THROUGHFALL AND LITTERFALL STUDIES

Throughfall and litterfall were monitored in the six moderately-inclined hill and foothill plots (GT-II, WT-I, GKV-I, ABT-I, ABT-II and GT-I). Stemflow, a further component of water and nutrient flow, was not measured. Consequently, rainfall interception is overestimated and nutrient inputs at the soil surface are underestimated in the present study. However, since stemflow was observed only once, and on a single tree, during the study period, errors resulting from the omission of stemflow inputs are likely to be small.

The higher branches of individual trees often extend beyond the limits of the sub-plot in which they are rooted. In order

to provide a more representative measure of the tree cover in the central sub-plot the basal area of tree trunks was weighted as follows:

$$\frac{\text{basal area in central sub-plot} - \text{mean basal area of adjacent sub-plots}}{2}$$

Values of this index, the 'weighted basal area' (w.b.a.), are given in table 46.

1. Interception

During the study period, the amount of rainfall intercepted in each plot varied considerably (table 47). For example, the 12 measurements made in ABT-I range from 8.2 to 47.4 per cent. This variability may reflect annual differences in the amount of rainfall intercepted. In part, however, it may be caused by sampling errors which arise in the collection of throughfall, and by the inaccuracies which result when rainfall data from nearby meteorological stations are used.

The mean rate of interception correlates positively with the weighted basal area index (r_s=0.714). Most interception (mean, 34.3 per cent) occurs in GKV-I, which has the densest canopy cover (w.b.a., 69.7 m^2/ha), whilst WT-I (w.b.a., 25.4 m^2/ha) has the lowest interception rate (mean, 18.9 per cent). ABT-II, with the sparsest tree canopy (w.b.a., 10.6 m^2/ha) has an interception rate of 24.6 per cent. This is only slightly less than in ABT-I (mean, 24.6 per cent), which has a weighted basal area of 37.9 m^2/ha. The results from Ak. Bt. Takun are somewhat difficult to explain, particularly as the density of undergrowth and number of

young trees (girth less than 15 cm) appeared to be similar in both plots. Two alternative hypotheses may be put forward to explain these results. First, ABT-II is in a comparatively sheltered location and may, therefore, receive less rainfall than the adjacent plot. The percentage interception, which is calculated from rainfall at Templer Park (2601), may thus be an overestimate of the true value. Secondly, the w.b.a. index may poorly reflect the density of the canopy cover in areas of disturbed forest (further discussion, Section E3.2.1, below).

Overall, the mean interception rates in the plots studied fall at the bottom end of the range recorded in the rain forest of the Main Range (18 to 42 per cent - Kenworthy, 1971; Low, 1972, 1974). Clearly, lower rates of interception would be expected in those areas with sparser vegetation covers, such as in the rocky hilltop and steep hill slope plots.

2. Chemical properties of throughfall

The chemical characteristics of throughfall in the six main sampling plots, and in ABT-III and ABT-IV (see Chapter V), are summarized in table 48. Considerable variability may be observed within each plot. For example, in ABT-I, where 28 samples were collected, pH ranges from 5.8 to 7.3, and specific conductance from 34.0 to 162.0 μmho. Marked variations also occur in the concentrations of individual cations: calcium (range, 1.5 to 14.0 ppm), magnesium (range, 0.85 to 3.50 ppm), potassium (range, 1.55 to 9.0 ppm) and sodium (range, 0.06 to 1.41 ppm). Since the collecting bottles were relocated after each sampling, the variability in the data has both a spatial and temporal component. Because of the relatively small number of samples, no attempt has been made to interpret within plot variations. Attention in the remainder of this section focuses upon the mean values recorded in each of the seven plots.

2.1. Interrelationships between the chemical properties

Correlations between the chemical properties of throughfall are shown in table 49. Specific conductance, which provides a measure of the total solute content, and the individual cations are generally closely interrelated. Throughfall with a high solute content thus tends to contain relatively high concentrations of all four cations. Statistically significant correlations ($p<0.05$) occur between specific conductance, calcium and magnesium. pH varies directly with specific conductance ($r_s=0.500$), but correlates most strongly with the sodium content ($r_s=0.893$, $p<0.05$). The latter relationship is difficult to explain, especially in view of the consistently low sodium concentrations in throughfall waters (range, 0.20 to 0.60 ppm), and may have occurred by chance.

2.2. Cation content of throughfall

The concentrations of individual cations in throughfall may be affected by a wide range of factors, including the cation exchange properties of soils, vegetation type and density, percentage interception and total volume of throughfall. In the present study, no single factor stood out as correlating significantly with the concentration of individual cations. This probably reflects the complexity of the environmental controls. Clearly a much more detailed study is required to elucidate these factors.

2.2.i. Calcium. The calcium content of throughfall in the seven plots averages 4.79 ppm, compared with 0.97 ppm in rainfall. The lowest mean concentration (2.43 ppm) was recorded in ABT-III/ABT-IV, which has a comparatively sparse vegetation cover. Highest values were recorded in ABT-I (7.04 ppm) and GKV-I (6.29 ppm), the plots with the smallest volumes of throughfall (1633.0 and 1329.6 mm, respectively; cf. other plots, range, 1687.7 to 2057.2 mm).

2.2.ii. Magnesium. Magnesium concentrations (mean, 1.16 ppm. cf. rainfall, 0.09 ppm) range from 0.30 ppm (ABT-IV/ABT-IV) to 1.99

ppm (ABT-I). The spatial variations in magnesium content correlate closely with those of calcium (r_s=0.964, $p<0.001$).

2.2.iii. Potassium. Potassium, like calcium and magnesium, is a vitally important plant nutrient. It differs however, from other macro-nutrients in that it is not a constituent of organic molecules in plants, but occurs in an ionic form (Sutcliffe and Baker, 1976, p. 17). Consequently, it is relatively mobile and readily leached from the vegetation canopy. Concentrations of potassium which average 5.27 ppm (cf. rainfall, 0.13 ppm), are higher than other cations in throughfall. Potassium contents range from 0.64 (ABT-III/ABT-IV) to 6.98 ppm (ABT-II).

2.2.iv. Sodium. Compared with above elements, sodium is a much less important nutrient in plant growth (Sutcliffe and Baker, 1976 pp. 12-13). Concentrations in throughfall average 0.52 ppm (cf. rainfall, 0.25 ppm) and range from 0.20 (ABT-III/ABT-IV) to 0.71 ppm (GKV-I).

2.3. Comparison with other studies

Comparable data on throughfall are available from the Ulu Gombak (Kenworthy, 1971) and Pasoh Forest Reserves (Manokaran, 1978). In these regions, concentrations of calcium average 1.38 and 0.26 ppm, respectively (cf. karst regions, 4.79 ppm). Corresponding levels of magnesium are 0.25 and 0.13 ppm (cf. karst regions, 1.16 ppm), and of potassium, 2.5 and 2.08 ppm (cf. karst regions, 5.27 ppm). Thus, in each case the solute concentrations are lower than those measured in the karst regions. This is probably a direct reflection of the relatively high nutrient status of the karst soils. Concentrations of sodium, a much less important plant nutrient, were lower in the karst regions (mean, 0.52 ppm) than at Pasoh where the average was 1.24 ppm. Data on sodium are lacking for the Ulu Gombak F.R.

3. Litterfall

The data discussed below are average figures based on

measurements made at four litter trays in each plot.

3.1. Temporal variations of litterfall

In all six plots, litterfall occurs continuously throughout the year. However, the weight of litterfall varies considerably from month to month, and the temporal patterns differ from one plot to the next (fig. 27). Elsewhere in the Peninsula, peaks in litterfall have been found often to correspond with the drier periods of the year (Koriba, 1958; Medway, 1972; Ogawa, 1978). It is most appropriate, therefore, to consider temporal variations on a regional (climatic) basis. Correlation coefficients were calculated for the relationships between monthly litterfall and the following rainfall data: actual monthly rainfall, actual rainfall of previous month, and the long-term monthly rainfall.

3.1.1. Selangor (ABT-I and ABT-II). In both plots litterfall varies considerably through the year, with amounts ranging from 0.485 to 1.007 t/ha in ABT-I, and from 0.294 to 0.634 t/ha in ABT-II. Although located less than 200 m apart, there is no direct correlation between the monthly pattern of litterfall in the two plots. The difference was most marked in August 1974 when litterfall in ABT-I was the lowest of the study period (0.485 t/ha) and litterfall in ABT-II was the highest (0.666 t/ha). In neither plot does litterfall correlate significantly with rainfall during the study period or with the mean monthly rainfall. Factors other than total precipitation input are clearly important. Storm characteristics may be significant. Wind velocity is often high during storms and this, combined with the extra weight of water retained in the canopy and the kinetic energy of raindrop impact, may accelerate the detachment of leaves, twigs and other forms of litter. Although litterfall was observed to increase visibly during particularly intense storms, daily measurements would be required to evaluate this effect.

3.1.ii. Kinta Valley (GT-I and GT-II). Marked variations in litterfall occur from month to month (fig. 27), especially in GT-I, where amounts range from 0.261 (September, 1974) to 0.802 t/ha (June, 1975). The temporal distributions of litterfall in the two plots are not significantly correlated. Also, in neither plot does litterfall correlate significantly with rainfall during the study period or with the mean monthly rainfall.

3.1.iii. North-west region (WT-I and GKV-I). Rainfall in the North-west region is more seasonal, and this is reflected in the patterns of litterfall. From January to March (1975), which are the driest months on average, litterfall in WT-I and GKV-I amounted to 1.660 and 4.042 t/ha, respectively. Corresponding values for September to November (1974), which are the wettest months on average, were 1.000 and 2.180 t/ha. Rainfall during the sampling period was somewhat exceptional (Chapter II, Section B3.3). At Kaki Bukit, for example, 314 mm of rain fell in January 1975, compared with an average of 55 mm. It is interesting to note, therefore, that litterfall during the study period varies inversely with the mean monthly rainfall (WT-I, r_s=-0.748; GKV-I, r_s=-0.734; fig. 27), the correlation coefficient in each case being significant at the 95 per cent confidence level, but is not related to the actual rainfall of this period. This result is important from an ecological viewpoint since it demonstrates that the vegetation of the sampling plots responds to long-term average periodicities in rainfall, and is relatively unaffected by short-term deviations.

3.2. Annual litterfall

3.2.i. Spatial variations. Litterfall during the one-year sampling period (table 50) was greatest in GKV-I (12.048 t/ha), the plot with the densest vegetation cover. Smallest amounts were recorded in the remaining moderately-inclined hill slope plots, GT-II (5.189 t/ha) and WT-I (5.734 t/ha). Overall, litterfall in the six plots correlates directly with the weighted basal area index (r_s=0.771, $p<0.05$). In

detail, however, the three foothill plots do not fit the general pattern. Thus, in GT-I, which has a relatively undisturbed forest cover with a weighted basal area of 53.8 m^2/ha, litterfall (6.411 t/ha/yr) is less than in ABT-I, which has a smaller w.b.a. index (37.9 m^2/ha). Also, the w.b.a. index of ABT-I is more than treble that of ABT-II (10.6 m^2/ha), yet the difference in annual litterfall between the plots is comparatively small, the amounts being 7.867 and 6.044 t/ha, respectively. Once again an indirect measure of canopy cover in the disturbed foothill plots is poorly reflected in the w.b.a. index (cf. interception, Section E1, above). On the assumption that fine litterfall is directly proportional to the total mass of leaves (hence, leaf area) and twigs which form the vegetation canopy, then the canopies of the three foothill plots are very similar (litterfall range, 6.044 to 7.867 t/ha/yr). This suggests that in areas with fairly complete soil covers, the leaf density in the canopy is as high where larger trees have been recently felled (ABT-I and, especially, ABT-II), as in relatively undisturbed forest, presumably due to the rapid growth of groundflora and tree saplings in canopy gaps. This finding is clearly of ecological significance, since leaf area is a fundamental factor in vegetation productivity (Jones, 1979, pp. 31-35).

3.2.ii. Components of litterfall. Litterfall was subdivided into three fractions: leaves, twigs, and a miscellaneous component, which comprises mostly bracts, scales, flowers and fruits. Over the six plots the proportions of leaves, twigs and miscellaneous material averaged 73.2 (range, 70.1 to 79.5), 13.7 (range, 9.2 to 16.1) and 13.1 (range, 9.2 to 20.3) per cent, respectively.

3.2.iii. Comparisons with other tropical areas. During the study period litterfall in the six plots averaged 7.216 t/ha/yr. This is lower than has been recorded in tropical rain forests, as for example in the Pasoh F.R., Malaysia (range, 7.5 to 10.2 t/ha/yr - Gong, 1972; Lim, 1978; Ogawa, 1978), the eastern part of the Amazon basin (range, 7.4 to 10.7 t/ha/yr - Klinge, 1978), and Columbia

(8.5 t/ha/yr - Jenny et al., 1949). This probably reflects the limitations imposed by the generally shallow and discontinuous soil covers of the limestone slopes. However, in GKV-I, where soil depth is least limiting, litterfall (12.048 t/ha) exceeds that of the above rain forests and also of semi-deciduous tropical forests, as for example in Ghana (range, 9.66 to 10.45 t/ha/yr - John, 1973; Nye, 1961). In this case the higher rate of litter production may be explained, in part, by the relatively high base content of the soil (mean pH, 5.5), and the abundance of calcium and magnesium at the weathering front (Sections E4 to E6, below). For comparison, the pH of soils at Pasoh ranges from 3.0 to 4.8 (Allbrook, 1973; Yoda, 1974).

The proportions of the various litter components in the Pasoh Forest Reserve, Malaysia, are in roughly the same proportion as in the study areas. The percentages of leaves, twigs and miscellaneous fragments (as defined above) average 72.1, 16.9 and 11.0 respectively (Lim, 1978).

4. Nutrient content of litter

4.1. Ash fraction

The percentage by weight of ash in the leaf litter is significantly higher than in the other litter components (randomization test; $p<0.05$). The leaf fraction contains, on average, 12.68 (range, 8.21 to 17.51) per cent ash, while the corresponding figures for the twig and miscellaneous fractions are 6.38 (range, 5.09 to 8.38) per cent and 6.60 (range, 4.86 to 8.59) per cent, respectively.

The ash content of leaf litter in the six plots correlates directly with the average percentage base saturation of the soil ($r_s=0.771$, $p<0.05$). Highest proportions of ash, 12.70 to 17.51 per cent, occur in the three foothill plots, which have base saturation levels in the range 64.8 to 74.5 per cent. The smallest ash content (8.21 per cent) was recorded in GKV-I, which has the most acid soils (mean base saturation, 20.2 per cent).

4.2. Nutrient contents

The concentrations of selected elements in litterfall are summarized in table 51.

4.2.i. Calcium. Calcium is the dominant cation in all three litter components, ranging from 2.43 (twigs/small branches) to 4.23 per cent by weight (leaves). The calcium content of leaf litter (fig. 28) is highest in the foothill plots (range, 4.78 to 5.48 per cent). Concentrations in the three hill slope plots range from 2.92 to 3.94 per cent, with the lowest being recorded in GKV-I, the plot with the deepest and most acid soil. The calcium content of leaf litter in the six plots varies directly with the average base saturation (r_s=0.771) and concentration of exchangeable calcium of the soil (r_s=0.771), the correlation coefficient in both cases being statistically significant ($p<0.05$). Calcium levels in the miscellaneous fraction correlate directly with those in leaf litter (r_s=0.886, $p<0.05$), the highest concentrations being in the foothill plots (range, 4.172 to 4.387 per cent). The calcium content of the twig/small branch fraction ranges from 1.26 (ABT-I) to 3.97 per cent (GT-I), but exhibits no regular spatial pattern.

4.2.ii. Magnesium. Concentrations of magnesium are highest in leaf litter (mean, 0.39 per cent) and lowest in twigs/small branches (mean, 0.20 per cent). As with calcium, magnesium levels in the leaf and miscellaneous fractions of the litter are greatest in the foothill plots (eg. leaf litter, range, 0.403 to 0.493 per cent) and smallest in GKV-I (eg. leaf litter 0.317 per cent).

4.2.iii. Potassium. Concentrations of potassium in litterfall are similar to those of magnesium, ranging from 0.17 per cent in twigs/small branches to 0.45 per cent in the miscellaneous fraction (table 51). Potassium in the leaf fraction (mean, 0.38 per cent) correlates directly with exchangeable potassium levels in the soil (r_s=0.543, $p<0.05$). Values range from 0.25 in WT-I, to 0.65 per cent in GT-I (fig. 28).

4.2.iv. Sodium. Sodium forms only a small component of litterfall, ranging from 0.014 (twigs/small branches) to 0.026 per cent (miscellaneous fraction). Concentrations in the leaf fraction range from 0.014 (ABT-I) to 0.056 per cent (WT-I), but exhibit no clear spatial pattern.

4.2.v. Other elements. Silicon has the second highest concentration, averaging 0.83 per cent overall. The amount of silicon in leaf litter (1.41 per cent) is more than treble that in twigs (0.38 per cent). The remaining elements, iron and aluminium, each average less than 0.1 per cent by weight. The aluminium content of twigs (0.145 per cent) is more than six times that of leaf litter (0.025 per cent).

4.3. Comparison with other studies

Data on the composition of leaves in limestone terrain in the humid tropics are available from Sarawak (Wilford and Wall, 1965). On the summits and flanks of the limestone hills the mean cation contents of leaves were as follows: calcium, 3.10 per cent; magnesium, 0.28 per cent; potassium, 0.46 per cent; sodium, 0.10 per cent. Apart from the concentration of sodium, which is four times greater than in the present study, these values are very similar to those reported above (cf. table 51).

In general, litter in the karst regions has a higher base content than in rain forests on non-carbonate lithologies. For example, the ash content of leaf litter in the Pasoh Forest Reserve (Lim, 1978) averages 4.8 per cent (cf. karst regions, mean, 12.68 per cent), and the proportions of calcium and magnesium in leaf litter average 0.70 and 0.22 per cent (cf. karst regions, 4.23 and 0.39 per cent, respectively). The mean potassium content of leaf litter at Pasoh and in the karst regions was 0.38 per cent.

5. Nutrient input at the soil surface

5.1. Components of nutrient input

Inputs of nutrients at the soil surface are derived

principally from four sources: rainfall, canopy leaching, litterfall and atmospheric dust. The average contributions made by the first three components are summarized in table 52. The soluble fraction of the last component is included in the rainfall data (see table 22).

5.1.i. Calcium. Calcium, which has the largest input (764.28 kg/ha/yr), is mostly derived from litterfall (75.89 per cent). Contributions from canopy leaching (18.13 per cent) and, especially, rainfall (5.89 per cent) are comparatively small.

5.1.ii. Magnesium. Inputs of magnesium are approximately one-eighth of those of calcium, averaging 46.85 kg/ha/yr. The input is derived chiefly from litterfall (53.81 per cent) and canopy leaching (41.64 per cent).

5.1.iii. Potassium. Potassium, which has the second largest input (130.23 kg/ha/yr), is derived mostly from canopy leaching (77.20 per cent). This reflects the mobility of potassium within plants, a feature which has been noted in other tropical regions (eg. Gombak, Malaysia - Kenworthy, 1971; Pasoh, Malaysia - Manokaran, 1978; Kade, Ghana - Nye, 1961). Litterfall (20.38 per cent) makes up the bulk of the remaining potassium input.

5.1.iv. Sodium. Of the four elements studied, sodium is least important (11.91 kg/ha/yr). Slightly more than half the sodium (50.38 per cent) is from rainfall. The total contribution of canopy leaching and litterfall is only 5.91 kg/ha/yr, with the former accounting for more than two-thirds of this.

5.2. Spatial variations in nutrient input

The various components of nutrient input during the study period in each of the six plots are summarized in fig. 29.

5.2.i. Calcium. ABT-I has the largest input of calcium at the soil surface (500.89 kg/ha/yr). 77.0 per cent is from litterfall. This reflects the large weight of litterfall (7.867 t/ha) and the high proportion of calcium in the litter (eg. leaf litter, 5.48

per cent). Large inputs of calcium also occur in GKV-I (423.10 kg/ha/yr), 80.2 per cent of which is supplied in litterfall. Here the very large weight of litter (12.048 t/ha/yr) compensates for the low concentration of calcium within the litter (eg. leaf litter, 2.92 per cent). Calcium inputs are lowest in WT-I (227.12 kg/ha/yr), where total litterfall (5.734 t/ha) and the calcium content (eg. leaf litter, 3.46 per cent) are both comparatively small. In GT-II litterfall is even smaller (5.189 t/ha/yr). Here, however, canopy leaching supplies relatively large amounts of calcium (32.1 per cent input), giving a total soil input of 308.06 kg/ha/yr.

5.2.ii. Magnesium. Magnesium inputs are highest in ABT-I (66.64 kg/ha/yr) and GKV-I (60.61 kg/ha/yr). This reflects the high proportion of magnesium in litterfall of the first plot (0.493 per cent) and the large weight of litterfall in the second (cf. calcium, above). Magnesium inputs are smallest in WT-I (35.30 kg/ha/yr), a result which is again consistent with those obtained for calcium.

5.2.iii. Potassium. Throughfall supplies the major proportion of the potassium input at the soil surface (table 52). Because rainfall is greater in the southern part of the Kinta Valley than in the other two field areas, volumes of throughflow are highest there. During the study period throughflow volumes in GT-II and GT-I were 2057.2 and 1907.0 mm, respectively, while potassium inputs totalled 158.64 and 153.99 kg/ha/yr. Throughflow in the remaining four plots ranged from 1329.6 (GKV-I) to 1688.4 mm (ABT-II) and inputs of potassium were correspondingly lower (range, 79.46 to 141.85 kg/ha/yr).

5.2.iv. Sodium. Inputs of sodium are mostly derived from rainfall and canopy leaching. Concentrations of sodium (mean, 0.28 ppm) in rainfall are slightly higher in the Kinta Valley than elsewhere (table 22). This, combined with the heavier rainfall of the area, supplies 7.97 kg/ha/yr to the two sampling plots on G. Tempurong. In GT-II canopy leaching (5.75 kg/ha/yr) also contributes signif-

icantly, giving a total input of 15.83 kg/ha/yr, the highest recorded in the six plots. By contrast, however, no sodium was leached from the canopy in GT-I and the total input was correspondingly small, 9.27 kg/ha/yr (86.0 per cent from rainfall). The data available indicate that a small amount of sodium (0.53 kg/ha/yr) may have been retained on, or absorbed by, the canopy in the latter plot. Sodium inputs in the remaining plots range from 9.38 (ABT-I) to 13.51 kg/ha/yr (WT-I).

5.3. Comparison with other studies

Estimates of surface inputs of nutrients at Pasoh may be calculated from published data. Mean annual rainfall in the area is 2054 mm (Soepadmo, 1978), of which an average of 39 per cent is lost by interception (Low, 1972, 1974), The nutrient contents of throughfall (Manokaran, 1978) and litterfall (Lim, 1978) are known. Stemflow inputs are excluded since no quantitative data are available. This omission does not seriously affect the results since it is likely that inputs from this source are relatively small. The total input of calcium at Pasoh is approximately 10.21 kg/ha/yr. This is about 1/35 th of the mean calcium input of the six karst plots (mean, 364.28 kg/ha/yr), a difference which may be explained by the far higher concentrations of calcium in the soils and bedrock of the karst regions. Inputs of magnesium (19.83 kg/ha/yr) and potassium (57.51 kg/ha/yr) are also lower at Pasoh than in the karst plots (46.85 and 130.23 kg/ha/yr, respectively). This again reflects the contrasted base status of soils in the two areas.

6. Interrelationships between surface inputs and soil characteristics

In order to facilitate comparisons of the cation inputs at the soil surface and the exchangeable cations of soils in the six plots, attention in this section focuses on three dimensionless molar ratios: Ca+Mg:K+Na, Mg:Ca and K:Na. In this way the effects of variations in C.E.C. and the absolute amounts of rain-

fall, throughfall and litterfall are eliminated.

6.1. Ca+Mg:K+Na ratio

Values of this ratio are presented in table 53.

6.1.i. Spatial variations. The Ca+Mg:K+Na ratio of rainfall ranges from 0.79 in the Kinta Valley (GT-I and GT-II) to 3.30 in Selangor (ABT-I and ABT-II). The exact cause of variations in the solute content of rainfall in the three field areas, as reflected in this and the other dimensionless ratios, is uncertain (Chapter II, Section B4). Figures for the total amounts of cations leached from the canopy range from 0.67 (WT-I) to 1.33 (ABT-I), while corresponding values for litterfall vary between 7.24 (GKV-I) and 20.32 (ABT-I). Positive correlations occur between the Ca+Mg:K+Na ratio of exchangeable cations and that of canopy leaching (r_s=0.371) and litterfall (r_s=0.600). No relationship was found between the ratio of exchangeable cations and that of either the total input from the vegetation canopy or the total surface input. The lack of correlation in the last case may be explained, in part, by variation in rainfall input, since a high proportion of sodium is derived from this source. In fact, the Ca+Mg:K+Na ratio of rainfall correlates directly with that of the total surface input (r_s=0.843, $p<0.05$).

6.1.ii. Average values of individual components of nutrient flow

If it is assumed that the total amounts of cations leached from the canopy and present in litterfall are broadly representative of nutrient up-take by plants, then dimensionless ratios, such as Ca+Mg:K+Na, provide a basis for identifying selective nutrient absorption by plant roots. The Ca+Mg:K+Na ratio of the total nutrient input from vegetation (mean, 3.04) is significantly less than that of exchangeable cations in the soil (mean, 44.6) at the 99.9 per cent confidence level.

Clearly, therefore, plants in the field areas absorb cations disproportionately from the soil. This process is well-established and is discussed in recent reviews (eg. Sutcliffe and Baker, 1976). Selective up-take of potassium by plants is probably the dominant factor which controls the Ca+Mg:K+Na ratio in the karst soils of West Malaysia.

6.2. Mg:Ca ratio

Values of the molar Mg:Ca ratio are presented in table 54.

6.2.1. Spatial variations. The Mg:Ca ratio of rainfall varies from 0.145 in the North-west region (WT-I and GKV-I) to 0.205 in the Kinta Valley (GT-I and GT-II). Corresponding values for solutes leached from the canopy range from 0.326 (GT-I) to 0.942 (WT-I). Spatial variations in the Mg:Ca ratio of canopy leaching correlate significantly with those of litterfall (r_s=0.829, $p<0.05$), with the latter ranging from 0.127 (GT-I) to 0.256 (WT-I). The range of the Mg:Ca ratio of cations derived from the vegetation canopy (0.162 to 0.267) is small compared with that of exchangeable cations in the soil (0.059 to 0.581). This provides further evidence of selective up-take by the vegetation cover. For example, in the relatively magnesium-rich soil of WT-I (Mg:Ca of exchangeable cations, 0.059), calcium is absorbed in disproportionately large amounts by plant roots (Mg:Ca of total input from vegetation, 0.240). On the other hand, where soils are comparatively deficient in magnesium, as in ABT-I (Mg:Ca of exchangeable cations, 0.059), magnesium appears to be absorbed preferentially (Mg:Ca of total input from vegetation, 0.219). This being the case, it explains why only weak correlations exist between the Mg:Ca ratio of exchangeable bases in soil and the corresponding ratios of canopy leaching (r_s=0.257), litterfall (r_s=0.429), and total input from the vegetation canopy (r_s=0.429).

6.2.ii. Average values of individual components of nutrient flow. Mg:Ca ratio of exchangeable cations in the six soils averages 0.245 (table 54). There is no statistically significant difference between this value, and that of the total input from vegetation (mean, 0.217). This finding supports the suggestion made earlier that plants take up magnesium selectively from calcium-rich soils, and calcium from magnesium-rich soils. The mean Mg:Ca ratio of cations leached from the canopy, 0.539, is significantly higher (randomization test; $p<0.05$) than that of cations in litterfall, 0.150. Magnesium thus appears to be more susceptible to canopy leaching than calcium.

6.3. K:Na ratio

Values of the molar K:Na ratio are given in table 55.

6.3.i. Spatial variations. Sodium occurs in relatively high proportions in rainwater, the K:Na ratio ranging from 0.259 in the Kinta Valley (GT-I and GT-II) to 0.421 in Selangor (ABT-I and ABT-II). The K:Na ratios of canopy leaching and litterfall are greatest in GT-I (n.d., see footnote of table 55, and 12.19, respectively) and GKV-I (12.57 and 12.06, respectively). These plots have the highest proportions of exchangeable potassium in the soil (3.46 and 3.47). The K:Na ratio of total inputs from the vegetation canopy in the six plots correlates positively with that of exchangeable cations in the soil ($r_s=0.886$, $p<0.05$) suggesting that, to some extent nutrient absorption is limited by soil conditions.

6.3.ii. Average values of individual components of nutrient flow. The K:Na ratio of the total cation input from the vegetation canopy averages 22.05 (table 55). This value is significantly higher than that of exchangeable cations in the soil (randomization test; $p<0.05$). Clearly, therefore, potassium is being absorbed from the soil in disproportionately large amounts. The K:Na ratio of canopy leaching, which averages 12.07 (excluding GT-I), is significantly higher than that of litterfall (randomization test; $p<0.05$). Thus, potassium

is more susceptible to canopy leaching than sodium. The K:Na ratio of the total input of nutrients at the soil surface (mean, 6.68) is significantly less than that of cations derived from the vegetation canopy (randomization test; $p<0.05$). This difference is attributable to the rainfall input, which has a relatively low ratio (mean, 0.325).

F. SUMMARY OF RESULTS

1. Range of meso-relief, soil and vegetation characteristics

1.1. Meso-relief and degree of surface cover vary greatly. In the plots studied, mean slope angles range from 15.1 to 67.3°, proportions of exposed bedrock from 0.3 to 83.3 per cent, and maximum soil depths from 15 to 240 cm.

1.2. Marked contrasts also occur in the vegetation cover, with the basal area of tree trunks varying from 2.2 to 70.2 m^2/ha.

2. Soil properties

2.1. Hill soils are predominantly clays (mean clay content, 40.6 to 71.7 per cent). Foothill soils are comparatively clay-deficient (range 33.1 to 46.9 per cent), but contain high proportions of sand (range, 19.8 to 37.8 per cent). Clay-rich B horizons are present in many soils.

2.2. Gravels occur in most soils but vary considerably in their amount, distribution, mineral composition and origin. In hill soils they are mostly lateritic concretions, whereas in foothill soils they are predominantly quartz grains.

2.3. Hill soils on pure limestones have low bulk densities (range, 0.68 to 0.93 g/cm^3 at 20 cm) compared with those developed on impure limestones and on foothill slopes (range, 1.02 to 1.14 g/cm^3 at 20 cm).

2.4. Concentrations of organic carbon are highest on rocky hilltops

(range, 4.78 to 11.03 per cent), and lowest in profiles deeper than 60 cm (range, 0.80 to 2.16 per cent at 10 cm).

2.5. C.E.C. varies directly with organic carbon. The observed range of rocky hilltop soils is 40.29 to 48.90 m.e./100 g, whereas deeper soils mostly have C.E.C. values of less than 20 m.e./100 g at 10 cm.

2.6. Calcium and magnesium are the main exchangeable bases. The Mg:Ca ratio of exchangeable cations correlates directly with that of the parent limestone. The Mg:Ca ratio generally decreases with depth. This trend is indicative of the preferential absorption of magnesium from soils which are relatively deficient in this element.

2.7. The average K:Na ratio of exchangeable cations ranges from 1.79 to 5.82. The ratio decreases with depth. Evidence suggests that plants absorb potassium in disproportionately large amounts.

2.8. The base content generally increases with depth, towards the weathering front. Additionally, in certain soils surface inputs of cations cause a higher pH in the uppermost horizons. Highest average values occur in the foothill soils (range, 7.14 to 7.23), whereas the lowest figures were recorded in the deeper soils of the hilltop depression and moderately-inclined hill slopes (range, 5.50 to 6.46). Organic acids may favour the leaching of bases from the shallow soils of the rocky hilltops and steep hill slopes (pH range, 6.61 to 6.73).

2.9. Average carbonate contents of 3.0 and 3.5 per cent were observed in two rocky hilltop plots. Elsewhere the mean concentration is generally less than 1.0 per cent (minimum, 0.21 per cent).

3. Rainfall interception

3.1. Average amounts of interception vary directly with the weighted basal area index. Values range from 18.9 (WT-I) to 34.3 per cent (GKV-I).

4. Chemical properties of throughfall waters

4.1. Concentrations of calcium (mean, 4.79 ppm), magnesium (mean, 1.16 ppm), and potassium (mean, 5.27 ppm) in throughfall waters of the study plots are considerably higher than in nearby rain forests. This probably reflects the higher base status of the karst soils.

5. Litterfall

5.1. In the North-west region (plots WT-I and GKV-I), litterfall during the year varies inversely with mean monthly rainfall. Elsewhere, no relationship between litterfall and rainfall has been firmly established.

5.2. Amounts in the moderately-inclined hill and foothill plots range from 5.189 (GT-II) to 12.048 t/ha/yr (GKV-I). Overall, annual litterfall correlates directly with the basal area of tree trunks. However, data from the foothill plots suggests that the leaf area of vegetation remains almost constant following the felling of larger trees, presumably due to rapid regrowth in canopy gaps.

5.3. The proportions of ash (mean, 12.68 per cent), calcium (mean, 4.23 per cent), and magnesium (mean, 0.39 per cent) in leaf litter vary directly with the base content of the soil, the average values in each case being considerably higher than in adjacent rain forests on non-carbonate rocks. Concentrations of potassium in leaf litter (mean, 0.38 per cent) correlate positively with levels of exchangeable potassium in the soil.

6. Inputs of nutrients at the soil surface

6.1. The total input of calcium at the soil surface in the six plots which averages 364.28 kg/ha/yr, is approximately 35 times greater than the corresponding figures from the Pasoh F.R. Inputs of magnesium (mean, 46.85 kg/ha/yr) and potassium (130.23 kg/ha/yr) are also much higher.

7. Nutrient up-take by plants

7.1. Studies of the relative proportions of the major cations in the various components of nutrient flow and storage demonstrate clearly that potassium is preferentially absorbed by plant roots in all six plots, and that magnesium is absorbed in preference to calcium from magnesium-deficient soils.

7.2. The results also suggest that disproportionately high up-take of calcium may occur where karst soils are relatively rich in magnesium.

CHAPTER V

SOLUTION ON EXPOSED ROCK OUTCROPS AND BENEATH SHALLOW ORGANIC-RICH SOILS AND LITTER COVERS

CHAPTER V. SOLUTION ON EXPOSED ROCK OUTCROPS AND BENEATH SHALLOW ORGANIC-RICH SOILS AND LITTER COVERS

A. INTRODUCTION

A high proportion of chemical denudation in karst terrain occurs close to the surface (Jennings, 1972b, 1972c; Pitman, 1978a; Pitty, 1968b; Williams, 1968). The surface and near-surface zone is subject to seasonal, diurnal and less regular fluctuations in temperature and moisture content. It also exhibits great spatial diversity. Thus, within an area of 1 ha, differences in aspect, degree of exposure, and soil and vegetation characteristics, may create a complex mosaic of weathering environments (Jakucs, 1977, p. 152). Within such an area, however, variations in macro-climate and lithology are usually so small that these factors may be regarded as constants. Consequently, bare limestone outcrops and soil/rock interfaces provide sites at which a range of geomorphological factors may be investigated under partially-controlled conditions. Rock outcrops and shallow soils are characteristic features of summits and ridge-crests in the three field areas, and afford ample opportunity for studying short-term, micro-scale variations in solutional activity. The present study is based entirely upon field observations and measurement. However, the interpretations advanced in this, and also in the following chapters, rely heavily upon established theory.

1. Literature review

1.1. Previous field studies

Bare and thinly-covered limestone surfaces, fretted by superficial solution features, have been described from many karst regions (Bögli, 1960; Cvijić, 1924; Fellows, 1965; Jennings, 1971; Pluhar and Ford, 1970; Smith and Albritton, 1941; Thomas, 1970; Wall and Wilford, 1966a; Wilford and Wall, 1965; Williams, 1966).

In several cases, morphometric procedures have been used to quantify their forms (Duncan, 1979; Goldie, 1973; Langridge, 1971; Lundberg, 1977b).

In many areas, relationships have been observed between the gradient of rock outcrops and the development of surface forms. For example, solution runnels (Ger. Rinnenkarren) in the Dinaric Alps occur mainly on moderately-steep slopes (Cvijić, 1924), whilst near the Gardesee, solution flutes (Ger. Rillenkarren) attain their greatest length at angles of between 50 and 60° (Kaaden, 1975 - cited by Watts and Trudgill, 1979). Increases in the length of solution flutes with gradient (up to 60°) have also been found in laboratory simulation experiments (Glew, 1977). The flow characteristics of surface runoff vary with gradient, and may significantly affect both the solution rate (Miotke, 1972b) and the type of solutional form which may develop. Shallow subcircular pits are thus characteristic of surfaces of less than 3° in the Sierra Blanca, Texas (Smith and Albritton, 1941), and ripple-like features in Sarawak are restricted to vertical and overhanging surfaces, up to 60° from the vertical, located beneath seepages (Wall and Wilford, 1966a, 1966b). In contrast, Lundberg (1977b) found no significant relationship between slope angle and the morphometric characteristics of solution flutes in the Chillagoe karst.

Lithological and structural variations in limestones significantly affect meso- and micro-relief forms in karst regions (Cvijić, 1924; Sweeting and Sweeting, 1969; Thomas, 1970). In Puerto Rico, for example, purer limestones generally give rise to sharply-etched solutional features, while the forms developed on impure limestones are smoother (Miotke, 1972). More rounded forms are also characteristic of soil-covered karst surfaces (Bögli, 1960; Pluhar and Ford, 1970). The roles of biological agents in the weathering of limestone surfaces have not been investigated

fully. However, the effects of fungi, bacteria and other terrestrial micro-organisms (Smyk and Drzal, 1964), molluscs and other marine organisms (Trudgill, 1976a), lichen (Fry, 1922; Jones, 1965; Sweeting, 1966) and tree roots (Wall and Wilford, 1966) have been reported in literature.

Data are available from several sources on the chemical quality of waters in pools on exposed rock outcrops (eg. Balázs, 1968; Jennings and Sweeting, 1963; Sweeting, 1966; Williams, 1968), and of waters flowing over bare and covered limestone surfaces (eg. Aubert, 1967; Bauer, 1964; Bögli, 1960; Gams, 1966; Jennings, 1978; Lehmann et al., 1956; Muxart et al., 1969; Smith, 1969b). In many of these studies, a small number of sites were sampled, often at a single point in time, and therefore the results do not cover completely the range of spatial and temporal variability present. On two occasions, however, waters were sampled at different stages during a storm (Bauer, 1964; Muxart et al., 1969). By monitoring the quantity and solute content of runoff over a two-year period, Jennings (1978) has shown that surface lowering at Cooleman Plain, New South Wales, is greater on bare than soil-covered surfaces, the difference being attributable to the higher rate of evaporational and evapotranspirational water loss in the latter case.

1.2. Theory of limestone dissolution

The reaction kinetics of the $CaCO_3$-CO_2-H_2O system are fairly well-established (Garrels and Christ, 1965; Krauskopf, 1967; Picknett, 1973; Roques, 1969). Although there are disagreements concerning the values of certain equilibrium constants (Picknett, 1973), and the significance of ion-pairs, especially $CaHCO_3^+$ and $CaCO_3^o$ (contrast work by Picknett (1973) and Wigley (1971) with that of Jacobson and Langmuir (1974a), Langmuir (1968), and Roques (1964, 1969)), the composition of dilute solutions saturated with calcite may be predicted to within 7 per cent

(Picknett et al., 1976). The reactions outlined in figure 30 and table 56 will provide the theoretical basis for interpretation in the present study. The effect of foreign ions upon calcite solubility, and the mechanisms which control the rate at which saturation proceeds from aggressive or supersaturated solutions are less well-known.

1.2.1. Effect of foreign ions. In addition to carbonic acid, natural waters frequently contain other acids which augment the dissolution process. Inorganic acids are often present in rainwater, especially in regions close to sources of atmospheric pollution (Likens et al., 1972). They may also originate from exchange reactions during the absorption of the anions of neutral salts by humic acids (Fetzer, 1946), and as weathering products of naturally occurring minerals. For instance, sulphuric acid may result from the oxidation of sulphide ores (O'Reilly and Bray, 1974). Organic and humic acids, produced during the decomposition of organic litter, may further increase calcite dissolution. The former are soluble in water and release H^+ ions directly into solution. Certain organic acids, such as tartaric acid and citric acid, are much more effective than others in dissolving calcium carbonate (Trudgill, 1977d). In contrast, humic acids are insoluble in water and affect calcite dissolution by the formation of complexes with Ca^{2+} ions (Fetzer, 1946; Roques, 1969). Furthermore, humic acids, and also organic acids of high molecular weight, are readily oxidized to form organic acids of lower molecular weight:

$$\text{HHumate} \longrightarrow n\text{HOrg} + x\text{CO}_2 + \text{"inactive fragments"}$$

The dissolutional potential of karst waters is augmented indirectly by the increase in the number of true organic acid molecules present and by the release of carbon dioxide (Bray, 1975, 1977a):

$$n\text{HOrg} + n\text{CaCO}_3 \longrightarrow n\text{Ca}^{2+} + n\text{HCO}_3^- + n\text{Org}^-$$

$$x\text{CO}_2 + x\text{H}_2\text{O} + x\text{CaCO}_3 \longrightarrow x\text{Ca}^{2+} + 2x\text{HCO}_3^-$$

(All equations after Bray, 1977a, p. 68).

Magnesium ions, which are present in most natural waters, may significantly affect calcite dissolution. Where they are derived from salts other than carbonates, the dissolution potential of water will be enhanced by the formation of the ion-pairs, $MgHCO_3^+$ and $MgCO_3^o$ (Garrels and Christ, 1965). More commonly, Mg^{2+} ions in karst waters originate from the dissolution of the carbonates, magnesite and dolomite, and usually calcite solubility will be reduced by the common ion effect (Ek, 1973; Picknett, 1977b). However, recent experiments have shown that calcium solubility is enhanced in the presence of small proportions of magnesium, up to a Mg:Ca molar ratio of 0.30 (Picknett and Stenner, 1978). The cause of this apparent anomaly is unknown. The incongruent dissolution of calcium and magnesium from dolomites, and from calcites with magnesium carbonate in solid solution, has been widely discussed in the literature (Berner, 1971; Bricker and Garrels, 1967; Chave and Schmaltz, 1966; Douglas, 1965; Langmuir, 1971; Nicod, 1972; Wigley, 1973; Yanat'yeva, 1954).

Other foreign ions may form ion-pairs or complexes with the dissolution products of calcite (Ek, 1973). The SO_4^{2-} ion, for example, combines with Ca^{2+} to form free undissociated molecules of calcium sulphate, $CaSO_4^o$ (Garrels and Christ, 1965; Reardon and Langmuir, 1976), and thus increases the level of calcite solubility to above that of the pure $CaCO_3$-CO_2-H_2O system. In contrast, traces of certain metal ions, including lead, copper, manganese and scandium inhibit calcite dissolution (Nestaas and Terjesen, 1969; Terjesen et al., 1961). Even a very pure calcite, semi-optical grade Iceland spar, has been found to release inhibiting ions which reduce the dissolution rate as equilibrium is approached (Plummer and Wigley, 1976). Trace metals also affect the crystalline form calcium carbonate which results from deposition. Aragonite, for instance, is favoured by the presence of lead, strontium and

copper (Eraso, 1977; Wray and Daniels, 1957). A further effect of foreign electrolytes is to increase the overall ionic strength of the solution and thereby reduce the activity of individual ions (Debye-Hückel effect, Akin and Lagerwerff, 1965a). However, the total concentration of salts in karst waters is usually so small that this effect is relatively unimportant (Picknett, 1977b).

1.2.ii. Rates of chemical reactions. The distribution of dissolution and deposition within karst outcrops is affected by the rate at which chemical equilibrium is attained in water which is under- or supersaturated with respect to calcite. This depends upon individual reaction rates, and upon the speed with which ions and molecules move through the water towards or away from the rock surface and air/water interface. The effects of temperature are well-known. At higher temperatures, chemical reaction and diffusion rates are both increased.

Certain reactions (labelled 'fast' in fig. 30) take place almost instantaneously and, therefore, are not rate-limiting. The rates at which carbon dioxide diffuses through the air, and at which it may be transferred across the air/water interface, are thought to be fast compared with the speed of diffusion through water (Picknett et al., 1976). The latter process is retarded somewhat by the slow conversion of CO_2 to H_2CO_3 (Kern, 1960). The response time (ie. the time taken to achieve 90 per cent of the total change in the system after a sudden change in conditions) for the relationship between gaseous carbon dioxide and carbonic acid is of the order of 4 hours (Roques, 1964, 1969). A response time of between 200 and 250 hours has been determined for the reaction between calcite and water in a system with P_{CO_2} equal to 0.008 atmospheres, and where the ratio of surface area of soil to volume of water was 5 m^{-1} (Roques, 1969). In the case of deposition from slightly supersaturated solutions, the response time is 2 to 4 months. It

would appear, therefore, that chemical exchange at the calcite surface is rate-limiting under conditions of solution and deposition.

Reactants and products migrate towards and away from the calcite surface by diffusion and advection. On purely theoretical grounds, Picknett et al., (1976) demonstrate that the rate of diffusion is more rapid than the rate of reaction at the calcite surface both in still water, and under conditions of streamline and turbulent flow. This conclusion is supported by the experimental studies of Howard and Howard (1967), by measured rates of calcite dissolution (Morse and Berner, 1972) and by the microscopic form of solution-etched calcite surfaces (Berner, 1978). In contrast, however, the results of other experimental studies suggest that mass transfer, controlled by flow characteristics, may be the principal rate-determining mechanism (eg. Kaye, 1957; Watts and Trudgill, 1979; Weyl, 1959). Thus, even under controlled laboratory conditions, it has not been established with certainty whether reaction rate or ionic/molecular movement is rate-limiting. Indeed, the results of recent simulation experiments, using 1 per cent HCl, suggest that the rate-limiting mechanism may vary according to flow velocity (Watts and Trudgill, 1979).

2. Outline of objectives

The aims of the project described in this chapter are threefold:

1. To report on and interpret the spatial and temporal variability of the solute content of runoff from bare and partially-covered rock surfaces within a small area (30 x 40 m).
2. To provide estimates of short- and long-term erosion rates within this area from solute measurements.
3. To compare these erosion estimates with those determined by the erosional weight-loss tablet technique.

B. FIELD AREA

The area selected for detailed investigation comprises plots III and IV on Ak. Bt. Takun, Selangor. The site was chosen because, although its meso- and micro-relief are typical of karst hilltops and ridge-crests in all three field areas, the summit of the hill is relatively low and accessible at an altitude of 170 m. The hill also lies close to the autographic rain gauge established in Templer Park, and within 30-minutes' drive of the main laboratory base in Kuala Lumpur.

The limestone is a finely-crystalline marble of low porosity (density, 2.65 g/cm^3) and high purity, the mean acid insoluble residue being 0.31 per cent by weight. It comprises 98.53 per cent $CaCO_3$ (range, 98.43 to 98.79 per cent) and 1.16 per cent $MgCO_3$ (range, 0.68 to 1.39 per cent), the molar ratio Mg:Ca averaging 0.014 (range, 0.008 to 0.017). The above values have been corrected for analytical error, on the assumption that the soluble component of the limestone is composed entirely of calcium and magnesium carbonate. Thus, the rock is composed mostly of calcite, and under these circumstances the magnesium is present as magnesium carbonate in solid solution within the calcite (Douglas, 1965). Slight variations in chemical composition are present, but these are so small that rock type may be regarded as constant within the area.

Almost two-thirds of the surface of the 40 x 30 m plot comprises bare rock (table 57; fig. 18; plate 14). Everywhere, except on overhanging sections, the rock is grooved by solution features (plates 15 and 16). In places, these are smooth rounded forms, similar to the solution runnels described by Bögli (1960). In others, they are irregular, jagged and sharply etched, being frequently separated by arête-like ridges (plate 18; cf. small-scale features of karst in Sarawak, Wilford and Wall, 1965).

Occasionally, as in Sarawak, the bases of the more sharply-fretted features contain small solution cups which fill with water during rains. Many bare rock surfaces, including certain vertical and overhanging cliffs, are pitted by small circular depressions (plate 17; detailed discussion, Section D2.2.i, below).

The remaining one-third of the surface comprises mainly organic-rich soils (table 57). True mineral soils are absent. Five samples each of organic-rich mineral soil and decaying organic litter had pH values (water, 1:2.5) of 6.7 (range, 6.1 to 7.2), and 5.6 (range, 5.2 to 5.9), respectively. In more favoured locations with deeper soils, small, stunted trees form a discontinuous canopy (detailed discussion, Chapter IV, Sections C1.2.iii and C1.3.iii). Elsewhere, low herbs and shrubs, rooted in rock crevices and small pockets of soil, are the dominant flora.

C. EXPERIMENTAL DESIGN

1. Important variables

Since the study was limited to a very small area (40 x 30 m) several factors, including rock type and macro-climate, were naturally held constant. Nevertheless, a wide range of variables had to be considered in the experimental design.

1.1. Aspect

In July, when the investigation was undertaken, the midday sun was almost directly overhead. Contrasts between sunny and shaded rock surfaces were therefore small. Differences in insolation receipt between east- and west-facing surfaces were minimized by selecting sites in canopy gaps (Section C1.2, below), since such areas were often in shade except at around noon. By chance, all the storms studied happened to begin between 11.00 and 16.00 hrs, thus further reducing insolation differences.

In areas where rain-bearing winds are from one dominant

direction, accelerated fretting may occur on the windward sides of rock outcrops due to their greater precipitation receipt (eg. N. Pennines, Sweeting, 1966, p. 198). On Ak. Bt. Takun, however, winds are characteristically of low velocity and irregular in direction. Consequently, the fall of rain drops deviates little from the vertical, and precipitation input may be assumed constant, irrespective of aspect.

1.2. Tree canopy

Tree canopies considerably modify precipitation inputs (cf. tables 22 and 48; Chapter IV, Sections E1 and E2) and, therefore, will affect solution processes. However, because of the difficulties of quantifying throughfall, and as such an input will vary irregularly beneath trees, sampling was restricted to those areas without canopy cover.

1.3. Surface characteristics

Almost the entire surface, apart from overhangs, is grooved by solution runnels (plates 15 and 16). During rain, diffuse runoff from rock surfaces becomes concentrated naturally into small trickles along the lower parts of the runnels. Such trickles are convenient to sample, and their 'micro-catchments' may be readily delimited. The morphologies of the runnels and their amounts of surface cover vary greatly within the sampling area.

2. Site selection

A completely random sampling framework was impracticable because of the sheer inaccessibility of some rock surfaces and the difficulty of relocating widely-scattered sites established in such terrain. The following procedure was adopted. Six accessible points, free from a canopy cover, were located by random co-ordinates. Within a radius of 5 m from each point, ten solution flutes were then selected so as to cover the range of micro-catchment types present. A number of variables, including length, width, depth,

gradient, and proportion of soil/litter cover (see footnotes, table 58), were measured in each catchment. Extreme care was taken not to disturb the surface. Previous work had suggested that catchment size, gradient and cover might be important geomorphic controls. In order to focus attention upon these, sites which differed with respect to one of the three factors, but which were otherwise similar, were selected for closer investigation. These sites, three for each factor, were taken mostly from the original 60, but it was necessary to include a further four sites. Table 57 provides an indication of the representativeness of the sampling sites. Except for the deeper soil category, there is a close correspondence between the 64 micro-catchments and the four transect lines.

3. Characteristics of the micro-catchments

The 64 micro-catchments sampled range from 0.01 to 2.52 m^2 in area, from 23.0 to 85.0^o in gradient, and from 0 to 95 per cent in terms of soil/litter cover (table 58). They therefore encompass a broad range of weathering environments.

In a number of micro-catchments, the choice of collection point was limited to the availability of suitable cross-sections in which the water sampling equipment (fig. 15) could be installed. In others, particularly on fallen blocks, runnels are truncated at their downstream end. Consequently, the micro-catchments sampled are not necessarily entire runnels, nor are they all fully-developed forms. For these reasons, relationships observed between morphometric variables must be interpreted with caution. A significant correlation was found between catchment gradient and the percentage soil/litter cover (r=-0.685, $p<0.001$). Apart from this, few significant relationships were found, other than those which reflect the overall scale component.

4. Sampling programme

Sampling devices were set up in each of the micro-catchments, and samples were collected immediately after each downpour. Five storms were sampled (table 59). During the third, which chanced to have the highest intensity, samples were taken at 10-minute intervals from nine of the catchments, After the final storm, the soil/litter cover of each catchment (sub-samples from larger sites) was collected to determine the mass of soil/litter present.

5. Assessment of precipitation input

Because storms were very localized, and because it was impossible to guarantee the safety of an autographic rain gauge on Ak. Bt. Takun, three temporary gauges (fig. 14) were set up on the top of the hill. The magnitude of inaccuracies which might otherwise have arisen in extrapolating storm data over a distance of 1 km from the main gauge in Templer Park, outweighed the precision sacrificed by using these gauges. The project was scheduled at a time when daily fieldwork was being undertaken in the area. In this way, the duration of storms could be established precisely, and samples could be collected immediately after each.

D. SPATIAL VARIATIONS IN RUNOFF CHARACTERISTICS

Since each micro-catchment received identical precipitation inputs, spatial variations in solute characteristics may be attributed to the different morphologies and surface covers of the sites. In order to minimize the temporal component, and thereby focus attention upon the inter-site variability, values have been averaged for each site over the five storms. Simple arithmetic means were used in preference to means weighted according to runoff volumes, for two reasons. First, the latter would mainly reflect the patterns of the two heavier downpours, rather than the full range of storms studied. Secondly, sampling errors are more likely in storms of

higher magnitude (Chapter III, Section B5.1).

1. Properties of surface runoff

The chemical properties of runoff waters are summarized in table 60. Perhaps the most striking feature of the data is the degree of inter-site variability. For example, total hardness ranges from 22.3 to 108.2 ppm, magnesium hardness from 0.5 to 18.7 ppm, and potassium from 0.1 to 5.3 ppm. If the values for individual storms were included, then these ranges would be extended further. Clearly, in view of this inherent variability, data are required for a large number of sites before reliable averages can be obtained. By chance, with a mean of 51.83 ppm and a standard deviation of 21.18 ppm, a sample size of 64 is required to estimate total hardness to within 10 per cent at the 95 per cent confidence level.

Runoff from all 64 micro-catchments is alkaline in reaction, the pH averaging 7.84. Total hardness (mean, 51.83 ppm) comprises mostly calcium hardness (48.01 ppm). Nonetheless, the magnesium hardness (mean, 3.82 ppm) is far higher than would occur if the dissolution of calcium and magnesium from the limestone was congruent. The molar ratio Mg:Ca, after correcting for the inputs of Ca^{2+} and Mg^{2+} from rainwater (table 22), averages 0.073. This is almost an order of magnitude greater than the molar ratio in the rock (table 31). Bicarbonate is the main anion in the waters, with alkaline hardness (mean, 46.79 ppm) accounting for more than 90 per cent of the total hardness. This should be qualified, however, by noting that in addition to bicarbonate ions, other weak acid anions, including some organic acid ligands, may associate during the standard titration for alkaline hardness (Johnson et al., 1977, p. 334). However, according to Picknett et al., 1976, p. 244), most organic acids which are important in limestone solution behave as strong acids.

The potassium content of runoff waters (mean, 1.50 ppm) is more than ten times that of rainfall during the study period (table 22), whereas the sodium content (mean, 0.36 ppm) is approximately double that of rainfall. The molar ratio K:Na in runoff averages 2.45, compared with 0.45 in rainfall.

Relationships between the various chemical properties are shown in table 61. Specific conductance, sodium, potassium, and total, alkaline and non-alkaline hardness are all positively related. Since calcium and magnesium usually form the bulk of cations in karst waters, the high correlation (r=0.955, $p<0.001$) between total hardness and specific conductance may be readily understood. Also, since the bicarbonate ion is usually the main anion in karst waters, the positive correlation between alkaline hardness and total hardness and specific conductance is expected. The interrelationships between non-alkaline hardness, sodium and potassium, and their relationships with specific conductance and total and alkaline hardness are thought to result mainly from their mutual correlation with the amount of soil/litter cover. As predicted from theory, pH varies inversely with water hardness.

Because of the inaccuracies associated with pH determination, no attempt has been made to calculate the precise degree of saturation of the runoff waters. However, some indication may be gained by reference to the equilibrium curves for saturated calcite solutions at 20 and 30°C (fig. 31 - after Picknett, 1972, fig. 2, p. 143). Since the experimental studies reported by Picknett (1972) do not consider the effect of magnesium upon the relationship between pH and calcium hardness for calcium concentrations of less than 1.2 millimoles/litre, no allowance may be made for magnesium content of surface runoff (approximately 5 per cent). However, errors arising from this source are likely to be small, being, at

most, 0.04 pH units (based on Picknett, 1972, fig. 5, p. 147). The points shown in figure 31 show the sample with the median calcium hardness at each of the 64 micro-catchments. Clearly, the majority of samples are undersaturated with respect to calcite, with only four lying above the 20°C curve.

2. Relationships between runoff properties and micro-catchment characteristics

Details of the correlations observed between the chemical properties of runoff and surface characteristics are presented in table 62. Apart from two relatively weak correlations with depth, the chemical properties studied appear to be unrelated to any of the main morphometric variables except gradient. The degree of surface cover, measured either as percentage area or as mass per unit area, is a key factor. Since gradient (Section C.3, above), and to a lesser extent, depth (r=0.284, $p<0.05$) are the only two morphometric variables which correlate significantly with the percentage soil and litter cover, it seems likely that their effect upon the chemistry of runoff waters is indirect. Thus, the accumulation of leaf litter and organic-rich soils is more favoured at lower angles and in deeper solution runnels, and runoff has a correspondingly higher solute content and lower pH. This interpretation is confirmed by the fact that when the bare rock and partially-covered micro-catchments are considered separately, then in neither grouping is there a significant relationship between gradient and water hardness.

2.1. Bare and partially-covered micro-catchments compared

In view of the above observations it is interesting to make more specific comparisons between bare rock sites (n=27; table 63) and those with a partial soil/litter cover (n=37; table 64). It should be noted that the first group includes sites with some lichen cover.

2.1.1. Water hardness properties. Total hardness and specific conductance are significantly higher, and pH significantly lower, in partially-covered than in bare micro-catchments (t-test, $p<0.001$). The range of total hardness at the former sites is 38.8 to 108.2 ppm, compared with 22.3 to 66.8 ppm at the latter. Of the bare rock sites, only one has an average total hardness in excess of 50 ppm, this being by far the largest catchment, with a length of 3.8 m. With increased length of flow, there will be a corresponding increase in contact time between water and the rock surface, and this will allow greater opportunity for the solutional potential of the rainwater to be expended. As noted previously, three sites were selected to allow closer investigation of the effect of surface cover. The micro-catchments chosen, sites 4, 17, and 15, were smooth, rounded solution runnels of similar size and gradient, but with different amounts of surface cover (table 65). Site 4 is bare, site 17 has a 5 per cent cover of leaf litter (pH, 5.7), and site 15 has a 70 per cent cover of soil and litter (pH, 6.5). The total dry weights of soil/litter in the last two micro-catchments are 5 and 190 g, respectively. Total hardness values, averaged over the five storms, are 28.6, 54.0 and 95.0 ppm for these sites, thus confirming the general trend observed above.

At bare rock sites, non-alkaline hardness is low, averaging -0.34 ppm. Negative values reflect a higher content of weak acid anions, principally bicarbonate ions, than may be accounted for by the dissolution of calcium and magnesium carbonate alone. However, the negative values are so small that they may be accounted for by analytical error. The consistently low non-alkaline hardness, with a maximum of 5.1 ppm, suggests that the anions of strong inorganic acids such as sulphate and chloride, and also of the stronger organic acids do not significantly affect solution. Therefore, chemical aggressiveness derived from sources other than carbon

dioxide is slight. In contrast, the non-alkaline hardness content of runoff from partially-covered sites averages 8.97 ppm (ie. 14.2 per cent of total hardness), with a maximum of 22.3 ppm. As anions other than bicarbonate were not determined, the non-alkaline hardness cannot be apportioned to organic or inorganic sources. However, since the input of strong acid anions from rainfall is likely to be quite small, and because the limestone contains few such anions, it seems likely that the non-alkaline hardness is derived principally from the organic fraction of soil and litter layers. Of the average increase in total hardness of 26.9 ppm in the covered, as compared with the bare rock, catchments, 9.3 ppm (34.6 per cent) is accounted for by an increase in non-alkaline hardness, and therefore the source of the remaining two-thirds (17.6 ppm) is likely to be mainly carbonic acid. It is well-established that carbon dioxide levels in the soil and litter layers are higher than in the free atmosphere as a result of microbial activity, and that the increased chemical aggressiveness of rainwater is derived largely from this source (Drake and Wigley, 1975; Rightmire and Hanshaw, 1973).

It should be noted that flow rates are reduced in covered, as compared with bare micro-catchments. The duration of contact between water and rock per unit volume of water and per unit length of flow is thus increased, thereby allowing more time for the solutional potential to be expended. Also, the soil and litter layers contain exchangeable calcium and magnesium ions which may be taken up by percolating waters (table 35). The surface area for ionic exchange is thus greatly increased in partially-covered micro-catchments.

2.1.ii. Mg:Ca ratio. The molar ratio Mg:Ca is higher in partially-covered than in bare micro-catchments. After correcting for the input from rainfall, the values are 0.091 and 0.026, respectively.

Since organic-rich soils in the plot contain relatively high proportions of magnesium (mean Mg:Ca ratio, 0.0220), and as leaf litter in the karst regions is also magnesium-rich (Chapter VI, Section E4.2), a high proportion of the magnesium in runoff from partially-covered micro-catchments is probably derived from these sources. Even at bare rock sites, however, the Mg:Ca ratio is about three times that of the limestone. In part, this may be caused by the incongruent dissolution of limestone. On theoretical grounds, Douglas (1965, p. 28) suggests that such a process may occur in solution runnels on limestones containing dolomite minerals. Also, it is well-established that magnesite is more soluble than calcite. Some magnesium in runoff from these micro-catchments may also be derived from the lichen cover, but further work is required to substantiate this.

2.1.iii. K:Na ratio. The potassium and sodium contents of runoff from bare rock catchments are low, averaging 0.68 and 0.23 ppm respectively. Corresponding figures for rainwater are 0.13 and 0.17 ppm. Since potassium was not detected in limestone from Ak. Bt. Takun, and sodium occurs only in trace quantities, the source of the additional 0.55 ppm potassium and 0.06 ppm sodium is conjectural. Moss, lichen, snails and other plant and animal organisms which may be present on the surface are the most probable sources. At partially-covered sites, potassium (mean, 2.10 ppm) and sodium (mean, 0.46 ppm) contents are significantly higher. Both elements are present in the soil and litter covers (Chapter IV). The molar ratio K:Na of exchangeable cations in the organic-rich soils on Ak. Bt. Takun and of total cations in leaf litter of the karst regions average 5.22 and 8.12, respectively. The molar K:Na ratio increases from 0.45 in rainfall, through 1.74 in bare micro-catchments, to 2.68 at partially-covered sites. In view of the low electrolyte content of the runoff waters, the ionic strength effect

will be negligible, and the overall positive correlations between sodium, potassium and water hardness characteristics noted above (Section D.1), are most probably due to auto-correlation with the amount of soil/litter cover.

2.1.iv. Water temperatures. The temperature of runoff from 5 bare and 15 partially-covered catchments was determined at the beginning of the final storm. Rain commenced at 13.30 hrs after much of the surface had been exposed to direct bright sunlight for several hours. Water temperature averaged 24.3°C, with a 22.7° to 27.1°C range, in the bare rock micro-catchments, and 31.26°C, with a 25.2° to 33.7°C range, at sites with a partial soil cover. Rock, soil and litter temperatures were not determined. It was found, however, that even the very light-coloured rock surfaces, when exposed to direct sunlight during the day, became so hot as to be uncomfortable to touch. The generally higher water temperatures of covered micro-catchments may be accounted for by the higher rates of heat absorbence by the shallow, dark-coloured soils and litter. The temperature of runoff will probably decrease markedly during the early part of storms, as heat is progressively absorbed from the rock and surface covers. Rainwater sampled elsewhere in Selangor was found to have temperatures in the range 20.2 to 23.4°C, and such values are likely to be representative of runoff temperatures during the latter parts of storms

2.2. Total hardness variability between bare rock micro-catchments

Solute variations over the 64 micro-catchments are clearly dominated by differences in amounts of surface cover. In order to eliminate this effect, and thereby assess the significance of other factors, attention is now focused on the 27 bare micro-catchments. In addition to those variables listed in table 58, several others, including the percentage of moss and lichen cover, and the number of snails, small circular depressions and solution

cups were measured. To simplify the discussion, only variations in total hardness are considered. No significant relationship was found between the latter and the amount of moss and lichen cover. The remaining factors are considered below.

2.2.1. Effect of snails. Of the surface variables measured, snail density (no. per 100 cm^2) was most strongly related to total hardness (r=0.406, $p<0.05$). Molluscs are known to be important erosional agents in coastal karst regions (Lundberg, 1977a; Trudgill, 1972b, 1976a), but their effects upon subaerial processes are largely unknown.

The snails on the rocky outcrops are principally of the species Alycaeus (identification by F.C. Naggs, British Museum). They are small, being less than 5 mm in height and diameter, but locally they occur in quite high densities (plate 17). In several micro-catchments their density exceeded 13 snails per 100 cm^2 of rock surface at the time of sampling. A random sample of 25 snails was selected for closer examination. The weights of the shells averaged 8.6 mg, with a 5.2 to 12.4 mg range. The shell material was completely soluble in 3N HCl. Assuming the calcium and magnesium to be present as carbonates, these account for 99.7 and 0.3 per cent by weight, respectively.

Many of the bare rock sites are pitted by small circular depressions, 2 to 5 mm in diameter, 2 to 4 mm in depth, and in places exceeding 150 per 100 cm^2. Such forms occur even on vertical and overhanging rock surfaces and are, therefore, not caused by rain-splash. There is a significant correlation between the density of depressions and of snails (r=0.828, $p<0.001$). Undoubtedly, the depressions have been caused by the snails.

It is well known that various types of mollusc are able to erode limestone either mechanically, by means of their radula, or chemically, by releasing acid secretions (see review by Trudgill,

p. 178, 1976). If the latter were the main mechanism by which snails enhance dissolution in the bare rock micro-catchments, higher values of non-alkaline hardness would be expected where larger numbers of snails are present. No such relationship was found. In view of the high calcium content of their shells, it would seem likely that the snails absorb much of the calcium they dissolve from the bedrock, and that this is only removed from these sites when the organisms die and their shells are washed away by mechanical action. The greater roughness of the pitted rock surfaces (plates 17 and 18) will increase the area of water/rock contact, and produce greater irregularity and turbulence in the flow. From theory, both effects are likely to increase the amount of limestone dissolved per unit length of flow, and per unit time.

2.2.ii. Effect of solution cups. Most of the bare rock catchments are pitted by small, sharply-etched solution cups, numbering more than 20 in certain cases. They range from 10 to 30 mm in diameter, and from 5 to 15 mm in depth. Total hardness is directly related to the number of cups within a micro-catchment ($r=0.323$, $p<0.05$). This relationship may be explained by the fact that the solution cups fill with water during storms, thereby retarding runoff and allowing greater contact time between rock and water per unit length of flow. It is also interesting to note that non-alkaline hardness varies directly with the density of solution cups ($r=0.358$, $p<0.05$), suggesting that the cups may be sources of organic or inorganic acids. As the cups often retain water for a day or so after rainfall, algae often cover the bottom of the pools, and may be the source of such acids. This would account for the preservation of these forms, and their progressive deepening.

2.2.iii. Effect of gradient. No significant correlation was found between total hardness and micro-catchment gradient. Three sites (numbers 12, 42 and 55) were selected for closer examination. Their

lengths, widths and percentage cover are very similar, but their gradients range from 46 to 66° (table 65). Total hardness increases from 37.6 to 44.6 ppm as the gradient steepens from 46° to 54°. This increase (18.6 per cent) is almost in the same proportion as the decrease in rainfall receipt for the two sites (15.4 per cent). With channel lengths virtually the same, it follows that the volume of rainfall input is smaller in steeper micro-catchments. Therefore the depth of runoff waters will be less at steeper sites, and the proportion of water in direct contact with the rock will be correspondingly higher. This mechanism may explain the higher hardness values observed at site 12. Some other control is clearly operative at site 55 (gradient, 66°), since total hardness is lower, averaging 32.2 ppm. At such steep angles, the above effect may be countered by the reduced contact time associated with greater flow velocities. The importance of transit time in bare rock catchments has been observed by Jennings (1978, p. 218). It is also interesting to note that the lengths of solution flutes have been found to increase with gradient up to an angle of 60°, and then decrease (Kaaden, 1975 - cited by Watts and Trudgill, 1979) suggesting that a critical process threshold is crossed at about 60°.

2.2.iv. Effect of micro-catchment size. There was no significant correlation between total hardness and micro-catchment size at the 27 bare rock sites. Sites 11, 2 and 10, were considered in more detail (table 65). They have similar gradients and proportions of surface cover, but vary in size, their lengths being 30, 70 and 190 cm, respectively. No clear pattern emerges from the results. Total hardness increases from 33.9 ppm in the smallest micro-catchment to 36.7 ppm in the largest, but is 29.9 ppm in the intermediate-sized micro-catchment. It thus seems that water flowing over only 30 cm of bare rock has sufficient time to attain a solute level which is barely exceeded where flow is almost 2 m in length.

Possibly depth of flow is an important variable. The volume of flow increases cumulatively down the solution flute, and hence the depth of flow will vary directly with micro-catchment size. Consequently, at larger sites, a smaller proportion of the runoff will be in direct contact with the rock surface, and this may counter the tendency for solute content to increase with longer contact time with the rock.

3. Comparison with other studies

Previous studies of surface runoff from bare rock outcrops in the humid tropics have reported higher hardness concentrations than those found on Ak. Bt. Takun, where the total hardness varies from 22.3 to 65.0 ppm. For example, Cooke (1971) records total hardness values in the range 49 to 107 ppm for runoff from steep limestone blocks in north-east Tanzania, and Lehmann et al. (1956) report a figure of 90 ppm for waters draining a 1 m long solution runnel in Cuba. However, since the chemical and physical characteristics of the limestones, and the duration and intensity of the storms sampled are not stated, the reasons for this difference cannot be isolated. Hardness values in the range 29 to 132 ppm have been reported from temperate regions (eg. Co. Clare, Ireland - Muxart et al., 1969; Cooleman Plain, N.S.W., Australia - Jennings, 1978; Jura, France - Aubert, 1969; Yugoslavia - Gams, 1966). The corresponding range recorded in colder environments is 15 to 88 ppm (eg. Alps, Switzerland - Bögli, 1960; Lapland, Norway - Muxart et al., 1969; Somerset Island, Canada - Smith, 1969b).

The total hardness of runoff from partially-covered micro-catchments in West Malaysia (range, 38.8 to 108.2 ppm), is similar to that observed beneath thin soil covers in other humid tropical regions. Muxart et al. (1969), for example, report values of between 48 and 105 ppm from Jamaica and Puerto Rico. Few comparable data are available from other climatic zones. However,

figures in the range 55 to 235 ppm have been recorded in temperate regions (Jura, France - Aubert, 1967; Cooleman Plain, N.S.W. - Jennings, 1978). Higher hardness concentrations from covered, as compared with bare rock, surfaces have been observed in the Dachstein Alps, Austria (Bauer, 1964) and at Cooleman Plain, N.S.W. (Jennings, 1978).

E. TEMPORAL VARIATIONS IN WATER HARDNESS

Analysis of temporal variations in the hardness of runoff provides further insight into the solution process. Here, comparisons are made between the five storms sampled, and between different stages of a single storm.

1. Inter-storm variations in water hardness

In view of the results obtained in Section D, data from the 27 bare rock and 37 partially-covered sites are grouped separately. The average total and non-alkaline hardness were determined for each of the five storms, for two groups of sites.

1.1. Bare micro-catchments

At bare rock sites total hardness varies inversely with rainfall intensity (r_s=-0.900; fig. 32A), but is unrelated to the amount of rainfall in a particular storm. As the solutional potential of runoff is essentially constant, being derived mainly from atmospheric carbon dioxide, variations in hardness probably reflect the extent to which this aggressivness has been expended. At higher rainfall intensities, greater velocities and depths of flow will reduce both the duration of water/rock contact per unit length of flow, and the proportion of water in direct contact with the rock at any instant in time. The inverse relationship with rainfall intensity is thus readily explained.

1.2. Partially-covered micro-catchments

In contrast to the bare rock sites, the total hardness

of runoff from the covered surfaces varies independently of storm intensity, but correlates inversely with storm magnitude (r_s=-0.700; fig. 32B). In regions with deeper soils than those found on the summit of Ak. Bt. Takun, water hardness values vary directly with rainfall amount (eg. Jennings, 1978; Rossi, 1976; see also Chapter VI). Under such circumstances, rainfall stimulates microbial activity in the soil. It also blocks soil pores in the surface horizons, impeding gaseous diffusion and thereby favouring the build up of carbon dioxide. A different mechanism of solutional control is clearly operative in the field area, since negative relationships have been found between rainfall amount and total hardness.

The following explanation is advanced. Since the soil and litter layers are shallow, the absolute volumes of carbon dioxide and organic acids contained within them are small. They provide, therefore, a limited source of chemical aggressiveness, which is progressively depleted through a storm. Consequently, the greater the magnitude of a downpour, the lower will be the solute content in the final stages of the storm. Though this interpretation is necessarily tentative when based on a sample of five storms, observations of hardness fluctuations during a single storm support such arguments (see below).

2. Intra-storm variations in water hardness

To investigate the temporal pattern of solute changes during a single downpour, water samples were taken at 10-minute intervals during the third storm (table 59) from the nine sites detailed in table 65. The storm chanced to have the highest intensity of the five studied. Rainfall in the five periods (fig. 33) ranged from 2.9 mm (17.4 mm/hr) to 9.2 mm (55.0 mm/hr).

At all nine sites, specific conductance and total hardness are highest at the beginning and end of the storm (fig. 33). The high solute content in the latter part of the storm corresponds

with the period of lowest rainfall intensity, and may be largely attributed to a reduction in the rate and depth of flow. A similar increase in hardness at the end of a storm was noted by Muxart et al., (1969) in their study of runoff from beneath a soil-covered site, Jamaica. The initial peak is not so easily explained. At each site total hardness is greater in the first than in the fourth period, yet the former has the higher rainfall intensity. At sites 15 and 17 (fig. 33A) this may be explained by the high initial solutional potential provided by biological activity in the inter-storm period, or by the washing out of salts from thin films of carbonate-saturated capillary water which may be held in the soil between storms. The development of such carbonate-rich water layers has been termed 'Vorverwitterung' or 'pre-weathering' (Miotke, 1968). However, some other factor is clearly augmenting solution at the bare rock sites during the earliest phase of the storm. There are two possible explanations. First, chemical reaction and diffusion rates may be higher at the beginning of the storm when runoff temperatures are highest (Section D2.1.iv, above). Secondly, a superficial layer of powdery calcite, such as that frequently developed on vertical or overhanging limestone cliffs (Crowther, 1978a) may form between storms. Such porous deposits would be readily dissolved by the first rains, and might result from evaporation at the end of the previous storm or from alternating phases of condensation and evaporation, which are common in the area.

2.1. Effect of a soil/litter cover

On bare rock outcrops (site 4, fig. 33A; fig. 33B/C), specific conductance and total hardness show clear correspondence with rainfall intensity. At site 4, typical of such micro-catchments, total hardness falls from an initial value of 20 ppm to 12 ppm during the period of maximum intensity, and rises to 26 ppm at the end of the storm. Non-alkaline hardness, in common with the

other six bare rock sites, varies little from zero and exhibits no recognizable temporal pattern.

In contrast, at site 15 (fig. 33A), which has a 70 per cent soil/litter cover, specific conductance and total hardness minima occur in the fourth period, and non-alkaline hardness shows a distinct trend, falling regularly until period three and increasing at the end of the storm. These results provide further evidence that the reserves of carbon dioxide and organic acids are gradually depleted, and the solutional potential reduced, through the storm. The slight increase in hardness during the last period is attributed to greater contact time between rock and water under conditions of dwindling flow.

2.2. Effect of micro-catchment size

During the first 40 minutes of the downpour, total hardness was consistently lower at sites 2 and 10 than at site 11, the smallest of the three micro-catchments (fig. 33B). This difference was most marked in the third period of the storm when total hardness at site 11 was 20 ppm, compared with 14 ppm at sites 2 and 10. At times of high precipitation intensity the thicker and faster flowing water films, characteristic of the lower parts of larger micro-catchments, may be sufficient to prevent much of the solutional potential of the rainwater from being realized. However, at lower intensities, during the final period, greater flow length and reduced flow velocity appear to favour relatively high solute levels at the largest site.

2.3. Effect of gradient

The trends in total hardness at sites 55, 42 and 12 are almost parallel throughout the duration of the storm (fig. 33C). Values are highest at site 42, the catchment of intermediate gradient, and this is consistent with the spatial pattern of average values considered above (Section D2.2.iv). The marked inverse

relationship between total hardness and rainfall intensity observed during the storm at these, and at other bare rock sites, lends further support to the suggestion that the influence of gradient upon solute concentration is mainly attributable to variations in volume and velocity of flow.

F. EROSION RATES ON THE SUMMIT OF AK. BT. TAKUN

Erosion rates on bare and partially-covered rock surfaces were estimated independently by two methods: first, from water hardness data, and secondly, by means of the erosional weight-loss tablet technique. As discussed previously (Chapter III, Section B7.4), the latter method is suitable only for assessing seasonal or annual erosion rates, whereas the former may also be used to measure erosion during single storms.

1. Estimation of erosion rates from solute measurements - bare and partially-covered sites compared

1.1. Erosional losses during individual storms

Since runoff from bare micro-catchments is extremely rapid, and the relative humidity is high, especially during storms, evaporation losses from such sites may be considered as negligible. The volumes and solute contents of rainfall, and the hardness of runoff waters are known, as is rock density (2.65 g/cm^3). From these data the average depth of surface lowering in each storm may be calculated (table 66A). Depths range from 0.13 microns to 0.37 microns, and correlate directly with storm magnitude (r_s=0.900). Thus, although rainfall intensity is the main factor which affects water hardness, storm magnitude is the overriding determinant of total erosional loss.

Some water loss through evaporation and evapotranspiration is inevitable from covered rock surfaces. Since, for practical reasons, total runoff volumes were not collected (Chapter III,

Section B5.1), the scale of water loss is unknown. However, since the micro-catchments studied have shallow and incomplete soil/litter covers, it seems likely that the magnitude of such loss is generally small. Even at sites with almost complete soil covers, runoff was observed within several minutes of the beginning of storms. The depths of surface erosion given in table 66B assume no water loss, and therefore represent the upper limits of erosion. A significant positive correlation was found between erosion depth and storm magnitude (r_s=1.00, $p<0.001$), even though water hardness at these sites varies inversely with the latter (Section E1, above). Douglas (1964, p. 468) has demonstrated that rates of chemical erosion in river basins "are largely dependent on the water discharge from the catchment concerned". The data from Ak. Bt. Takun show clearly that this relationship also obtains at a much smaller scale in both bare and partially-covered micro-catchments on exposed rock outcrops.

Three further points should be noted with respect to the 37 sites with partial covers. First, some solutes in runoff are derived from the soil/litter layers, and depths of lowering therefore relate strictly to the soil/litter surface and not to the underlying rock surface. Secondly, since soil/litter covers are only partial, erosion rates obtained include a bare rock component. Finally, mechanical removal of soil particles, and of the shells of dead snails (bare rock micro-catchments), may represent significant, yet undetermined, erosional losses.

1.2. Long-term erosion rates

Records from Templer Park provide data on the magnitude and intensities of storms from 1.9.74 to 31.8.75. The amount of erosion on the summit of Ak. Bt. Takun during this period may therefore be estimated using the regression lines in figure 32. Rainfall totalled 2234 mm, and the average depths of lowering are estimated

to have been 0.0300 mm in bare micro-catchments, and 0.0503 mm at sites with a partial cover. These figures are subject to two further qualifications, in addition to those discussed above (Section F1.1) in relation to single storms. First, data on storms in 1974/75 relate only to average intensities and total magnitudes, as do the regression lines in figure 32. As was seen earlier (Section E2) marked variations in rainfall intensity and runoff characteristics may occur in individual storms. The figures calculated may, therefore, be regarded only as approximations. Secondly, sampling was confined to one month, and the results may not be representative of conditions through the year. However, errors from this source are thought to be small, since seasonal variations of temperature and rainfall are of limited magnitude in the field area.

The pattern of daily rainfall at Templer Park during 1974/75 corresponded closely with that of Kepong (r=0.428), the nearest station for which long-term records are available. Also the total amount of rainfall during the study period was similar at both stations (2234 and 2318 mm, respectively). The long-term, mean annual rainfall at Kepong is 2441 mm (table 19). If it is assumed that the latter figure is representative of rainfall at Ak. Bt. Takun, and that the magnitude, intensity and frequency of storms recorded in 1974/75 are typical of the field area, then estimates of longer-term surface lowering may be made. Under the conditions which obtain on the summit of Ak. Bt. Takun, the rate of surface lowering on bare rock surfaces is approximately 32.8 mm/1000 yrs, and beneath a partial cover of shallow organic-rich soils and litter is, at most, 54.9 mm/1000 yrs.

2. <u>Long-term erosion rates estimated by the erosional weight-loss tablet technique</u>

Six rock tablets were sited on the summit of Ak. Bt.

Takun, three on bare limestone surfaces and three within shallow organic-rich soil and leaf litter. They were installed on 1.9.74 and remained in situ for exactly one year (rainfall, 2234 mm). Each tablet was laid horizontally. Consequently, erosion was concentrated on the top and sides, the effective surface area being 38.81 cm^2. Erosional loss over this period, calculated as the depth of surface lowering, averaged 0.89 and 1.32 microns, respectively, for the bare and partially-covered sites. Assuming that the depth of weathering is directly related to the amount of rainfall (mean annual rainfall, 2441 mm), then the corresponding rates of erosional loss are 0.97 and 1.45 mm/1000 yrs. Groups of three tablets each were also left on rock surfaces and beneath shallow litter covers in plot WT-I over the period 21.9.74 to 9.5.75, when 1281 mm of rain fell. With a mean annual rainfall at Kaki Bukit of 1859 mm, then surface lowering at the bare and covered sites amounted to approximately 0.94 and 1.10 mm/1000 yrs, respectively. The significance of the lower erosion rates estimated by the erosional weight-loss tablet method is considered in Chapter VI, Section F4.

3. Comparison with other studies

In view of the above results, erosion estimates based on the erosional weight-loss tablet technique are excluded from discussion. The rate of surface lowering on exposed rock surfaces on Ak. Bt. Takun (32.8 mm/1000 yrs) is similar to that reported for bare outcrops of massive brecciated limestones in South Wales (range, 25.0 to 35.0 mm/1000 yrs - Thomas, 1970), of Carboniferous limestone in Yorkshire (42.0 mm/1000 yrs - Sweeting, 1966) and of calcarenites on Aldabra Atoll (39.0 mm/1000 yrs - Trudgill, 1976b). In contrast, values of 9.3 and 10.0 mm/1000 yrs, respectively have been determined for rock outcrops in Co. Clare, Ireland (Williams, 1966) and at Cooleman Plain, N.S.W. (Jennings, 1978), and a figure of 90.0 mm/1000 yrs has been determined for coral fragments on Aldabra (Trudgill, 1976b). Few data are available for erosion rates beneath shallow soil covers, and these display a wide variability.

The lowest figure, 5.3 mm/1000 yrs, was reported from beneath a 25 to 30 cm rendzina soil, Cooleman Plain (Jennings, 1978), whereas the highest, 12500 mm/1000 yrs, was measured beneath a deep organic soil on Aldabra (Trudgill, 1976b). The above data base is inadequate to permit detailed interpretation of inter-regional contrasts.

G. SUMMARY OF RESULTS

1. Spatial variations in properties of runoff from bare and partially-covered limestone surfaces

1.1. Within an area of 40 x 30 m, the properties of surface runoff from solution flutes exhibit much variability.

1.2. Runoff waters are mostly undersaturated with respect to calcite.

1.3. Most of the variability may be accounted for by differences in surface cover. Specific conductance, total, alkaline and non-alkaline hardness, and sodium and potassium concentrations are significantly higher in runoff from covered as compared with bare rock surfaces. Molar ratios Mg:Ca and K:Na are also higher.

1.4. The Mg:Ca ratio of runoff waters is consistently higher than that of the limestone. In partially-covered micro-catchments, magnesium may be derived from the overlying soil and litter. On bare rock surfaces, incongruent dissolution of the limestone is the most probable cause.

1.5. At bare rock sites water hardness varies directly with surface roughness, this being caused principally by small circular pits, enlarged by snails, and solution cups. The effects of flow length and gradient have not been fully established. However, in a set of more specific comparisons, total hardness in one 30 cm-long micro-catchment was almost as high as at a site of 180 cm in length, and was higher in one micro-catchment of 54° than at otherwise similar sites with gradients of 46 and 66°.

2. Temporal variations in water hardness

2.1. Total hardness differs considerably between storms, varying inversely with rainfall intensity on bare rock surfaces, and inversely with rainfall amount where a soil/litter cover is present.

2.2. Similar relationships were observed within a single storm. Additionally, hardness values were higher at the beginning of the storm, due either to higher water temperatures or to the water readily dissolving calcite deposited during the inter-storm period, and also at the end of the storm during dwindling water flow.

3. Erosion rates

3.1. In both bare and partially-covered micro-catchments, erosional loss during individual storms varies directly with the rainfall amount.

3.2. Long-term erosional loss is estimated to be 32.8 mm/1000 yrs on bare rock surfaces, and, at most, 54.9 mm/1000 yrs beneath a partial cover of shallow organic-rich soil and litter.

3.3. Erosional loss recorded by the weight-loss tablet technique is approximately 1/35 th of that determined by the solute method.

3.4. Surface erosion rates in West Malaysia are similar to those reported from other karst regions in temperate and humid tropical environments.

CHAPTER VI

SOIL CARBON DIOXIDE AND THE SOLUTION OF LIMESTONE BENEATH MINERAL SOIL COVERS

CHAPTER VI. SOIL CARBON DIOXIDE AND THE SOLUTION OF LIMESTONE BENEATH MINERAL SOIL COVERS

A. INTRODUCTION

1. General background

Mineral soils developed on the karst outcrops vary considerably in depth and in their physical and chemical properties (Chapter IV). Such differences may cause spatial variations in the carbon dioxide concentrations of the soil atmosphere, thereby affecting rates of chemical denudation. In order to test this hypothesis the carbon dioxide content of soil air was monitored regularly at sites with different pedological characteristics, and the chemical properties of runoff at the soil/rock interface and of groundwaters (Chapter VII) were determined. Additionally, the relative solutional potentials of these sites were assessed by means of erosional weight-loss tablets. The spatial and temporal variations observed are explained by reference to published studies of root and microbial respiration and established theories of soil aeration.

2. Literature review

2.1. Soil respiration

Plant root respiration (Grable, 1966) and the activity and population dynamics of micro-organisms in soil are dependent upon a wide range of environmental factors (see reviews by Gray and Williams, 1975; Stotzky, 1972). Bunnell et al. (1977) have shown that between 70 and 96 per cent of variations in the response rates in soil respiration may be accounted for by temperature, moisture, oxygen and an empirically-determined 'substrate' factor. The number of soil micro-organisms, and hence their productivity, varies directly with temperature (Anderson, 1973; Monteith et al., 1964; Reiners, 1968).

Soil moisture has two effects. First, water absorption by plants and micro-organisms is facilitated at lower moisture tensions, as is shown by the positive relationship observed in many studies between microbial growth and moisture change (Bhaumik and Clark, 1947; Campbell and Biederbeck, 1976; Shameemulla et al., 1971). Secondly, and acting counter to the first, gaseous diffusion is reduced at higher moisture levels and therefore microbial activity may decline as the available oxygen reserves are depleted (Griffin, 1972; Miotke, 1974). Other factors, such as pH, nutrient availability and oxidation-reduction potential, are also important variables (Stotzky, 1972).

As a result of the enormous range of habitats present in the soil, populations of micro-organisms exhibit marked spatial and temporal variability. For example, micro-organisms are generally more abundant in the organic-rich surface horizons and in the rhizosphere than in the remainder of the soil (Rovira and Davey, 1971). The microbial population is extremely sensitive to changing environmental conditions. Thus, in addition to seasonal fluctuations in soil respiration (Anderson, 1973; Currie, 1975), responses to sudden weather changes (Campbell and Biederbeck, 1976) and to diurnal temperature rhythms (Currie, 1965, 1971; Peterson and Billings, 1975) have also been observed.

2.2. Soil aeration

Baver (1956), Grable (1966), Miotke (1974) and Stolzy (1971) present comprehensive reviews of soil aeration theory. Diffusion is the main mechanism of soil gas exchange (Buckingham, 1904; Romell, 1922, 1935), though locally, air turbulence (Kimball and Lemon, 1971) and other types of mass flow may be significant. Gaseous diffusion in soils generally varies with air porosity (Blake and Page, 1948; Currie, 1960, 1961, 1962; van Bavel, 1951, 1952). When the latter falls below approximately 10 per cent, air-

filled passages become blocked by water and diffusion effectively ceases (Baver and Farnsworth, 1940; Vershinin and Poyasov, 1954 - cited by Zborishchuk, 1977; Wesseling and van Vjik, 1957).

2.3. Field studies of soil carbon dioxide levels

Several studies of carbon dioxide have been made in karst regions. In Jamaica, for example, carbon dioxide levels are higher in depressions than on adjacent slopes (Day, 1976) and the average concentrations of carbon dioxide at 30 cm depth range from 0.72 per cent in a cultivated loam soil to 1.94 per cent in a forested clay soil (Nicholson and Nicholson, 1969). Observations in the Yorkshire Wolds (Pitman, 1978a) and in Madagascar (Rossi, 1974) reveal marked seasonal variations in soil atmosphere composition - these being temperature dependent in the first case and moisture dependent in the second. Jakucs (1971) has presented detailed carbon dioxide concentration curves over several daily cycles for sites of different aspect within a doline near Karlovac, Croatia.

By far the greatest volume of published data, however, originates from agricultural and horticultural studies. The effects of different soil types (Vine et al., 1943; Vugakov and Popova, 1968), of different cultivation practices (Pritchard and Brown, 1979) and of seasonal (Boynton and Compton, 1944; Boynton and Reuther, 1938, 1939; Cooper, 1975; de Jong and Schappert, 1972; Vugakov and Popova, 1968) and shorter-term climatic controls (Pritchard and Brown, 1979; Zborischuk, 1977) have been clearly identified.

3. Outline of objectives

The objectives of this chapter are:

1. To report on the carbon dioxide and moisture contents of a range of mineral soils characteristic of those present in the three karst regions and to interpret the spatial and temporal variations observed.

2. To provide estimates of limestone erosion rates immediately

beneath these soils under a range of environmental conditions, and to interpret the spatial variations identified.

B. EXPERIMENTAL DESIGN

Studies were undertaken in the six main plots (table 29).

1. Soil carbon dioxide

Previous studies afforded little guidance in the choice of sampling depths. When sampling chambers were inserted less than 10 cm into the soil they fitted loosely and the surrounding soil was easily disturbed, thereby allowing air to be drawn from the surface during sampling. To minimize such errors, the shallowest depth sampled was 15 cm. Pockets of soil in all six plots were known to be at least 60 cm deep and this was taken as the lowest depth. Changes in the vertical distribution of carbon dioxide through the soil have been found to be most marked in the upper horizons (eg. de Jong and Schappert, 1972). Intermediate samples were therefore taken at 30 cm, a depth for which comparable data were available from Jamaica (Nicholson and Nicholson, 1969).

A set of three sampling chambers was located randomly within each plot. The chambers were sited 20 cm apart. The samples obtained were regarded as being from a single profile. Additional chambers were installed in three plots to assess local variability. Samples were taken from ABT-I and ABT-II at approximately three-week intervals. Elsewhere, because of technical difficulties and the logistics of the fieldwork schedule, fewer samples were obtained.

2. Soil moisture

Ideally, soil moisture profiles would have been determined each time soil air samples were obtained. However, since moisture was determined by gravimetric means, frequent sampling at depth close to the chambers would have significantly disturbed the site.

As a compromise, the following procedure was adopted. Whenever soil air samples were obtained, and also on a number of other occasions, three soil samples (15 cm) were taken at random at a distance of between 20 and 70 cm from the soil chambers. Additionally, in October, January, April and July, samples were taken at 15 cm and at 10 cm intervals down the main profile in each plot, at a distance of between 70 and 100 cm from the chambers. Average moisture profiles were calculated from the latter data.

3. Runoff from the soil/rock interface

Soil water in contact with bedrock was collected in, or adjacent to, four of the plots. At three sites, ABT-I, ABT-II and GT-I, pits were excavated to bedrock and a throughflow collection device was installed. Water was piped through the soil, via PVC tubing, to a collection point 2 m downslope and the original hole was infilled. Samples were collected in 3-litre plastic bottles. The fourth site, located near WT-I, is a semi-natural exposure of the soil/rock interface and uppermost layers of weathered rock. Water was sampled where it seeps from the exposure.

4. Additional measurements

Three sets of erosional weight-loss tablets were installed in each plot at depths of 15, 30 and 60 cm. Also, soil temperature was monitored regularly at 7.6, 15.2 and 30.5 cm in ABT-I. Detailed sampling and analysis of the soils (Chapter IV) was undertaken upon completion of the above sampling programme.

C. SPATIAL AND 'WITH DEPTH' VARIATIONS IN SOIL CARBON DIOXIDE AND MOISTURE

Average concentrations of carbon dioxide and soil moisture were found to vary considerably between plots and at different sampling depths. The aims of this section are to describe and suggest expl-

anations for the variability observed.

1. Variations in the average carbon dioxide contents of soil air

Carbon dioxide concentrations at 15, 30 and 60 cm in the six plots average 0.65, 1.15 and 1.85 per cent, respectively. A more detailed breakdown of the data is presented in table 67.

1.1. Variability within sampling plots

Soil air was sampled from more than one profile in ABT-I, ABT-II and GT-I. Differences in mean carbon dioxide concentrations are generally small. At 60 cm, for example, values average 1.26 and 1.28 per cent in ABT-II and 1.94, 1.78 and 1.81 per cent in GT-I. The largest difference (1.62 per cent) occurs at 60 cm in ABT-I, where the concentrations in the two profiles average 2.96 and 1.34 per cent. Of the nine pairs of sampling depths from which replicate data were obtained, only in the last case given above was the difference significant at the 95 per cent confidence level (randomization test). Clearly the small number of profiles sampled in each plot forms an inadequate basis for estimating intra-plot variability. However, since eight of the nine pairs of means tested do not differ significantly, this provides some reassurance, if only of a qualitative nature, as regards the representativeness of data obtained from a single profile.

1.2. Variations with depth

Mean carbon dioxide concentrations increase with depth in each plot. This may be explained by the longer flow paths for gaseous diffusion from the deeper soil horizons. The magnitude of the increase varies considerably between plots. Thus, in GT-II concentrations rise from 0.35 per cent at 15 cm to 0.73 per cent at 60 cm, a difference of 0.38 per cent, whereas the corresponding figures for GKV-I are 1.32 and 4.21 per cent, a difference of 2.89 per cent. Data from the several profiles in ABT-I, ABT-II and GT-I were combined to provide an overall average for each depth

within each plot. Differences between the mean carbon dioxide contents at the three depths within each of the six plots have been assessed by randomization tests (table 68). In all cases except three the differences are statistically significant at the 95 per cent confidence level.

1.3. Differences between sampling plots

At a given depth, concentrations of carbon dioxide vary considerably between the plots. At 15 cm, for example, the range is from 0.28 (WT-I) to 1.32 per cent (GKV-I), while at 60 cm the range is from 0.73 (GT-II) to 4.21 per cent (GKV-I). According to their mean concentrations, the plots are ranked in the following order:

15 cm: GKV-I>ABT-I>ABT-II>GT-I>WT-I>GT-II

30 cm: GKV-I>ABT-I>ABT-II>GT-I>GT-II>WT-I

60 cm: GKV-I>ABT-I>GT-I>ABT-II>WT-I>GT-II

The statistical significance of the differences between pairs of mean values from the various plots is presented in table 69. Carbon dioxide concentrations at all three sampling depths are significantly higher in GKV-I than elsewhere ($p<0.05$), while at the other extreme, concentrations in WT-I are significantly lower than in all plots except GT-II. Spatial variations in carbon dioxide are thus generally consistent throughout the range of depths sampled. Of the 15 pairs of plots, the number with significantly different mean values increases from 9 at 15 cm depth, to 13 at 60 cm. This suggests that differences in soil aeration become more marked at depth.

2. Variations in soil moisture

2.1. Variability within sampling plots (15 cm)

In ABT-I, ABT-II and GT-I soil moisture at 15 cm was monitored in each of three profiles. Within these plots the range

of mean values is small, the highest being 5.02 per cent (ABT-I). Differences between pairs of replicate profiles in each plot are not significant at the 95 per cent confidence level (randomization test). Overall, soil moisture in the six plots averages 31.0 per cent by volume.

2.2. Differences between sampling plots (15 cm)

Soil moisture varies considerably between the plots (table 70), with mean values ranging from 24.7 per cent in WT-I to 40.2 per cent in ABT-II. The plots may be ranked in the following order according to their moisture content:

ABT-II>ABT-I>GKV-I> GT-I>GT-II>WT-I

Moisture levels in ABT-II are significantly higher ($p<0.05$) than in any other plot (table 71) and, while differences in soil moisture between GT-I,GT-II and WT-I are not significant, the soils in the latter plots are each significantly drier than the others.

The total volume of rainfall which reached the ground during the sampling period was greatest in the Kinta Valley (GT-I and GT-II) and least in the North-west region (WT-I and GKV-I). Since the wettest soils occur outside the Kinta Valley, factors other than the amount of throughfall must be important. Of the six plots, those which have the driest soils, WT-I (24.74 per cent) and GT-II (25.72 per cent), also have the best-developed soil structure and lowest bulk densities (Chapter IV, Section D). The free-draining nature of these soils probably explains their characteristically low moisture contents. Observations in the North-west region support this explanation. The amount of throughfall is greater in WT-I than GKV-I. Also, because a much higher proportion of WT-I is rock covered (41.4 per cent; cf. 6.1 per cent in GKV-I), runoff from rock outcrops will further increase the input of moisture at the soil surface. Despite these factors, the soil is significantly wetter in GKV-I (mean, 34.0 per cent; bulk density at 15 cm, 1.18 g/cm^3) than in WT-I (mean, 24.7 per cent; bulk density at 15 cm, 0.76g/cm^3).

However, this pattern is not repeated in the two plots studied on G. Tempurong. The moisture content in GT-I (mean, 26.1 per cent) does not differ significantly from that in GT-II (mean, 25.7 per cent), despite the soils of the former plot having a much higher bulk density (1.10 g/cm^3 at 15 cm; cf. 0.68 g/cm^3 in GT-II). As noted previously (Chapter IV, Section C5.3), GT-I is located on a small spur, a water-shedding site, and this probably explains the drier soils found here. On Ak. Bt. Takun topographic and geological factors appear to be important. ABT-II is located on the floor of a corridor, a water-receiving site, and may also be underlain by impermeable shales. As a result, the moisture content of soils is significantly higher in ABT-II (mean, 40.2 per cent) than in ABT-I (mean, 35.5 per cent).

2.3. Variations with depth

Soil moisture variations down the main profile in each plot were determined at 3-month intervals. The results, summarized in table 72, show that moisture content generally increases with depth. In GKV-I, for example, moisture levels increase from 34.6 per cent at 10 cm to 45.5 per cent at 100 cm. The correlation (r_s) between soil moisture and depth is positive in all six plots and in four the relationship is statistically significant ($p<0.05$). Evapotranspirational losses probably explain the relatively dry surface and near-surface horizons.

The average moisture contents of the four 3-monthly measurements at 15 cm do not differ significantly from the more frequent determinations made when soil air was sampled. Consequently, the vertical distribution of soil moisture recorded at 3-monthly intervals, when adjusted in the same proportions as that between the two 15 cm means, provides an estimate of the average soil moisture distribution down the main profile in each plot during the sampling period.

3. Variations in air-filled pore space

The porosities of the six soils have been discussed

previously (Chapter IV, Section D2.3). Volumes of air space in the soil may be calculated directly from the moisture and porosity data. In all six soils pore space (tables 40 to 45) diminishes, and soil moisture (table 72) increases, with depth. Consequently, as is seen in figure 34, there is much less air space in the lower horizons. From this graph, the soils may be divided broadly into two groups. Soils of the first group, comprising GKV-I (on impure limestone), ABT-I, ABT-II and GT-I (foothill soils), have relatively small volumes of air-filled space. The average for the four profiles falls from 27.5 per cent at 10 cm to 14.7 per cent at 60 cm. By comparison, volumes of air-filled pore space in the second group, which comprises GT-II and WT-I (hill soils on pure limestones), are greater and the decrease with depth is less marked. The average volumes range from 50.0 per cent at 10 cm to 43.3 per cent at 60 cm.

4. Interpretations of spatial variations in soil carbon dioxide

4.1. Effect of the volume of air-filled pore space

The average volume of air space (percentage) at, and above, each sampling depth was calculated for each of the six main profiles. Relationships between these data and soil carbon dioxide concentrations are shown in figure 35. Negative correlations (r_s) were found at each depth. Thus, compared with the profiles in GT-II and WT-I, soil aeration in the remaining four plots appears to be impeded by the relatively small volumes of air space available for gaseous diffusion. Additional factors are clearly important, however, since two sites (GKV-I and ABT-II, labelled in fig. 35), with similar volumes of air space, have widely different carbon dioxide concentrations at all depths.

4.2 Other factors

In ABT-II the average air-filled pore space falls below 10 per cent, the critical threshold for gaseous diffusion, at depths of 40 cm and more (fig. 34), yet the concentrations of carbon

dioxide are relatively low (mean at 60 cm, 1.26 per cent). One possible explanation is that the average moisture distribution down the profile is not representative of those conditions which limit carbon dioxide production and diffusion at this particular site. During periods of intense or prolonged rainfall the zone of impeded diffusion will rise through the soil, and at times the whole profile may become saturated. For example, on 30.9.74 the soil at 15 cm was found to be saturated following a heavy storm on the previous day when 83 mm of rain fell. The profile in ABT-I was not saturated at this time. Since storms of this magnitude are common in the field area (table 10), waterlogged conditions may frequently obtain in ABT-II, with two consequences. First, at such times the number of aerobic micro-organisms will be reduced, and this may limit microbial numbers and rates of microbial activity during more favourable periods. The relatively high organic carbon content of this soil as compared with ABT-I (tables 44 and 43; see also Chapter IV, Section D3.1), and the temporal pattern of soil carbon dioxide over the sampling period (Section D, below), supports this explanation. Secondly, as the soil drains, following saturation, air will be drawn into the soil by mass flow.

At GKV-I, like ABT-II, volumes of air space are small, yet the mean carbon dioxide concentrations in the former (eg. 60 cm 4.21 per cent) are two to three times those of the latter (1.26 per cent at 60 cm), and are also considerably higher than in ABT-I and GT-I. In contrast with ABT-II, GKV-I is located on a small spur. Thus, although the laterite accumulations below 30 cm in this soil are indicative of temporarily waterlogged conditions at depth, it is unlikely that the upper parts of the profile will become saturated. Under these conditions microbial activity and plant root respiration will not be greatly inhibited. The high concentrations of soil carbon dioxide may be largely attributed to the high levels of pro-

ductivity within this ecosystem (fine litterfall, 12.048 t/ha/yr; cf. 7.867 and 6.411 t/ha/yr, respectively, in ABT-I and GT-I).

5. Comparison with other studies

5.1. Variations of carbon dioxide concentration with depth

In the six soils studied carbon dioxide increases with depth. While similar results have been obtained from soils on a wide variety of non-calcareous parent materials in the tropics (Cooper, 1975; Vine et al., 1943; Wong, 1974) and elsewhere (Boynton and Compton, 1944; Boynton and Reuther, 1939; de Jong and Schappert, 1972; Pritchard and Brown, 1979), existing data from tropical karst regions do not fit this pattern. For example, carbon dioxide levels in a clay soil developed on Yellow Limestone in Jamaica (Site C, Nicholson and Nicholson, 1969) increase from 0.64 per cent at 15 cm to 1.33 per cent at 30 cm, and then decrease to 1.02 per cent at 60 cm. Also, in Madagascar, values in the range 0.04 to 0.06 per cent were recorded at 50 cm in the lower mineral horizons of a limestone soil, whilst the observed range at 20 cm, in the upper, organic-rich horizons, was 0.5 to 3.2 per cent (Rossi, 1974).

5.2. Average carbon dioxide concentrations of karst soils

Inter-regional comparisons of soil carbon dioxide must necessarily be regarded with caution when different methods of soil air extraction have been used (discussion, Section D6, below). Nonetheless, it is interesting to compare the West Malaysian data with results obtained from the same depths in other karst soils. The concentration at 15 cm in the present study (mean, 0.65 per cent; range, 0.28 to 1.32 per cent) is very similar to that recorded by Nicholson and Nicholson (1969) in Jamaica (mean, 0.83 per cent; range, 0.64 to 1.03 per cent). These values are, however, considerably lower than the average figure of 2.7 per cent (range, 2.2 to 3.2 per cent) reported from S. Thailand (Pitman, 1978b).

In Jamaica, carbon dioxide concentrations in the range 0.5 to 2.4 per cent have been found at 30 cm depth on hillslopes (Day, 1976), while an average figure of 1.19 per cent may be calculated from the data presented by Nicholson and Nicholson (1969) for the same depth. The latter authors provide the only data for 60 cm depth (mean, 1.02 per cent), which is based on three measurements from a single profile. These figures are similar to the average values recorded in the present study (30 cm, 1.15 per cent; 60 cm, 1.83 per cent). Carbon dioxide concentrations in karst soils in temperate environments appear to be more variable than those in the humid tropics. On the one hand, a mean concentration of 0.13 per cent (range, 0.065 to 0.32 per cent) has been observed at 15 cm depth beneath grassland at Givendale, Yorkshire, between February and May (Pitman, 1978a). On the other, Jakucs (1977, p. 164) reports a value of 9.0 per cent at 5 to 10 cm depth beneath oak forest in the Bukk Hills, Hungary, in spring.

One of the most striking features of the Malaysian data is the high degree of variability observed between different soils (eg. observed range at 30 cm, 0.37 to 2.48 per cent). Spatial variations of the same order of magnitude have been reported from other karst regions in the humid tropics. In certain cases, as in West Malaysia, the variations are largely attributable to differences in soil texture and structure. In Jamaica, for example, carbon dioxide concentrations at 30 cm are significantly higher in a clay soil with a high humus content and under natural woodland (mean, 1.94 per cent), than in a loam soil with a lower humus content and which has been cleared for cultivation (mean, 0.71 per cent - Nicholson and Nicholson, 1969). Other studies have highlighted the effects of topographic position. Concentrations of carbon dioxide at 25 to 30 cm depth are higher, for instance, in the wetter soils of karst depressions in Jamaica (range, 2.5 to 4.5

per cent) than in the better-drained soils of the adjacent slopes (range, 0.5 to 2.3 per cent - Day, 1976). Relatively high values have also been reported from soils on the alluvial plains at Maros, Sulawesi (McDonald, 1976a). Jakucs (1977) reports even wider variations from Hungary, where carbon dioxide concentrations in April and May in the Bukk Hills and North Borsod region ranged from less than 2.0 to 9.0 per cent.

In view of the magnitude of local variability observed in West Malaysia and other karst regions, it is clear that far more data are required before reliable estimates of carbon dioxide can be made for individual climatic zones.

D. TEMPORAL VARIATIONS IN SOIL CARBON DIOXIDE AND MOISTURE

Wide variations in soil carbon dioxide (table 67) and moisture (table 70) were recorded in each of the six plots during the study period. For example, the carbon dioxide content of soil air at 15 cm in ABT-I varies from 0.15 to 1.38 per cent (s. dev., 0.43 per cent), and soil moisture at 15 cm in ABT-II varies from 34.2 to 69.1 per cent (s. dev., 8.51 per cent). The observed range of soil temperature is small, being 1.0^{o}C at 15 and 30 cm depth in ABT-I (table 18). Consequently, temperature may be regarded as constant. Rainfall (Chapter II, Section B2.1) and litterfall (Chapter IV, Section E3), however, both exhibit marked seasonal, and less regular, fluctuations through the year. These two factors, which directly cause changes in soil moisture and decomposable organic matter, may affect soil carbon dioxide concentrations. In order to assess their significance, the relationships between soil carbon dioxide and antecedent rainfall and litterfall, and those between soil moisture and antecedent rainfall, have been established. Statistical analysis is confined to the three foothill plots (ABT-I, ABT-II and GT-I), since only 4 or 5 samples were obtained from the remain-

ing sites. First, the temporal patterns of moisture and carbon dioxide changes within individual plots are examined.

1. Temporal variations in the carbon dioxide content of soil air

1.1. Variability within sampling plots

Since the number of paired samples obtained from the replicate profiles in GT-I is small, only data from ABT-I and ABT-II are considered in this section. In both plots one matched set of carbon dioxide measurements was made at each of the two depths, 30 and 60 cm (table 67). Positive correlations (r_s) were found between each pair of samples, and in three cases the relationship was statistically significant at the 95 per cent confidence level (eg. at 30 cm in ABT-II, r_s=0.893). These results indicate that carbon dioxide fluctuations in adjacent profiles within a single plot are controlled mostly by a single, or perhaps several, common factor(s).

1.2. Variations with depth

In ABT-II and GT-I carbon dioxide concentrations at the three sampling depths correlate positively during the study period (table 73). In the first plot the correlation coefficients for all three pairs of depths are statistically significant ($p<0.05$). Thus, carbon dioxide concentrations at different depths fluctuate in unison, a pattern which may be explained by gaseous diffusion between soil horizons. The results from ABT-I are, however, in marked contrast. Whilst values at 15 and 30 cm are positively correlated (table 74), the correlations between those sampling depths and 60 cm are negative. The reason for this apparent anomaly is uncertain.

2. Temporal variations in soil moisture at 15 cm

2.1. Variability within sampling plots

In ABT-I and ABT-II soil moisture at 15 cm was determined in three profiles, giving six matched sets of data in all (see table 70). Without exception, the moisture contents of each pair

of profiles within the two plots vary directly, and the correlation coefficients (r_s) are statistically significant at the 99.9 per cent confidence level.

2.2. Relationship between soil moisture and antecedent rainfall

2.2.i. Method of analysis and data presentation. In this section, and in Section D4 (below), reference is made frequently to figures 36 to 39. These show the Pearson correlation coefficient (r) for the relationships between antecedent rainfall and soil moisture and carbon dioxide. The parametric correlation coefficient is shown here since the principal aim is to compare the relative strengths of several correlation coefficients, rather than to assess the absolute significance of particular relationships. The x-axis refers to rainfall during the stated antecedent period. Thus, the seven correlation coefficients in position 1 in figure 36 relate to relationships between soil moisture at the time of sampling (13.00 to 14.00 hrs) and rainfall of the previous day (effectively from 7.30 a.m. on the previous day to 7.30 a.m. on the day of sampling) for each of the seven profiles studied (3 in ABT-I; 3 in ABT-II; 1 in GT-I). Those in position 1-2 refer to the correlation with the combined rainfall of the previous days, and so on. In all cases shown the antecedent period ends on the day before sampling. Correlation coefficients for antecedent periods which end prior to the day before sampling (eg. the period 8-15 days) have been calculated, and the more interesting of these are discussed in the text. Where confidence levels are given in the text, these relate to non-parametric correlation coefficients (r_s).

2.2.ii. Results. With few exceptions, the correlation coefficients (r) for the seven profiles are positive over the range of antecedent periods studied. The results from the main profile in each plot are shown in figure 36. In all cases, soil moisture correlates most closely with rainfall 1-4 or 1-8 days before sampling and the correlation coefficients (r_s) for these periods are statistically

significant at the 95 per cent confidence level. Clearly, therefore, soil moisture is strongly influenced by recent rainfall. In general, however, only relatively weak correlations exist between soil moisture and antecedent rainfall of 1 and 1-2 days, and also of 2-3 and 4-7 days, before sampling. This probably reflects the fact that rainwater from most storms is retained by the soil in sufficient quantities that it significantly affects soil moisture at 15 cm for about 4 to 8 days.

If one reverses the above arguments and assumes that rainfall is the principal determinant of soil moisture, and that the latter has been accurately measured, then an important corollary of the above results is that the rainfall data used in the analysis are reliable. This finding is particularly important since rainfall records from the two stations used here, Templer Park (2601) and Sg. Siput South (3601), form the basis of the investigations of temporal variations in groundwater characteristics at two of the main groups of water sampling sites (Chapter VII).

3. Relationship between soil moisture and soil carbon dioxide

In general, soil moisture and carbon dioxide are only weakly correlated (table 74). Although statistically significant positive relationships occur in two cases, in three others the correlation coefficient is slightly negative. Thus, whilst soil moisture exerts a significant influence over the average concentrations of soil carbon dioxide (Section C, above) it appears to be a less important determinant of temporal variations. The implication of this result is considered further in later discussion (Section D6, below).

4. Relationship between soil carbon dioxide and antecedent rainfall

4.1. ABT-I

4.1.1. 15 cm (profile 1). Soil carbon dioxide at 15 cm correlates

only weakly with rainfall during the week before sampling (fig. 37), and thus appears to be relatively unaffected by recent storms. For example, values of r for antecedent periods of 1, 1-2, 1-4 and 1-8 day(s) are -0.041, 0.062, 0.174 and 0.300, respectively. Whilst a minor peak (r=0.449) occurs at 1-16 days, carbon dioxide correlates most closely with the combined rainfall of the period 1-64 days before sampling, the correlation coefficient (r_s) being significant at the 95 per cent confidence level. The highest carbon dioxide concentration (1.38 per cent) was recorded on 28.1.75, and followed a particularly wet December, when 328.5 mm of rain fell at Templer Park (2601). In contrast, the four lowest concentrations, each of less than 0.55 per cent, occurred in June and July 1975. Rainfall in April (154.1 mm), May (143.8 mm) and June (83.7 mm) 1975 was generally low.

4.1.ii. 30 cm (profiles 1 and 2). Variations in carbon dioxide at 30 cm in profile 1 parallel closely those at 15 cm (r_s=0.849; $p<0.05$). Thus the highest value (1.68 per cent) was recorded on 28.1.75, whilst four samples taken in June and July each had concentrations of less than 1.10 per cent. Measurements of carbon dioxide at 30 cm were made in two profiles. Peak correlations with antecedent rainfall occur in the period 1-64 days, and in both profiles the correlation coefficient (r_s) is significant at the 95 per cent confidence level.

4.1.iii. 60 cm (profiles 1 and 3). Results from the two profiles show generally weak inverse correlations with antecedent rainfall (fig. 37). In no case is the correlation coefficient (r_s) statistically significant.

4.2. ABT-II

4.2.i. 15 cm (profile 1). The correlations between carbon dioxide and rainfall are mostly positive and increase in strength as the length of the antecedent period increases from 1 to 1-128 days

(fig. 38). The peak positive correlation, with 1-128 days, is statistically significant at the 95 per cent confidence level. A minor peak (r=0.573) occurs at 1-16 days. As in ABT-I, the overall effects of storms during the week prior to sampling are relatively unimportant.

4.2.ii. 30 cm (profiles 1 and 2). In the two profiles studied, the relationships between carbon dioxide and antecedent rainfall are mostly positive and follow a broadly similar pattern (fig. 38). The strongest correlations, which in both profiles are statistically significant at the 95 per cent confidence level, are associated with total rainfall of the period 1-32 days before sampling. A secondary peak at 1-128 days is present in profile 1.

4.2.iii. 60 cm (profiles 1 and 3). In contrast to ABT-I, carbon dioxide levels at 60 cm generally correlate directly with antecedent rainfall (fig. 38). The closest relationship in both profiles occurs with the period 1-128 days before sampling, and in each case the correlation coefficient (r_s) is statistically significant at the 95 per cent confidence level.

4.3. GT-I

Attention here is confined to profile 1, since the number of carbon dioxide samples obtained from profiles 2 and 3 is small, being less than six.

4.3.i. 15 cm (profile 1). Although carbon dioxide at 15 cm in GT-I varies directly with soil moisture (r_s=0.857, $p<0.05$), it is only weakly correlated with rainfall during the week or so before sampling. Carbon dioxide values correlate most closely with antecedent rainfall in the period 1-32 days (fig. 39). The highest concentration of carbon dioxide (0.96 per cent) was recorded on 24.9.74. During the previous 32 days rainfall at Sg. Siput South (3601) totalled 489.0 mm. In contrast, October 1974 and June 1975 were exceptionally dry. Rainfall during the 32 days prior to

sampling on 2.11.74 and 28.6.75 totalled 56.1 mm and 108.3 mm, respectively, and carbon dioxide values were 0.42 and 0.38 per cent, the lowest recorded in GT-I.

4.3.ii. 30 cm (profile 1). Relationships between carbon dioxide and antecedent rainfall are mostly weak (fig. 39) the only exception being that with rainfall in the period 1-64 days before sampling. In this case the correlation coefficient (r_s) is significant at the 99.9 per cent confidence level.

4.3.iii. 60 cm (profile 1). Relationships between the carbon dioxide content of soil at 60 cm and antecedent rainfall are mostly weak. However, whilst many of the correlation coefficients calculated indicate a slightly negative relationship, the closest correlation, that with rainfall in the period 1-32 days before sampling, is positive.

5. Relationships between soil carbon dioxide and litterfall

Total litterfall during the periods 1-30 and 1-60 days before sampling was estimated for each plot. No significant correlations exist between soil carbon dioxide and litterfall in the 1-30 day period. Correlations with the 1-60 day period (table 75) are also generally weak. However, two significant positive correlations do occur, these being at 15 cm (profile 1) and 30 cm (profile 2) in ABT-II. Clearly, the results are highly irregular, and variations in litterfall through the year thus appear to exert little influence over measured carbon dioxide concentrations.

6. Discussion of results and implications for soil air sampling

This section of the study has set out to examine the relationship between temporal variations in soil carbon dioxide and three possible causal factors. Although two of these factors, soil moisture at the time of sampling and antecedent litterfall, exhibit significant positive correlations with carbon dioxide

concentrations at certain sampling points, no consistent relationship has been identified. In contrast, the effects of the third variable, antecedent rainfall, have been clearly established. Statistically significant positive correlations occur at 11 of the 13 sampling points at which temporal variations in soil carbon dioxide were investigated. The closest correlation is with antecedent period 1-32 days at four points, with 1-64 days at four, and with 1-128 days at three (table 76).

The strong positive correlations with antecedent rainfall imply that soil moisture is an important factor controlling the carbon dioxide content of soil air. However, the relationship between soil moisture and measured carbon dioxide levels is complex. Correlations between carbon dioxide and moisture at the time of sampling are mostly weak (table 74), as are those between soil moisture at 15 cm and rainfall during the periods 1-32, 1-64 and 1-128 days before sampling (fig. 36). Thus, whereas soil moisture varies directly with rainfall in the week or so before sampling, carbon dioxide levels appear to respond more slowly, being affected most by the total rainfall of the previous 1 to 4 months. The effects of short-term moisture changes upon soil carbon dioxide, which may be predicted from theory, are not detectable in the present study. One possible reason for this lies in the sampling technique adopted. The soil chambers (Chapter III, Section B2.2) have a cross-sectional area of only 9.1 cm^2, whereas the internal volumes of the 15, 30 and 60 cm chambers are approximately 135, 270 and 540 cm^3, respectively. Consequently, changes in the gaseous composition of the chamber will take place only slowly compared with those of the air-filled pore spaces in the soil. Thus, while carbon dioxide in the soils studied may fluctuate considerably from day to day, or from hour to hour, the sampling method is relatively insensitive to such changes. From the point of view of temporal

variations, the measured carbon dioxide concentrations, therefore, have a low degree of resolution. With the short-term effects of moisture upon gaseous diffusion being 'smoothed out', the observed trends in carbon dioxide are of a longer-term nature. The positive correlations with antecedent rainfall (table 76) may be interpreted as reflecting the higher average levels of biological activity in the soil, stimulated by the greater availability of moisture, and the generally reduced levels of gaseous diffusion, during the wetter parts of the year.

The above findings, together with the observation that soil air is extracted from different sizes of pores under varying tensions (see discussion, Chapter III, Section B2.6), demonstrates clearly that measured carbon dioxide concentrations are dependent, in part, upon the sampling method used. Values obtained by different methods of extraction are not, therefore, strictly comparable - a fact that has often been overlooked in recent intra- and inter-regional comparisons. The need for the world-wide adoption of a single sampling method is apparent.

7. Comparison with other studies

In the karst region of N. Madagascar, which has a humid tropical climate (mean annual rainfall, 1900 mm) with a marked seven month dry season, carbon dioxide concentrations are considerably higher during the wetter months (Rossi, 1974). At 20 cm in the humic horizons of a deep mineral soil values are almost constant in the dry season at about 0.5 per cent, but increase at the beginning of the wet season to a peak of between 3.0 and 3.2 per cent. Similar results have been reported from soils on other lithologies in the seasonally humid tropics (eg. W. Kenya - Cooper, 1975; Trinidad - Vine et al., 1943). Hitherto, however, no comparable data exist for soils in the permanently humid tropics.

In the Pasoh F.R., where a marked peak in litterfall was

observed between February and May, 1971 and 1972, no seasonal pattern in soil respiration was found (Ogawa, 1978). This result is in accordance with the generally weak relationship found in the present study between the concentration of carbon dioxide in the soil atmosphere and antecedent litterfall.

E. SOLUTE CONTENTS OF THROUGHFLOW AND SUBCUTANEOUS FLOW

Throughflow at the soil/rock interface was collected from pits in ABT-I, ABT-II and GT-I, and subcutaneous flow was sampled from a semi-natural exposure in Wang Tangga, close to WT-I.

1. Results from throughflow pits in ABT-I, ABT-II and GT-I

Throughflow collection devices were installed during September 1974 in soils which ranged from 55 (ABT-II) to 72 cm (ABT-I) in depth. Samples were obtained through until January or February 1975 when the collection equipment failed (Chapter III, Section B5.2). Consequently, only small numbers of samples (n=3 to 5) were obtained from these sites, and the data are inadequate for the investigation of temporal variations in solutes. It should also be noted that samples were collected at two- to three-week intervals and therefore the pH values recorded do not provide a reliable basis for assessing the degree of chemical saturation with respect to calcite. The results are summarized in table 77.

1.1. Water hardness properties

Total hardness averages 160.8 ppm (range, 136.3 to 199.0 ppm) at the three sites. Differences in total hardness, and also in alkaline hardness (range, 130.9 to 170.2 ppm) and specific conductance (range, 273.4 to 362.0μmho/cm/sec, 25°C), between pairs of sites are not statistically significant at the 95 per cent confidence level (randomization test). However, non-alkaline hardness is significantly higher in ABT-II (mean, 28.8 ppm) than in ABT-I (mean, 5.4 ppm). Thus, whereas non-alkaline hardness accounts for

16.9 per cent of the total hardness of waters in ABT-II, it accounts for only 4.0 per cent of the total hardness in ABT-I. Immediately below ABT-II, in the underlying cave system, Gua Anak Takun, groundwaters also contain notably high concentrations of non-alkaline hardness (Chapter VII, Section D3.1.ii). Two hypotheses may be put forward to explain these results. First, inorganic acids may be produced during the chemical weathering of the metamorphosed shales which ABT-II is thought to overlie. Secondly, higher concentrations of organic acids may occur in ABT-II than in ABT-I. Organic acids frequently form as intermediate products of organic decomposition. Thus, where rates of microbial activity are relatively low, as is thought to be the case in ABT-II (Chapter IV, Section D3.1), organic acids are likely to be present in significant amounts. The relative importance of these two processes cannot be determined with certainty from the data available.

1.2. Mg:Ca ratio

The molar ratio Mg:Ca ranges from 0.053 (ABT-I) to 0.193 (GT-I). In all three plots the mean value lies within the range of the Mg:Ca ratio of exchangeable cations observed in the soil profiles studied (tables 43 to 45). In ABT-II, for example, the Mg:Ca ratio of exchangeable cations ranges from 0.131 at 10 cm to 0.026 at 70 cm, whereas the corresponding value for water at the soil/rock interface is 0.053. Since the Mg:Ca ratio of bedrock in these plots is less than 0.020 it would appear either that magnesium is being leached from exchange sites in the soil or that it is being dissolved at a disproportionately high rate at the rock surface. The results of the weight-loss tablet experiments (Section F3, below) provide indirect evidence of the first process, but this does not preclude the second process also being effective. Although the throughflow pits were each less than 5 m from the soil profile investigated, it should be noted that considerable differences in

soil properties may occur over such short distances. The comparisons made above, and in the following section, must therefore be regarded with caution.

1.3. K:Na ratio

The molar K:Na ratio ranges from 0.92 (ABT-I) to 3.43 (GT-I). Since potassium and sodium occur in only trace quantities in the limestone, these elements are mostly derived from the soil. In each plot the mean K:Na ratio of throughflow is within the observed range of exchangeable cations. Thus, potassium and sodium appear to be leached from the soil in proportion to the number of exchange sites they occupy.

2. Results from semi-natural seepages in Wang Tangga

The site studied in Wang Tangga lies approximately 100 m south of WT-I. It is unusual in that a combination of natural collapse, caused by gully erosion during storm runoff, and small-scale, tin-mining activity have exposed a section of the soil and upper part of the bedrock. The soil behind the exposure is very similar to that in WT-I, being a dark reddish brown (2.5 YR 3/4) clay, with a well-developed crumb structure. On the whole, however, the soils appear to be slightly deeper (maximum depth observed, 110 cm) and rock outcrops less frequent than in WT-I. The exposed face, which is approximately 3.5 m high and 25 m long, is mantled so thickly by secondary deposits (plate 56) that the precise position of the soil/rock interface cannot be determined with certainty. It seems likely that the seepage points sampled lower down the exposure correspond with flow lines in the upper, most weathered layer of bedrock (ie. subcutaneous flow). All the seepages are ephemeral and flow only during wetter months or after particularly heavy downpours. A total of 92 samples was obtained from 16 seepage points. The pH of each sample was determined within two hours of collection. Present discussion focuses upon the average

solute characteristics of each site. The results are summarized in table 78.

2.1. Water hardness properties

Total hardness at the 16 seepage sites averages 341.9 ppm (range, 263.6 to 395.9 ppm). Figure 40 shows the relationship between pH and calcium hardness for the sample with the median total hardness at each of the 16 sites. The curve is that for a saturated calcite solution at 25°C without magnesium present. It is based on the graph presented by Picknett (1972, fig. 2, p. 143), and assumes that the curve for 25°C lies midway between the curves for 20 and 30°C. Since the proportion of magnesium in the seepage waters averages 22 per cent, the effect of magnesium upon the solubility curve is likely to be small, being at most 0.04 pH units (see Picknett, 1972, fig. 5, p. 147). Clearly, therefore, all the seepage sites are supersaturated with respect to calcite, and this explains the formation of the secondary deposits.

Non-alkaline hardness, which has a mean concentration of 1.24 ppm (range, -2.54 to 5.87 ppm), accounts on average for only 0.26 per cent of the observed total hardness. This clearly indicates that dissolved carbon dioxide is the major source of chemical aggressiveness in these soils. It also confirms that organic decomposition proceeds rapidly to completion in the well-aerated hill soils formed on pure limestones (cf. foothill soils, Section E1.1, above).

2.2. Mg:Ca ratio

The molar Mg:Ca ratio averages 0.219 (range, 0.180 to 0.260). Since WT-I lies approximately 100 m beyond the catchment of the seepage sites, specific comparisons cannot be made between water quality and soil characteristics of the two areas.

2.3. K:Na ratio

Sodium concentrations are consistently high at all

points, averaging 4.25 ppm (range, 3.68 to 4.83 ppm). The K:Na ratios are correspondingly low (mean, 0.074; range, 0.03 to 0.13).

2.4. Mg:Ca ratio of secondary deposits

Three samples of the secondary deposits which encrust the surface of the exposure were analysed. They were found to comprise on average: 4.57 per cent acid insoluble residue, 94.9 per cent calcium carbonate, and 0.21 per cent magnesium carbonate. The Mg:Ca ratio of the deposit (mean, 0.0026) is significantly less than that of the seepage waters (mean, 0.219), indicating that calcium carbonate is re-precipitated almost exclusively from the saturated waters. This result is discussed more fully in Chapter VII, Section D5.1.iii.

3. Comparison of results from throughflow pits with those from exposures in Wang Tangga

Although the carbon dioxide concentrations of soil air at 60 cm in ABT-I, ABT-II and GT-I are significantly higher than in WT-I (tables 67 and 69), the mean total hardness of the seepages in Wang Tangga is more than double that of throughflow in the former plots. The different circumstances under which samples were obtained provide the most likely explanation of this result. Because of the talus-like nature of the foothill slopes it is likely that vertical or subvertical seepage between the limestone boulders will be dominant, and that lateral flow over the discontinuous surface of the boulders will be limited. Throughflow collected in the foothill plots may, therefore, only have flowed a short distance over a single boulder before being diverted to the sampling bottle. In contrast, the waters sampled in Wang Tangga appear to be mostly of subcutaneous origin. Such waters have necessarily been in contact with limestone for a longer period of time and over a greater distance. Consequently, they are more likely to be saturated, or supersaturated, with respect to calcite than the throughflow waters

sampled in the foothill plots.

4. Relationship between solutional potential and measured soil carbon dioxide concentrations (Wang Tangga)

Since the samples obtained from the foothill plots are thought to be undersaturated with respect to calcite, and as their pH values are not known accurately, the solutional potential of these throughflow waters cannot be estimated. Consequently, only the results from Wang Tangga are considered in this section.

The seepage waters collected in Wang Tangga are supersaturated with respect to calcite (fig. 40). While the pH may change considerably in response to the fall in the partial pressure of carbon dioxide as the waters seep towards the exposed face and emerge at the surface, changes in water hardness are thought to be small. Confirmation of this is seen in the fact that no detectable decrease in hardness or specific conductance was found in water which had flowed as a thin film for a distance of 1.5 m from a seepage point. It is assumed, therefore, that the observed hardness values provide a reliable estimate of the solutional potential of soil waters at this site.

The relationship between the calcium content of saturated calcite solutions and partial pressures of carbon dioxide in the pure $CaCO_3$-CO_2-H_2O system is well established (Picknett, 1973). A curve which relates calcium hardness and the partial pressure of carbon dioxide at 25°C (cf. mean temperature of seepages, 24.4°C) may be plotted using the data presented by Picknett (1973, table 5, p. 72). From this curve the concentration of carbon dioxide which corresponds with the mean calcium hardness of 280.8 ppm is approximately 2.6 per cent.

Clearly, because of the complex nature of natural karst waters (Chapter V, Section A1.2), and, especially in this case, the presence of magnesium ions, the latter figure is only approximate.

Nonetheless, it is interesting to note that the value obtained is about three times the measured carbon dioxide concentration at 60 cm in WT-I (mean, 0.78 per cent). To some extent the difference may be explained by the slightly deeper soils which occur in the vicinity of the seepage site. These may, on average, have higher carbon dioxide contents in the lowest horizons than those in WT-I.

A second factor, related to the method of carbon dioxide collection, must also be considered. In using the gas reservoir method it is assumed that the air in the chamber is in equilibrium with that of the air-filled soil pores. However, as noted previously (Chapter III, Section B2.6) the carbon dioxide content of larger soil pores is likely to be less than that of the smaller pores. Since the rate of diffusion between the soil and the gas reservoir will be greater in the case of larger, interconnected pores than of smaller, often isolated, ones, the air sampled will more closely resemble that of the former pores. When throughflow occurs in soils it is mostly through the larger macro-pores. At such times air pockets within these pores become isolated, permitting an increase of carbon dioxide. Also, chemical exchange will occur freely between the gravitational and capillary waters, both by diffusion and mass flow. Throughflow waters thereby come more closely into equilibrium with the carbon dioxide levels of the capillary pores and sealed macro-pores. Consequently, the measured carbon dioxide concentrations are probably less than the amounts present in the pores with which throughflow waters are in direct contact. This being the case, then the relationship between the observed carbon dioxide levels and soil and groundwater chemistry cannot therefore be established with certainty in the present study.

F. RATES OF LIMESTONE EROSION BENEATH DEEPER MINERAL SOILS

In the present study, independent estimates of erosion rates beneath

mineral soils have been derived from two sources: first, from the hardness of soil waters in contact with limestone, and secondly, from the results of erosional weight-loss tablet experiments.

1. Assumptions made in estimating erosion rates

Ideally, measurements of throughflow volumes would be required to determine erosion rates beneath mineral soil covers from solute and rock tablet data (Section F2, below). However, because of the pronounced and highly irregular meso- and micro-topography of the field areas, throughflow measurement was considered to be beyond the scope of this study. The amount of throughflow in the six moderately-inclined plots has, therefore, been estimated as:

$$\frac{\text{Mean annual rainfall (mm)}}{S} - \text{Estimated annual evapotranspirational loss (mm)},$$

where 'S' is the proportion of soil cover in a particular plot. In using this equation the following assumptions are made:

1. It is assumed that surface runoff to swallets is negligible. Although runoff was observed on certain occasions during particularly intense downpours, the limited occurrence of gullies on the soil-mantled hill slopes indicates that surface runoff is of relatively minor importance. In many cases storm waters flow over only short distances before infiltrating, often at the centres of small soil pockets. Typically these are 4 to 10 m in diameter, as for example, in WT-I (Chapter IV, Section C4.2.1). Further evidence of the importance of diffuse percolation, as compared with swallet flow, is seen in the narrow range of water hardness fluctuations of cave streams draining the Setul Boundary Range (Chapter VII, Section D4.3).

2. The factor 'S' is included since it is assumed that water falling on a rock outcrop will mostly flow directly on to adjacent soil. The volume of water reaching the soil surface is augmented

in inverse proportion to the amount of soil cover present.

3. It is assumed that the combined water loss through the evaporation of rainfall intercepted by the vegetation canopy, evaporation at the soil surface and evapotranspiration equals the estimated potential evapotranspirational loss. Water balance data for the three field areas, based on an arbitrary soil water storage capacity of 6 inches (15.24 cm), are presented by Nieuwolt (1965). These show that while on average no soil moisture deficit occurs in the area of west coast climate (which includes Selangor and the Kinta Valley), a total deficit in excess of 100 mm may be expected in the North-west region during January and February. Soil moisture deficit is here taken as the difference between the potential and the actual evapotranspirational loss. The maximum moisture content recorded in most soils studied varies between 30 and 45 per cent by volume (table 70). Taking 40 per cent by volume as a conservative estimate of the water holding capacity of these soils, then Nieuwolt's data relate to karst soils which are approximately 38 cm deep. If those sampling points without soil cover along the transect lines in the study plots are excluded from computation (cf. table 30), then the average soil depth in five of the six moderately-inclined foothill and hill slope plots is greater than 36.0 cm. Only ABT-II is exceptional in having a value of 28.8 cm. The assumption made above is, therefore, justified in the case of ABT-I, GT-I and GT-II. It may be argued that the additional input of water created by runoff from rock outcrops is sufficient to compensate for the low water storage capacity and low rainfall input, respectively, in ABT-II and WT-I. Also, in GKV-I the greater depth of soil, and hence greater moisture storage capacity, may considerably reduce the magnitude of soil moisture deficits.

Any inaccuracies which may arise in calculating throughflow volumes by means of the above equation are thus thought to be small.

2. Erosion rates estimated from water hardness data

In calculating erosion rates beneath mineral soils from the water hardness of throughflow and subcutaneous flow, two further corrections are necessary. First, allowance must be made for the input of calcium and magnesium in rainwater (table 24). Secondly, account must also be taken of the hardness of runoff from bare and litter-covered rock surfaces, since that proportion of throughfall which falls on such outcrops will have already dissolved some limestone before entering the soil. These waters are assumed to have a total hardness of 51.8 ppm, this being the mean value observed on Ak. Bt. Takun (table 63). The estimated rates of erosion beneath mineral soils in the four plots are presented in table 79. Highest rates occur in GT-I (182.9 mm/1000 yrs) and WT-I (173.7 mm/1000 yrs). In the first plot, the high rate is attributable to the large amount of throughflow (3819 mm), which is caused by the combined effects of heavy rainfall and a relative low proportion of soil cover. In the second, it reflects the high hardness values of the seepages. The lowest erosion rate recorded, 65.2 mm/1000 yrs (ABT-I), is almost double that on bare rock outcrops at the summit of Ak. Bt. Takun (32.8 mm/1000 yrs). It is, however, only 19 per cent higher than the 54.9 mm/1000 yrs recorded on rock surfaces with a partial soil/litter cover (Chapter V, Section F).

3. Erosion rates estimated by the erosional weight-loss tablet technique

Tablets were installed at 15, 30 and 60 cm in three profiles in the six main study plots (ABT-I, ABT-II, GT-I, GT-II, WT-I and GKV-I). In order to estimate long-term erosion rates from the weight loss observed during the study period, the following procedure has been adopted. The volume of throughflow which occurred while the tablets were in situ has been estimated from the actual rainfall and the average monthly evapotranspiration rates. The

measured amount of erosion has then been adjusted proportionately to give a value which corresponds with the estimated mean annual volume of throughflow (Section F1, above).

At most depths in the six plots weight loss varied considerably between the three profiles studied (table 80). At 60 cm in GT-II, for example, the estimated erosion rates are 3.36, 3.95 and 4.96 mm/1000 yrs, a range of 1.60 mm/1000 yrs. However, the mean values display a discernible pattern. In four of the six plots the mean erosion rate increases with depth. This result is consistent with the generally higher carbon dioxide concentrations recorded at depth. However, differences between the various plots at, for example, 60 cm do not correlate closely with measured carbon dioxide concentrations. Thus, while the highest rate of erosion (5.03 mm/1000 yrs) occurs in GKV-I, the plot with the greatest concentrations of soil carbon dioxide, high rates were also recorded in GT-II (4.09 mm/1000 yrs) and WT-I (2.63 mm/1000 yrs), the plots which have the lowest carbon dioxide concentrations. In contrast, the lowest erosion rates at 60 cm occur in ABT-I (1.58 mm/1000 yrs), ABT-II (2.01 mm/1000 yrs) and GT-I (2.26 mm/1000 yrs), the three foothill plots. In view of the observed relationship between the Mg:Ca ratio of throughflow at the soil/rock interface and exchangeable cations in the latter plots (Section E1.2, above), the most likely explanation of the low erosion rates estimated by the present technique is that the chemical aggressiveness of soil waters towards the limestone is satisfied, in part, by the leaching of bases from the neutral to slightly alkaline soil. In these soils, therefore, the measured weight loss of the rock tablets poorly reflects the net removal of calcium and magnesium. Leaching of exchangeable cations may also occur in GT-II, WT-I and GKV-I. Here, however, the effectiveness of this process is thought to be limited since the soils have relatively low levels of base saturation (tables 40 to 42).

Some indication of the effects of other factors upon erosion rates may be gained by comparing the results from different pairs of plots. For example, because of differences in the extent of rock outcrops, the two plots in the North-west region, WT-I and GKV-I, have different amounts of throughflow, 1526 and 823 mm, respectively. However, despite this, measured erosion rates at 60 cm in GKV-I are more than double those in WT-I. This result reflects the importance of the carbon dioxide content of soil air as a determinant of solutional potential. Interesting comparisons may also be made between GT-II and WT-I since the soils in both plots are very similar in their structure, chemical properties and carbon dioxide content. At 60 cm the measured erosion rates in GT-II (4.09 mm/1000 yrs) are almost double those in WT-I (2.44 mm/1000 yrs). This difference may be largely attributed to the greater volumes of throughflow in the former as compared with the latter plot, the amounts being 4017 and 1526 mm, respectively.

4. Comparisons of erosion rates estimated from water hardness and weight-loss tablet data, and some methodological implications

Erosion rates estimated from water hardness data in ABT-I, ABT-II, GT-I and WT-I range from 65.2 to 182.9 mm/1000 yrs, whereas those based on the weight-loss tablet technique range from 1.58 to 5.03 mm/1000 yrs at 60 cm. The latter values vary between 1.2 (GT-I) and 2.4 per cent (ABT-I) of the former (cf. 2.9 per cent for bare and thinly-covered rock surfaces, Chapter V, Section F2). While the leaching of exchangeable bases from the soil is undoubtedly a contributory factor (Section F3, above), additional factors are clearly required to explain the results from the bare rock catchments, and the overall magnitude of the differences observed. One possible explanation is that the small rock tablets receive very little lateral flow. Consequently, the volume of water which flows over them is extremely limited, being confined mostly to

direct rainfall at surface sites, and to vertical seepage where a soil cover is present. In contrast, under natural conditions the volumes of water which flow over exposed and covered rock surfaces are augmented greatly by lateral flow derived from the 'catchment area' up-slope. This increases not only the potential for solution, but also the duration of contact between water and the rock surface. This result is significant in that it precludes meaningful inter- or intra-regional comparisons of erosion estimates which have been obtained by the two different methods.

In view of these observations, and of the contrasts noted previously between the water hardness and solute methods (Chapter III, Section B7), solute data are though to provide the most reliable basis for estimating surface and sub-soil erosion rates.

G. SUMMARY OF RESULTS

1. Spatial and 'with depth' variations in soil carbon dioxide

1.1. Carbon dioxide concentrations increase with depth in the six soils studied.

1.2. At a given depth, carbon dioxide levels vary inversely with the volume of air-filled pore space. Lowest concentrations were thus recorded in GT-II and WT-I (0.73 and 0.78 per cent, respectively, at 60 cm), which have the driest and most porous soils. Elsewhere, average concentrations at 60 cm range from 1.27 to 4.21 per cent.

1.3. Higher levels of microbial activity, and hence of soil carbon dioxide, occur in more highly productive ecosystems (eg. GKV-I). Lower levels may result from frequent waterlogging in soils with impeded drainage (eg. ABT-II).

1.4. The average concentrations at 15, 30 and 60 cm in the six soils (0.65, 1.15 and 1.85 per cent, respectively) are similar to those reported from Jamaica.

2. Temporal variations in soil carbon dioxide

2.1. At a given depth in a single profile, soil carbon dioxide exhibits marked variability.

2.2. In general, carbon dioxide concentrations correlate most closely with rainfall in periods 1-32, 1-64 or 1-128 days before sampling. This is thought to reflect the higher levels of biological activity in the soil and the reduced levels of gaseous diffusion during the wetter parts of the year.

2.3. Whilst soil moisture at 15 cm correlates strongly with rainfall during the week before sampling, no significant relationship was found between carbon dioxide and short-term moisture changes. This may be attributable to the method of sample extraction used.

3. Solute contents of throughflow and subcutaneous flow

3.1. The total hardness of throughflow waters (mean, 160.8 ppm; range, 136.3 to 199.0 ppm) is significantly lower than that of subcutaneous flow (mean, 341.9 ppm). The longer period of water/rock contact is probably a key factor.

3.2. At the point of emergence, the subcutaneous waters (near WT-I) are supersaturated with respect to calcite.

3.3. From known theoretical relationships, the subcutaneous waters appear to have equilibriated with soil air which has a mean carbon dioxide content of approximately 2.6 per cent. This is about three times greater than that observed at 60 cm in WT-I, and indicates that measured carbon dioxide levels may be considerably less than the concentrations present in the capillary pores and isolated macro-pores with which soil water comes into equilibrium during storms.

4. Rates of limestone erosion beneath deeper mineral soils

4.1. Erosion rates estimated from water hardness data, range from 65.2 to 182.9 mm/1000 yrs. Highest rates occur in areas of heavy

rainfall, especially where volumes of throughflow are increased by runoff from rock outcrops.

4.2. Results obtained from erosional weight-loss tablet experiments suggest that where mineral soils are base-rich, as on the foothill slopes of Selangor and the Kinta Valley, leaching of exchangeable calcium and magnesium from the soil forms an important component of total erosional loss.

5. Methodological implications of the results

5.1. Soil pores of different size ranges are sampled by different methods of soil air extraction. Consequently, results obtained by different methods are not directly comparable.

5.2. Erosion estimates based on the erosional weight-loss tablet technique are between one and two orders of magnitude smaller than those derived from water hardness data. Results obtained by the two methods are, therefore, incompatible.

CHAPTER VII

AN INVESTIGATION OF KARST WATERS

CHAPTER VII. AN INVESTIGATION OF KARST WATERS

A. INTRODUCTION

Karst waters in the Malay Peninsula are extremely varied. On the one hand they range from deep groundwaters in the limestone aquifers beneath the karst alluvial plains, to shallow-depth, diffuse-flow seepages in abandoned, high-level cave passages in tower karst hills. On the other, they range from surface streams with a considerable groundwater contribution, to those which rise either on the alluvial plains or in mountainous terrain on non-carbonate lithologies. Previous studies have shown that calcium, derived from the weathering of limestone, is an important component of the dissolved load in the lower part of the Sg. Gombak catchment (Selangor), which is underlain by limestone (Douglas, 1968c), and that the total hardness of this river varies inversely with discharge (Douglas, 1968a). Seepages in the uppermost caverns of Bt. Batu have also been found to have higher calcium hardness concentrations than those in lower caves (Douglas, 1968a). However, with the exception of these investigations in Selangor, and the observation that the pH of swamp waters in the Kinta Valley may be as low as 3.5 (Paton, 1964), no data on karst water quality in West Malaysia have been published. Attention in the present investigation focused, therefore, upon those properties of karst waters which could be determined expediently in the field and laboratory, thereby allowing frequent sampling at a large number of points. In all, more than 2500 water samples were analysed from a total of 324 sites.

The aim of this chapter is to examine the systematic, spatial and temporal variations in karst water properties observed in the three field areas. Literature related to individual karst water characteristics is considered in the appropriate section.

B. EXPERIMENTAL DESIGN

1. Selection of water sampling sites

The principal objective in selecting the sampling points was that they should encompass the diverse types of karst water present in the three regions. A clear-cut sixfold division of the waters into underground diffuse-flow seepage, karst spring, underground stream, surface stream, alluvial seepage and deep groundwater may be made.

Although the limestone hills are invariably permeated by innumerable cave passages, most of the 30 or more caves explored during the preliminary field survey are either completely dry, or contain only a small number of seepages with discharges sufficient to yield an adequate sample for analysis over a 2-hour period. Many caves are inaccessible, with entrances in precipitous cliff faces, often 20 m or more above the contemporary plains. Others, though accessible, either are in an extremely dangerous condition, such as the many abandoned, underground tin mine workings in the Setul Boundary Range, or are of religious significance, such as the cave temples of Bt. Batu and of the hills around Ipoh (plate 27). In all, four caves were investigated, and 230 seepages, chosen so as to be representative of the flow types in each cave, were sampled.

By comparison, the sampling of springs, underground streams and surface waters was more complete. Karst springs, and cave streams with autogenic catchments, are almost entirely restricted to the Setul Boundary Range. Most of the accessible waters of these types which lie between Wang Tangga and Kangar were investigated. Samples were taken at the cave entrance or rising. In addition, with the notable exception of the Sg. Pari in the Kinta Valley, all the major rivers draining the karst regions were sampled, together with several smaller streams which typify particular

catchment types (details, Section C4, below). In the present report these data are supplemented by unpublished figures for the Sg. Klang and its main tributaries (Selangor), which were provided by the Department of Agriculture and Rural Development, Malaysia.

Two further types of groundwater sites were studied. First, waters seeping through the alluvium of the karst plains were sampled at points where they emerged in alluvial mine workings. In this case the choice of sites was necessarily determined by the location of active mines. However, since a total of 48 sites were sampled in nine different mines, the results are broadly representative of these waters. Finally, deep groundwater from the limestone aquifer beneath the karst alluvial plain was obtained at one location in the Kinta Valley.

2. Sampling frequency

The sampling strategy adopted was essentially two-tiered. First, 156 stations were sampled regularly at approximately 3-week intervals in Selangor and the Kinta Valley, and at 6-week intervals in the North-west region. All of the karst spring, cave stream and surface stream sites, the deep groundwater site, and approximately half of the underground seepages were in this category. Secondly, single samples were obtained from the alluvium and from remaining underground seepages. Sampling at this second level was undertaken at times when fieldwork was concentrated in one locality, as for example during the survey of Gua Tempurong. Except during flood pulse studies, underground seepages were sampled at between 9.00 and 13.00 hrs, and karst springs, and cave and surface streams at between 11.00 and 14.00 hrs, thereby minimizing the effects of diurnal temperature changes. The flow rate of a number of underground seepages was inadequate to permit reliable temperature (Chapter III, Section B6.1) or pH determination (Chapter III, Section B5). The number of sites considered with respect to these properties is, therefore,

less than for the other karst water properties.

3. Additional measurements

Cave air temperatures were measured regularly at a total of ten locations in the four cave systems.

C. DESCRIPTION OF SAMPLING SITES

1. Underground seepages

Regular sampling was undertaken in Gua Batu, Gua Anak Takun (both in Selangor), Gua Tempurong (Kinta Valley), the subterranean passage which links Kaki Bukit and Wang Tangga, and in the cave at the south-eastern corner of Wang Tangga (North-west region). The last two caves, which are unnamed on topographic maps, are here referred to as the Subway Tunnel and Perlis Mines' Cave (owned by Perlis Mines Ltd.), respectively. Since the latter is effectively a tributary of the Subway Tunnel (see fig. 46), results from the two have been combined in the following analysis. The locations of the caves are shown in figures 3 to 5.

1.1. Gua Batu (Batu Caves)

1.1.1. Morphology. Gua Batu is located at a height of 92 m on the southern side of Bt. Batu (plates 4 and 7). It comprises two separate caves: (1) Temple Cave, which since 1891 has been the site of the Sri Subramania Swamy Temple, the principal shrine of the Hindus in West Malaysia (Veeriah and Wycherley, 1971), and (2) Gua Gelap (Dark Cave), which comprises five interconnected caverns, totalling almost 1.5 km in length (fig. 41). A detailed survey of the cave has been published by Heynes-Wood and Dover (1929).

Gua Batu is probably of phreatic origin (cf. caves in Sarawak and Sabah - Wilford, 1964), but shows evidence of subsequent vadose modification (plate 8). It is rectilinear in plan. The caverns are straight, parallel-sided and, in Gua Gelap, terminate in sheer

bedrock walls. The walls are indented by large shallow scallops and blind passages, and the roof is very irregular, with marked height discontinuities, especially at the junctions of the major caverns. Locally, the roof exceeds 80 m in height. In several places it is breached by open shafts or chimneys (plate 9), the largest being 30 to 40 m in diameter. These connect directly with the top of the hill and coincide with the line of a transverse northwest-southeast trending ravine. In detail, however, these features are independent of, and appear to pre-date, the present surface topography. Their formation and role in the development of the cave is problematical. The chimneys provide the only visible inlets into the cave system, but do not always occur near the end of the caverns. It is difficult, therefore, to conceive that water from these sources could have been responsible for solutional enlargement of the necessary scale unless other outlets previously allowed the passage of large volumes of water. The irregular floor of the cave comprises undetermined thicknesses of guano and angular rock debris, and both the floor and bedrock walls are frequently covered by extensive deposits of secondary calcite. Flowstone and dripstone deposits are particularly prominent, and often of massive proportions (plates 8 to 10). One column in Temple Cave, for instance, is 6.5 m in diameter and 18 m in height (plate 9). It seems probable that the superficial covers conceal passages formerly continuous, and developed contemporaneously, with those extant today, and that the chimneys once formed part of such a system.

1.1.ii. Hydrology. Although there are no permanent streams in Gua Batu, parts of Gua Gelap, notably the lower part of Cavern C, often become flooded to depths of 30 cm or more during heavy rainfall. At these times considerable volumes of water spill into the main caverns from air-filled conduits in the roof of the cave. The latter appear to be very old, abandoned, high-level passages, which are closely connected with the surface of the hill. Between storms they

yield only small trickles of water, presumably from diffuse-percolation sources. However, during storms they convey surface run-off and subcutaneous flow rapidly into the underlying caverns. One site, for instance, which has an average discharge of less than 20 ml/min between storms, yielded more than 10 l/min during one downpour. Other seepages in the cave are mostly from stalactites, fed by tight bedrock joints. In many cases the thickness of lime-stone through which the water has percolated is in excess of 130 m. Typically the flow of these seepages responds very little to seasonal rainfall patterns or single storms.

Sampling was undertaken at 56 seepage points, mostly located in Caverns B and C. Of these, 48 have mean discharges of less than 100 ml/min (fig. 48). Where six or more discharge measurements were obtained at a site, the coefficient of variation (c.v.) was calculated as an index of flow variability. The average c.v. for 36 such sites, 76.7 per cent, falls in the middle of the range rec-orded in the four caves, with three-quarters of the sites having values of less than 100 per cent (fig. 48). As in the case of soil moisture and carbon dioxide (Chapter V1, Section D2.1.i) correlation coeffic-ients (r) for the relationships between discharge and rainfall during pre-set antecedent periods were determined for each of the 36 seepages. Where rain fell on the sampling day, prior to sample collection, rainfall data for the previous day have been adjusted to include this. For 29 sites the highest positive correlation coefficient calculated was significant at the 95 per cent confidence level.The distribution of the peak correlations at these 29 seepages is shown in figure 48. At almost half of the seepages the peak corresponds with the period 1-16 days before sampling, and in 26 cases with the periods 1-16 days or more before sampling. The apparently sluggish response of the majority of the sites, and a lack of significant response at seven, probably reflects the length,

and tightness of the fractures through which much of the water percolates from the top of the hill. If it is assumed that the flow rate is dependent upon the pressure of water within the joint network supplying a particular seepage, then the modal group of 1-16 days suggests that water is retained within many of the main feeder joints in sufficient quantities that it significantly affects seepage discharge for about 16 days. In view of the observed rapid response of the high-level conduits to storms, the absence of close correlations with periods of 1 and 1-2 days requires explanation. Two factors appear to be important in this respect. First, the surges of water which result from individual downpours often pass through the conduits so quickly that discharges return almost to base flow levels several hours after the end of the storm. Consequently the immediate effects of a storm which occurred on a day prior to sampling may no longer be detectable at such sites by 11.00 am the following day. Secondly, only on a small number of occasions did regular sampling coincide with a day following one with heavy rainfall. For both these reasons, the flashy responses of the open conduits are poorly represented in the overall data.

1.2. Gua Anak Takun

1.2.1. Morphology. Gua Anak Takun is a comparatively small cave which lies approximately beneath plots ABT-IV, -III and -II on Ak. Bt. Takun (plate 12). The entrance is level with the alluvial plain (95 m - Wycherley, 1975). A survey is shown in fig. 42 (after Dunn, 1965), and more detailed plans of two sections, surveyed by the author, are shown in figures 43 and 44. The passages are generally low and, in contrast with most caves in the Peninsula, are wide in relation to their height. Gobbett (1965) noted Gua Anak Takun as an example of an horizontal cave developed in vertically-bedded limestone. However, the thickness of the soft mud and guano which form the present floor of the cave is not known and it is possible that the bedrock floor is at a considerable depth. Two features of the

cave are of particular interest from the point of view of karst water quality. First, the limestones in the south-west corner of the cave are interbedded with steeply dipping, metamorphosed shales (fig. 43). Secondly, the passages occur at two levels. North Point, of the less accessible upper passage, directly overlies Bench Corridor (lower passage) at one location, the two being separated by 12 m or more of limestone bedrock (Wycherley, 1975, p. 41). The significance of these aspects of the cave is considered in later discussion (Section D, below).

1.2.ii. Hydrology. The floor of the lower passages remains mostly dry, except following particularly heavy storms when water pouring through openings in the roof of the upper passages, spills down the Upper Chimney into the Dome Room. A total of 63 sites were investigated in the following sections of the lower cave: Pipe Corridor, Snake Room, Liphistius Room, Bench Corridor and the Dome Room. Diffuse percolation flow in the cave differs from that of Gua Batu in three respects. First, there is a much higher frequency of drips per unit area of cave roof and wall. In consequence, the flow rates of individual seepages tend to be low, with 86 per cent of the seepages having mean discharges of less than 10 ml/min (fig. 48). Secondly, joints in the roof of the lower cave are all very tight, with the result that there are no seepages equivalent to the high-level conduits of Gua Batu which yield large volumes of water during storms. Thirdly, flow rates tend to be more variable, with an average c.v. of 106.4 per cent (fig. 48), the highest recorded in the four caves. This probably reflects the small size of the outcrop. The maximum thickness of limestone through which water percolates before entering the cave is about 70 m. However, because of the overall morphology and deeply-dissected form of the hilltop (Chapter IV), it seems more likely that groundwater flow paths are generally in the order of 30 to 40 m. Under these circumstances,

seepages are likely to respond to seasonal variations in rainfall, and possibly to single storms. The fact that the flow rate at four of the sites correlates most closely with rainfall of the day prior to sampling (fig. 48), and that more than two-thirds of the sites have peak correlations with rainfall in antecedent periods of 1-16 days or less, is indicative of this. Unlike the conduit seepages in Gua Batu, there is a lagged response of several hours at those sites in Gua Anak Takun which react to single storms. Consequently the effects of rainfall on the day prior to sampling are more readily detected in the data.

1.3. Gua Tempurong

1.3.1. Morphology. Gua Tempurong, which passes through the G.Gajah-Tempurong massif, is the largest and most spectacular cave passage in the Kinta Valley. A survey of the cave, undertaken by the author, is shown in figure 45. The Sg. Tempurong, a small stream which rises on G. Chante in the Main Range, presently follows a 1.5 km, east-west, course through the hill. The central sections of the cave comprise three caverns of enormous dimensions. Alam Cavern, the largest, is more than 130 m wide and reaches a maximum height of approximately 72 m. The upper parts of the cavern appear to be phreatic in form. They are straight in plan and, like the surface topography (Gobbett, 1971), are guided by northwest-southeast tectonic lineaments. They are contiguous, but have no visible outlet. As in Gua Batu, former passages may have collapsed or be blocked or buried by cave deposits. In contrast, the lower parts, within 15 m of the stream, are continuous throughout the entire length of the system and exhibit a meandering plan, typical of vadose development. Pronounced stream-cut notches recur at heights of 3.5 and 15 m above the present channel, and in places the walls are undercut to depths of 30 to 40 m (plates 34 and 35). It is interesting to note that the highest notch at the eastern entrance is at 79 m,

an elevation which corresponds closely with that of benches and other features in the Kinta Valley and North-west region (Chapter II, Section C1.4.1). The bed of the Sg. Tempurong comprises sands and gravels of unknown depth, and in several places, notably in Tin Miners' Cavern, alluvial fill occurs to heights of 8 m above the present channel. These deposits may be contemporaneous in origin with the Old Alluvium. A more detailed description of the cave has been published elsewhere (Crowther, 1978a).

1.3.ii. Hydrology. Despite the size of the cave and the high rainfall of the region, no detectable increase in the discharge of the Sg. Tempurong was observed on four occasions when measurements were made at both ends of the system. Whilst some stream water may be absorbed by the alluvial bed, the apparent absence of a significant groundwater input most probably reflects the scarcity of major seepages within the cave. The cave roof is unbreached and at no point was seepage flow seen to increase rapidly in response to single storms. Of 79 sites investigated along the length of the system, only four had average flows of more than 100 ml/min. (fig. 48). As in Gua Anak Takun, seepages in certain sections of the cave, especially those parts undercut by recent stream action, are closely spaced, often occurring in lines where the roof intersects tightly-closed rock fractures (plates 36 and 37). However, in marked contrast to Gua Anak Takun, flow variability is low. Of 34 seepages at which six or more discharge measurements were made, 31 have c.v. values of less than 100 per cent (fig. 48), with an overall mean of 47.9 per cent. At seven sites the observed range of discharge through the year was less than 0.5 ml/min. Since measurement error may account for a significant proportion of the variability of these seepages, their discharge may effectively be regarded as constant. Significant correlations between flow rate and antecedent rainfall occur at 17 of the remaining 27 sites. The modal group of peak correlations is with the period 1-32

days before sampling, and at 11 of the sites peak correlations are associated with antecedent periods of 1-32 days or longer (fig. 48). Overall, the response times in Gua Tempurong are longer than in the other caves investigated. This finding is consistent with the low variability of discharge and clearly reflects the long paths of diffuse percolation flow, which are in the order of 200 to 400 m.

1.4. Perlis Mines' Cave and Subway Tunnel

The limestones of the Wang Tangga area differ fundamentally from those of the caves considered previously in that they have been affected less strongly by metamorphism and consequently retain many of their original sedimentary characteristics. This has significantly influenced cave development, and is reflected in the hydrological characteristics of the outcrop.

1.4.1. Morphology. Perlis Mines' Cave (figs. 5 and 46) is typical of caves in the Setul Boundary Range in that its form, in contrast with caves in the tower karst hills, is largely controlled by sedimentary structures (for general description of caves, see Jones, 1965). It also contains a permanent vadose stream. In plan the cave essentially comprises a long and generally straight passage which is guided by major bedding planes with dominant north-south strike. The passage is mostly less than 9 m wide and 10 m high and extends well beyond the mine workings, which at the time of sampling were located about 1.5 km from the entrance. The walls are often irregular due to variations in the susceptibility of the different beds to erosion. Typical sections, which are indicative of vadose stream action are shown in figure 47. Several larger chambers, approximately 30 m wide and 50 m long, occur along the length of the cave. These formerly contained deep pockets of alluvium. The latter has been mostly removed by mining activity, and the caverns now contain deep underground lakes.

Subway Tunnel is the natural underground passage of the

Sg. Pelarit through the narrow ridge of limestone which forms the south-eastern rim of Wang Tangga (fig. 46). It varies from 9 to 15 m in width.

1.4.ii. Hydrology. The most striking feature of groundwater seepage into Perlis Mines' Cave and Subway Tunnel is that it tends to be concentrated at a relatively small number of sites which characteristically have high flow rates. Nineteen sites were sampled in the former cave, and five in the latter.All the sites have mean discharges of more than 10 ml/min, and one-third have values in excess of 1000 ml/min. (fig. 48). The key hydrological factor appears to be the presence of horizontal or subhorizontal bedding planes. These cause discontinuities in the vertical movement of water through the rock, leading to lateral flow. Over time, certain fractures above the more pronounced bedding planes will become widened by solution, and integrated systems of groundwater flow patterns will focus on these. Diffuse percolation seepages in both passages appear to be mostly of this type, emerging where the cave wall intersects major joints which lie above marked discontinuities in the bedding. In many cases seepage occurs from air-filled conduits. From an hydrological viewpoint, these are similar in function to the high-level conduits in Gua Batu. They are, however, generally less flashy in their response to storms. Overall, the hydrological properties of the Setul limestones contrast markedly with the regionally metamorphosed limestones of Selangor and the Kinta Valley. Since the latter bear few traces of the original bedding, downward flow along largely uninterrupted vertical or subvertical fractures predominates. As a result, large numbers of discrete flow lines have developed, with comparatively little integration.

The variability of discharge of seepages in Perlis Mines' Cave/Subway Tunnel is quite marked (mean c.v., 70.2 per cent), but lies well within the range observed in the other three caves (fig. 48).

However, the greater sensitivity of these sites to rainfall at the surface distinguishes them from other percolation waters. Of the 24 seepages investigated the discharges of all except three correlate significantly ($p<0.05$) with rainfall during one or more pre-set antecedent periods. The modal group, the 1-8 day period (fig. 48C), is lower than that of the previous caves, and more than three-quarters of the sites have peak correlations with periods 1-8 days or less. This finding is indicative of rapid transmission of water through the bedrock, a result which is consistent with the ground-water flow patterns postulated above.

1.5. Systematic variations in diffuse-flow seepages

In addition to the broad regional variations described above, underground seepages may also be classified systematically into four groups. First, and under somewhat exceptional circumstances (see Section D3.1), seepages may be sampled where they emerge from bedrock joints. More usually, however, seepage waters percolate through stalactitic deposits before emergence. Some, forming the second group, drip from small stalactites less than 1.5 m in length. Characteristically they have low flow rates. Others, which form the third group, are associated with large stalactite forms. Where discharges are higher water may flow over as well as through these features. The final grouping is of sites where water spills from high-level, air-filled conduits. Seepages of this type have been described above, and are confined entirely to Gua Batu and Perlis Mines' Cave/Subway Tunnel.

2. Karst springs

Three small karst springs were studied (table 81). Their locations are shown in figure 5. Sites 4306 and 4312, with mean discharges of 0.0031 and 0.0021 cumecs, respectively, appear to drain small catchments of relatively low relief. Much of their water is probably derived directly from shallow-depth, subcutaneous flow.

In contrast, the spring near Kg. Gtg. Kabu (site 4317) is located at the base of the steep north-eastern slopes of the Gelang Kerbau Valley and probably receives a significant contribution from deeper groundwater flow. Some indication of this is seen in the relatively narrow discharge fluctuations of the latter site (c.v., 56.0 per cent). This compares with values of 116.3 and 107.1 per cent at the other springs.

3. Underground streams

A total of 11 cave streams were sampled in the Setul Boundary Range (table 82). Nine, including the stream which drains Perlis Mines' Cave (site 4304), are shown in figure 5. The two remaining stations are natural underground tributaries of the latter stream which were sampled at a distance of between 750 m and 1 km from the cave entrance. At certain sites, indicated in table 82, the cave stream emerges from a water-filled constriction. At others, the cave entrance is open and allows free entry of air into the vadose passages. Often the alluvium in the latter grouping of caves has been disturbed by mining activities, and consequently, as in Perlis Mines' Cave, artificial pools may be present underground. These are usually separate from the cave stream, and are not thought to markedly affect groundwater properties. At the time of sampling, only Perlis Mines' Cave was being worked. Thus the main stream in this cave (site 4304), and also Kaki Bukit cave stream (site 4307), which appears from mine surveys to be linked with Perlis Mines' Cave, may be affected by current mining operations.

4. Surface streams

Data were obtained at 31 sampling stations (figs. 3 to 5). Details of the catchment areas of these streams are presented in table 83. Because of the disposition of the limestones and their advanced stage of erosion, few surface streams have a significant

groundwater contribution from the limestone hills. The main exceptions are the Sg. Pelarit (sampled at three locations) and the Sg. Gelang Kerbau (sampled approximately 350 m downstream from the main rising), which drain the Setul Boundary Range. Limestone hills account for between 45.7 and 100 per cent of the catchments of the four stations investigated on these rivers. Elsewhere, they cover less than 20 per cent of the catchments studied, and many streams derive almost all, or a significant part, of their flow from the lowland plains. Such streams rise either on the lowlands, or on adjacent granite or sandstone/shale outcrops. In Selangor and the Kinta Valley limestones underlie much of the alluvial plain. In the North-west region, however, the lowland plain of the Sg. Tasoh is mostly underlain by non-carbonate lithologies. A third group, comprising seven streams which rise in mountainous terrain on non-carbonate rocks, were sampled at points before they reach the karst alluvial plain (eg. the two headwater streams of the Sg. Tempurong, sites 3301 and 3302). These streams are, therefore, representative of purely allogenic waters flowing into the karst regions.

5. Seepages in the alluvium of the karst alluvial plains

A total of nine mines were investigated in the Kinta Valley (fig. 4). In all, 48 seepages were sampled. Of these 37 were from the sub-alluvial karst surface, the rest being entirely confined within the alluvium at various heights above the underlying bedrock.

6. Deep groundwater

Deep groundwater was monitored regularly at Pinji Fish Farm (site 3311), which is located close to the point where the Simpang Pulai-Lahat road crosses the Sg. Pinji (fig. 4). Here water is pumped from a depth of approximately 20 m beneath the sub-alluvial karst surface.

D. SYSTEMATIC AND SPATIAL VARIATIONS IN KARST WATER PROPERTIES

In this section the following karst water properties are considered: average water temperatures, water temperature variability, average water hardness properties, total hardness variability, calcium, magnesium, Mg:Ca ratios, sodium, potassium, and K:Na ratios.

1. Average water temperatures

Two types of heat transfer affect the average temperatures of karst groundwaters. First, heat moves by convection, as when heat in water is moved along with water flow. The effects of groundwater flow upon subsurface temperatures have been recognised in several studies (eg. Diment and Robertson, 1963; Domenico and Paliauskas, 1973; Hackbarth, 1978; Schneider, 1964). Secondly, there is conduction in response to temperature gradients within a rock mass. Thermal conductivities of limestones of between 6.0 and 8.2×10^{-3} cal/(cm s ^{o}C) have been reported (Birch and Clark, 1940). However, thermal conductivities are much greater if water is the saturating fluid rather than air (Woodside and Messmer, 1961). Heat transfers resulting from contact with circulatory air must also be taken into consideration. Many factors are important in determining underground air temperatures. These include air movement, external conditions, the configuration of underground passages, distance from the surface and the liberation and absorption of latent heat by condensation and evaporation (Wigley and Brown, 1976). When groundwaters emerge as surface flows, or when only surface streams are considered, diurnal air temperature changes and other factors such as direct insolation receipt become important.

Water temperature data were obtained for 214 sampling points. At 138 of these the mean value is based on six or more measurements. Overall, the average water temperature is 25.02^{o}C (fig. 49). Values range from less than 23.0^{o}C at six diffuse-flow seepages in Gua Batu to an average of 27.5^{o}C at the 48 seepages in

alluvium of the karst plains. Despite the diverse and complex controls operative, certain fundamental systematic and regional differences in mean water temperature may be observed in the three field areas.

1.1. Underground diffuse-flow seepages

Of the six groups of karst waters investigated, underground diffuse-flow seepages generally have the lowest mean temperatures. The average temperature of 121 sites is 23.74°C. This is considerably less than the mean annual surface temperatures, which range from 26.3°C in Selangor to 26.9°C in the North-west region. This result may be explained by two factors. First, soil temperatures, at least on slopes with a fairly continuous vegetation canopy are approximately 2°C lower than the mean annual temperature (Chapter II, Section B2.4.iv). At 30.1 cm in ABT-I, for example, the soil temperature averages 24.3°C. Percolating soil waters thus initially have a relatively low temperature, and the effect of this may be transmitted underground by convection as heat is carried along with water flow. Parallels are seen in groundwater temperatures in Alberta, Canada, where it has been shown that shallow groundwater temperatures are controlled more by temperatures at times of major infiltration than by mean annual air temperatures (Hackbarth, 1978). Secondly, cave air temperatures are consistently low, the observed range being from 22.84 to 24.52°C. The absence of direct insolation receipt may be a key factor here. The average cave air temperature of 23.90°C lies remarkably close to that of underground diffuse-flow seepages (23.74°C). Thus in outcrops with vadose and abandoned high-level cave passages, groundwater and, presumably, rock temperatures seem to be in equilibrium with ambient soil and cave air temperatures.

1.1.i. Systematic variations. Of the 121 seepages, all except 16 could be classified with certainty into the four groups considered

above (Section C1.5). The lowest mean temperature (23.49°C) was recorded at a group of 30 points where water was collected after it had flowed a distance of more than 1.5 m over, or through, stalactitic deposits. Sites of this group have greatest contact with circulatory cave air. Their lower temperatures may be explained, in part, by cooling induced by evaporation from the thin films of water which often coat the surface of such secondary deposits. A further contributory factor could be endothermic calcite formation as carbon dioxide is lost from the water. However, the effects of this mechanism are thought to be small (Gascoyne, 1974). By contrast, seepages from bedrock joints and from stalactites less than 1.5 m in length have very little, if any, direct contact with cave air after emergence. They have notably similar mean temperatures of 23.78 and 23.80°C, respectively. The highest average temperature was recorded at the final group of 11 sites where water spills from air-filled conduits. The rapid response of these sites to rainfall, especially in Gua Batu, indicates a relatively free and open connection with the surface. Compared with the other groups of seepages, these waters are therefore more likely to be affected by temperatures outside the cave, and this probably accounts for their higher temperature.

1.1.ii. Comparison between the four caves. Interesting inter-regional differences in water temperature may be observed in the data. The 36 seepages in Gua Batu and 27 in Gua Anak Takun, have mean temperatures of 23.27 and 23.26°C, respectively. Corresponding values for 34 sites in Gua Tempurong (Kinta Valley) and 24 in Perlis Mines' Cave/Subway Tunnel are 24.01 and 24.63°C. Differences between the three regions are statistically significant at the 99.9 per cent confidence level. This result may be explained largely by the progressive increase in mean annual temperature northwards along the western coast from 26.3°C in Selangor to 26.9°C in the North-west region. A further factor, namely the proportions of the

different types of underground seepages, is not thought to be an important control in the caves studied for the following reason. The two caves with numerous high-level seepages from air-filled conduits, Gua Batu and Perlis Mines' Cave/Subway Tunnel, were found also to have a high proportion of seepages from large stalactites. Consequently the slightly warmer waters associated with the former seepages are counterbalanced by the generally less warm waters of the latter. Waters in Gua Anak Takun and Gua Tempurong on the other hand are mostly from stalactites less than 1.5 m in length, and these typically have temperatures in the middle of the range.

1.2. Karst springs

The temperature of the three springs sampled averages 24.94°C. The highest mean temperature (25.51°C) was recorded at site 4317, the spring which is thought to receive a significant contribution from deeper groundwater. The hillslope behind the rising, though steep, has a fairly continuous soil cover and few enlarged joints are evident in rock outcrops. The slightly warmer waters at this site probably reflect the deeper flow and restricted contact with circulatory air in the underground passages. The relatively high proportion of waters at the two other springs which is derived directly from throughflow or subcutaneous flow may account for the lower temperatures recorded at these sites (minimum, 24.00°C at 4312).

1.3. Underground streams

The mean temperature of 11 cave streams which drain the Boundary Range is 24.86°C (range, 24.37 to 25.37°C). No difference was observed between caves with open entrances and those where water fills to the cave roof. This suggests that air enters the stream passages of the latter group of sites through cave entrances and other openings, such as widened joints and bedding planes, located at higher levels. Dry, high-level caves occur along much of the eastern front of the Boundary Range. The importance

of air flow through solution-widened fractures in rock outcrops has been demonstrated in other studies (eg. Cedar Spring, Missouri - Aley, 1970).

Specific comparisons between underground diffuse-flow seepages and cave streams may be made in Perlis Mines' Cave. Seepage waters in this cave and Subway Tunnel have a mean temperature of 24.63°C (range, 23.67 to 25.29°C). This is almost identical to the temperatures of the two tributary streams sampled underground (sites 4019 and 4025) which have mean values of 24.66 and 24.70°C, respectively. By comparison, the water emerging at the cave entrance (site 4304) has a slightly lower mean temperature of 24.37°C. This difference suggests that some cooling occurs along the underground streamway. Similar results have been found along certain cave streams in Venezuela (Gascoyne, 1974).

1.4. Surface streams

The temperature of the 31 surface streams sampled averages 26.03°C (fig. 49). This is significantly higher than the corresponding values of underground diffuse-flow seepages, karst springs and cave streams at the 99.9 per cent confidence level (t-tests). Contact with warmer surface air and direct insolation receipt provide the most likely explanation of the observed temperature differences.

Where surface waters receive substantial groundwater recharge, their temperature generally remains fairly low. Thus the mean temperature of the four stations on streams draining the Setul Boundary Range is 25.01°C, with a 24.57 to 25.22°C range. This is very close to the mean values of 24.90 and 24.86 °C recorded for karst springs and cave streams in the same area. A low mean temperature of 24.57°C (range, 23.57 to 26.88°C) was also observed in the seven streams with densely-forested mountainous catchments on non-carbonate lithologies. The shade afforded by the vegetation canopy is probably the main factor here.

As noted previously (Section C4), many surface streams crossing the karst plains derive much of their flow from the alluvium. The 20 sites sampled in this category have a mean temperature of 26.97°C. This is significantly higher than that of the two previous groups of surface sites (t-test; $p<0.001$). The lowest temperatures were recorded for the Sg. Tasoh (North-west region), which has a mean value of 25.07°C. The alluvial plain of this river is used intensively for padi and other types of agriculture, and the river banks are stable and mainly wooded, providing shade along much of its length. In marked contrast, vegetation is absent from large areas of the alluvial flats of Selangor and the Kinta Valley which have been much disturbed by tin-mining activities. Highest mean temperatures were recorded for the Sg. Kinta (station 3306, 28.30°C), the Sg. Johan (28.81°C), the Sg. Gopeng (29.60°C) and the Sg. Tekah (30.40°C). All four streams drain intensively-mined areas, and in the last case approximately two-thirds of the flow is derived from the alluvial plain. In addition to the lack of shade on these plains, two further facets of mining operations may cause higher river temperatures. First, extensive areas of the surface are occupied by shallow mining pools which may become excessively heated during the day. Water from such pools is released periodically to the rivers. Secondly, the sediment carried by the rivers, being dark in colour, may directly absorb solar radiation and conduct heat to the water. Significantly, the Sg. Tekah had the highest sediment load during the sampling period.

1.5. Seepages in alluvium of the karst plains

The temperature of 48 seepages sampled within the alluvium averages 27.50°C (range 24.60 to 32.30°C). This is approximately 1.0°C higher than the mean annual temperature. The warm waters may simply reflect the general lack of shade on the alluvial plains. Intense heating of sands and gravels in the immediate vicinity of

the seepages may also be a contributory factor. This could cause an increase of temperature as waters emerge at the surface.

1.6. Deep groundwater

The temperature of the borehole water in the Kinta Valley averages 26.76°C. This is very close to the mean annual temperature of the region (26.5°C). This finding is consistent with basic texts on groundwater which suggest that as a general principle groundwater temperatures should be approximately equal to the mean annual temperature (eg. Domenico, 1972; Walton, 1970). In turn, the closeness of these two temperatures confirms the value of water temperature data.

2. Water temperature variability

Short-term temporal variations in the heat transfers discussed above may cause small fluctuations in groundwater temperatures. For example, the thermal characteristics of inflowing water may vary according to diurnal or seasonal temperature waves and with rainfall events. Once the water enters the ground, both rock and water begin to drift towards an equilibrium temperature, but with the rock tending to act as a small source or sink of heat, depending on the persisting influence of the thermal characteristics induced in rock by preceding water inflows. Variations in the temperature of circulatory air may also be important. Surface waters, in contrast, are markedly affected by direct insolation, and diurnal and seasonal temperature rhythms. Recent studies in temperate regions suggest that water temperature variability provides an important index by which the flow characteristics of karst risings may be differentiated (Jacobson and Langmuir, 1974b; Shuster and White, 1972) and by which groundwater contributions to surface flow may be assessed (Pitty, 1979).

In common with other studies, the standard deviation (s.d.) of observed temperatures at a particular site is used here as an index of water temperature variability. Discussion is confined to 139 locations at which six or more measurements were made. With the

exception of alluvial seepages, each main group of sampling sites is represented. Overall, temperature fluctuations have an average s.d. of 0.515°C. Values range from 0.05°C in deep groundwater to greater than 2.0°C at seven stations on surface streams (fig. 50).

2.1. Underground diffuse-flow seepages

Underground percolation waters generally have very restricted water temperature variabilities, with an average s.d. of 0.26°C at 96 sites (fig. 50). However, the minimum value observed of 0.08°C distinguishes this group from the deeper groundwaters of the borehole. Temperature fluctuations of these seepages might be anticipated to be interpretable by the depth of groundwater flow (variations being moderated at greater depth), and by the degree to which water comes in contact with circulatory cave air.

2.1.1. Systematic variations. Consideration of the four sub-types of underground seepage (Section C1.5, above) provides some insight into the mechanisms which may be operative. Fourteen sites could not be classified with certainty and are not considered further. Of the remaining 96 sites, those which exhibit least variation are the group of five seepages which emerge directly from rock joints (average s.d., 0.17°C; range, 0.12 to 0.19°C). Fluctuations are generally greater where water percolates through stalactitic deposits before emergence. For 41 such seepages from smaller stalactites of less than 1.5 m in length, an average s.d. of 0.21°C (range, 0.08 to 0.28°C) was recorded, whereas for 26 seepages from larger depositional forms, the mean s.d. was 0.30°C (range, 0.18 to 0.44°C). However, the widest fluctuations were observed at points where water spills from high-level, air-filled conduits. The mean s.d. of ten such sites is 0.40°C, with a maximum value of 0.62°C.

The above results suggest that the degree of contact with cave air considerably affects temperature variability. The average s.d. for cave air temperatures at ten sites was 0.48°C

(range, 0.29 to 0.94°C). This value, which is significantly higher (t-test; $p<0.05$) than that of all except the last group of seepage sites, indicates that the variability of water temperatures is a characteristic more inherited from contact with bedrock rather than induced by rapid adjustment to circulatory air temperatures. Only at those sites where water flows from air-filled conduits does water temperature variability fall within the range of observed air temperature measurements, and even here the mean s.d. of 0.40°C is considerably below that of the latter. Also, since flow in many of the air-filled conduits responds quickly to storms, the temperature characteristics of these waters are more likely to be affected by surface temperatures. Evidence of this is provided by the generally higher temperatures of this group of diffuse-flow seepages (Section D1.1.i, above).

2.1.ii. Comparisons between the four caves. If one assumes that the sites sampled are broadly representative of water flow within the four caves, then notable differences in water temperature variability may be observed between the caves. Highest variability occurs in Gua Batu, with a mean s.d. for diffuse-flow seepages of 0.29°C, whereas the lowest variability is in Gua Anak Takun (mean s.d., 0.22°C). This difference, which is statistically significant at the 99.9 per cent confidence level, reflects the varying proportions of different types of diffuse seepages in the two caves. In Gua Batu, like Perlis Mines' Cave (mean s.d., 0.27°C), many of the seepages are from high-level, air-filled conduits and large stalactitic deposits, both of which characteristically exhibit relatively wide temperature fluctuations. By contrast, most seepages in Gua Anak Takun, and also in Gua Tempurong (mean s.d., 0.24°C), are from stalactites of less than 1.5 m in length.

2.2. Karst springs

Sufficient measurements for the assessment of water temp-

erature variability were made at sites 4312 and 4317. Smallest temperature fluctuations (s.d., 0.14°C) were observed at the latter site. This finding is consistent with the deep-groundwater flow postulated for this spring. The shallower groundwater emerging at the second spring has a wider temperature variability of s.d. 0.24°C.

2.3. Underground streams

Temperature fluctuations of nine cave streams for which six or more measurements were made are also restricted (average s.d., 0.27°C), with more than half the sites having variations of less than s.d. 0.20°C (fig. 50). The average s.d. lies close to that of the 96 underground diffuse-flow seepages, but is notably less than the average values of those seepages draining air-filled conduits (mean s.d., 0.40°C) and of cave air (mean s.d., 0.48°C). It thus seems that the higher ratio of water volume to surface area in the underground streams, as compared with the shallower flows of the conduit seepages, limits the rate of adjustments to ambient cave air temperatures.

2.4. Surface streams

Once groundwater emerges as surface flow, a progressive increase in water temperature fluctuations can be observed (fig. 50). The s.d. of the 31 surface sites sampled averages 1.40°C, with values varying from 0.23 to 2.28°C. Some indication of the rate at which temperature variability increases upon emergence is seen in the data from the headwaters of the Sg. Gelang Kerbau, close to Abi Head water works (fig. 51). Measurements were made at the main rising and also at five points at distances of between 25 and 345 m downstream. Temperature variability increases from s.d. 0.34°C to 0.66°C over this distance.

In many limestone regions with substantial groundwater recharge directly into surface streams, water temperature fluctuations may remain within narrow limits. The four stations of this type,

on streams draining the Setul Boundary Range, have an average s.d. of 0.66°C (fig. 50). The lowest value, s.d. 0.23°C, was recorded for the Sg. Pelarit (site 4303). Temperature fluctuations are somewhat higher in streams with forested mountain catchments on non-carbonate lithologies. However, their variability, which averages s.d. 1.01°C (range, 0.78 to 1.54°C), is small compared with that of rivers draining the alluvial karst plains (mean s.d., 1.68°C). This is probably due to the mountainous river courses occupying deep valleys and to the shade added by dense vegetation canopies.

Of the rivers draining the alluvial plains, the Sg. Tasoh has the smallest temperature fluctuations with an average s.d. of 0.69°C. This is consistent with the low mean temperature recorded for this stream and is accounted for by the stable and mainly wooded banks which provide shade along much of its length. By contrast, the smallest variability recorded for the rivers on the alluvial plains of Selangor and the Kinta Valley is s.d. 1.19°C, and seven sites have values in excess of s.d. 2.0°C. The higher variability of temperature is attributed partly to the lack of shade. These rivers will readily absorb solar radiation periodically because of their generally heavy sediment loads. A further effect of mining activites is the irregular release of waters from surface pools to the rivers. This may cause marked and rapid changes of river water temperature.

2.5. Seepages in alluvium of the karst plains

No temperature variability data were obtained.

2.6. Deep groundwater

The low variability of water temperatures of s.d. 0.05°C at the borehole is comparable to that of 0.035°C reported for artesian wells and springs at the lower end of hydraulic gradients in highly permeable basalts in Oahu, Hawaii (Mink, 1964, p. 5288). It is slightly less than the narrowest fluctuations recorded for

over 300 sites in northern England where an underground stream emerges into Peak Cavern Derbyshire, from a natural siphon with water temperature fluctuations of s.d. 0.08°C (Pitty et al., 1979).

2.7. Discussion of results

Studies in temperate regions suggest that temperature data give the best indication of the seasonal response of karst springs (Jacobson and Langmuir, 1974b; Shuster and White, 1971). Hitherto, comparable data on temperature variability have been lacking for shallow-depth groundwaters in humid tropical karst environments. In the Malaysian data, the deeper flows have fluctuations closely similar to those observed elsewhere, whilst the cave seepage data give a clear demonstration of the variability that is encountered in an aquifer with some circulatory air. For the surface streams, striking modifications appear to have been induced by tin mining. Clearly, any geohydrological characteristics, such as water temperature variability, which relates even indirectly to velocity, volume and depth of groundwater flow could provide an important index for hydrological studies, for two reasons. First, the methods used for logging boreholes, although numerous, yield little data describing directly such fundamental aquifer characteristics as porosity and permeability. Furthermore, it is difficult to test the utility of analytical equations since the basic data cannot be obtained readily by direct measurement. Secondly, a comprehensive index, such as water temperature variability, which identifies a groundwater component in surface flow has a broader potential utility. As R.A. Freeze (1972, p. 622) has remarked, the prominent role played by subsurface mechanisms in the generation of surface runoff is not recognised in certain approaches to hydrology.

3. Average water hardness properties

Water hardness properties are taken here to include pH, specific conductance, and total, alkaline and non-alkaline hardness. The average water hardness properties of a given autogenic karst water are dependent essentially upon four factors: (1) the initial chemical aggressiveness of percolating waters towards the limestone; (2) the rate at which the potential to dissolve limestone is expended; (3) the degree of contact with circulatory cave air which has a lower partial pressure of carbon dioxide than that of soil air; and (4) the rate at which reprecipitation proceeds from waters which are supersaturated with respect to the dominant aquifer mineral, usually calcite. A further factor, namely dilution, must also be considered when karst waters receive significant contributions from non-carbonate terrain. Since each of the above factors in turn is governed by a complexity of natural controls (see, for example, review by Atkinson and Smith, 1976), the influence of one specific variable on water hardness properties is difficult to isolate. However, by sampling at a large number of points the total spatial variability in water hardness characteristics is encompassed, and the effects of certain of the more important variables may be established.

Total hardness varies widely over the 324 stations sampled, ranging from 4.2 to 1306 ppm. The most extreme values were recorded at three groups of sites. Two of these are non-karstic in character: first, the seven stations draining non-carbonate catchments (mean total hardness, 18.7 ppm), and secondly, the 11 seepages in alluvium sampled at points above the sub-alluvial karst surface (mean total hardness, 83.9 ppm). The third group, comprising 28 underground seepages which are thought to be contaminated by bat guano, are atypical karst waters in that a high proportion of their hardness

is associated with anions other than bicarbonate (Section D3.1 - II, below). These sites have a mean total hardness of 367.3 ppm (maximum, 1306 ppm). The remaining 278 sampling points, which may be regarded as true karst waters, have a mean total hardness of 151.2 ppm, with a range of 15.5 to 373.5 ppm (table 84; fig. 54). Alkaline hardness (fig. 55) and non-alkaline hardness (fig. 56) average 143.4 and 7.8 ppm, respectively. Overall, the latter accounts for only 4 per cent of the total hardness (fig. 57). This, combined with the highly significant positive correlation observed between alkaline and total hardness (r=0.976, $p<0.001$) indicates that carbonic acid is the major agent of limestone solution in the field areas. A highly significant positive correlation also occurs between specific conductance, which averages 273.7 μmho, and total hardness (r=0.965, $p<0.001$; fig. 58). This result is consistent with the findings of workers in temperate regions (eg. Junee Resurgence, Tasmania - Goede, 1973; S. Wales - Bray, 1969) and is important in that it shows that specific conductance provides a reliable basis for estimating the total hardness of tropical karst waters.

3.1. Underground diffuse-flow seepages - I. Sites unaffected by guano

Characteristics of the 202 underground seepages which appear to be uncontaminated by guano are summarized in table 84. The waters are all alkaline, the range of pH values being from 7.28 to 8.19. Total hardness averages 140.7 ppm, with almost two-thirds of the sites (66.3 per cent) falling between 100 and 200 ppm (fig. 54). Non-alkaline hardness (mean, 8.0 ppm) accounts on average for 5.4 per cent of the latter. Relationships between the various water hardness properties are presented in table 86. Specific conductance, total hardness and alkaline hardness correlate directly, and each

is inversely related to pH. The relationships with pH are consistent with the theory of saturated calcite solutions. The weaker positive correlation observed between pH and non-alkaline hardness is less readily explained and may simply be a result of the inverse relationship between non-alkaline and alkaline hardness.

Approximate equilibrium curves for saturated calcite solutions in the presence of 0, 5, and 30 per cent magnesium at 25°C are shown in figure 59 (after Pickmett, 1972). Those waters with Mg:Ca ratios in the range 0 to 30 per cent are plotted. Of these, 85 per cent fall either within, or above, the area enclosed by the curves. Clearly, therefore, the majority of the sites are close to the saturation point or supersaturated with respect to calcite.

The total hardness concentrations of waters at the soil/rock interface and in the subcutaneous zone were observed to range between 130.9 and 395.9 ppm (Chapter VI). Corresponding values for runoff from bare and partially-covered rock surfaces range from 22.3 to 108.2 ppm (ChapterV). It would thus seem that the total hardness contents of the underground seepages, which range from 65.3 to 362.6 ppm, may be derived almost entirely from the surface and near-surface weathering zones, with the presence of different proportions and types of bare and soil-covered surfaces accounting for the observed spatial variability.However, a further aspect of karst waters, that of redeposition of limestone, must also be considered. The effects of the latter process may be seen more clearly when the different groups of diffuse-flow seepages are considered separately.

3.1.i. Systematic variations. Of the four groups of diffuse-flow seepages, those from air-filled conduits have the highest hardness values, with a mean value of 182.7 ppm (range, 125.9 to 251.3 ppm). This is significantly higher (t-test; $p<0.05$) than seepages from bedrock joints and from stalactites greater than 1.5 m in length, which average 132.4 and 126.5 ppm, respectively.

It is also higher than seepages from smaller stalactites (mean, 147.9 ppm). In general, open conduit flow is more rapid than that of other seepages studied, and responds to single rainfall events. Water temperature data (Section D1.1.1) indicate that these sites are closely connected with the surface of the outcrops. It may be assumed, therefore, that a high proportion of their waters are derived directly from surface runoff or subcutaneous flow, with only a relatively short flow path through bedrock joints. Opportunities for deposition are thought to be limited during the rapid transit through the air-filled passages. Consequently, it is suggested that the solute contents of this type of seepage are most closely representative of the capacity of precipitation and soil waters to dissolve limestone in the uppermost few metres of bedrock.

In general, drips from stalactites have lower total hardness contents, with a minimum value of 65.3 ppm. Evidence that deposition is the primary cause is seen in the fact that seepage waters from the 99 smaller stalactites (<1.5 m long), which have restricted opportunity for deposition, have significantly higher total hardness concentrations than those from 53 larger stalactites (t-test; $p<0.05$). Percolation waters which emerge directly from bedrock joints are distinguished from those from stalactites by the absence of very low total hardness values (minimum, 101.9 ppm). However, their mean value of 132.4 ppm is not significantly different. As suggested previously, these sites occur only under exceptional circumstances, and it seems likely that at a number of the eight seepages of this type the absence of depositional forms reflects the low hardness of emergent waters. This could occur, for instance, where a seepage is fed by a joint, or discrete set of joints, which connects with a predominantly bare rock catchment at the surface of the outcrop. The fact that sites of this type are lacking in Perlis Mines' Cave/ Subway Tunnel supports the hypothesis that groundwater flow in lime-

stones of the Setul Formation is of an integrated type, with individual seepages being fed by larger, and presumably more diverse, surface catchments.

3.1.ii. Comparisons between the four caves. Water hardness characteristics of the four main cave systems are summarized in table 87. Highest hardness values occur in Perlis Mines' Cave/Subway Tunnel, where the 24 seepages have a mean total hardness of 230.2 ppm (range, 136.5 to 362.6 ppm). This is significantly higher than in the other caves (t-test; $p<0.001$). The reasons for the relatively high hardness values are probably twofold. First, the surface topography above the caves is mostly of the steep and moderately-inclined hill slope types, as investigated in WT-II and WT-I (Chapters IV and VI). Thus most groundwater passes through pockets of mineral soil before entering the bedrock, thereby developing a high level of aggressiveness towards the limestone (cf. semi-natural seepage, Wang Tangga - Chapter VI, Section E2). Secondly, since flow rates are generally high, much of the limestone dissolved in the upper part of the outcrop will be transported through the system without being deposited.

By comparison, a much higher proportion of the surface relief above the remaining cave comprises morphological units of the rocky hilltop type. This is especially true of Gua Anak Takun (mean total hardness, 111.0 ppm) which for the most part underlies relief typified by plots ABT-IV and ABT-III. Average total hardness concentrations of 153.4 and 124.2 ppm, respectively, were recorded in Gua Batu and Gua Tempurong. The slightly higher hardness of seepages in the former cave probably reflects the generally higher contents of seepages from the high-level conduits.

The proportion of total hardness accounted for by non-alkaline hardness differs between the four caves, ranging from -0.3 per cent in Perlis Mines' Cave/Subway Tunnel to 21.8 per cent in

Gua Batu. The absence of appreciable non-alkaline hardness in the former passages (maximum, 5.7 ppm) suggests that chemical aggressiveness derived from sources other than carbon dioxide is small. This is consistent with data from the subcutaneous flow sampled near WT-I (Chapter VI) and supports the hypothesis advanced earlier that organic acids are relatively unimportant solutional agents in well-aerated soils. In contrast, runoff from rock surfaces with thin organic-rich soils and accumulations of organic litter often contains appreciable amounts of non-alkaline hardness (eg. ABT-III/IV; Chapter V). The comparatively high concentrations of non-alkaline hardness in Gua Batu (mean, 21.8 ppm; range, 3.9 to 62.9 ppm) may partly originate from this source. However, as the following observations in Gua Anak Takun indicate, inorganic anions must also be considered as possible components of non-alkaline hardness.

A detailed survey by the author of Pipe Corridor, Snake Room and Liphistius Room in Gua Anak Takun is presented in figure 43. This shows the vertically dipping shale beds which outcrop in the cave, and the location and solute contents of seepages sampled. Although total hardness and potassium exhibit no obvious spatial pattern, the ratio of non-alkaline:total hardness (per cent) appears to be notably higher in the vicinity of the shales. Seven seepages sampled in this part of the cave emerge from the shale/limestone junction. The average proportion of non-alkaline hardness at these sites is 16.2 per cent (range, 8.5 to 34.9 per cent). This compares with 2.3 per cent (range, -5.6 to 11.6 per cent) at the remaining sites shown. It seems likely, therefore, that the shales are a source of inorganic anions. The relatively high concentrations of non-alkaline hardness in the throughflow waters of ABT-II, which lies about 10 m to the north of Windgate, may thus be partially derived from this source (see Chapter VI, Section E1.1). Shale partings of this type may contribute to the generally high non-alkaline

hardness levels of Gua Batu, though none is evident in cave walls and roof.

3.1.iii. Comparison with other studies. Drips and seepages in caves in other parts of the humid tropics mostly lie within the total hardness range of 65.3 to 362.6 ppm observed in the present study. Cave seepages in the Mkulumuzi Gorge, N.E. Tanzania, for example, have values in the range 150 to 300 ppm (Cooke, 1971), whilst the corresponding range at three sites in Finim Tel, New Guinea is 121 to 140 ppm (Brook, 1976). The latter sites lie within the 100-150 ppm modal group recorded in West Malaysia. A mean calcium hardness of 95 ppm, with a 82 to 105 ppm range, has also been reported from several sites in the Caribbean (Corbel and Muxart, 1970). Although magnesium concentrations are not stated in the last study it seems likely that these sites will also fall in the 100-150 ppm range.

Studies in many extra-tropical regions have also yielded values which lie within, or close to, the above range. For example, under warm temperate maritime conditions in the Waitomo area, North Island, New Zealand, Gunn (1977a) reports calcium hardness levels of between 80 and 117 ppm, whilst under cool temperate conditions, New South Wales, Jennings (1972b) records a mean total hardness value of 152 ppm for 35 cave drip samples in Murray Cave. However, diffuse percolation waters in the British Isles tend to cluster towards the upper end of the range recorded in the present study (eg. five caves in the Mendip Hills, mean total hardness, 187 ppm - Atkinson, 1971; Coole Cave, W. Ireland, total hardness range, 216 to 416 ppm - Williams, 1968).

3.1. Underground diffuse-flow seepages - II. Sites affected by guano

During preliminary sampling in Gua Batu and Gua Anak Takun certain seepages were found to have markedly different solute characteristics from the majority of the sites. In particular they had a high specific conductance, total hardness and a low alkaline hardness. In Gua Batu waters of this type were either from high-level

conduits where bats were observed to congregate in large colonies, or from large depositional features, located in such positions that bat excrement would fall on them from the roosts. Two samples from the cave had nitrate concentrations of 84 and 120 ppm (analysis by D.I.D., Malaysia). These figures confirm the above interpretation, since Hem (1970, p.118) finds high nitrate concentrations are characteristic of pools in caves frequented by bats.

In Gua Anak Takun the link with bats is less obvious, since few bats are seen in that part of Bench Corridor where sites of this type were located. It should be noted that in 1974/75 guano deposits in the lower passages of Gua Anak Takun were much less extensive than those indicated in the survey by Dunn (1965; fig. 42). Seepages with the characteristics identified above do, however, occur in a fairly clearly defined zone (fig. 44), the boundary of which suggests a relationship with the passages of the upper cave. Certain sites are directly beneath North Point (as mapped by Dunn, 1965) and all lie within an area which could receive seepage waters from East and North Points. It is suggested therefore that the solutes in these waters are derived largely from guano on the floor of the upper cave.

Seepage waters which were known to be affected by guano were found to have a non-alkaline:total hardness of more than 30 per cent and potassium concentrations in excess of 2 ppm. These limits were used as criteria in identifying contaminated samples where contact with guano was less certain. One further seepage which had an average of 47.1 per cent non-alkaline hardness, but contained only 1.02 per cent potassium, was also classified as being contaminated since it occurred less than 1 m from a site which fulfilled both criteria. 28 seepages of this type were investigated, and their water hardness properties are summarized in table 88. These sites are distinguished from pure seepages (table 85) by their higher levels of specific conductance (mean, 767.4 μmho), total hardness (mean, 367.3 ppm), non-alkaline hardness (mean, 282.6 ppm) and

ratio of non-alkaline:total hardness (mean, 68.5 per cent), and by their lower concentrations of alkaline hardness (mean, 84.7 ppm). In each case the difference is statistically significant at the 99.9 per cent confidence level (t-test). Little difference was observed between the pH of guano-affected sites and that of the pure karst seepages. This contrasts with data reported from Finim Tel, New Guinea (Brook, 1976, p.188), where one drip which is thought to be affected by guano has an exceptionally low pH of 6.5, but otherwise differs little from other seepages.

The consistently low alkaline hardness contents of guano-affected waters in Gua Batu and Gua Anak Takun (maximum, 130.0 ppm) explains why secondary deposits were generally poorly developed or absent, at these sites. The remainder of the seepages from bed-rock joints were of this type.

The results from Gua Anak Takun show that where an outcrop is frequented by bats, patterns of karst water quality in a cave system may provide a fairly reliable indication of the presence and approximate location of open, higher level passages. This finding has clear practical applications in the field of speleological exploration in that guano-affected seepages in a known part of a cave system may indicate those areas in which extensions are most likely to be made.

3.2. Karst springs

The absence of low total hardness values, minimum 280.6 ppm (fig. 54), and the overall mean value of more than 300 ppm (table 89) distinguishes the karst springs from the other groupings of karst waters. The waters of all three springs have Mg:Ca ratios in the range 6 to 9 per cent and from figure 60 it may be seen that they are either slightly aggressive or saturated with respect to calcite, but not supersaturated. As with the subcutaneous flow and underground percolation seepages in the Setul Boundary Range, non-

alkaline hardness forms an insignificant component of the total hardness. The highest total hardness, 332.0 ppm, was recorded at the supposed deep groundwater spring (site 4317).

Site 4317 probably provides the best measure of the solutional potential of percolation waters in the outcrops of the peninsula for two reasons. First, available evidence suggests that the emergent water is mainly from deep groundwater sources, and presumably therefore its average residence time in the aquifer is longer than that of the majority of sites investigated. Studies in Britain have shown that mean hardness values increase in proportion to the length of flow-through time (Ede, 1975). However, it should be noted that the total hardness of the subcutaneous flow in Wang Tangga (mean, 341.9 ppm) is very close to that recorded at this spring. A very high proportion, if not all, of the solutional capacity of soil waters thus appears to be expended close to the surface of the outcrop. Secondly, and in contrast to most other groundwaters sampled, waters at this spring are not quite saturated with respect to calcite. This suggests, therefore, that flow is mostly confined within tight water-filled fractures (cf. water temperature fluctuations of c.v. 0.14^{o}C), and that no deposition will have occurred before emergence. The absence of tufa deposits below the spring is also indicative of the slightly undersaturated nature of the emergent waters.

If the above assumptions are correct then the difference between the total hardness of this spring (332.0 ppm) and that of the percolation waters in Perlis Mines' Cave/Subway Tunnel (mean, 230.2 ppm) should provide some measure of the amount of underground deposition which may already have taken place before waters emerge as seepages in the cave passages. This difference amounts to 101.8 ppm. It seems, therefore, that as much as one-third of the limestone taken into solution in the Setul Boundary Range may be redeposited before the waters emerge into the major vadose passages.

3.2.1. Comparison with other studies. Exact comparisons with springs in other areas are often somewhat difficult to make since precise descriptions of the catchment characteristics are not always given. However, data are available for several autogenic springs. In Jamaica, for example, Dromily Clear Stream spring, near Maroon Town, has a mean total hardness of 168 ppm (Smith, 1969a), whilst several springs in the Browns Town area have a corresponding figure of 172 ppm (Day, 1976). A mean calcium hardness concentration of 223.1 ppm has been recorded for the Tjikolo Well spring in the Kalapanunggal karst region of W. Java (Balázs, 1971). Considerable local variability of spring waters has been reported from Puerto Rico where the observed range is 150 to 328 ppm (Giusti, 1978), and Thailand, where values vary between 93 and 262 ppm (Pitman, 1978b). Outside the humid tropics, Williams (1977) reports a mean total hardness of 127 ppm at Flat Creek rising in the Pikikiruna Range, South Island, New Zealand, and Smith and Mead (1962) give values of 235 and 290 ppm, respectively, for risings in the Mendips and the Cotswolds. Clearly, the springs observed in the Setul Boundary Range, and especially site 4317, are at the uppermost end of the range recorded at other sites. However, without more detailed information concerning the flow characteristics of these other springs, and especially the opportunity for deposition within the aquifer, the reasons for this finding cannot be isolated.

The observation that limestone solution is mostly concentrated near the surface of limestone outcrops has been widely reported both from the humid tropics (eg. several areas in the Caribbean - Corbel and Muxart, 1970; Cockpit Country, Jamaica - Smith et al., 1972) and from temperate regions (eg. Cooleman Plain, New South Wales - Jennings, 1972; East Yorkshire - Pitman, 1978a; Derbyshire - Pitty, 1968b; W. Ireland - Williams, 1968). However, whilst the presence of massive secondary deposits in caves of the humid tropics has often

been cited as indirect evidence of rapid underground deposition in this environment (eg. Jennings, 1971, p.174), karst water studies have hitherto provided no direct measure of the rate of this process.

3.3. Underground streams

The total hardness of the 11 cave streams is notably less than that of the karst springs, with concentrations varying between 180.2 and 280.8 ppm (table 90). The overall mean value of 227.1 ppm is very similar to that of seepages in Perlis Mines' Cave/Subway Tunnel and suggests that very little deposition occurs along the underground streamways. More exact comparisons may be made in the case of the stream which issues from Perlis Mines' Cave (site 4304). Total hardness at the cave entrance averages 186.1 ppm. Since this value lies well within the observed range of diffuse-flow seepages (136.5 to 362.6 ppm), the solute data provide no definite evidence for deposition along the course of the stream. This is consistent with a lack of secondary depositional features on the stream bed. It thus seems that the structural characteristics and disposition of the limestones of the Setul Formation, by permitting the development of integrated diffuse-flow seepages and active vadose stream passages, also create rapid underground flow conditions which favour the removal of dissolved limestone from the outcrop. However, as noted above (Section 3.2), even under these conditions considerable redeposition may occur before waters enter the cave streams.

3.3.1. Systematic and regional variations. Differences in the water hardness properties of cave streams with open and water-filled entrances are small, the mean total hardness values being 224.8 and 229.8 for these groupings of sites. However, when the spatial distribution of water hardness values is examined a much clearer trend may be seen. Cave streams with mean total hardness concentrations of less than 220 ppm are confined to two zones (fig. 5): (I) Wang Tangga/Kaki Bukit (sites 4304, 4305 and 4307), and (II) near Kangar

(site 4318). The limestones in both areas are relatively pure, the average acid insoluble residue contents of the bedrock in the two regions being 2.35 and 4.17 per cent, respectively. Sites with higher hardness values are similarly confined to two zones: (III) the eastern front of the Setul Boundary Range between 1 km south of Kaki Bukit and 2 km south of Bt. Wang Mu (sites 4309, 4310 and 4311), and (IV) the area around Abi Head water works (sites 4315, 4316). Total hardness concentrations in these two zones average 249.6 and 228.0 ppm, respectively. The limestones are notably less pure, with the corresponding figures for the acid insoluble residue content being 8.94 and 7.08 per cent. This relationship between rock purity and water hardness is consistent with the contrast noted previously between rock purity and soil vegetation characteristics in the Boundary Range (cf. plots WT-I and GKV-I). However, whilst the contrasts between the shallower, better-aerated soils of the purer limestone beds and the deeper and poorly-aerated soils of the impure limestones are reflected in the water hardness of the risings, the difference is less marked than would be predicted from the observed soil carbon dioxide concentrations (Chapter VI, Section C). Underground deposition, which will tend to be more rapid where percolation waters have initially equilibriated with soil air of a high partial pressure of carbon dioxide, may explain this finding.

3.3.11. Comparison with other studies. Exact comparisons with results from other areas are somewhat difficult to make, especially as many cave streams receive some input from non-carbonate rocks. However, the hardness concentrations of underground streams in the Boundary Range are markedly higher than those reported by Smith and Atkinson (1976, fig. 13.2, p. 375) for 34 river and spring sites in various tropical regions, which have a mean of 174 ppm. Only three of the sites in the latter study have total hardness contents of more than 200 ppm, and these all fall in the range 200-240 ppm.

3.4. Surface streams

The total hardness of the 24 karst streams averages 68.9 ppm, which is significantly lower than the other groups of karst waters at the 99.9 per cent confidence level. These waters are also distinctive in that they exhibit a significant positive correlation between total hardness and pH ($r=0.614$, $p<0.001$). This relationship is characteristic of waters with hardness concentrations well below those of saturated calcite solutions.

Overall, total hardness correlates positively with the proportion of the drainage basin occupied by limestone hills ($r_s= 0.735$, $p<0.001$) and by karst alluvial plains ($r_s=0.332$, $p<0.05$), and negatively with the proportion occupied by non-carbonate rocks ($r_s= -0.639$, $p<0.001$). Thus, groundwater contributions from the limestone outcrops are the main sources of total hardness. Similar relationships between water hardness and the percentage of limestone in catchment areas have been reported from temperate regions (eg. Mendip Hills - Newson, 1971). The positive correlation observed between water hardness and the proportion of karst alluvial plain in the drainage basin is also important in that it supports the observation made by Douglas (1968c) that water hardness is considerably augmented by waters from areas underlain by limestone. This finding was based on detailed investigations of a single river basin (Sg. Gombak) where alluvium overlies more than 97 per cent of the catchment area which is of limestone lithology.

A minimum total hardness of 156.8 ppm distinguishes the four sampling points on rivers with significant groundwater contribution from the remaining streams (table 91). The highest value, mean 203.8 ppm, was recorded at station 4334 on the Sg. Gelang Kerbau. Significantly, this is the only catchment investigated which is composed entirely of limestone hills. Of the surface streams this is also the only one which shows evidence of active deposition. Along

steeper reaches the stream bed is formed entirely of thick tufa deposits (plate 57), whilst tufa dams are present at lower gradients (plate 58). Some indication of the magnitude of deposition may be gained by comparing the average total hardness concentrations of the stream water with corresponding figures for the three risings sampled. The main rising (station 4315), which provides more than three-quarters of the stream flow has a mean total hardness of 226.8 ppm. Values for the other risings (sites 4316 and 4312) are 229.1 and 280.9 ppm, respectively. The sampling station on the river is located 345 m downstream of the main rising, and over this length of flow total hardness decreases by approximately 50 ppm.

Two further features distinguish the streams with significant groundwater contribution. First, the waters are alkaline in reaction, the range of pH being from 7.34 to 7.90. Secondly, the proportion of non-alkaline to total hardness is close to zero (mean, -0.2 per cent). The close correspondence between total and alkaline hardness is in agreement with results from diffuse-flow seepages, springs and cave streams in the Boundary Range.

The average total hardness of those streams draining catchments which contain small outcrops of limestone and relatively large areas of karst alluvial plain is 48.6 ppm. This compares, for example, with calcium hardness values in the range 70 to 80 ppm for streams in Sarawak draining alluvial plains with residual hills of the Melinau Limestones (Sweeting, 1979). In the present study, total hardness concentrations in excess of 100 ppm were recorded at 3 stations, all in the Kinta Valley. Highest levels occur in the Sg. Sengat (station 3310) and its small east-bank tributary (station 3309), with mean concentrations of 127.9 and 124.7 ppm, respectively. The catchments of both streams include relatively high proportions of limestone hills (10.3 and 16.3 per cent) and karst alluvial plains (77.5 and 83.7 per cent). However, this fact alone

does not completely explain the high hardness of the waters since the catchments of the Sg. Pinji at stations 3307 and 3308 have similar percentages of limestone outcrops and plains, but yet have total hardnesses of 32.9 and 48.5 ppm, respectively. The distinctive feature of the area drained by the Sg. Sengat is the shallowness of the alluvial cover. This is particularly the case in the vicinity of G. Kroh where the limestone bedrock is frequently exposed at the surface in a pavement-like form (plate 39). Under these circumstances a significant proportion of the waters derived from the karst plain will have been in contact with the limestone (Section D3.5, below).

Water quality variations in the headwaters of the Sg. Tempurong demonstrate most clearly the significance of water hardness contributions from a shallow-depth, sub-alluvial karst surface. The main headwater streams draining the granites of G. Chante (sites 3301 and 3302), with catchment areas of 0.6 and 0.7 km^2, have mean total hardness concentrations of 26.2 and 9.3 ppm, respectively. However, at the eastern entrance of Gua Tempurong (site 3303) total hardness averages 105.5 ppm. Runoff from G. Tempurong, which forms an estimated 7 per cent of the third catchment, may partly account for the increased water hardness. However, field observations suggest that waters derived from the rock outcrops exposed in mine workings and from the shallow-depth, sub-alluvial karst surface of the plain (15.3 per cent of area) are a more likely source.

The karst alluvial plains of the three streams with high hardness concentrations contrast markedly with those of the Sg. Pinji and other surface streams where the sub-alluvial karst surface is for the most part buried to some depth. In the Kinta Valley, for example, the alluvium is mostly between 3 and 15 m in thickness (see Ingham and Bradford, 1960, figure 8, p.79). Consequently, runoff from the plains is often derived mostly from sediments well above the alluvium/rock interface. Such waters typically have low hardness

values (Section 3.5, below). This probably explains the low total hardness contents of the remaining 17 surface karst streams which lie in the range 15.5 to 66.4 ppm.

The mean ratio of non-alkaline to total hardness for the 20 surface streams of the second group (-11.1 per cent), is significantly less than that recorded for the four stations on streams with high proportions of groundwater contribution from limestone hills (t-test; $p<0.05$). The negative values in this case provide a measure of the bicarbonate anions in excess of those required to balance the calcium and magnesium ions in solution. The difference between the two groups of streams may be explained by two factors. First, chemical weathering of non-carbonate rocks often yields relatively high proportions of cations additional to calcium and magnesium. Secondly, amounts of non-alkaline hardness 'deficiency', which are small in absolute terms, are exaggerated by the lower total hardness of such waters when expressed in relative terms. As may be anticipated, streams draining purely non-carbonate terrain have much lower non-alkaline to total hardness ratios, with a mean value of -49.1 per cent. Such streams also have low total hardness values (mean, 18.7 ppm), the maximum being 29.6 ppm (table 91).

3.5. Seepages in alluvium of the karst plains

Seepages at the alluvium/rock interface have widely varying total hardness contents (fig. 54), ranging from 109.2 to 373.5 ppm. As in the case of the surface streams, pH varies directly with total hardness ($r=0.440$, $p<0.001$). This suggests that some of the waters sampled are very much undersaturated with respect to calcite. The aggressiveness of the 37 seepages sampled (method, Chapter III, Section C4), averages 32.6 ppm $CaCO_3$, with values varying from 0 to 130 ppm. The majority of waters at the alluvium/rock interface are thus capable of dissolving more limestone. Some of this aggressiveness may well be expended within the limestone aquifer.

Evidence that solutional processes are active below the watertable is provided by the high frequency of land subsidence in the Kinta Valley. This phenomenon, according to F.S. Chong of the Geological Survey, Malaysia (pers. commun.), is caused by collapse in the sub-alluvial karst where the roofs of caverns are weakened by solution.

By augmenting the mean total hardness of these waters (224.5 ppm) by their measured aggressiveness, the final value of 257.1 probably represents the average dissolutional potential of alluvial seepages. This is considerably lower than the corresponding figure of 332.0 ppm suggested for groundwater recharge in the Setul Boundary Range. It is nevertheless quite high in view of sparse vegetation cover of large areas of the plains. The observed hardness contents of these waters substantiates the claim, based purely on morphological evidence of the deeply-dissected form of the sub-alluvial karst surface, that considerable solution occurs beneath the alluvium (Ingham and Bradford, p. 30; plates 42, 43 and 49).

The total hardness of 11 seepages sampled above the limestone surface averages 83.9 ppm. This is significantly lower than the concentrations recorded at the alluvium/rock interface (t-test; $p<0.001$). However, the aggressiveness of these waters is notably higher, with a mean value of 72.7 ppm $CaCO_3$.

3.6. Deep groundwater

Water in the limestone aquifer beneath the alluvial plain of the Kinta Valley has four main sources: direct seepage from the overlying alluvium, groundwater from the tower karst hills, channel leakage from surface streams, and groundwater from adjacent non-carbonate terrain. The total hardness of the borehole water, which averages 216.9 ppm (table 93), may be accounted for by the varying contributions from these sources. Waters seeping through the alluvium in the Kinta Valley generally have the highest total hardness (mean, 257.1 - after correction for aggressiveness). Corresponding

figures for seepages in the tower karst hills (eg. Gua Tempurong) and surface streams are 136.3 and 50.1 ppm, respectively. The water hardness of groundwater in non-carbonate terrain is probably of the same order of magnitude as that of surface streams in those areas (mean, 18.7 ppm). These observations suggest that a high proportion of the groundwater in the aquifer originates as percolation through the alluvium, and that contributions from other sources cause relatively little dilution. This result is consistent with the location of the borehole in the centre of the karst alluvial plains of the Kinta Valley, at some distance from the larger limestone towers and the adjacent granites of the Main and Kledang Ranges.

3.6.1. Comparison with other studies. The total hardness of the deep groundwater lies well within the 130 to 339 ppm range recorded for deep artesian wells in Puerto Rico (Giusti, 1978). Trainer and Heath (1976), in a synthesis of the bicarbonate content in carbonate groundwaters in eastern North America, report alkaline hardness values of between 138 and 446 ppm.

4. Total hardness variability

Fluctuations in the hardness of karst waters may result from changes in the aggressiveness of percolating waters, in the rate at which the solutional potential is expended, in the degree of contact with air in the aquifer and, finally, in the rate at which secondary deposition occurs. Usually these variations are induced by seasonal temperature rhythms (eg. Derbyshire and Yorkshire - Pitty, 1966, 1974; Gams, 1962), by seasonal rainfall patterns and by single rainfall events. Whilst the variables which cause water hardness fluctuations are often difficult to isolate, the magnitude of water hardness variability of a karst water provides a useful index of carbonate aquifer type (Shuster and White, 1971). Diffuse percolation systems, for example, generally display a lower order of hardness variation than conduit systems which receive a

high proportion of their waters from swallets. Applications of water hardness variability indices, usually the coefficient of variation (c.v.) of calcium or total hardness, may be seen in several recent studies in temperate karst regions (eg. Cooper and Pitty, 1977; Halliwell, 1979; Paterson, 1972; Ternan, 1972). Hitherto, however, comparable data are lacking for karst regions in humid tropical environments. In West Malaysia, as in other humid tropical regions, temperature variations are characteristically small in magnitude compared with seasonal and day to day variations in rainfall. Attention in this section focuses, therefore, upon relationships observed between rainfall, discharge and total hardness.

As in the case of discharge and water temperature variability, those sites from which six or more measurements were obtained are considered here. To simplify discussion the guano-affected seepages and non-karst waters are excluded in this section. Overall, the total hardness fluctuations of 137 karst waters averages 10.05 per cent. Values range from less than 2.0 per cent in deep groundwater and several percolation seepages, to more than 40 per cent in three surface streams (fig. 61).

4.1. Underground diffuse-flow seepages

Water hardness fluctuations of the underground seepages are notably small (mean c.v., 7.12 per cent). More than half of the sites have c.v. values of less than 5.0 per cent (fig. 61), the arbitrary figure suggested by Shuster and White (1971) as distinguishing diffuse from conduit flow. All but three of the 101 sites have a c.v. of less than 20 per cent. For comparison, presumed diffuse-flow systems in Poole's Cavern, Derbyshire (Pitty, 1966) and the High Mark area, Yorkshire (Ternan, 1972) have variabilities of up to 13.2 and 10.6 per cent, respectively.

At one site in Gua Tempurong the total hardness of 12 samples varied between 95.3 and 100.5 ppm, a range of 5.2 ppm.

Equally small ranges of specific conductance (7.0 μmho) and alkaline hardness (5.0 ppm) were recorded at this site. Since analytical error may account for a significant proportion of the observed variability, the water hardness of this seepage, and of others where the measured range is less than 12 ppm, may effectively be regarded as constant. 35 of the 101 seepages fall into this category.

4.1.i. Seasonal trends in total hardness. Discharge appears to be the key variable which influences water hardness fluctuations of the remaining 66 seepages. At 25 locations statistically significant positive correlations were observed between flow rate and total hardness (r_s; $p<0.05$). Of these, 14 correlations are significant at the 99.9 per cent confidence level. Reduced opportunity for deposition per unit volume of water over a given flow length provides the most likely explanation of this finding. Temporal variations in the Mg:Ca ratio of seepage waters (Section 5.1, below), and the fact that none of the guano-affected sites displayed a positive correlation with discharge, substantiate this. A positive correlation between water hardness and discharge has been reported for cave drips in N.E. Tanzania (Cooke, 1971).

At 52 of the sites, statistically significant positive correlations between total hardness at the time of sampling and rainfall during pre-set, antecedent periods have been confirmed (for method of analysis, see Chapter VI, Section D2.2.i). The period with which total hardness correlates most closely at the various sites is indicated in table 94. In exactly half of the cases, the closest correlation is with one of the first nine periods. These are the periods which were considered in relation to discharge variations (Section C, above), and relate to the total rainfall over given time periods immediately prior to sampling. The total hardness variability of this group of seepages would appear therefore to be mostly affected by discharge. The highest number of sites in this category are from Perlis Mines' Cave/Subway Tunnel, and nine

of the 13 seepages correlate most closely with rainfall in periods 1-16 days or less. This finding is consistent with the hydrological characteristics of the Setul limestones. Equally, the fact that none of the seepages in Gua Tempurong fall in this group is in keeping with the low variability of discharge and the lagged response time of discharge to rainfall in this cave.

The total hardness fluctuations of six seepages in Gua Tempurong correlate most closely with lagged antecedent rainfall periods (table 94). At four the closest relationship is with the period 99-112 days before sampling. Since no significant positive correlation was observed between discharge at the time of sampling and rainfall during this period, it seems likely that the water hardness variability observed at these sites reflects changes in the solutional potential groundwater recharge induced by rainfall events. The positive correlations observed in the three foothill plots between soil carbon dioxide and rainfall in antecedent periods of 1-32, 1-64 and 1-128 days before sampling (Chapter VI, Section D4) may be significant in this respect. However, the link with the 99-112 day water hardness correlation is difficult to establish. Although no short-term variations in soil carbon dioxide were detected in the present study (see discussion, Chapter VI, Section D6), it seems likely that carbon dioxide concentrations will increase during storms as the porosity of the surface soil horizons is reduced. Soil waters will therefore become more aggressive during storms than at other times. If this assumption is correct, then the 99-102 day antecedent period at the four seepage sites in Gua Tempurong is interpretable in terms of the flow-through time of

these groundwaters. The wide range of flow-through times predicted for the other caves (table 94), especially Gua Batu and Gua Anak Takun, is perhaps indicative of discrete networks of fractures, each with their own hydrological characteristics.

4.1.ii. Effects of single storms. Marked short-term fluctuations in total water hardness were observed at those seepages with flow rates which respond to single storms. Two contrasted types of seepage, site 2001 in Gua Anak Takun and site 4001 in Subway Tunnel, illustrate the magnitude of water hardness variability which may occur over a period of several days. Unfortunately, continuous recording/sampling equipment was unavailable. Consequently, the trends observed may not be completely representative of the full range of flow conditions which occurred during the periods studied. Nevertheless, certain clear trends emerge from the data.

1. Site 2001 is a seepage from a small stalactite 20 cm in length, which is located approximately 8 m from the entrance of Gua Anak Takun. Discharge and total hardness averaged 10.63 ml/min and 147.6 ppm, respectively, during the year. The seepage was sampled at midday on ten occasions from 1 to 12.7.75 following a period of two weeks in which less than 10 mm of rain fell. Rainfall, discharge and total hardness data for this period are shown in figure 62. The flow rate (0.22 ml/min.) and total hardness (98 ppm) on 1.7.75 were both well below average. A storm of 39.1 mm on the morning of 2.7.75 resulted in a marked increase of discharge to 3.0 ml/min by midday, by which time the total hardness had increased to 157 ppm. The peak discharge of 11.0 ml/min on 3.7.75 coincided with a peak total hardness of 164 ppm. Discharge then fell progressively to 0.55 ml/min on 6.7.75 (total hardness, 115.9 ppm). A similar trend was observed between the 7th and 12th of July, following a heavy storm of 36.3 mm on the afternoon of 6.7.75 (fig. 62).

2. Site 4001, in contrast to the previous site, is of the open

conduit type. It is located in the roof of Subway Tunnel. During the year the observed discharge ranged from 0.3 to 180 l/min, with a mean of 13.5 l/min. The seepage was sampled on 12 occasions during the period 16 to 26.11.74 (fig. 63). Light rainfall, amounting to 21.0 mm, fell in three showers during the previous week. The discharge at 12.00 hrs on 16.11.74 was 0.5 l/min and the total hardness 140 ppm. Following an intense downpour of 17.3 mm during the afternoon of the 16 th, the discharge increased to 1.7 l/min by 18.30 hrs. Over this period, the total hardness increased to 144 ppm. A short shower of rain (4.4 mm) on 17.11.74 had no detectable effect upon the flow rate, which dropped to 0.7 l/min (total hardness, 140 ppm) by 18.00 hrs on that day. Rain, totalling 64.8 mm, then fell almost continuously between 20.00 hrs on 18.11.74 and 16.00 hrs on the following day. The highest discharge (16.4 l/min) was recorded during the storm at 10.00 hrs on 19.11.74, whereas the peak total hardness (225 ppm) occurred at 19.00 hrs, shortly after the end of the storm when the flow rate decreased to 11.3 l/min. Apart from 2.3 mm of rain on 25.11.74, the remainder of the study period was dry. Discharge and total hardness declined progressively to 0.9 l/min and 142 ppm, respectively, at midday on 26.11.74.

One notable difference between this seepage and site 2001 is that above a certain discharge the relationship between total hardness and discharge becomes negative. This is confirmed by the fact that samples obtained at higher discharges of 54 and 180 l/min following other storms had total hardness concentrations of 202 and 190 ppm, respectively. The rapid movement of water through the soil and bedrock, and the consequent reduced opportunity both for absorption of carbon dioxide by soil waters and for dissolution within the bedrock, probably accounts for the inverse relationship between water hardness and discharge observed at certain sites under exceptionally high flow conditions.

4.1.iii. Comparison between the four caves. Differences in water hardness variability between the four caves are consistent with

the above interpretations. The widest fluctuations (mean c.v., 10.49 per cent) occur at seepages in Gua Anak Takun, which also have the most variable discharges (fig. 48). Sites in Perlis Mines' Cave/Subway Tunnel and Gua Batu are of intermediate variability, with average c.v. values of 8.96 and 6.78 per cent, respectively. At the other extreme, the narrowest fluctuations (mean c.v., 3.56 per cent) occur in Gua Tempurong, the cave with lowest variability of discharge (fig. 48).

4.2. Karst springs

The magnitude of total hardness fluctuations varies markedly between the two springs for which the c.v. index was determined (fig. 61). Narrowest fluctuations, of c.v. 5.96 per cent, were recorded at site 4317, the presumed deep-groundwater spring. The total hardness at this site, which ranged between 306 and 363 ppm, varies directly with discharge (r=0.438). In contrast, wide fluctuations (c.v. 20.03 per cent) were observed at site 4312. Here, total hardness exhibits a weak inverse correlation with discharge, and it is suggested that varying contributions of throughflow and subcutaneous flow under different antecedent rainfall conditions may account for the greater variability observed at this shallow-groundwater spring.

4.3. Underground streams

The water hardness fluctuations of cave stream waters in the Boundary range remain notably within narrow limits. At nine sites for which the c.v. was determined, a maximum value of 11.05 per cent was recorded, with a mean of 7.95 per cent. The latter figure is slightly less than that of 8.96 per cent measured at percolation seepages in Perlis Mines' Cave/Subway Tunnel. This finding clearly suggests that whilst many of the underground streamways are large open conduits, a high proportion of the cave stream waters are derived from diffuse percolation sources. This confirms the assumption made earlier (Chapter VI, Section F1) that surface runoff to swallets during heavy downpours is of limited magnitude.

4.4. Surface streams

Of the groups of karst waters studied, the surface streams, on the whole, exhibit the widest water hardness variability (fig. 61). Total hardness fluctuations at the four stations on streams with significant groundwater contributions do, however, remain notably within narrow limits, the range of c.v. values being 8.55 to 11.47 per cent. The mean figure of 10.49 per cent recorded at these stations is somewhat higher than that of cave streams (mean, 7.95 per cent) and deep-groundwater springs (5.96 per cent) in the Boundary Range. Nevertheless, it is equal to that recorded for diffuse-flow seepages in Gua Anak Takun, and is slightly less than the upper limit of 10.6 per cent for diffuse-flow systems in Yorkshire (Ternan, 1972).

The total hardness variability index of the remaining sites, which receive considerable inputs from the karst alluvial plain, ranges from 8.71 to 62.97 per cent, with a mean value of 25.80 per cent. Temporal variations in the proportion of water contributed from allogenic parts of the drainage basins probably explain much of the variability, both at individual river stations and between different catchments.

4.5. Seepages in alluvium of the karst plains (no data obtained).

4.6. Deep groundwater

Deep groundwater from the borehole in the Kinta Valley is distinguished by its low water hardness variability (fig. 61). During the study period total hardness varied between 212 and 220 ppm, with a c.v. of 1.64 per cent .

5. Calcium, magnesium and the Mg:Ca ratio

Hitherto, no quantitative comparisons have been made in karst regions between the rates at which calcium and magnesium are cycled by vegetation and the rates at which these elements are removed by groundwater. However, as a broad generalization, based on

an assimilation of data on different components of nutrient cycling and rock weathering from separate studies, Trudgill (1977, p. 102) has suggested that in areas of calcareous bedrock "cycling is self-contained and tends to be small relative to the larger weathering and leaching processes".

Most attention in karst geomorphology has focused on the differential removal of calcium and magnesium in solution from limestone outcrops. Theoretical considerations related to the effects of magnesium ions on limestone dissolution have been discussed earlier (Chapter V, Section A1.2.1). In general the Mg:Ca ratio of natural karst waters corresponds closely with that of the carbonate rocks which they drain (see, for example, Atkinson and Smith, 1976, fig. 5.4, p. 157). However, studies of karst springs emerging from the calcareous Dolomite Formation in N.E. Transvaal, S. Africa, have revealed marked seasonal patterns in the Mg:Ca ratio (Marker, 1973b). Here, higher levels of biological activity during the summer wet season favour more rapid solution, and concentrations of calcium increase relative to magnesium. During the drier winter months less calcium is dissolved, but concentrations of magnesium, which is slightly more soluble, either remain constant or increase. Field evidence also suggests that deposition of calcite from karst waters which contain some magnesium may produce a measurable increase in the Mg:Ca ratio. This process has been reported in the case of cave seepages (eg. Indian Echo Cave, Pennsylvania - Holland et al., 1964), shallow-groundwater aquifers (eg. Puerto Rico - Giusti, 1978) and tufa depositing surface streams (eg. N.E.Transvaal - Marker, 1973a).

Calcium concentration of the 278 karst waters investigated in the present study, guano-affected seepages excluded, averages 49.9 ppm (expressed as Ca^{2+}). The magnesium content, expressed as Mg^{2+}, averages 6.39 ppm. Overall, the mean molar Mg:Ca ratio is 0.223, with a 0.01 to 1.09 range. The corresponding figure for limestones in the three field areas, based on 197 rock samples, averages

0.220 (range, 0.002 to 1.02). This close correspondence clearly suggests that the Mg:Ca ratio of the karst waters is very much influenced by the chemical composition of the limestone.

5.1. Underground diffuse-flow seepages - I. Sites unaffected by guano

5.1.i. Calcium. Calcium concentrations of the 202 diffuse-flow seepages average 48.5 ppm, with values ranging from 41.4 ppm in Gua Anak Takun to 79.2 ppm in Perlis Mines' Cave/Subway Tunnel. Since the calcium content, or more strictly calcium hardness, is the major component of total hardness and correlates closely with the latter (r=0.917, $p<0.001$), readers are referred to Sections 4 and 5 above for discussion of water hardness variability.

If surface runoff is assumed negligible (Chapter VI, Section E1.5.i), and the mean annual effective precipitation is assumed to equal the difference between mean annual rainfall and estimated evapotranspirational loss (table 14), then volumes of underground flow in the three field areas may be calculated (table 95). From these data, and the solute contents of percolation waters, the amounts of the various elements removed in groundwater may be estimated. Two points should be noted with respect to these calculations. First, whilst percolation waters in caves provide an appropriate basis for estimating rates of solute removal from limestone outcrops in Selangor and the Kinta Valley which lack springs and permanent underground streams, data from the latter types of site (Section 5.2/5.3, below) are used here to estimate rates of solute loss for the Setul Boundary Range. Secondly, the net removal of calcium, and to a lesser extent magnesium (Section D5.1.iii, below), is considerably less than gross removal from the near-surface zone since redeposition often occurs within the limestone prior to the emergence of seepage waters.

The amount of calcium removed by groundwater in the three field areas averages 443.1 kg/ha/yr (table 95). This compares with a mean annual input of 21.79 kg/ha/yr in rainfall (table 52), giving a net loss of 421.3 kg/ha/yr. The latter figure is only slightly greater than the amount recycled by vegetation in the six study

plots (mean, 342.5 kg/ha/yr; table 95). Thus, in the case of moderately-inclined hill and foothill slopes, almost as much calcium is absorbed by plant roots as is lost from the rock outcrops in groundwater.

5.1.ii. Magnesium. The magnesium content of the 202 diffuse-flow seepages averages 4.74 ppm. Highest values were recorded in Gua Batu and Perlis Mines' Cave/Subway Tunnel, where the mean concentrations are 8.42 and 7.85 ppm, respectively. The average rate of magnesium removal from the limestone outcrops studied is 37.8 kg/ha/yr (table 95), compared with mean input of 2.13 kg/ha/yr from rainfall (table 52). The net loss of magnesium (mean, 35.7 kg/ha/yr) is slightly less than the combined weight in canopy leaching and litterfall in the six main study plots. As noted previously, the Mg:Ca ratio of exchangeable cations in soils indicates that magnesium is retained preferentially within the soil/vegetation system where the bedrock is relatively deficient in this element (Chapter IV, Section E6.2). It seems likely, therefore, that most of the magnesium in the groundwaters of magnesium-poor limestones is derived from bedrock below the normal rooting zone of plants.

5.1.iii. Mg:Ca ratio. The molar Mg:Ca ratio of the 202 seepages averages 0.168. Marked differences may be observed between the four caves (fig. 64). Highest proportions of magnesium occur in Gua Batu where the mean Mg:Ca ratio of 48 sites is 0.319 (range, 0.12 to 1.09). The wide variability of waters sampled in the cave is paralleled closely by rock analytical data. Of 12 rock samples analysed, nine have values of less than 0.2, whilst the remaining three fall in the range 0.7 to 0.9. The observed mean value of 0.255 is slightly less than that of the seepages.

High proportions of magnesium also occur in Perlis Mines' Cave/Subway Tunnel, where the Mg:Ca ratio of percolation waters averages 0.203 (range, 0.06 to 0.70). Correspondence with bedrock composition in this case is very close, the mean Mg:Ca ratio being

0.217 (range, 0.016 to 0.711). In Gua Anak Takun and Gua Tempurong magnesium accounts for a much smaller proportion of the total hardness. The Mg:Ca ratios of seepage waters in the two cases average 0.077 and 0.120, respectively. As in Gua Batu, these figures are higher than those for bedrock, which average 0.024 and 0.068.

Although the rock samples analysed are taken mostly from the roof and walls of the caves, and may, therefore, not be completely representative of the sequence of limestones drained by percolation waters, the fact that waters in three of the caves have higher Mg:Ca ratios than the limestone merits further consideration. As discussed previously, magnesite is slightly more soluble than calcite. This may be important in two respects. First, dissolution of limestone may be incongruent, with magnesium ions being released into solution more readily than calcium ions. Evidence of this process is seen in the solute content of runoff from bare rock surfaces (Chapter VI, Section D2.1.ii). It should be noted, however, that magnesium is absorbed preferentially from soil waters, and presumably from subcutaneous flow, by plant roots. Clearly the latter process will tend to counter the tendency for disproportionate loss of magnesium ions from the uppermost layers of bedrock. This will be especially the case where a dense vegetation cover has developed on soils with a low Mg:Ca ratio of exchangeable bases. Indeed, the fact that many hill soils have higher Mg:Ca ratios than the underlying limestone suggests that the net removal of calcium may be disproportionately high.

The second, and probably most important, consequence of the differential solubilities of calcite and magnesite is that the former will be precipitated first from saturated solutions. Two independent lines of evidence suggest that this mechanism is operative in the field areas. First, the Mg:Ca ratios of secondary deposits are consistently low, the maximum value of 42 samples being 0.025.

Mean values based on between 8 and 16 samples from each cave were 0.011, 0.004, 0.016 and 0.007, respectively, for Gua Batu, Gua Anak Takun, Gua Tempurong and Perlis Mines' Cave/Subway Tunnel. Secondly, statistically significant inverse correlations ($p<0.05$) were found between the Mg:Ca ratio and discharge at 31 of the 101 sites from which six or more samples were obtained. Nine of these correlations are significant at the 99.9 per cent confidence level. There are no significant positive correlations. Clearly, therefore, rates of deposition of calcium carbonate per unit volume of water are inhibited at higher discharges. The fact that seepages in Perlis Mines' Cave/Subway Tunnel are distinguished by having Mg:Ca ratios which are similar to those of the bedrock further supports the mechanism of depositional control, since rates of deposition are thought to be lower here than in the other caves investigated.

5.1. Underground diffuse-flow seepages - II. Guano-affected sites

The Mg:Ca ratios of seepage waters in Gua Batu and Gua Anak Takun contaminated by guano are similar to those recorded at uncontaminated sites. In Gua Batu, for example, 11 such waters have a mean Mg:Ca ratio of 0.252 (cf. 0.319 for pure karst seepages). It is interesting to note that this value is almost identical to that of bedrock in Gua Batu (mean, 0.255). Since deposition from these waters is negligible, these results lend further support to the idea that the amount of deposition is a key factor which influences the degree of incongruity between rock and water.

In Gua Anak Takun the Mg:Ca ratio of guano-affected waters averages 0.044, which is slightly less than that of the uncontaminated seepages. This result may be explained by the spatial distribution of Mg:Ca values within the cave (figs.43 and 44). The seepages sampled may be divided broadly into two groups according to whether their Mg:Ca ratio is greater or less than 0.10. Sites of the former category are almost entirely restricted to that part of the

cave between, and adjacent to, the shale outcrops. It is suggested that the higher magnesium contents of these waters result either from the weathering of the shale, or from the occurrence of limestones with a higher Mg:Ca ratio in the vicinity of the shale beds. However, more detailed sampling of bedrock than was undertaken in the present study would be required to isolate the exact cause. Waters with Mg:Ca ratios of less than 0.10 occur throughout the rest of the cave, including those sites affected by guano. This result suggests that the calcium and magnesium in the contaminated samples are mostly derived from the dissolution of limestone.

5.2/5.3. Karst springs and underground streams

The mean calcium and magnesium concentrations of the 14 spring and cave stream waters in the Setul Boundary Range are 87.8 and 5.92 ppm, respectively. The average Mg:Ca ratio of 0.121 (range, 0.06 to 0.22) compares with a value of 0.108 for 37 limestone samples from the Setul Formation. Thus, as in the case of underground diffuse-flow seepages in Perlis Mines' Cave/Subway Tunnel there is close correspondence between the Mg:Ca ratios of karst waters and bedrock.

5.4. Surface streams

The four sampling stations on surface streams draining the Boundary Range have an average Mg:Ca ratio of 0.125, with a 0.09 to 0.14 range. Since the mean value is very similar to that of spring and cave stream waters in the same area, it would appear that the Mg:Ca ratio of surface streams with significant groundwater contributions is largely determined by the input from the limestone hills.

Streams draining non-carbonate terrain, and seepages in alluvium from locations above and at the junction with the underlying limestone, typically have much higher Mg:Ca ratios. The observed mean values for these three groups of sites are 0.401 (range, 0.03 to 0.85), 0.268 (range, 0.12 to 0.71), and 0.489 (range, 0.05 to 0.85), respectively.

The varying contributions from these sources, and from the limestone hills, probably account for the wide range of Mg:Ca ratios (0.12 to 0.68) observed at the 20 stations on surface streams which receive considerable input from the karst alluvial plains.

5.5. Seepages in alluvium of the karst plains

The Mg:Ca ratio of limestones beneath the alluvial plain of the Kinta Valley has a very pronounced bi-modal distribution, with 23 of the 40 samples analysed having values of less than 0.12 and the remainder being greater than 0.65. A mean Mg:Ca ratio of 0.535 was recorded for 33 seepages from the alluvium/bedrock interface in this region.

5.6. Deep groundwater

Groundwater from the borehole at Pinji Fish Farm has a Mg:Ca ratio of 0.080, the lowest recorded for any of the main types of karst water. As noted above, more than half of the limestones sampled beneath the alluvial plain of the Kinta Valley have Mg:Ca ratios of less than 0.12. The small proportion of magnesium at this site may, therefore, simply reflect the absence of magnesium-rich limestone or dolomite in the vicinity of the extraction point. Clearly, a more detailed geological investigation than that undertaken in the present study would be necessary to confirm this. Certainly there is no evidence of selective precipitation of calcium within the aquifer, as has been found, for example, in Puerto Rico (Giusti, 1978).

6. Potassium, sodium and the K:Na ratio

Rapid nutrient cycling within the soil/vegetation system, with comparatively little loss to groundwaters and surface streams, is a characteristic feature of many rain-forested areas on non-carbonate lithologies (see, for example, review by Douglas, 1969). Deficiencies of calcium and, to a lesser extent, magnesium are unlikely to limit plant growth on limestone outcrops in the humid

tropics, since both elements are freely available from the bedrock. However, the potassium and sodium contents of limestones are generally small compared with those of the other rocks. It might be anticipated, therefore, that these elements will be of major ecological significance in the karst regions, and that their levels within the soil will be actively maintained by vigorous plant uptake. The potassium and sodium concentrations observed in underground seepage waters in West Malaysia, combined with the measured rates of cycling of these elements (Chapter IV, Section E), provide a unique opportunity to investigate the complete nutrient balance of these two elements. Additionally, studies by Christopher (1975, 1977) in Derbyshire have suggested that the K:Na ratio might provide a useful index for characterizing groundwaters, the K:Na ratio tending to decrease with increased residence time of waters within a carbonate aquifer. The present study affords an opportunity for testing this relationship in a humid tropical environment.

The potassium and sodium contents of the 278 karst waters sampled are summarized in figures 65 and 66. Overall, the concentrations of potassium (mean, 0.84 ppm) are lower than those of sodium (mean, 1.41 ppm), with the molar K:Na ratio averaging 0.450 (range, 0.01 to 2.65; fig. 67).

6.1. Underground diffuse-flow seepages - I. Sites unaffected by guano

6.1.1. Potassium. The potassium content of the 202 seepages sampled averages 0.52 ppm, with a 0.01 to 2.70 ppm range. Estimates of the rate of removal of potassium in groundwater are presented in table 95. Values vary from 2.76 kg/ha/yr in the North-west region (based on karst springs and underground streams) to 9.02 kg/ha/yr in the Kinta Valley. The net loss of potassium from the limestone outcrops, after correcting for rainfall input (table 52) averages 1.80 kg/ha/yr in three field areas. Potassium is barely detectable in most of the limestones sampled, the maximum concent-

ration observed being 0.02 per cent by weight. Consequently, soil rather than bedrock is the most likely source of this element. However, even if it is assumed that the entire potassium content of underground seepages originates from this source, this amounts to only 1.4 per cent of the average total input of 130.23 kg/ha/yr observed in the six study plots. This result is consistent with findings made in Chapter IV which suggest that there is very active cycling of potassium by the vegetation cover. Thus, where the reserves of an important plant nutrient, such as potassium, are relatively small, comparatively little is lost by leaching.

6.1.ii. Sodium. The sodium content of the 202 underground seepages averages 0.92 ppm. Highest concentrations occur in Perlis Mines' Cave/Subway Tunnel, where the mean value is 2.70 ppm. Corresponding values for Gua Batu, Gua Anak Takun and Gua Tempurong are 0.70, 0.92 and 0.53 ppm, respectively. In contrast to potassium, sodium is detectable in all the limestone samples analysed. Since the highest concentrations were recorded in the limestones of the Setul Formation, where values range from 0.01 to 0.25 per cent by weight (mean, 0.063 per cent), this may well account for the higher sodium contents of percolation waters in Perlis Mines' Cave/ Subway Tunnel. The observed mean concentrations of sodium of the Kuala Lumpur and Kinta Limestones are 0.033 and 0.021 per cent, respectively, the maximum in each case being 0.05 per cent.

The estimated rates of sodium removal are similar in all three regions, varying from 6.81 kg/ha/yr in the Kinta Valley to 12.59 kg/ha/yr in the North-west region (based on spring and cave stream data; table 95). The net loss of sodium from the karst outcrops, calculated as the difference between rainfall input and groundwater output, averages approximately 3 kg/ha/yr. Since throughflow and subcutaneous waters contain between 0.81 and 4.83 ppm sodium (cf. 0.51 and 4.72 ppm, potassium), and as this

element is absorbed by plant roots in small amounts, compared with potassium (Chapter IV), it is reasonable to suppose that a significant proportion of the sodium in underground seepage waters is derived from the soil. The amount of sodium leached from the soil is therefore probably equivalent to between one-quarter and one-third of that cycled by vegetation in the moderately-inclined hill and foothill plots (table 52).

6.1.iii. K:Na ratio. The molar K:Na ratio averages 0.470 at the 202 sites. Highest ratios occur in Gua Tempurong, with a mean value of 0.722. This is significantly higher (t-test; $p<0.001$) than in the other caves, which have mean ratios in the range 0.180 (Perlis Mines' Cave/Subway Tunnel) to 0.394 (Gua Batu). Since sodium is more variable than potassium in the seepage waters, it seems likely that the contrasts observed between the caves largely reflect the different sodium contents of the limestone bedrock (Kinta Limestone, 0.021 per cent; Setul limestone, 0.063 per cent).

The average K:Na ratio of karst groundwaters is notably lower than that of exchangeable cations in the soil which, for example, vary from 1.79 to 3.47 in the six main plots investigated. This result provides further evidence of the preferential uptake of potassium by plants and the disproportionately high rate at which sodium is leached from the soil.

6.1. Underground diffuse-flow seepages -II. Guano-affected sites

As noted previously, sites contaminated by guano are distinguished by their high potassium content. The mean potassium concentration of the 28 sites is 14.09 ppm (cf. 0.53 at uncontaminated sites), with a maximum of 69.3 ppm. Sodium concentrations, which average 2.32 ppm (range, 0.55 to 6.80 ppm), and the K:Na ratio (mean, 2.98) are also significantly higher (t-test; $p<0.001$).

6.2. Karst springs

The three karst springs are distinguished by having

particularly high sodium contents (fig. 66), the mean concentration being 4.63 ppm, and a correspondingly low K:Na ratio (mean, 0.087). The lowest sodium concentration (3.92 ppm at site 4317) lies within the range recorded for cave streams in the Boundary Range. Concentrations at sites 4306 (5.55 ppm) and 4312 (4.41 ppm) are notably higher. This may reflect the fact that a high proportion of their waters are of shallow-groundwater origin (cf. mean sodium content of subcutaneous flow, Wang Tangga, 4.25 ppm).

6.3. Underground streams

The potassium and sodium contents of the 11 cave streams sampled average 0.67 and 2.68 ppm, respectively, with an overall K:Na ratio of 0.154. The generally high sodium content is consistent with the relatively high proportion of sodium in the limestones of the Setul Formation (Section 6.1, above). Specific comparisons between cave stream and underground percolation waters may be made in the case of the stream draining Perlis Mines' Cave. Correspondence between the potassium and sodium contents of the cave streams, which average 0.81 and 2.62 ppm, respectively, and the underground seepages (0.71 and 2.70 ppm, respectively), is extremely close, indicating that very little change occurs along the underground streamway. The average K:Na ratio of both waters is, in fact, identical (0.180).

6.4. Surface streams

6.4.1. Potassium. The consistently high potassium content of the surface streams (mean, 2.39 ppm) distinguishes them from the other groups of karst waters (fig. 65). Some indication of the source of the potassium may be gained by comparing the three main groups of surface streams. The four stations on streams draining catchments with significant groundwater contributions have characteristically low potassium levels (fig. 65), the average concentration being 1.49 ppm. Of these sites, station 4334, with a catchment composed

entirely of limestone hills, has a potassium content of 0.58 ppm, which is the lowest of the surface streams. This lies close to the mean value of 0.68 ppm recorded for springs and cave streams in the Boundary Range and indicates clearly that karst groundwaters contribute only small amounts of potassium to surface streams.

The mountain streams draining non-carbonate terrain generally contain more potassium. The seven streams sampled have a mean concentration of 1.81 ppm, with a 0.84 to 2.46 ppm range. The bedrock of these drainage basins, which is chiefly granite, often contains significant amounts of potassium. The granites of the Kinta Valley, for example, contain between 4.50 and 5.52 per cent K_2O (Ingham and Bradford, 1960, table VI, p.60). However, the highest concentrations of potassium occur in those streams with significant inputs from the karst alluvial plains (fig. 65). The mean concentration of the 20 stations sampled is 2.59 ppm (range, 1.21 to 4.32 ppm). The alluvium, like the weathered mantle of the mountainous granitic catchments from which it is mostly derived, contains clay minerals, grains of felspar and flakes of mica, both biotite and muscovite (Ingham and Bradford, 1960, p.81). However, unlike the rain-forested catchments, a high proportion of the surface is unvegetated. Consequently, in the absence of significant plant uptake, a much higher proportion of the potassium released by chemical weathering is likely to be leached. This is reflected in the high potassium content (mean, 2.69 ppm) of the alluvial seepages sampled above the limestone bedrock, and probably accounts for the higher potassium content of the karst streams with significant inputs from the alluvial plains.

6.4.11. Sodium. Concentrations of sodium in the 24 karst streams average 3.80 ppm, which is significantly higher ($p<0.001$) than underground diffuse-flow seepages (mean, 0.89 ppm). However, differences between the various types of surface stream water are small. The

streams draining the Setul Boundary Range have an average sodium content of 3.35 ppm. This lies well within the 1.94 to 5.55 ppm range of the spring and cave stream waters sampled, and suggests that the sodium yield from the limestones of the Setul Formation is similar to that from the other lithologies, which occupy more than 40 per cent of three of the catchments. It should be noted, however, that the limestones of the Boundary Range contain higher proportions of sodium than the other limestones of the Peninsula. Consequently the sodium contributions from the limestone outcrops in Selangor and the Kinta Valley are likely to be less, probably of the order of 0.5 to 1.0 ppm (the range of mean values of groundwater seepages in Gua Batu, Gua Anak Takun and Gua Tempurong). The sodium in the 20 remaining surface karst waters, which averages 3.80 ppm, is therefore likely to be mostly derived from non-carbonate terrain or from the alluvium. The granites of the Kinta Valley, which contain between 2.84 and 4.75 per cent Na_2O (Ingham and Bradford, 1960, table VI, p.60), are the most likely source. As with potassium, the forested mountain streams on the granites yield less sodium than the alluvial plains. The seven streams sampled on the granite, for example, have a mean sodium concentration of 2.76 ppm, whereas the corresponding figure for 11 shallow-depth seepages within the alluvium is 3.42 ppm. Again it is suggested that recycling of sodium by vegetation in the former areas is the key variable which distinguishes these two types of site.

6.4.iii. <u>K:Na ratio</u>. The molar K:Na ratio of the 24 surface streams averages 0.412. This figure lies close to the mean values of several other groups of sites, notably: underground diffuse-flow seepages (0.494), alluvial seepages at the alluvium/bedrock interface (0.486), alluvial seepages above the limestone bedrock (0.458) and rivers draining non-carbonate lithologies (0.409). This clearly indicates

that sodium is being removed from the landscape more rapidly than potassium. To a certain extent this may reflect the low potassium content of the limestones in the Malay Peninsula. However, in view of the K:Na ratio of waters which have had little or no contact with limestone, it is suggested that more general controls are also operative. Most probably these are the preferential cycling of potassium by vegetation and the disproportionate absorption of potassium by clay minerals (Black, 1968, pp. 689-708).

6.5. Seepages in alluvium of the karst plains

See discussion in Section 6.4, above.

6.6. Deep groundwater

The borehole water contains 0.24 ppm potassium (K:Na, 0.20), which is considerably less than the average concentrations of other groups of karst waters (fig. 65). The reasons for this are uncertain. One possibility is that potassium is absorbed from groundwater by the clay minerals which often remain behind as residual products of limestone solution. For example, layers of clay, several cm in thickness, frequently line pockets in the sub-alluvial karst surface (Mohammed Ayob, 1970). Also, clay residues are a characteristic feature of phreatic cave passages, As noted previously land subsidence in the Kinta Valley is thought to be indicative of contemporary solution beneath the watertable.

The sodium content of the deep groundwater (0.72 ppm) lies well within the range of percolation seepages in Gua Tempurong (0.30 to 1.40 ppm), but is notably less than that of waters sampled within the alluvium of the karst plains and at the alluvium/rock interface (overall mean, 1.94 ppm). Possibly the absorption of sodium by clays may be a significant factor.

6.7. Discussion of results

As may be seen in figure 67, the K:Na ratio varies consid-

erably within the main groups of sites sampled. The K:Na ratio of underground diffuse-flow seepages, for example, ranges from <0.01 to 2.65. In this case, a considerable proportion of the observed variability may be explained by inter-regional differences. Thus the mean K:Na ratio of 0.722 for seepages in Gua Tempurong contrasts with that of 0.180 recorded at seepages in Perlis Mines' Cave/ Subway Tunnel. Because regional variations are marked, systematic differences between the major groupings of sites are difficult to isolate. For example, the overall mean values for underground diffuse-flow seepages, surface karst waters, seepages in alluvial mines, both at and above the sub-alluvial karst surface, and surface streams draining non-carbonate terrain all lie in the range 0.41 to 0.49. However, as is seen in the comparison between underground seepage and stream waters in Perlis Mines' Cave (Section 6.3, above), systematic differences between different types of karst water cannot be identified with certainty, even where regional effects are eliminated.

The results of the present study suggest that local and regional scale variations in this ratio, which reflect differences in soil and vegetation characteristics and bedrock mineralogy, are considerably greater in magnitude than systematic differences between different types of karst water. The K:Na ratio appears, therefore, to provide an inadequate basis for categorizing karst waters in the Malay Peninsula. This finding, combined with the difficulties posed in certain areas by culturally-induced changes in potassium and sodium (eg. sewage outfalls, fertilizer usage - Manley and Taylor, 1978), suggests that extreme care should be used when using the K:Na ratio as index of groundwater flow.

E. RATES OF CHEMICAL DENUDATION

1. Limestone outcrops

1.1. Results from West Malaysia

It is reasonable to suppose that observed mean hardness concentrations of underground seepages, karst springs and cave streams are broadly representative of long-term average conditions, for two reasons. First, total hardness fluctuations of these karst waters remain notably within narrow limits, with 109 of the 112 sites at which six or more determinations were made having c.v. values of less than 20 per cent (fig. 48). Secondly, samples were obtained under widely differing flow conditions. Consequently, these data, combined with estimated amounts of effective precipitation, provide a sound basis for calculating net chemical denudation rates in the three field areas (table 96). In making these calculations, corrections have been made for cation inputs from rainfall, and a rock density of 2.65 g/cm^3 is assumed.

Highest rates, of 81.0 mm/1000 yrs, were recorded at G. Tempurong. Here the relatively low hardness concentrations of underground, diffuse-flow seepages in Gua Tempurong are compensated by the high effective precipitation (table 96). In Selangor, denudation rates at Bt. Batu (54.6 mm/1000 yrs) are higher than at Ak. Bt. Takun (38.4 mm/1000 yrs). This difference reflects the generally lower hardness contents of seepage waters in Gua Anak Takun, which mostly have low discharges. Small rates of chemical denudation, amounting to 38.5 mm/1000 yrs, also occur in the Setul Boundary Range. Thus, whilst the morphology and hydrological properties of the outcrop Setul limestones favour high solute contents in emergent groundwaters (mean total hardness, 244.0 ppm), the low effective precipitation of the North-west region considerably reduces the scale of net denudational loss.

1.2. Comparison with other studies

The rates of chemical denudation observed in West Malaysia are less than those in certain humid tropical karst regions. Smith (1969a), for example, reports values of between 89 and 96 mm/1000 yrs in Jamaica, and Rossi (1976) gives figures in the range 109 to 135 mm/1000 yrs for Madagascar. The 38.4 to 81.0 mm range recorded in the present study does, however, overlap considerably with the values of between 63 and 99 mm/1000 yrs for several Indonesian region (Balázs, 1971). It also encompasses the mean erosion rates of 45.5 and 56.9 mm/1000 yrs, respectively, for tropical and temperate karst regions, given by Smith and Atkinson (1976, fig. 13.4, p. 384). Clearly, in view of the degree of local, inter- and intra-regional variability observed in West Malaysia and other karst areas, the specific effect of climate upon world-wide variations in net denudation rates cannot be isolated.

2. Sub-alluvial karst surface

Rates of chemical denudation of the sub-alluvial karst surface in the Kinta Valley may be estimated from the hardness of waters at the alluvium/rock interface (table 96). The calculated rate ranges from 112.9 mm/1000 yrs, to a maximum of 129.4 mm/1000 yrs. Measured levels of chemical aggressiveness have been incorporated in the latter estimate. These figures compare with values in the range 30 to 300 mm/1000 yrs for denudation beneath 'blanket sand' deposits of up to 30 m in thickness in Puerto Rico (Monroe, 1969).

The Malaysian data suggest that chemical weathering rates beneath the alluvial plains are more than double those of many of the limestone hills. However, it should be noted that since substantial amounts of secondary deposition occur within the hills, the net rate of limestone removal is considerably less than the gross movement of material within these outcrops.

F. SUMMARY OF RESULTS

1. Morphology and hydrology of the caves

1.1. Caves of the tower karst hills are mostly phreatic in form, though some passages show signs of vadose modification. Autogenic waters in these hills are confined almost entirely to diffuse-flow seepages, fed by discrete sets of tightly-closed joints. Vertical flow paths probably predominate, and discharge at the majority of these sites correlates most closely with rainfall during antecedent periods of 1-16 days, or longer, before sampling. High-level, air-filled conduits, which are closely connected with the hilltop, are also present in Gua Batu.

1.2. Autogenic vadose streams are restricted to the Setul Boundary Range.

1.3. The Setul limestones retain much of their original bedding, with the result that groundwater flow has a dominant lateral component. Consequently, underground seepages are fed by a more integrated network of fractures than those in the tower karst, and the response of discharge to rainfall is more rapid. The flow rate of the majority of seepages correlates most strongly with rainfall in antecedent periods of 1-8 days, or less.

2. Water temperature

2.1. The Malaysian data demonstrate that average water temperature and, especially, temperature variability (s.d.), provide a reliable basis for distinguishing karst water types.

2.2. Diffuse-flow seepages have the lowest mean temperatures (23.74°C), and restricted water temperature variabilities (mean s.d., 0.26°C). In general, water temperature decreases, and temperature fluctuations widen, as contact with circulatory cave air increases. The mean temperature varies directly with surface temperature, reaching a maximum of 24.63°C in Perlis Mines' Cave/Subway Tunnel.

2.3. The mean temperatures and temperature fluctuations of karst springs and underground streams are similar to those of diffuse-flow seepages in the Boundary Range. Narrowest fluctuations (s.d., 0.14°C) occur at the deep-groundwater spring.

2.4. Surface streams mostly have higher temperatures and display wider temperature fluctuations than groundwaters especially in Selangor and the Kinta Valley. In the latter areas, groundwater contributions to surface flow are small, and the sparsely-vegetated river banks provide little shade.

2.5. Deep groundwater in the Kinta Valley has the narrowest temperature fluctuations (s.d., 0.05°C).

3. Water hardness properties

3.1. Systematic variations of total hardness within the karst landscape indicate that a very high proportion of solution occurs near the surface, and that much deposition takes place within the outcrops.

3.2. Underground diffuse-flow seepages (mean total hardness, 140.7 ppm) are mostly saturated or supersaturated with respect to calcite. The ratio of non-alkaline:total hardness is generally low (mean, 5.4 per cent), though higher values (mean, 16.2 per cent) have been observed where groundwaters are in contact with shale beds. Hardness fluctuations of the seepages remain within narrow limits, the average c.v. of total hardness being 7.12 per cent. At individual sites, total hardness generally increases with discharge, since deposition per unit volume of water over a given length of flow is reduced at higher flow rates.

3.3. Diffuse-flow seepages which are contaminated by bat guano are distinguished by having a non-alkaline:total hardness ratio of more than 30 per cent, and potassium concentrations in excess of 2 ppm.

3.4. Karst springs have characteristically high hardness values (mean, 305.9 ppm). Total hardness is highest (332.0 ppm), and

hardness fluctuations smallest (c.v., 5.96 per cent), at the deep-groundwater spring.

3.5. The total hardness of surface streams varies directly with the proportion of limestone hills in the catchment. Four stations on streams draining the Setul Boundary Range have a mean value of 170.6 ppm (variability - mean c.v., 10.49 per cent). This compares with 48.6 ppm (mean c.v., 25.80 per cent) for streams in Selangor and the Kinta Valley, where a much higher proportion of the water is derived from karst alluvial plains and non-carbonate terrain. Notably high hardness levels occur in those catchments where the alluvial cover is thinner.

3.6. Waters at the alluvium/rock interface, which have a mean total hardness of 224.5 ppm, are mostly capable of further dissolution (mean aggressiveness, 32.6 ppm $CaCO_3$).

3.7. Deep groundwater in the Kinta Valley has a total hardness of 216.9 ppm, suggesting that much of the water is derived as seepage through the alluvium. Hardness fluctuates within narrow limits (c.v., 1.64 per cent).

4. Calcium, magnesium and the Mg:Ca ratio

4.1. The Mg:Ca ratio of the 278 karst waters investigated (mean, 0.223) corresponds closely with 197 limestone samples (mean, 0.220). Preferential deposition of calcium from seepage waters is seen in the low Mg:Ca ratio of 42 samples of secondary deposits (mean, 0.025), and in the fact that the Mg:Ca ratio at many seepage sites varies inversely with discharge.

4.2. Almost as much calcium per unit area is taken up by plants in the moderately-inclined foothill plots as is lost in groundwaters.

4.3. Most of the magnesium, especially in the groundwaters of magnesium-deficient limestones, is probably derived from bedrock beneath the normal rooting zone of plants.

5. Potassium, sodium and the K:Na ratio

5.1. The potassium content of karst groundwaters is generally low (eg. diffuse-flow seepages: mean, 0.52 ppm). The total loss of potassium from limestone outcrops, per unit area, amounts to less than 1/50 th of the surface inputs in the moderately-inclined hill and foothill plots. By comparison, the potassium concentrations of streams draining the karst alluvial plains are higher (mean, 2.59 ppm), this being attributable to the greater potassium content of the alluvium, and the reduced levels of nutrient cycling of the often unvegetated surfaces.

5.2. The limestones of the Peninsula, especially those of the Setul Formation, contain higher proportions of sodium than potassium, and this is reflected in the higher sodium content of diffuse-flow seepages (mean, 0.89 ppm). The amount of sodium leached from the soil is probably between 1/4 and 1/3 of that taken up by plants in the six main study plots.

5.3. The K:Na ratio of the diffuse-flow seepages (mean, 0.470) compares with values in the range 1.79 to 3.47 for exchangeable cations in the soils of the karst outcrops, demonstrating clearly that sodium is leached preferentially from these soils.

6. Rates of chemical denudation

6.1. Estimated rates of chemical denudation of the karst outcrops vary from 38.4 mm/1000 yrs at Ak. Bt. Takun, to 81.0 mm/1000 yrs at G. Tempurong.

6.2. A figure of 112.9 mm/1000 yrs is given for the sub-alluvial karst surface.

CHAPTER VIII

CONCLUSIONS

CHAPTER VIII. CONCLUSIONS

"Large global schemes of landform development will not advance our knowledge; it is detailed studies set in the context of a wider field which will"

(M.M.Sweeting, 1976, p.4)

A. INTRODUCTION

The principal results and conclusions related to specific aspects of karst waters and environment in West Malaysia have been summarized at the end of each chapter and require little further elaboration. This final chapter focuses, therefore, upon several important general themes which may be evaluated in the broader context of an overall synthesis of the results. Finally, some directions for future research are suggested.

B. MAJOR THEMES

1. Variability of karst water and environment within the Malaysian karst landscape

The chemical, physical and hydrological characteristics of karst waters and rates of chemical denudation in the field areas investigated are mostly influenced by four main groups of variables: (1) climate, (2) surface topography, and soil and vegetation characteristics, (3) properties of the groundwater aquifer, and (4) the degree to which the properties of groundwaters are affected upon emergence, either directly, by contact with the atmosphere, or indirectly, by inputs of water from allogenic sources. Each, in turn, is controlled by a complex set of environmental factors.

1.1. Climatic factors

Since all three field areas lie within narrow altitudinal

and latitudinal limits, inter-regional variations in mean annual temperature are small. Thus, whilst differences in mean groundwater temperature are detectable, they are of such small magnitude that mean annual temperature may effectively be regarded as constant. Seasonal variations in temperature are also limited, the range of mean monthly temperature being less than 2°C.

In contrast, variations in rainfall are quite marked, particularly between the North-west region and Selangor and the Kinta Valley. Rates of leaching and chemical denudation are significantly reduced in the first area by the low effective rainfall, which is less than half that in the other regions. The occurrence of a pronounced dry period in the North-west region between January and March is reflected in the semi-deciduous character of trees in the Setul Boundary Range, where peak litterfall in plots WT-I and GKV-I corresponds with the drier months. Studies in Selangor and the Kinta Valley have shown significant positive correlations between soil carbon dioxide and antecedent rainfall. It is reasonable to suppose, therefore, that temporal variations of soil carbon dioxide will be even more pronounced in the North-west region, with maximum levels occurring during, or shortly after, the wettest months (usually September to November). Some evidence of this is seen at three diffuse-flow seepages in Perlis Mines' Cave/Subway Tunnel, where peak positive correlations between total hardness and antecedent rainfall are with periods of 43-56, 57-70 and 71-84 days, respectively. Discharge and water hardness vary markedly at certain underground seepage points in response to individual rainfall events, whilst over an even shorter time-scale, fluctuations in karst water quality have been observed during an individual downpour on exposed rock outcrops.

1.2. Surface topography, soil and vegetation

The relief of the limestone outcrops comprises a mosaic

of distinct topographic units, which differ in overall gradient, meso-topography, degree and depth of soil cover, and vegetation. Six major morphological units have been identified. On the one hand, they range from precipitous rock cliffs, 200 m or more in height, to gentler slopes, which characteristically have an almost continuous soil cover, locally exceeding 1 m in depth. On the other, they range from rocky hilltops with deeply-dissected meso-relief, to large enclosed depressions, floored by alluvium or limestone residual soils. In general, cliffs and rocky hilltops are less frequent, and moderately-inclined slopes more frequent in the Setul Boundary Range than in the tower karst hills. Locally, however, the detailed distribution of these components is determined by factors such as recent or contemporary basal undercutting, relative relief, and the size of the individual outcrop. Local factors also affect the form of areas which lie within the same broad topographic grouping. In the Setul Boundary Range, for example, rock purity is an important variable. Deeper and more continuous soils, and a denser vegetation cover, occurs on limestones which contain a high proportion of acid insoluble residue (cf. WT-I and GKV-I).

Investigations of surface runoff from bare and partially-covered rock outcrops and of throughflow and subcutaneous flow beneath deeper mineral soils, have demonstrated clearly that the potential for chemical weathering is increased considerably by the presence of a soil cover. Variations in the depth of soil cover may therefore produce marked, local variations in denudation rate. Differences in soil type may also be important. Shallow, organic-rich mineral soils and decaying litter, which accumulate at the bottom of solution runnels and in rock crevices, are characteristic of rocky hilltops and steep hill slopes. Drainage is often impeded, thereby reducing the rate of organic decomposition and favouring

the production or organic acids. This is reflected in the relatively high proportion of non-alkaline to total hardness in runoff from beneath these soils. The deeper mineral soils of the hilltop depressions and moderately-inclined hill slopes are mostly very friable clays, with a well-developed crumb structure. Consequently, they are porous and well-aerated. Carbon dioxide concentrations in these soils are relatively low, averaging 0.76 per cent at 60 cm. However, deeper hill soils, of higher bulk density and lower porosity, have developed on less pure limestones of the Setul Formation. Concentrations of carbon dioxide in these soils are characteristically high, the mean value being 4.21 per cent at 60 cm. Within the Setul Boundary Range, differences in soil carbon dioxide and soil depth, which result from variations in rock purity, are reflected in the spatial pattern of the total hardness of cave stream waters. Soils of relatively low porosity and high carbon dioxide content also occur on foothill slopes in Selangor and the Kinta Valley. Small differences in topographic position, and possibly the influence of underlying shales, cause notable differences in the moisture and carbon dioxide contents of the foothill soils studied. Impeded drainage conditions in ABT-II, for example, tend to reduce levels of microbial activity, resulting in lower carbon dioxide levels and, possibly, increased concentrations of organic acids.

Variations in the vegetation cover of the hilltops and hill slopes are determined largely by the depth and extent of soil cover. On the rocky hilltops, for example, stunted trees form a low and discontinuous canopy, with the basal areas of trunks ranging from 2.2 to 26.0 m^2/ha. A taller and more luxuriant vegetation cover is characteristic of the moderately-inclined hill slopes. The densest cover (basal area, 70.2 m^2/ha), occurs on deeper soils of the impure Setul limestones. Whereas the vegetation

described above is controlled almost entirely by natural edaphic factors, that of the foothill slopes in Selangor and the Kinta Valley has been modified by the selective felling of larger trees. However, greatest human modification has occurred on the plains. Here the climax rain forest vegetation has either been replaced by rubber, padi and other types of agricultural land-use, or has been laid waste by tin-mining activity.

Vegetation may affect limestone weathering and karst water quality in four main ways. First, where a dense tree canopy is present, one-third or more of the rainfall may be intercepted before it reaches the ground surface, thereby reducing rates of soil leaching and of chemical erosion on exposed rock surfaces. Secondly, by absorbing and cycling nutrients the vegetation reduces the rate of leaching loss. This is especially the case with major plant nutrients such as potassium and, to a lesser extent, magnesium, of which there are limited reserves within the soil. Thirdly, plant root respiration and organic litter production are vital to rock weathering since carbon dioxide and organic acids in the soil are derived almost entirely from these sources. Finally, plant roots may act directly as agents of mechanical weathering and corrosion.

1.3. Properties of the groundwater aquifer

Where limestones are at an advanced stage of erosion, as in Selangor and the Kinta Valley (also in the case of the Chuping and Kodiang limestones), the catchment areas afforded by individual limestone outcrops are mostly too small to support permanent vadose streams. Also, the watertable generally lies beneath the level of the adjacent plains. Joints and faults, formed during or after metamorphism, provide the main lines of weakness in the tower karst hills. At present groundwater flow within the hills is mostly through discrete sets of tight, vertical or subvertical fractures. The latter are frequently intersected by the walls of otherwise

dry and inactive cave passages. The points of emergence of diffuse-flow seepage waters are often loci of large-scale underground deposition. Since these waters are mostly absorbed by the cave earth and guano which floors most of the cave passages, surface streams receive little groundwater contribution from the tower karst hills. In detail, notable differences may be observed in the hydrological characteristics of individual outcrops. Gua Batu, for example, is much more closely connected with the surface than is Gua Tempurong. In consequence, many seepages in the former cave respond to single storms whereas none in the latter do.

The hydrological characteristics of the Setul Boundary Range differ fundamentally from those of the tower karst hills for three main reasons. First, since the outcrop is much more extensive, it is able to maintain vadose flow, even though effective rainfall is considerably less than in the other regions. Secondly, the original sedimentary structures have been largely preserved. Bedding planes favour the lateral movement of water and the development of integrated systems of diffuse-flow seepage. Finally, the main body of the Setul limestones, which is disposed on the eastwards-dipping limb of a major anticline, is underlain by less permeable deltaic deposits of the Machinchang Formation. Conditions, therefore, favour the rapid lateral movement of water through the aquifer. As a result, opportunities for underground deposition within the Setul limestones are limited compared with those in the other limestones of the Peninsula, and adjacent streams receive a significant groundwater contribution.

1.4. Surface karst waters

The physical and chemical properties of a karst groundwater change appreciably once it emerges at the surface. These changes may occur without allogenic inputs. For instance, water temperature variability increases from s.d. 0.34 to 0.66°C, and

total hardness decreases by approximately 50 ppm, over a distance of 345 m downstream from the main rising at Abi Head water works (site 4315). More usually, however, waters from non-carbonate lithologies exert a strong influence over the properties of surface karst streams. In West Malaysia these are mainly streams from mountainous granitic terrain, or from the alluvium of the karst plains. Both groups of waters have lower hardness contents than karst groundwaters. The second group also have higher temperatures and exhibit greater temperature variability, principally because of their periodic exposure to direct insolation. Most surface karst streams in the Peninsula are a variable admixture of karst groundwaters from the limestone hills and from bedrock at or near the surface of the alluvial plain (eg. Sg. Sengat), and of allogenic waters. Inevitably, the chemical properties of these waters vary considerably from one catchment to the next. Total hardness, for example, ranges from 15.5 to 278.1 ppm. These variations are largely interpretable in terms of the proportions of the drainage basins occupied by carbonate and non-carbonate rocks.

2. Distribution of limestone solution within the karst landscape

The average total hardness contents of the major hydrological components of the karst landscape in West Malaysia are shown in figure 68. Rainfall generally contains very little calcium and magnesium (mean total hardness 2.8 ppm), though on occasions notably high values may occur (maximum, 6.8 ppm). Appreciable amounts of these elements may be leached from the vegetation canopy, the mean total hardness of throughfall being 16.8 ppm.

Where rain falls directly on to a bare rock surface, total hardness increases rapidly to between 22.3 and 66.8 ppm after flowing on average a distance of 1 m at a gradient of 50°. Higher hardness values, in the range 38.8 to 108.2 ppm, were observed where runoff is from rock exposures with a partial cover of organic-

rich soil and/or decaying organic litter. Since the shallow soils of the rocky hilltops tend to have relatively high levels of base saturation and contain some free carbonate, a considerable proportion of the total hardness observed at the latter sites may be derived from the surface cover. The total hardness of throughflow at the soil/rock interface in the foothill soils of Selangor and the Kinta Valley was observed to vary between 136.3 and 199.0 ppm. Evidence suggests that the chemical aggressiveness of these waters is satisfied, in part, by the leaching of bases from the neutral to slightly alkaline soil. Subcutaneous flow, and spring waters which derive a high proportion of their flow from the soil and uppermost few metres of bedrock have high total hardness values (range, 263.6 to 359.9 ppm). The total hardness of the one deep-groundwater spring sampled, 332.0 ppm, lies well within the range of the last grouping of sites. If it is assumed that the solute content of this spring is representative of the solutional potential of groundwaters in the Boundary Range, then clearly a very high proportion of the total limestone dissolution occurs at the surface and in the near-surface zone.

Whilst the total hardness of the deep-groundwater spring lies within the observed range of diffuse-flow seepages in the Setul Boundary Range (136.5 to 362.6 ppm), it is approximately 100 ppm higher than the mean value of the latter group of sites (230.2 ppm). This clearly indicates that a considerable proportion of the limestone which is initially taken into solution is redeposited within the aquifer. Exact comparisons cannot be made between the spring waters of the Boundary Range and diffuse-flow seepages in the tower karst hills. However, since the maximum total hardness of 178 such sites is 230.9 ppm, and the mean total hardness of seepages in the three caves investigated is 129.5 ppm, the process of underground deposition appears to be even more active here. This

finding is consistent with the sharply-contrasted hydrological characteristics of the Kuala Lumpur/Kinta limestones as compared with the limestones of the Setul Formation. Two further points should be noted, however, with respect to chemical denudation within the limestone hills. First, some groundwaters will inevitably be confined within water-filled fractures during their passage through the outcrop, and hence will retain their original capacity to hold limestone in solution. Secondly, seepage waters in caves may gain renewed chemical aggressiveness as they percolate through the guano which floors many of the high-level cave passages (eg. Gua Anak Takun).

Deposition also occurs where groundwaters which are saturated with respect to calcite emerge at the surface. This is seen, for example, where subcutaneous waters emerge at the semi-natural seepage in Wang Tangga (plate 23), and in the external stalactite forms which adorn many of the tower karst cliffs (plate 24). However, where emergent groundwater becomes diluted by allogenic waters, a much higher proportion of the dissolved limestone will be removed in solution. This occurs, for instance, in Wang Tangga and along the eastern front of the Setul Boundary Range, where effluent karst waters become admixed with the surface waters of the Sg. Pelarit. The latter rises on the granites of Bt. China and receives east-bank tributaries, such as the Sg. Tasoh, which have catchments of predominantly non-carbonate, sedimentary lithologies. Allogenic waters in Selangor and the Kinta Valley fulfil a similar function.

Waters at the alluvium/rock interface have an average total hardness of 224.5 ppm (mean aggressiveness, 32.6 ppm). Though the solutional capacity of these waters is slightly less than that of groundwater recharge in the Boundary Range, it is nevertheless quite high. In contrast to the groundwaters sampled in the limestone

hills, those beneath the alluvium have little, if any, contact with circulatory air after entering the aquifer. Under these circumstances, amounts of deposition are likely to be small compared with the hills, and the difference between net and gross solution rates is likely to be smaller. Equally, deposition is likely to be limited in the deep-groundwater aquifer beneath the karst alluvial plains. Consequently, the 216.9 ppm recorded at the deep-groundwater site is likely to represent a net solutional loss.

3. Implications for karst landform development in West Malaysia

Sharply-etched and deeply-runnelled bedrock outcrops, and dissected and highly irregular meso-topographies are characteristic features of the limestone hills in all three regions. The first phenomenon is largely attributable to the rapid rate at which rainwater dissolves limestone upon contact with bedrock at the high temperatures which obtain in the Peninsula. However, other solutional agencies, such as snails, may also be important. The second is usually caused by solution being concentrated along joint planes. The effects of the latter process are self-reinforcing since particulate residues are readily washed from projecting rock outcrops during heavy downpours. More rapid solution in the hollows is thereby favoured by the deeper soils, and correspondingly denser vegetation cover. As is seen in GT-IV and ABT-III/IV, oversteepening and the mechanical pressures of plant roots may cause the detachment of blocks of limestone from rock pinnacles.

When broad intra- and inter-regional variations in topography are considered, geological factors, especially rock structure, disposition and tectonic history, are key variables. The most striking contrast in the Peninsula is between the tower karst hills and the Setul Boundary Range.

3.1. Tower karst hills and adjacent plains

The limestones which form tower karst hills in West Malay-

sia are distinguished by their lack of pronounced bedding planes. Consequently, the dominant direction of groundwater movement within them is downwards towards the watertable. Under these circumstances flow will become concentrated along the major vertical and sub-vertical joints and faults, resulting in the progressive dissection of the limestone into separate blocks. The importance of the main tectonic lineaments is seen in the major axes of the tower karst outcrops in the Kinta Valley which show marked north-south and northwest-southeast orientations (fig. 69). The straight cliff faces and narrow, steep-walled corridors (plates 24, 25 and 28) are common features of many of the hill masses, and provide further evidence of joint and fault control.

The widening of existing corridors is mostly through stream, swamp and soil water action, and periodic rockfall. Using Bt. Batu as an example, Jennings (1976) has demonstrated the importance of cliff-foot caves and notches in tower karst development. At locations where there is no evidence of undercutting at the base of cliffs, notches are often present beneath the level of the alluvial plain at the junction between the sub-alluvial karst surface and the hills (plate 21). Since such tower walls clearly pre-date the Old Alluvium, this implies that slope retreat is extremely slow. This may be explained by the high mechanical strength of the limestones and by the fact that redeposition, in the form of external stalactites, is the main process operative on precipitous cliff faces which receive little direct rainfall. However, the presence of scattered rock debris on the surface of the alluvium indicates that rockfall is still active today. Several falls of rock have been recorded since the beginning of the century, notably in 1927, 1940 and October 1973 at G. Cheroh, near Ipoh, Kinta Valley. In the last instance a triangular slab of rock measuring 43 m in both height and basal length, averaging 6 m in thickness, and weighing

approximately 24,000 tonnes (plate 29; Shu and Lai, 1973) fell on to a longhouse killing about 50 people (Straits Times, 2nd Nov., 1973). Recent falls have occurred at G. Kroh and G. Tasek, also in the Kinta Valley (T. M. Lee, personal communication).

Soil-mantled slopes of between 30 and 60° often extend virtually from the hilltops to the surrounding plains. They are almost invariably covered in angular rock debris derived from higher cliffs. Such slopes, which are thought to conceal bedrock ramps (Jennings, 1976), are probably attributable to progressive cliff retreat under conditions of limited basal removal.

In Selangor and the Kinta Valley there is a notable change in soil type at approximately 130 m. Below this height the soils become more alkaline in reaction, contain higher proportions of quartz gravel, and are less well-aerated. The presence of limestone blocks and rock fragments within these soils may explain their higher pH. However, the source of the quartz is less certain. It may originate from quartz veins within the limestone. Another possibility is that it is derived from the adjacent granites, for whilst the Old Alluvium has been widely reported at a height of 70 m, there is no apparent reason why vestiges may not remain at higher altitudes in the immediate vicinity of the Main Range. Wycherley (1975) has noted that the floors of many of the larger caves in the Peninsula occur at heights of 92 to 122 m or more. These may correlate with the change in soil properties noted above. The generally higher carbon dioxide contents of the foothill soils combined with the basal removal of limestone by allogenic waters around the towers, favours relatively high net rates of denudation around the hill margins. By comparison, the rates of surface lowering beneath the characteristically shallower and better-aerated soils of the hilltops are likely to be lower. Also, since a significant proportion of the limestone eroded from the summit

areas is subsequently redeposited within the outcrop, the central body of rock which forms the tower karst hills will become progressively more massive as the edges retreat.

Since the relief of the central part of the Kinta Valley is generally low and the karst alluvial plain lies close to sea level, the hydraulic gradient within the karst aquifer is likely to be small. Consequently, groundwater flow will be relatively slow, and it may be anticipated that cave development within the phreatic zone will be limited at the present time. Nevertheless, two field observations suggest that large and interconnected passages occur beneath the watertable. First, the collapse phenomena, which have been widely reported in the Kinta Valley (Chapter VII, Section D3.5), indicate the presence of large, water-filled cavities within the aquifer. Secondly, since the Kinta Limestones have a low primary porosity, the fact that large volumes of water may be extracted continuously from the aquifer suggests that interconnected cave passages are present. Clearly, if sizeable passages have formed below the present watertable, it is reasonable to suppose that larger and more extensive phreatic systems would have existed at an earlier stage in the erosion of the Kinta Limestone, when the watertable would have been higher and the hydraulic gradient steeper.

The major caverns in Gua Tempurong (fig. 45), like those in most caves of the limestone towers, are primarily of phreatic origin. It seems likely that major phreatic passages would also have occurred along the lines of many of the wider corridors which separate the present towers. The mechanism of corridor development that is envisaged may be explained most easily by reference to Gua Tempurong. Whilst the roof of this cave is presently sound, there is little doubt that it will eventually be breached, thereby dissecting the G. Gajah-Tempurong massif. The chimneys in Gua Batu probably represent the earliest phase in this evolution. Two inde-

pendent erosional mechanisms favour breaching: first, sub-aerial denudation at the top of the hill, and secondly, rockfall within the cave systems. The latter may be caused either by the solutional widening of joints in the bedrock roof, or by instability induced by the undercutting of cave walls by stream action. The boulder-strewn slopes which form the southern side of Alam Cavern, Gua Tempurong (fig. 45), and the angular rock debris which covers the floor of the Central Hall, Gua Batu, indicates the effectiveness of underground rockfall. Clearly, once an open corridor forms, subsequent widening will take place by the range of sub-aerial processes discussed previously. The gross outlines of the present tower karst topography are thought, therefore, to have been determined largely by phreatic caves which developed early in the denudation of what must formerly have been an extensive karst plateau area. This model of tower karst development in the Peninsula differs from that put forward by Paton (1964) in that it places greater emphasis upon the part played by underground solution.

3.2. Setul Boundary Range

The strikingly-contrasted landscapes of the Setul Boundary Range and the tower karst hills have been attributed to differences in rock purity. Paton (1964), for example, suggests that the residue which results from the dissolution of the less pure limestones of the Setul Formation, as in the case of silicate rocks, forms a mantle of weathered insoluble rock and soil which favours a relatively even distribution of weathering, thereby giving rise to a generally smooth and rounded topography. This interpretation is certainly applicable to those parts of the Boundary Range, such as the western flank of the Gelang Kerbau Valley (eg. GKV-I), where the limestones contain a high percentage of impurity and the soil cover is almost complete. The presence of high proportions of impurity will also tend to disrupt the crystalline structure of

the limestone, thus reducing the mechanical strength of the rock and thereby precluding the formation of precipitous cliffs and large cave passages. However, in many parts of the Boundary Range (eg. Wang Tangga) the limestones are relatively pure, and the depth, extent and type of soil cover is similar to that found at comparable gradients in the tower karst hills. The only marked differences lie in the higher frequency of moderately-inclined slopes, and lower frequency of cliffs, in the former area. Clearly, therefore, differences in rock purity cannot account entirely for the variability observed.

The results of the present investigation have shown that lateral flow along bedding planes forms a major component of groundwater movement within the Setul limestones. This has two important implications for landform development. First, chemical erosion which is chiefly confined to the uppermost several metres of bedrock, is likely to have a more even spatial distribution than in the other limestones studied, where solution is concentrated along the lines of the main vertical joints. The amplitude of relief is therefore likely to be less pronounced. Secondly, the whole outcrop will tend to evolve as an integrated hydrological and geomorphological unit in which the evolution of the surface topography is closely linked with contemporary cave development and underground drainage. Dissection of the landscape into independent residual blocks, a diminution of vadose flow and lowering of the watertable, will, therefore, occur later in the evolution of the Setul Boundary Range than it did in the tower karst hills. Active vadose flow, which is an important medium for the removal of dissolved limestone, will consequently persist in the former area until a very advanced stage of erosion has been reached.

It is suggested therefore that the sharply-contrasted topographies of the Setul Boundary Range and the adjacent towers of

the Chuping and Kodiang limestones (fig. 2) are not simply attributable to differences in rock purity and stage of erosion. The presence of near-horizontal bedding planes in the Setul limestones is held to be the key factor which precludes tower karst development. Indeed, high, cliffed walls usually only occur where more massive beds of the Setul Formation are exposed and where basal undercutting has been recently active. Such circumstances obtain, for example, at either end of the Subway Tunnel. Finally, it should be noted that offshore, in the archipelago of Pulau Langkawi, marine erosion maintains precipitous cliffs in both the Chuping and Setul limestones.

4. West Malaysian karst in a world-wide context

4.1. Chemical denudation in karst terrain

Net chemical denudation rates are controlled principally by two factors: the mean total hardness of effluent waters from karst outcrops and the mean annual runoff. The first has often been interpreted purely in terms of ambient temperatures and the carbon dioxide content of the soils through which groundwater recharge waters percolate before entering the aquifer (eg. Smith and Atkinson, 1976, pp. 374-381). Indeed, the morpho-climatic debate (Chapter I, Section D4) still centres very much upon the relative importance, under field conditions, of the inverse relationship between temperature and carbon dioxide solubility, and the frequently-assumed positive relationship between temperature and the partial pressure of carbon dioxide in soil. The latter relationship has an apparently sound theoretical basis. It is supported by data from certain spring waters in N. America and Mexico (Harmon et al., 1975), and also by findings from non-carbonate lithologies which show that bicarbonate ion concentrations in soil waters are higher in areas of tropical rain forest than in areas of temperate and sub-alpine conifer woodland (Johnson et al., 1977). However, the relationship between

temperature and the carbon dioxide content of soil air is by no means clear-cut, since factors other than levels of organic matter production, notably soil depth and volume of effective pore space, often exert an overwhelming influence (Miotke, 1974). The Malaysian data demonstrate this most clearly. Carbon dioxide concentrations increase from an average of 0.65 per cent at 15 cm to 1.83 per cent at 60 cm, and at the latter depth the observed range is from 0.73 to 4.21 per cent. As discussed previously (Chapter VI, Section C5) there is a considerable overlap between these figures and those reported from karst soils elsewhere in the humid tropics, and in temperate regions.

In addition to variations in the chemical aggressiveness of soil waters, differences in rates of underground deposition may considerably affect the hardness of emergent karst groundwaters. The results from West Malaysia indicate the significance of this process in humid tropical regions (Section 2, above).

The Malaysian data also show that, even over a latitudinal range as small as 3°, variations in mean annual runoff, the second major determinant of chemical denudation, are sufficiently wide to cause marked differences in the rate of erosion. The contrast between the Kinta Valley and the Setul Boundary Range, which are separated by less than 250 km, is most striking. Mean annual runoff in the former area is almost four times that of the latter. This more than compensates for the lower total hardness of diffuse-flow seepage waters in the tower karst hills of the Kinta Valley, resulting in a net chemical denudation rate of 81.0 mm/1000 yrs, which is more than double that of the Boundary Range. Since effective precipitation is more variable than the hardness of karst groundwaters at a global scale, and as organic matter production, microbial activity and soil porosity are directly dependent, in part, upon soil moisture levels, the effects of latitudinal variations in temp-

erature upon chemical denudation rates are largely overridden (Chapter VII, Section E).

4.2. Landform development

The strikingly-contrasted form of the Setul Boundary Range compared with that of the adjacent tower karst hills of the Chuping and Kodiang limestones, and the topographic variations observed within the former outcrop, establish beyond doubt that rock structure and lithology exert a significant control over landform development in tropical karst terrain (Section 2, above). Since the correlation of tower karst with humid tropical environments is central to the morpho-climatic debate (Chapter I, Section D4.2), it is important to consider whether climatic factors play a significant part in the development of these forms.

Clearly, as the karst landscape of West Malaysia has been subject to sub-aerial denudation since the late Triassic period, it is the distribution of denudation within the landscape, and not the overall rate of surface lowering, which provides the key to understanding tower karst formation. Two aspects of climate distinguish the equatorial zone from middle- and high-latitude regions, and both may affect landform development. First, the intensity of rainfall is characteristically higher. As a result, soil particles will be washed more readily into depressions, thereby favouring further accentuation of the relief by differential chemical weathering of the slopes and floors of depressions. Also, there will be a greater tendency for surface, and particularly underground, flow to become increasingly concentrated along certain larger channels and cave passages, since these will allow the most efficient and rapid transmission of high-discharge, flood pulses.

The second distinguishing feature of climate in the equatorial zone is the uniformly high temperature. This has two important consequences for karstification. First, as seen in West Malay-

sia, both solution and redeposition of limestone occur rapidly within tropical karst terrain. Surface lowering is thus accompanied by a progressive infilling of presently inactive cave passages, thereby rendering existing outcrops more massive. Secondly, once precipitous cliff walls have formed, their preservation is favoured by surface deposition, even where basal undercutting is no longer active. By comparison, at the lower temperatures characteristic of middle- and higher-latitudes, deposition, both underground and on exposed rock faces, is likely to be less effective, and cliffs are likely to retreat more rapidly due to frost action. In fact, the absence of frost action is probably a key factor in the overall development of tropical karst terrain. Thus, whilst the jagged, pinnacle-like forms characteristic of tower karst hilltops in West Malaysia (above) and Sarawak (Sweeting, 1979), and the fragile outcrops noted by Jennings and Sweeting (1963) in the Fitzroy area, north-west Australia, are undoubtedly of solutional origin, their preservation and degree of development may be partly explained by the frost-free conditions which obtain in these regions.

In conclusion, it appears that massive limestones with dominant vertical and subvertical fractures are normally prerequisite to tower karst development. Whilst the exact role of climate is somewhat less certain, karstification processes under humid tropical conditions, such as those which obtain in West Malaysia, seem particularly favourable for the formation and preservation of this type of topography. On the other hand, denudation processes in temperate and cold environments are held to be less favourable, and may act counter to tower karst development. It is not surprising, therefore, that tower karst hills are more frequent in the humid tropics than elsewhere. Where, however, the structural characteristics of a particular limestone are especially favourable for vertical dissection (eg. Nahanni region, Northwest Territories, Canada - Brook and Ford, 1978), or where basal undercutting of slopes is

exceptionally active, then topographies resembling the tower karst forms of the humid tropics may develop in other climatic zones.

C. DIRECTIONS FOR FUTURE RESEARCH

The results of the present investigation suggest a number of possible directions for future research:

1. Marked variations have been observed in the soil type and vegetation cover of the five main morphological units studied: rocky hilltops, hilltop depressions, steep hillslopes, moderately-inclined hill slopes, and moderately-inclined foothill slopes. Studies of rainfall interception, throughfall chemistry, litterfall, soil moisture and soil carbon dioxide, have, however, been confined to the last two. In order to obtain a more complete picture of nutrient cycling and soil aeration in the karst regions, the present study could profitably be extended to include plots typical of the three remaining topographic groupings.

2. Observations of runoff from bare and partially-covered rock surfaces on Ak. Bt. Takun have demonstrated the value of solute studies in estimating rates of chemical denudation and in assessing the significance of a range of variables. In future studies a more exact measure of surface lowering, and an indication of rates of evaporational and evapotranspirational loss, may be obtained by collecting total volumes of runoff. Also, replicate measurements undertaken elsewhere in catchments of similar size and on rocks of similar chemical composition and physical characteristics as those in the present study, would enable the climatic factor in the surface weathering of limestones to be isolated with some confidence.

3. Now that an overall impression has been gained of the systematic, spatial and temporal variations in karst water quality

within the Malay Peninsula, more detailed investigations may be undertaken at the more interesting or unusual sites, and at those sites which are known to be typical of a particular type of karst water. For example, continuous monitoring of discharge and solute contents of cave streams and springs, and of diffuse-flow seepages which display a wide variability of flow, will provide greater insight into the hydrological and denudational processes operative in the landscape. Also, from the firm basis of directly-measured parameters, such as those determined in the present study, derived measures of karst water quality could be assessed. These include the saturation index with respect to the dominant aquifer mineral, and the theoretical partial pressure of carbon dioxide of a gas phase that would be in equilibrium with a water sample (see Jacobson and Langmuir, 1972).

4. A survey of mechanical erosion in the three field areas would complement the present investigation, thereby providing a complete picture of denudation processes in a tropical karst environment.

5. Finally, since the character of natural environmental processes in the karst regions of the Malay Peninsula has now been specified, it would also be profitable to consider specifically the cultural modifications introduced in these areas.

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KARST WATER STUDIES AND ENVIRONMENT IN WEST MALAYSIA

VOLUME II

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FIGURES

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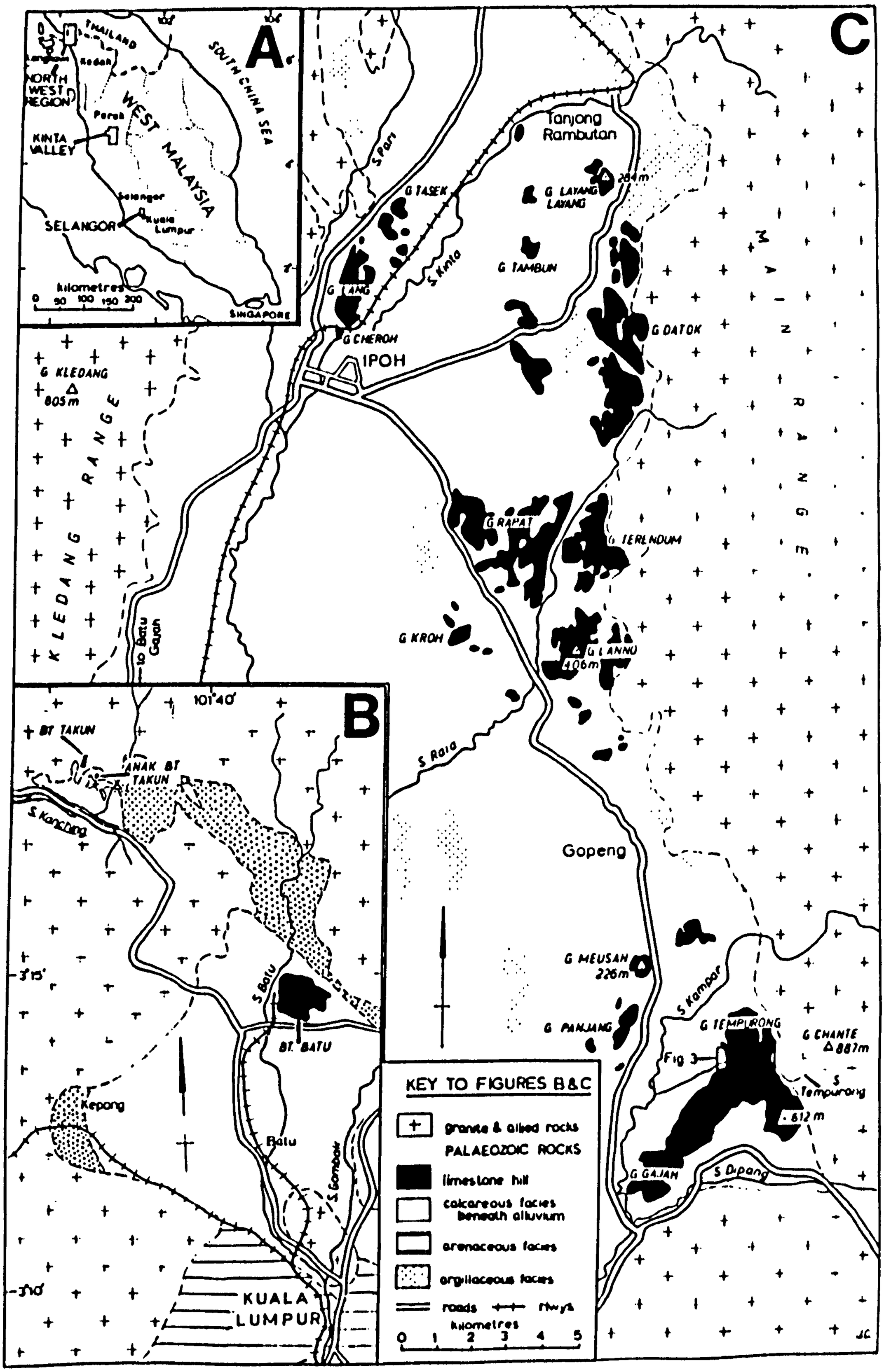


Figure 1 A. Location of the three field areas.
B. Selangor karst region (after Gobbett, 1964 and Roe, 1953).
C. Kinta Valley (after Ingham and Bradford, 1960).

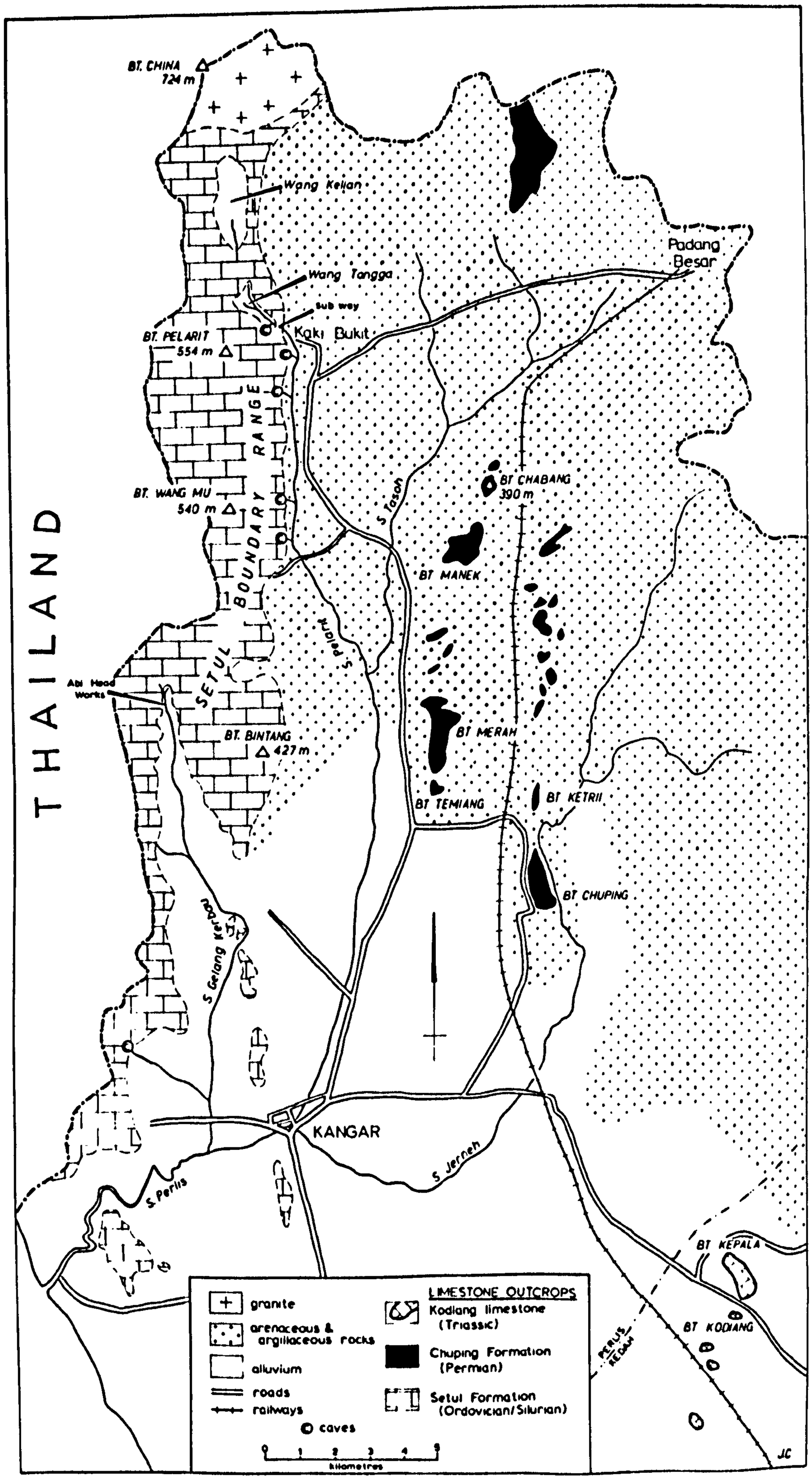


Figure 2 North-west region.

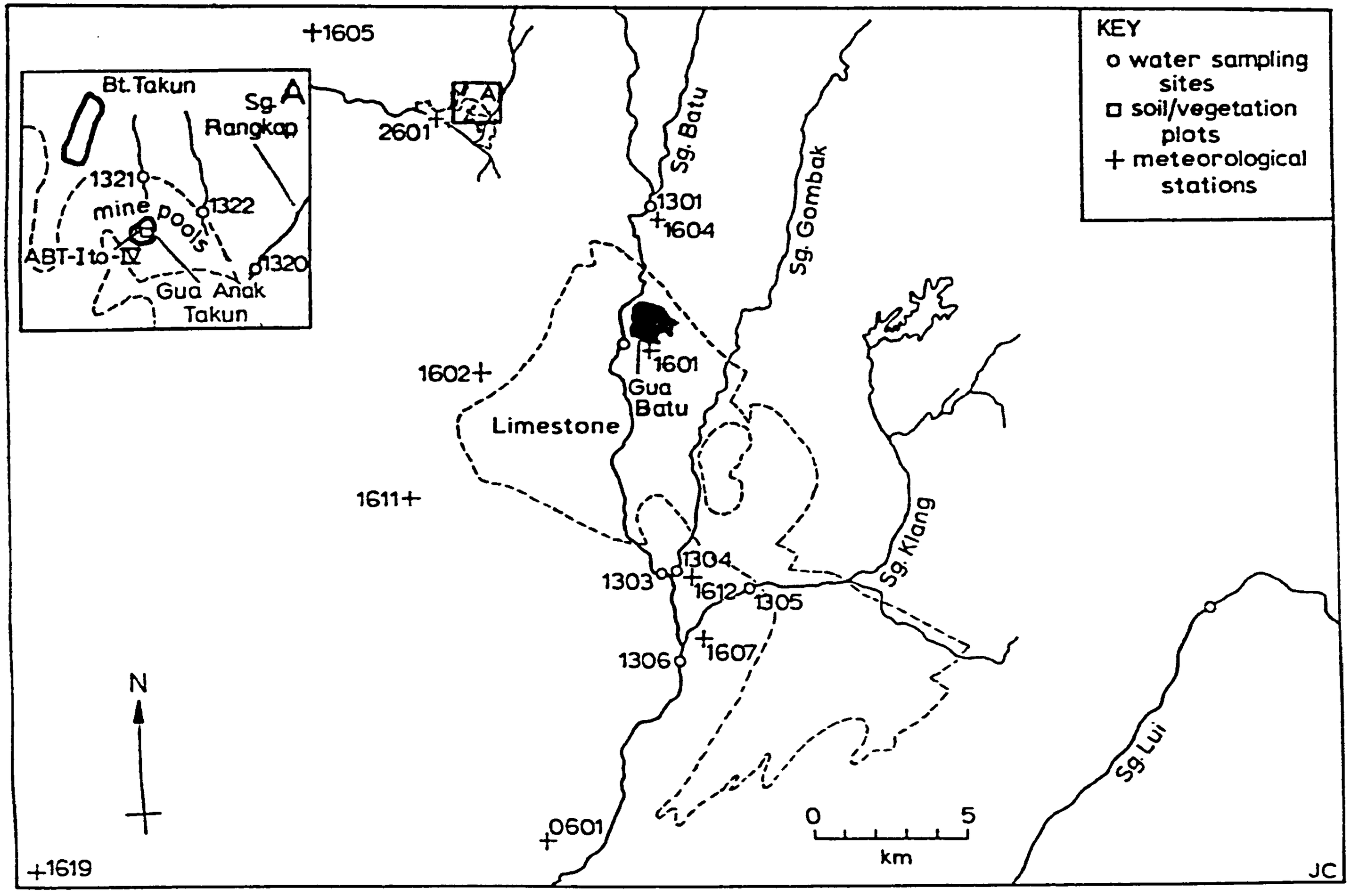


Figure 3 Key sites in the Selangor region.

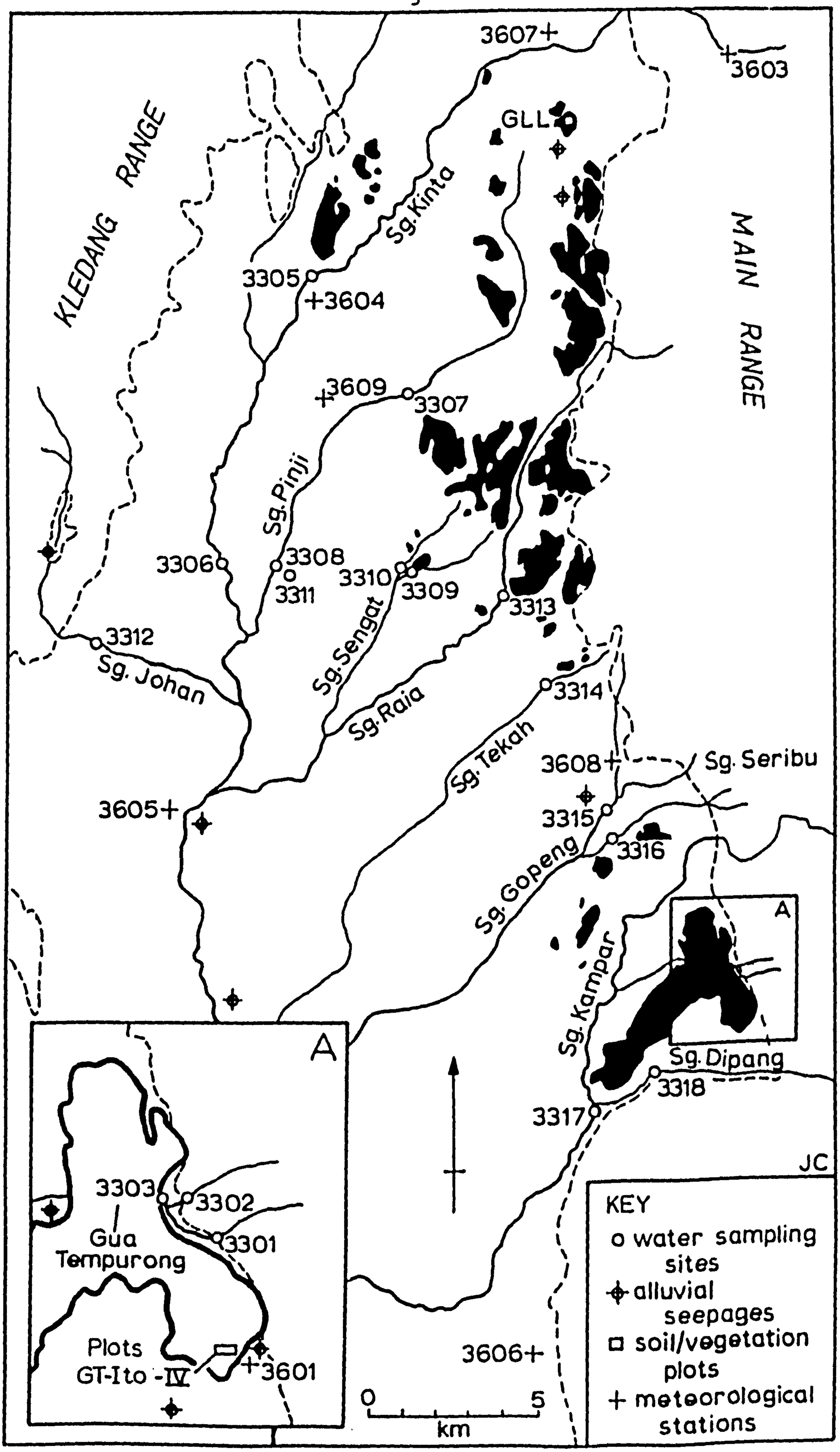


Figure 4 Key sites in the Kinta Valley.

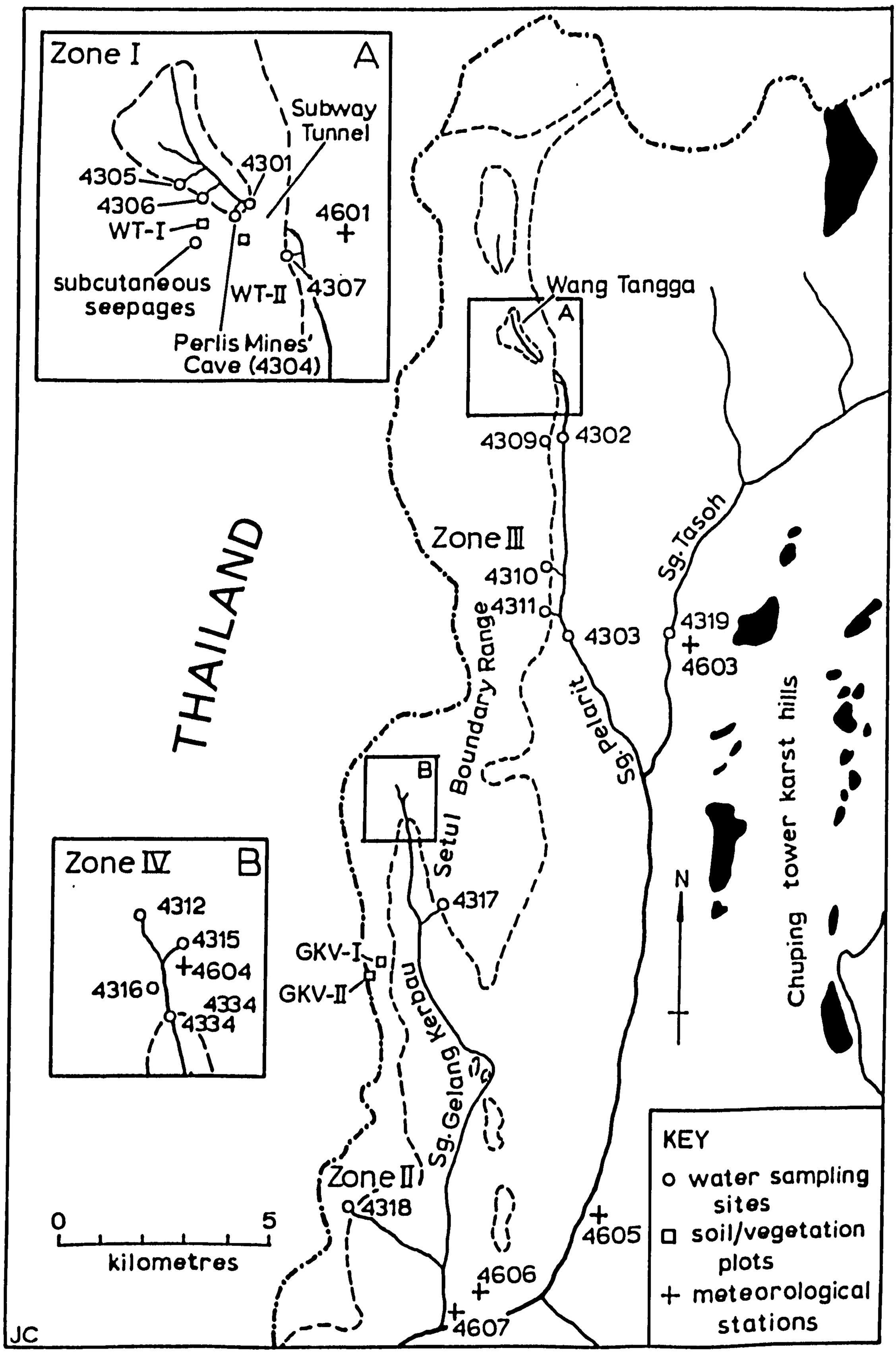


Figure 5 Key sites in the North-west region.

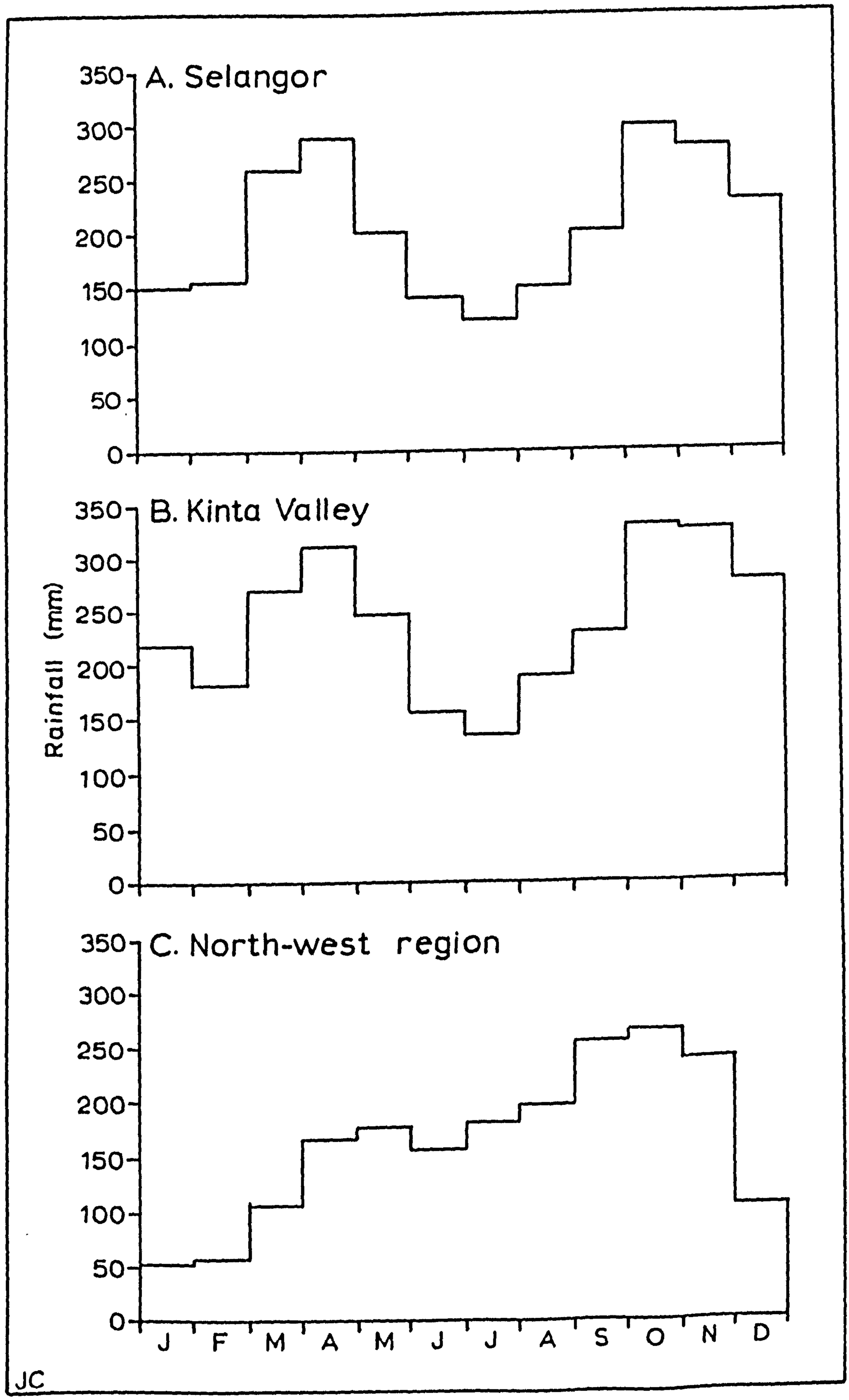


Figure 6 Monthly distribution of rainfall in the three field areas.

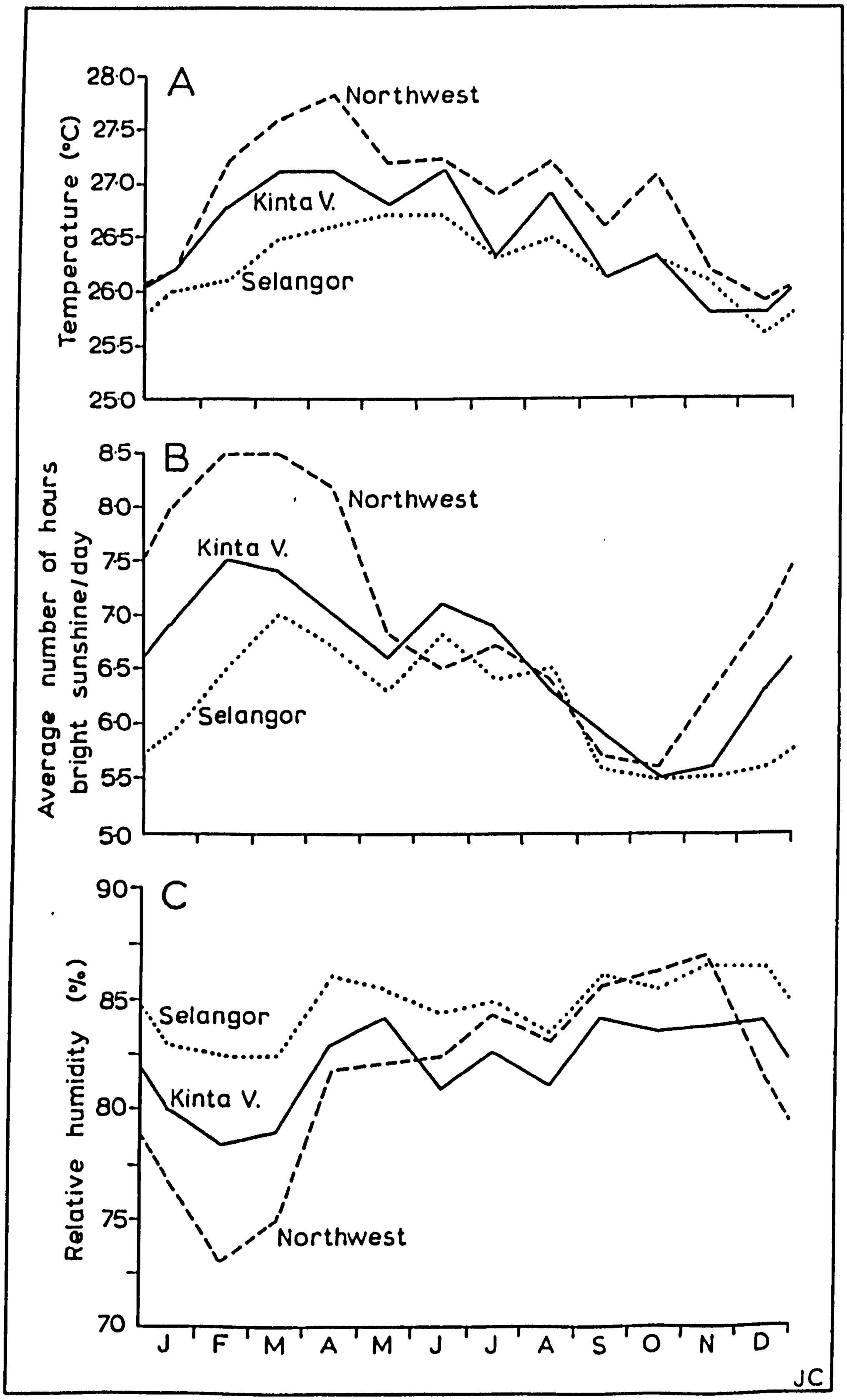


Figure 7 Monthly variations in climatic variables: A. Mean temperature; B. Sunshine hours; C. Relative humidity.

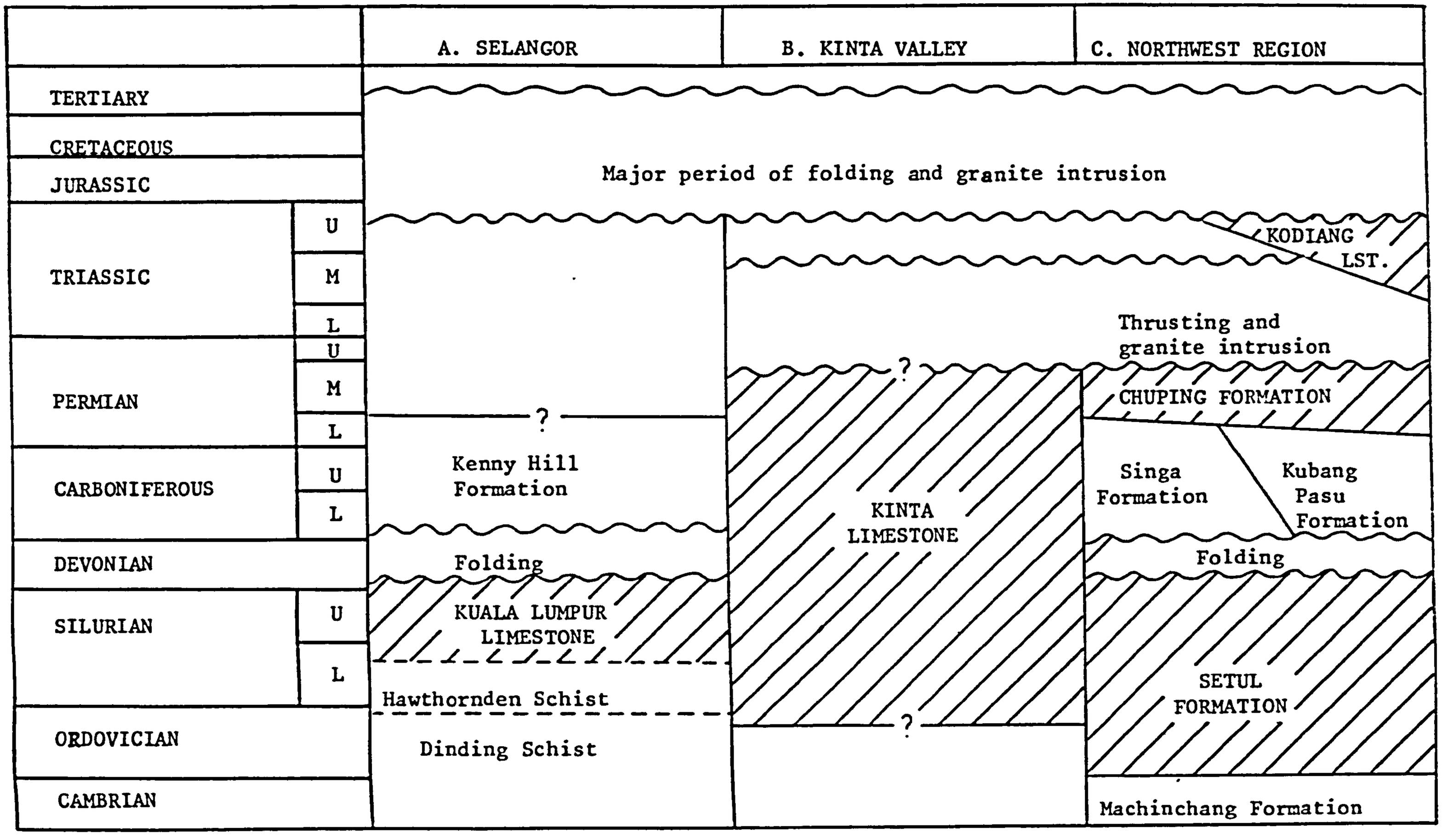


Figure 8 Time correlations of the main rock formations in the three field areas.

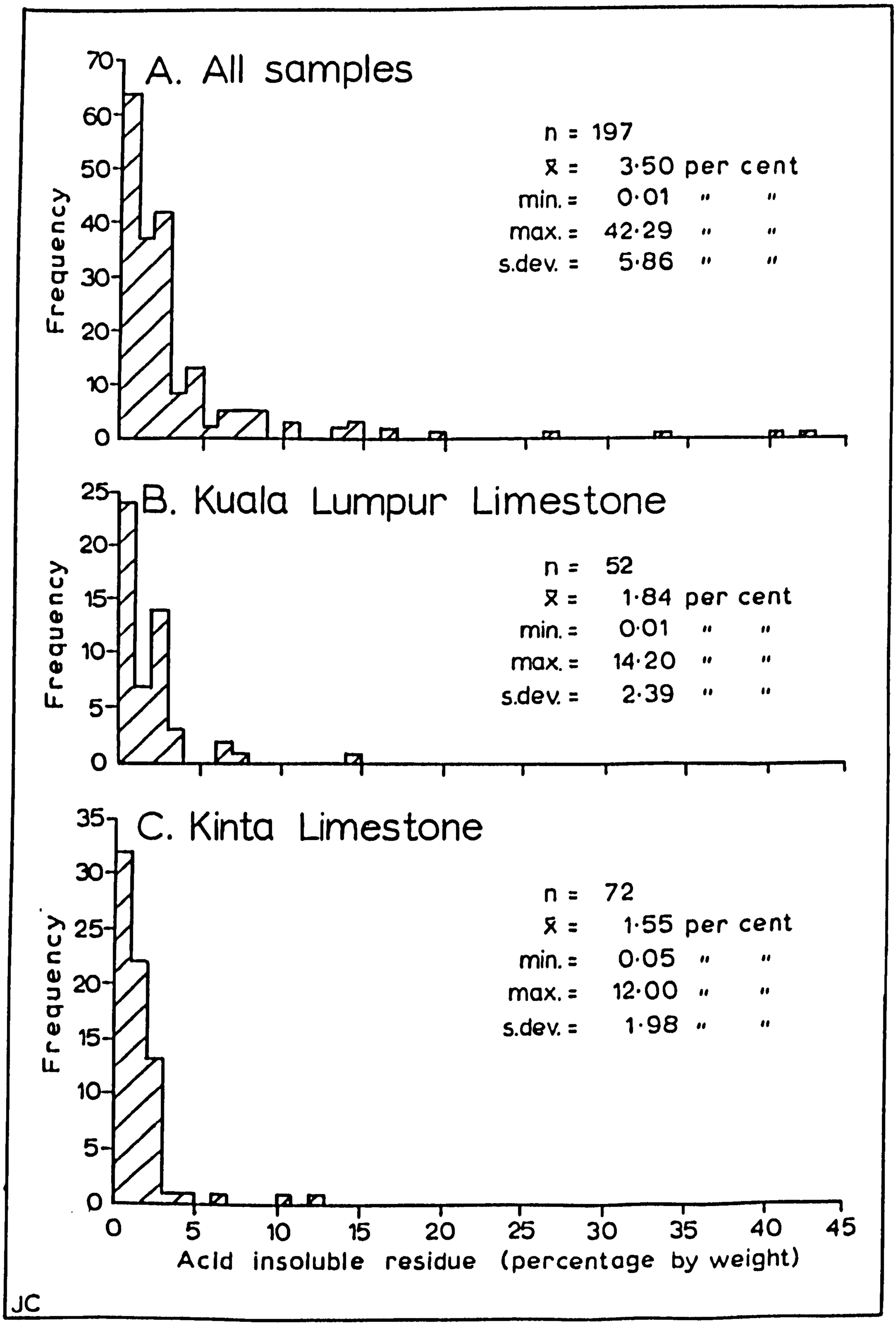


Figure 9 Acid insoluble residue content of limestones in the three field areas (figure continued).

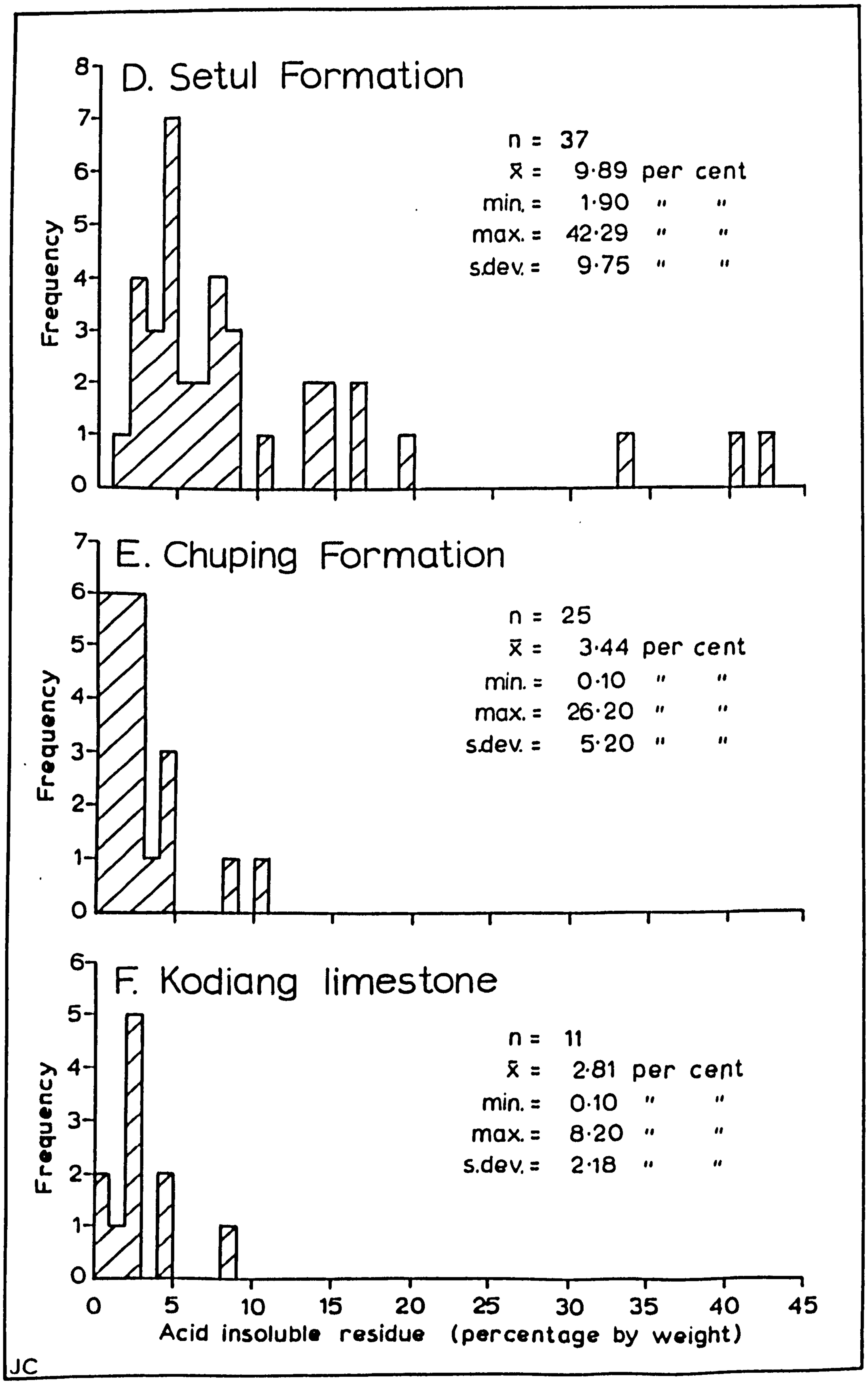


Figure 9 (ctd.) Acid insoluble residue content of limestones in the three field areas.

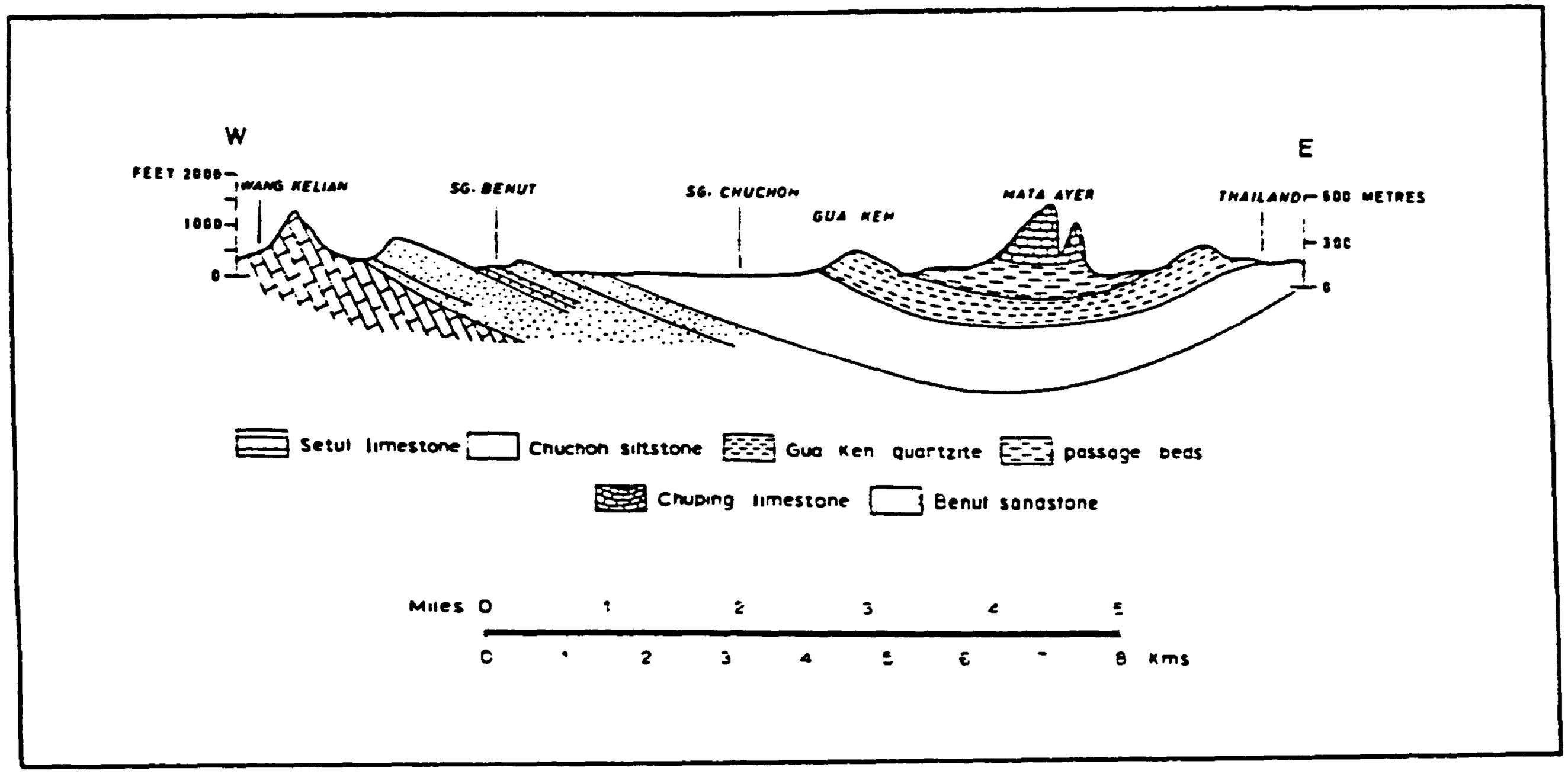


Figure 10 Disposition of the major rock formations in the North-west region (from Gobbett, 1973b, figure 4.6, p. 67).

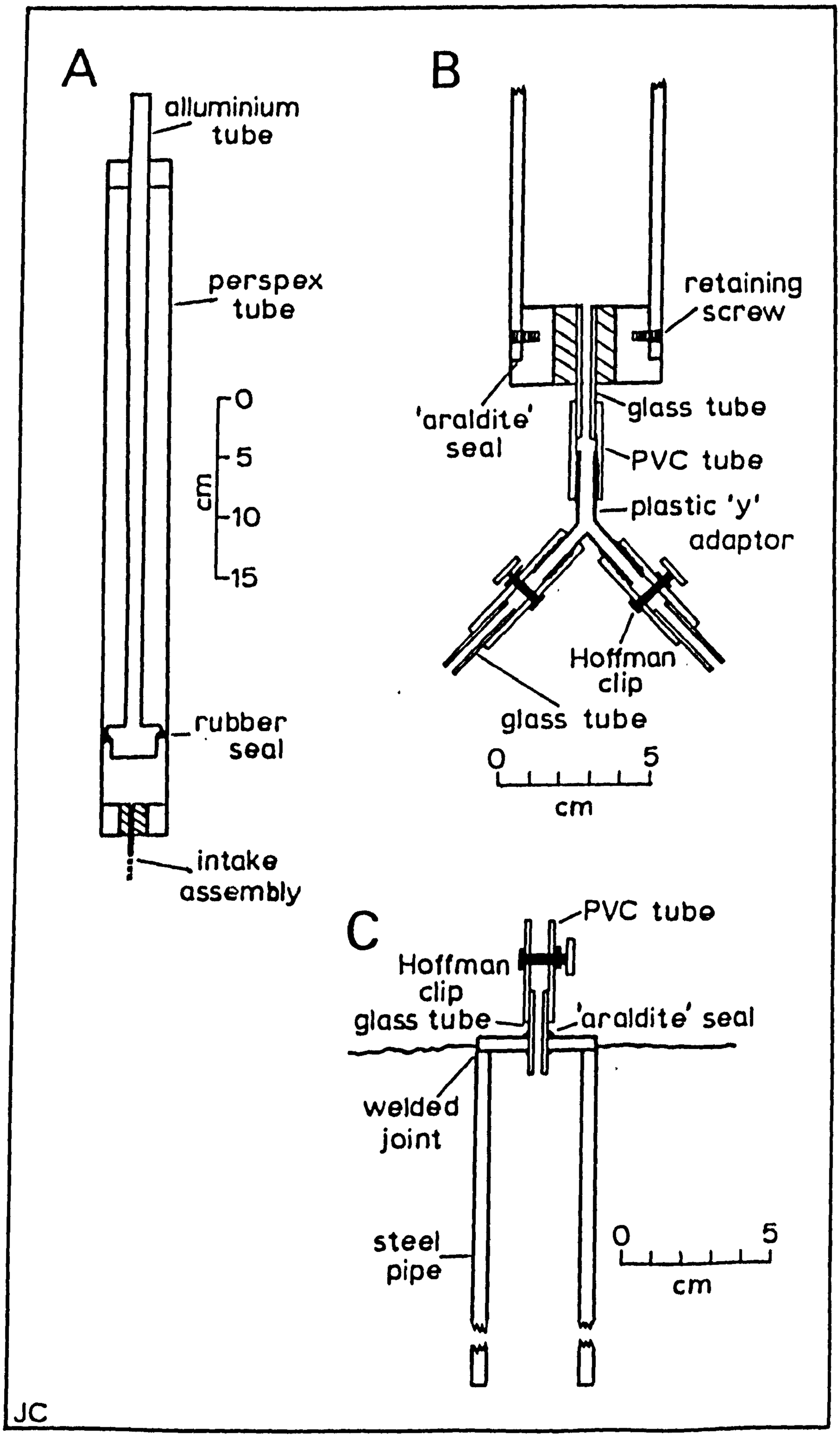


Figure 11 Equipment used in sampling soil air:
A. Syringe,
B. Intake assembly,
C. Soil chamber.

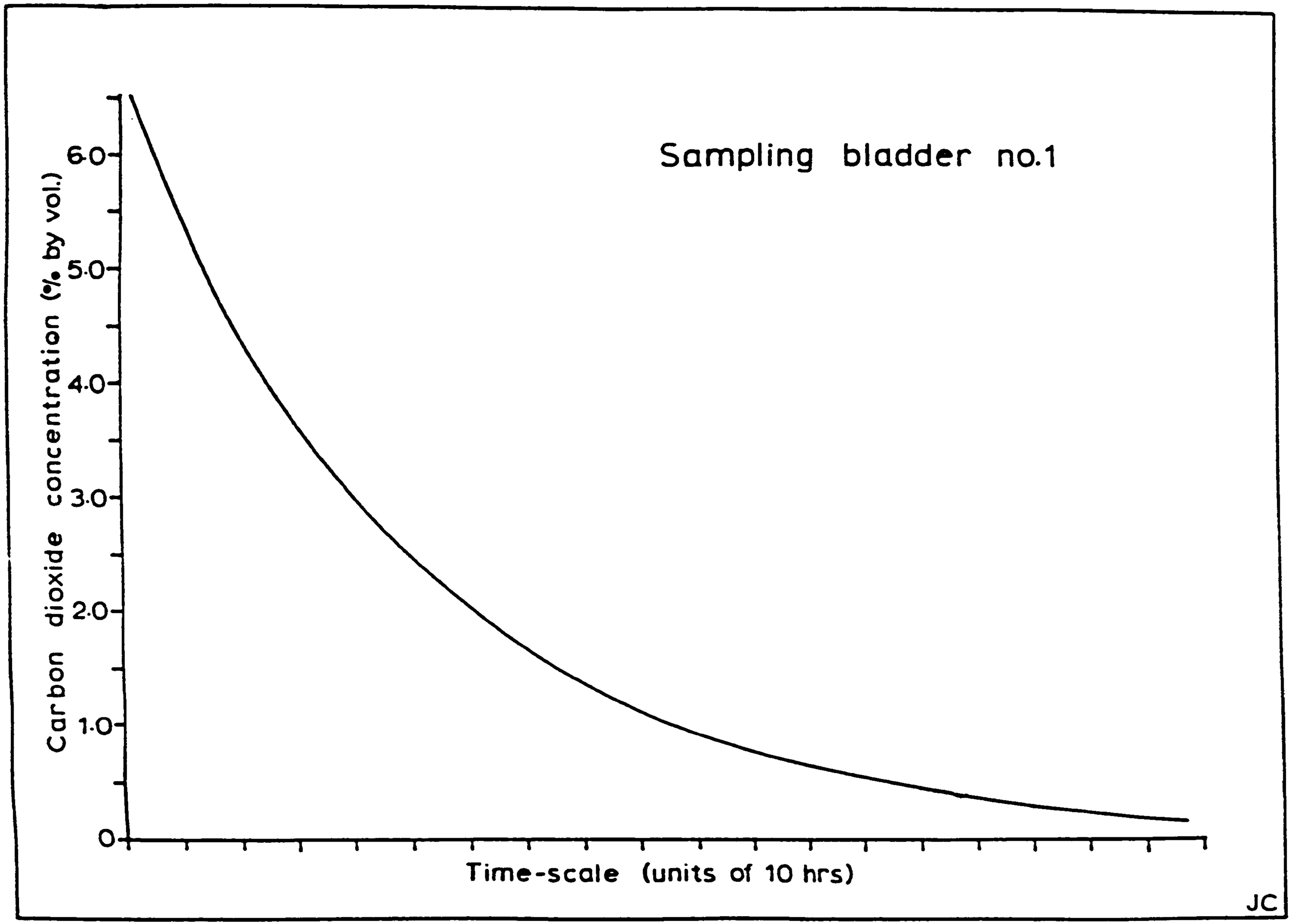


Figure 12 Curve showing rate of carbon dioxide loss from a sampling bladder under conditions typical of sample storage.

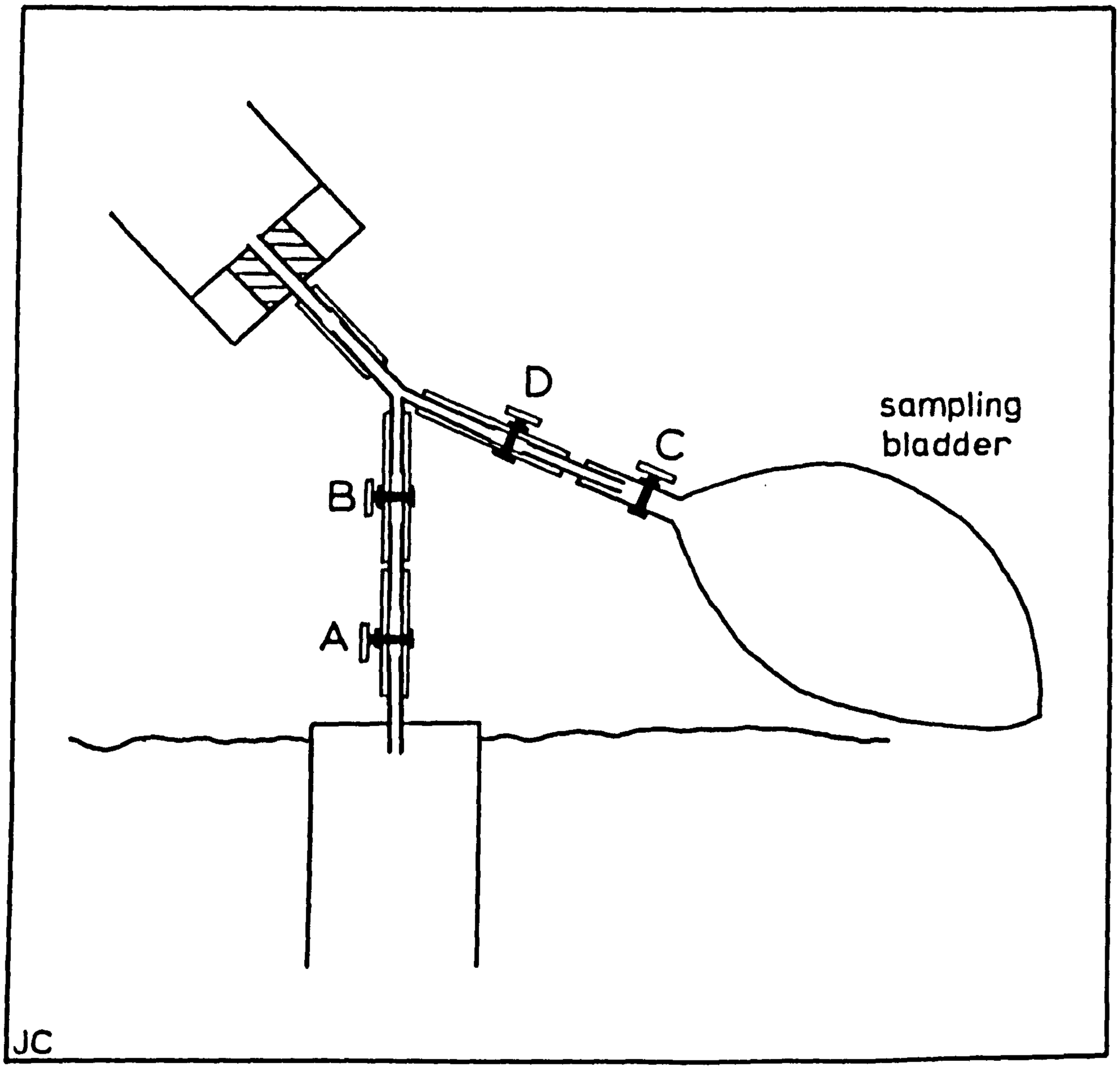


Figure 13 Procedure for sampling soil air. Letters relate to clips referred to in text.

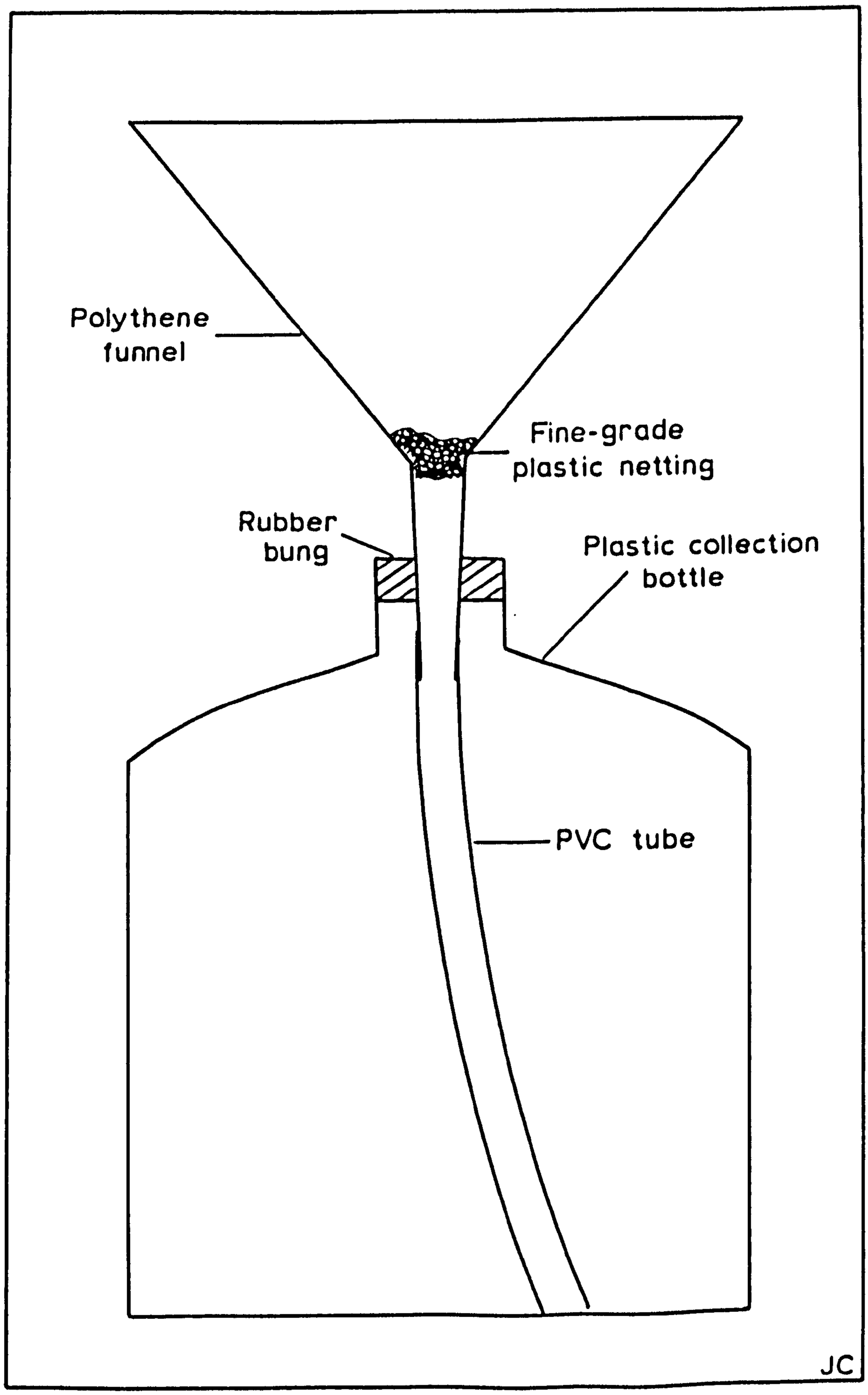


Figure 14 Throughfall collection device.

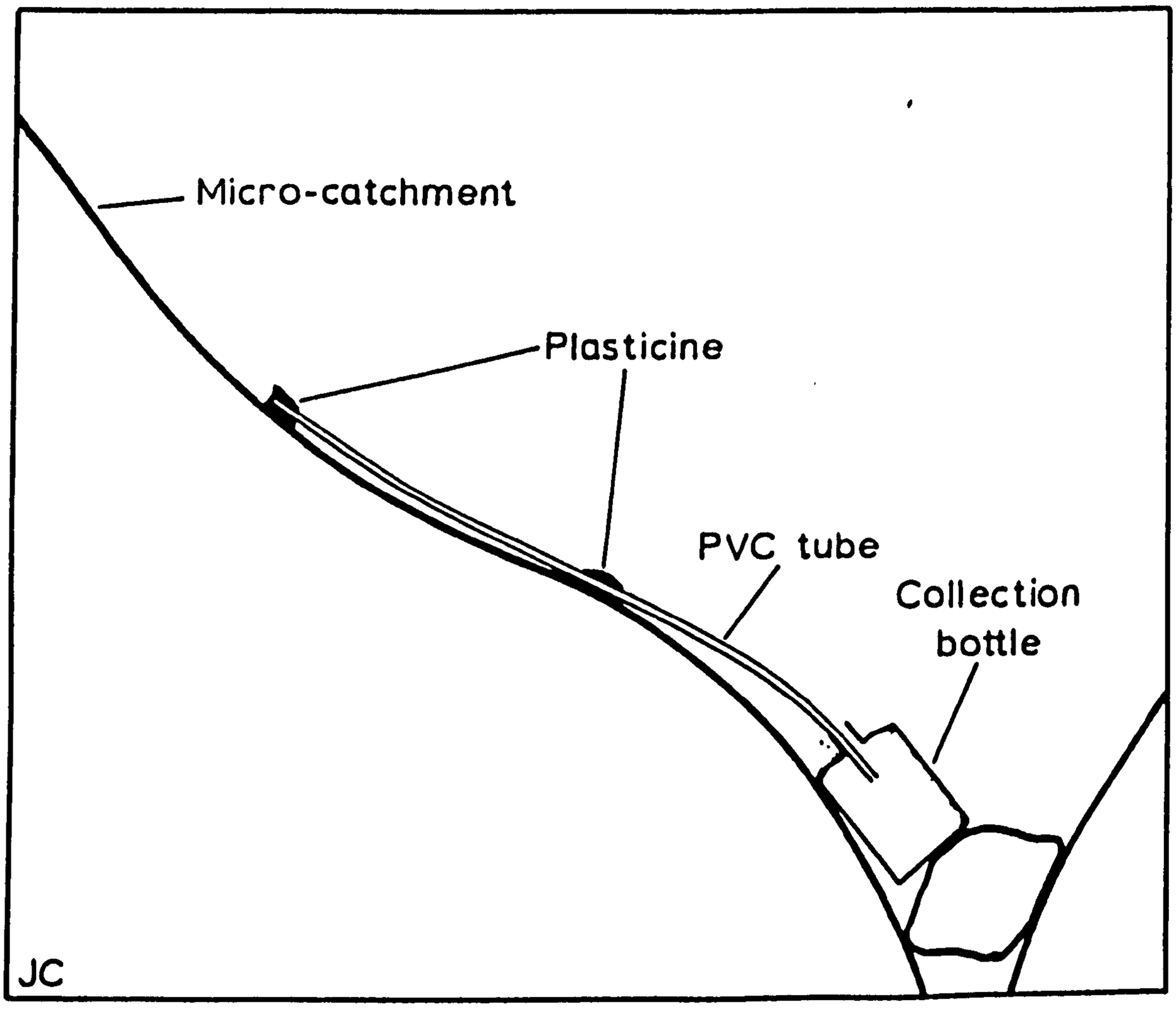


Figure 15 Equipment used to collect surface runoff from solution runnels.

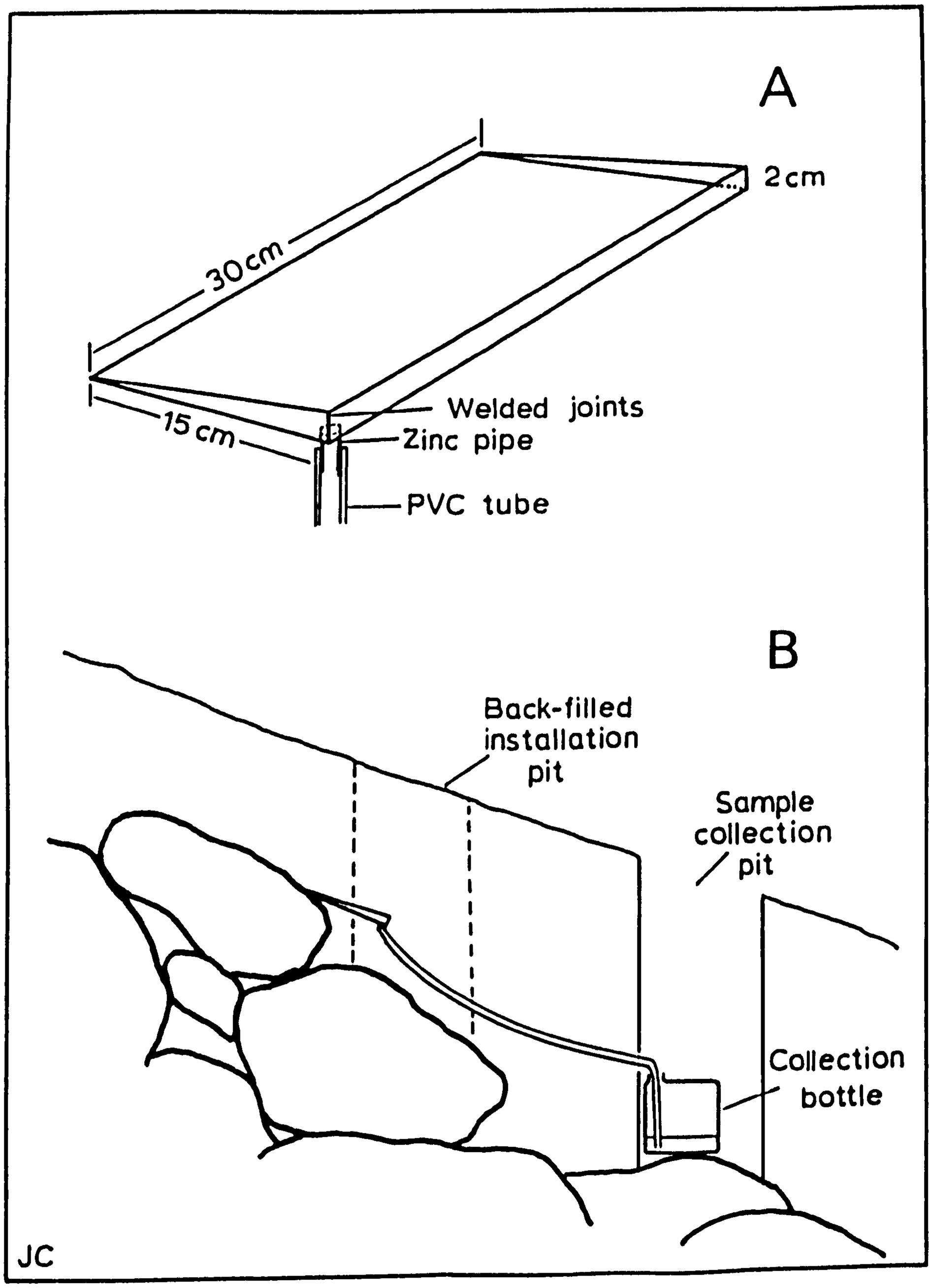


Figure 16 Collection of throughflow:
A. Collection device,
B. Typical field installation.

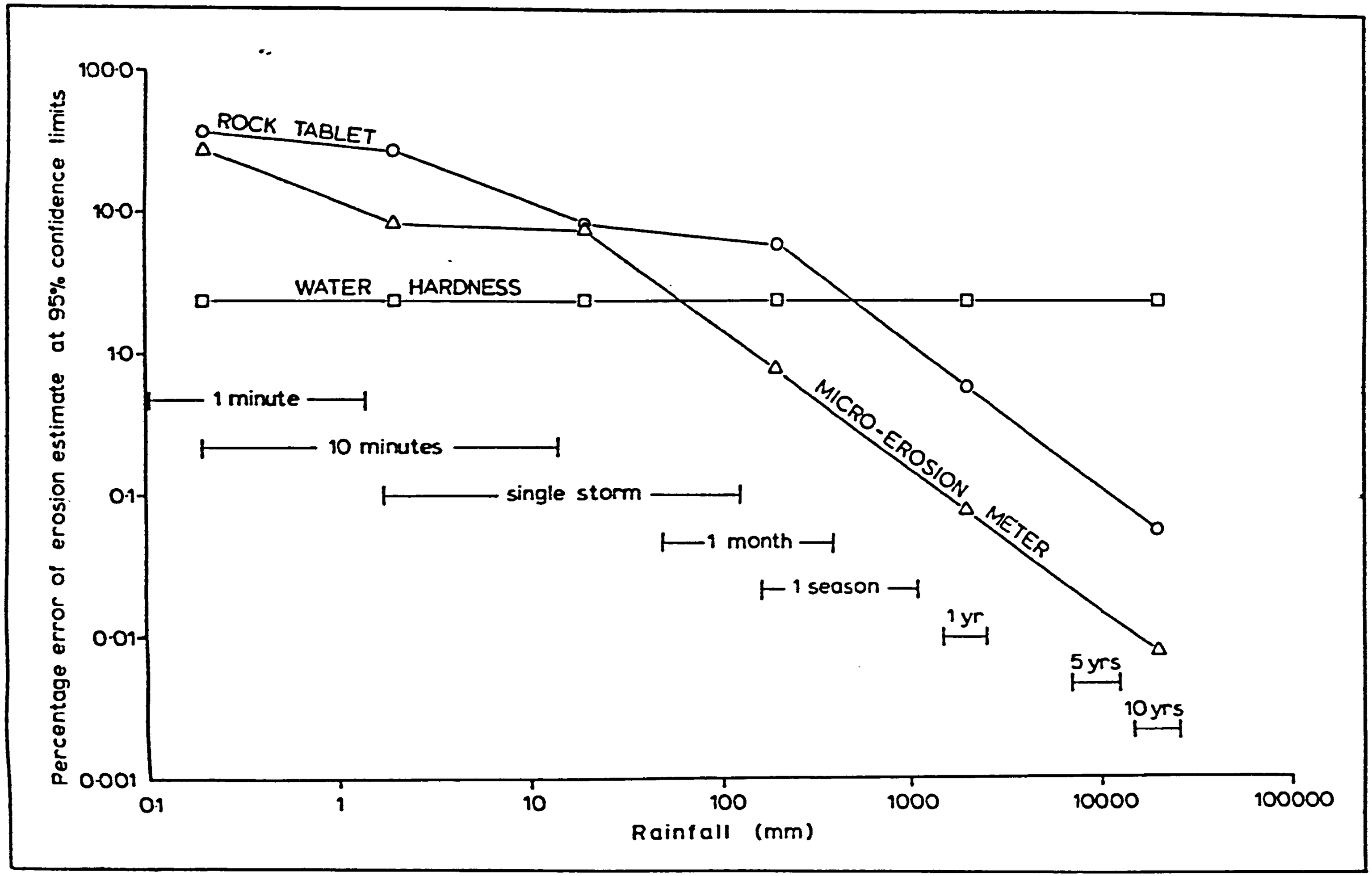


Figure 17 Percentage errors of erosion estimates for three techniques of micro-erosion measurement in relation to precipitation events of given magnitudes. Curves refer to a bare rock outcrop on which runoff is assumed to have a total hardness of 40 ppm over all time periods. A time-scale appropriate for West Malaysian conditions is given.

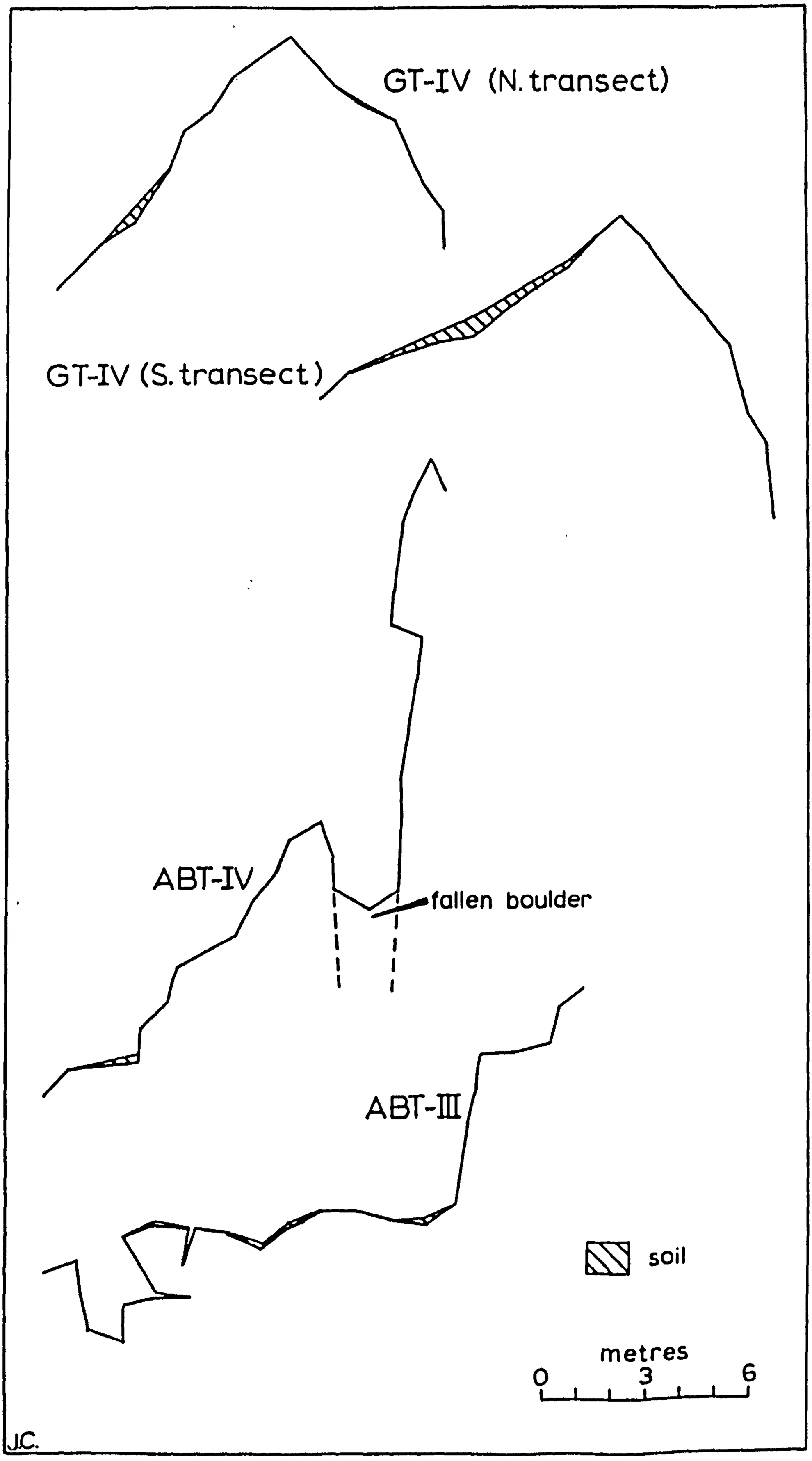


Figure 18 Slope transects in the rocky hilltop plots.

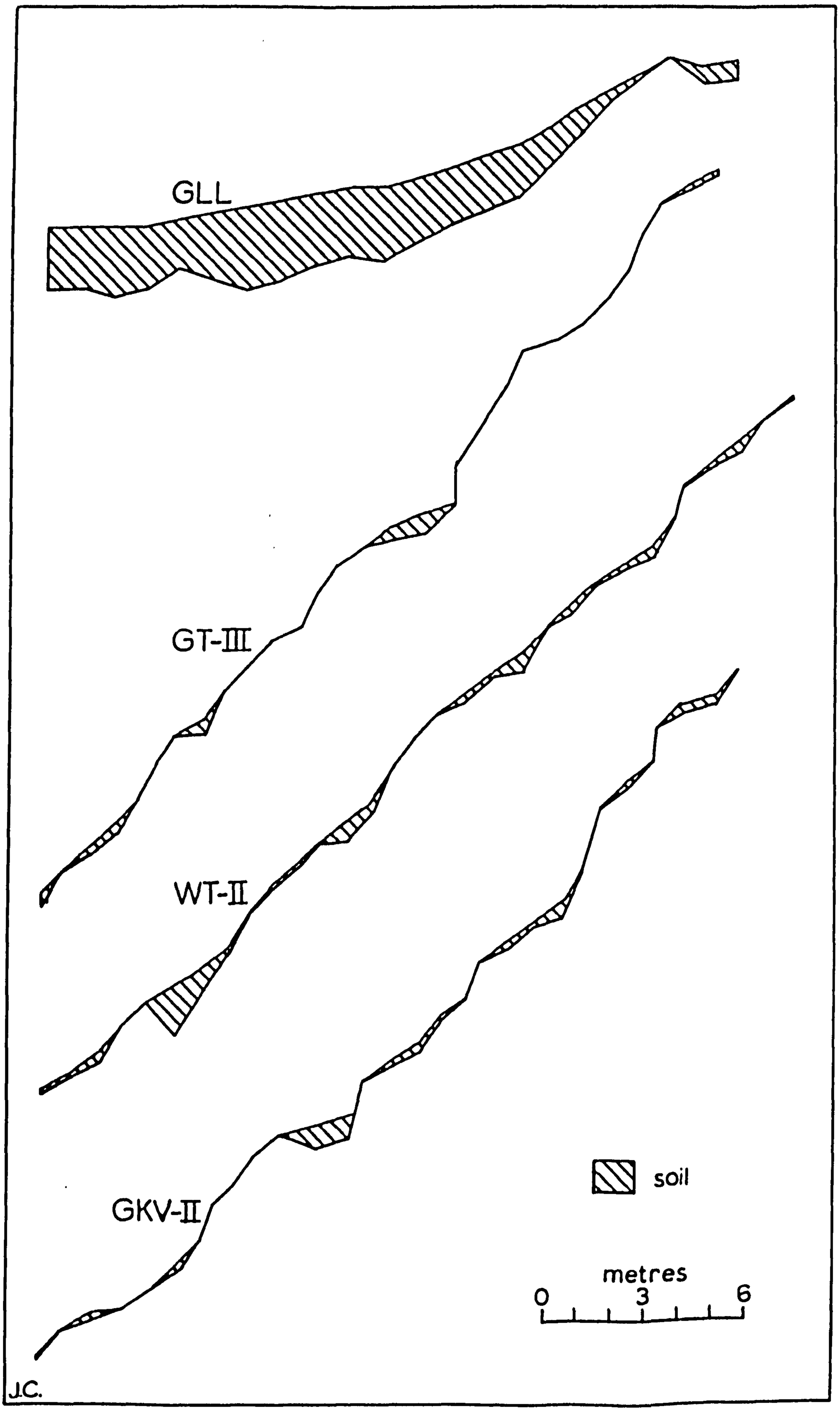


Figure 19 Slope transects in a hilltop depression (GLL) and in the three steep hill slope plots.

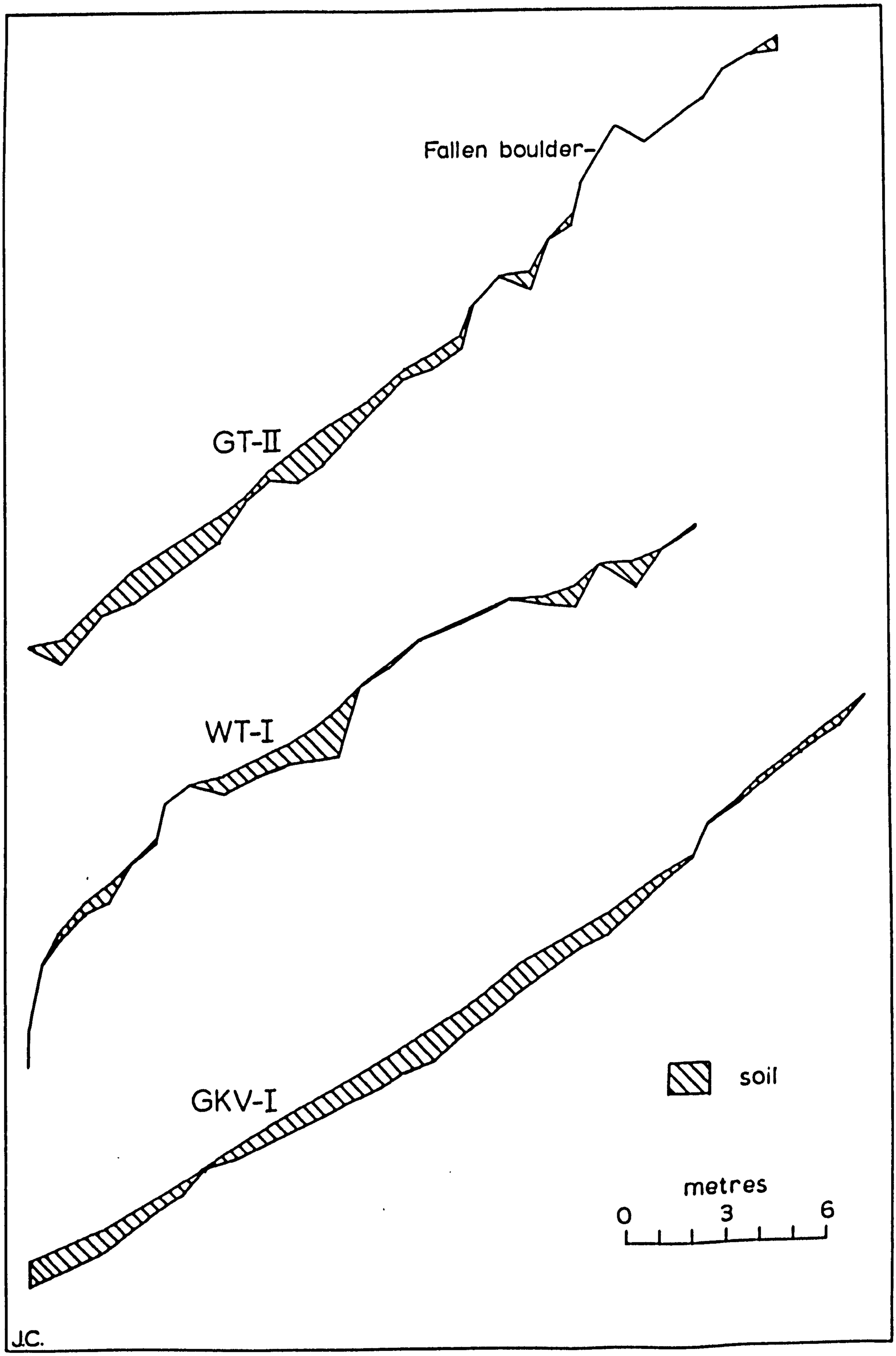


Figure 20 Slope transects in the three moderately-inclined hill slope plots.

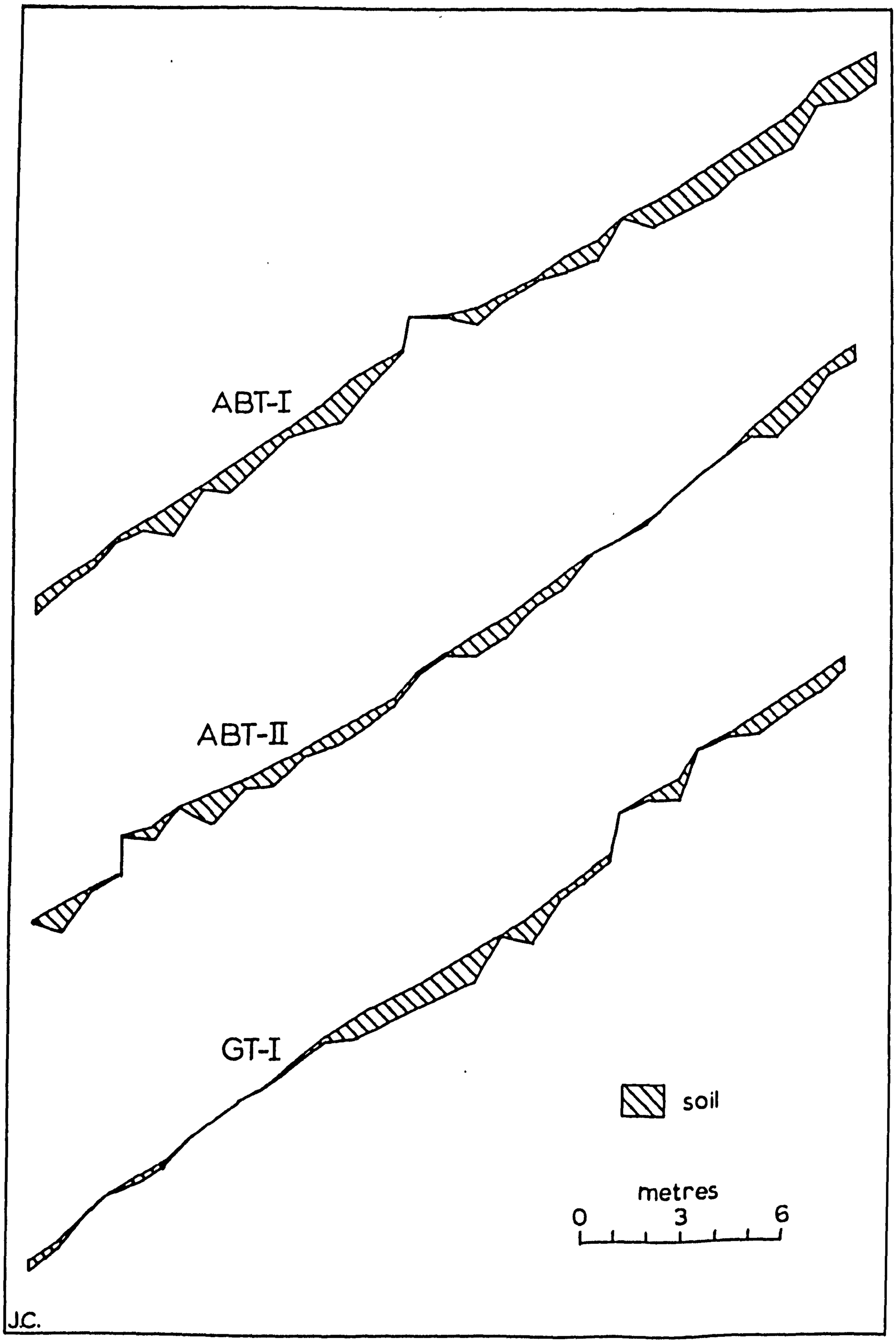


Figure 21 Slope transects in the three moderately-inclined foothill plots.

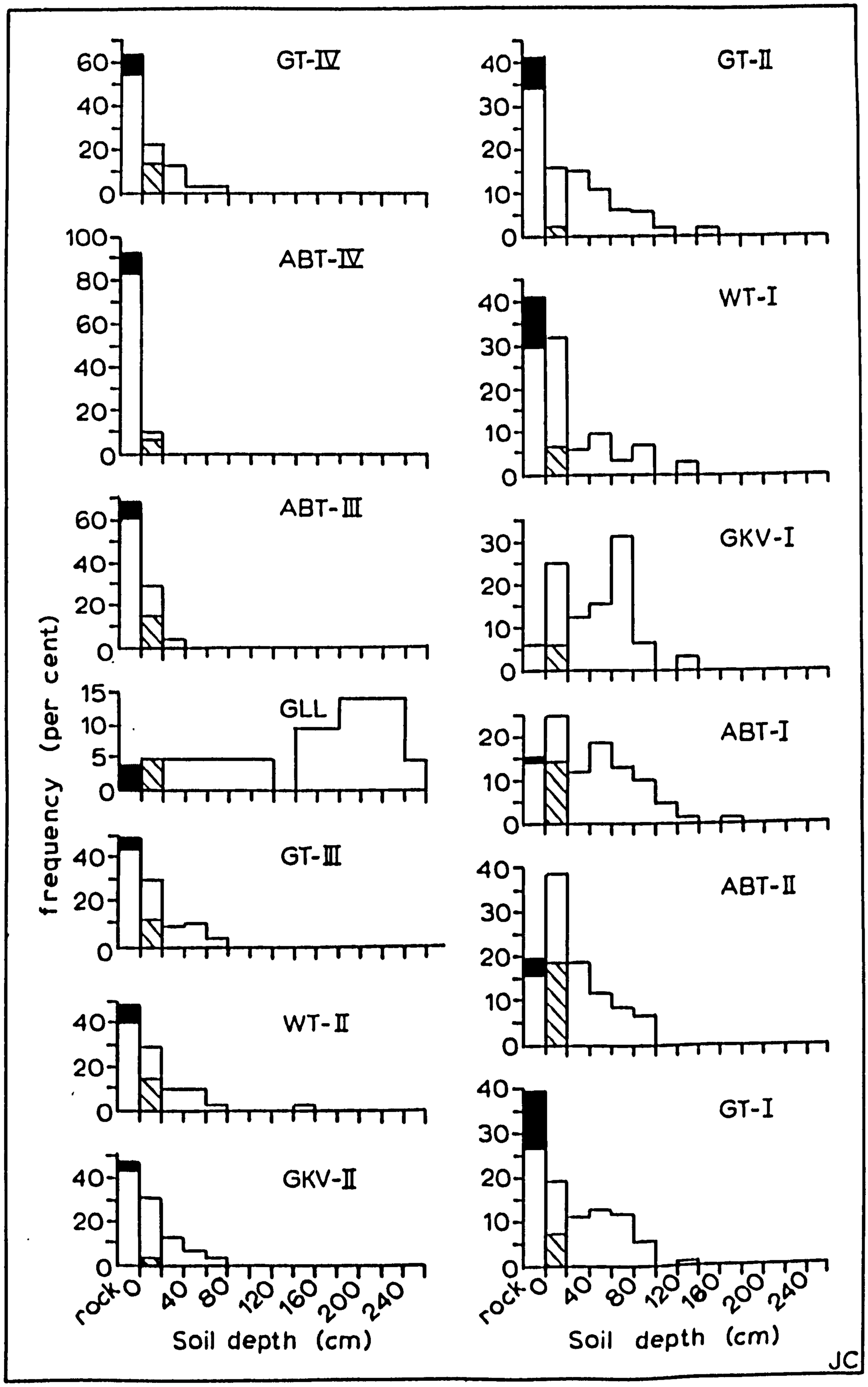


Figure 22 Soil depth distribution in the 13 study plots. In column 1, (rock outcrops), the black part represents those areas covered by leaf litter and decaying organic matter. In column 2, the shaded part represents soils less than 5 cm deep.

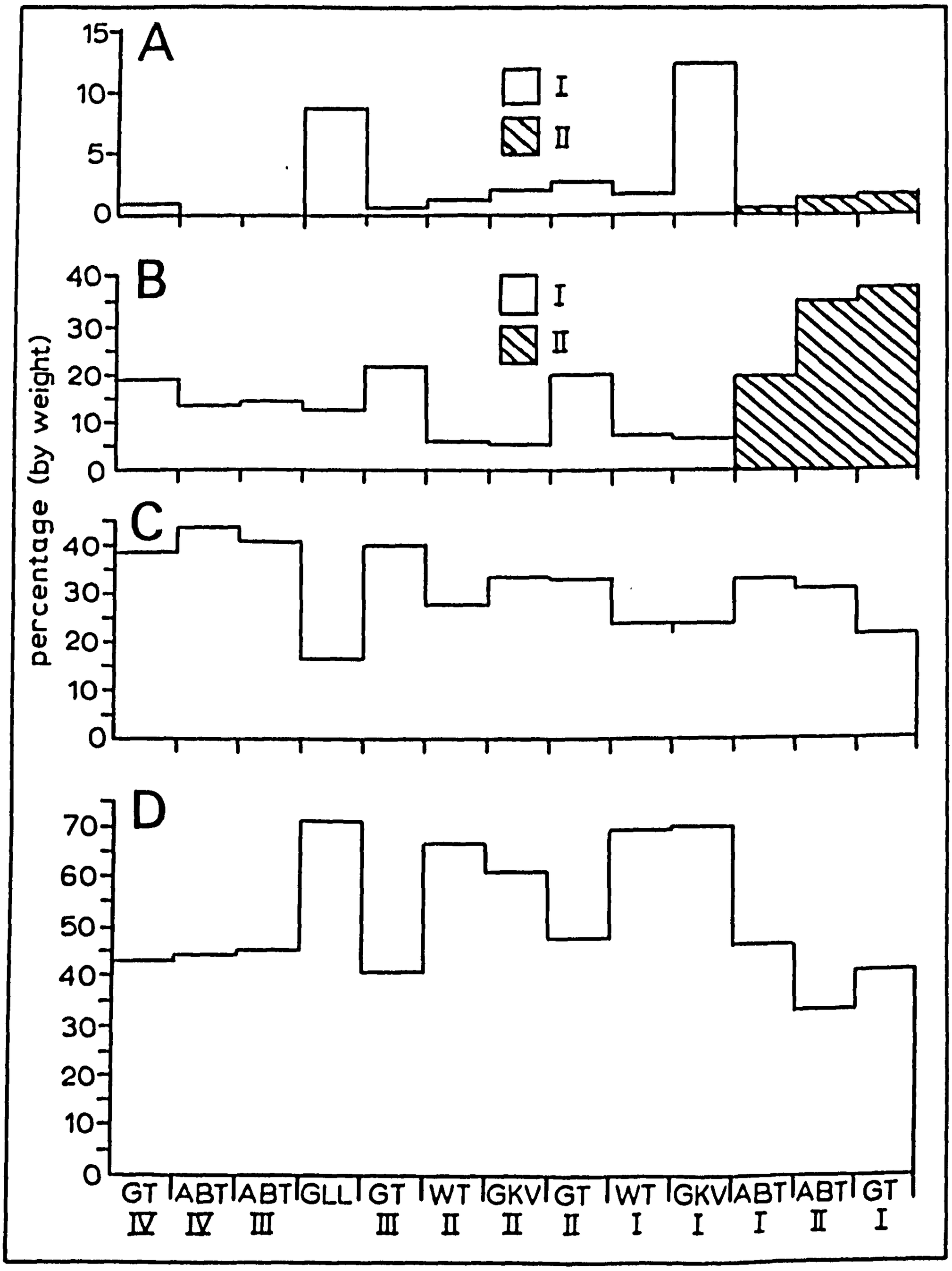


Figure 23 Particle size of soils in the 13 plots:
A. Gravel - I. Predominantly lateritic; II. Mostly quartz,
B. Sand - I. Predominantly lateritic; II. Mostly quartz,
C. Silt,
D. Clay.

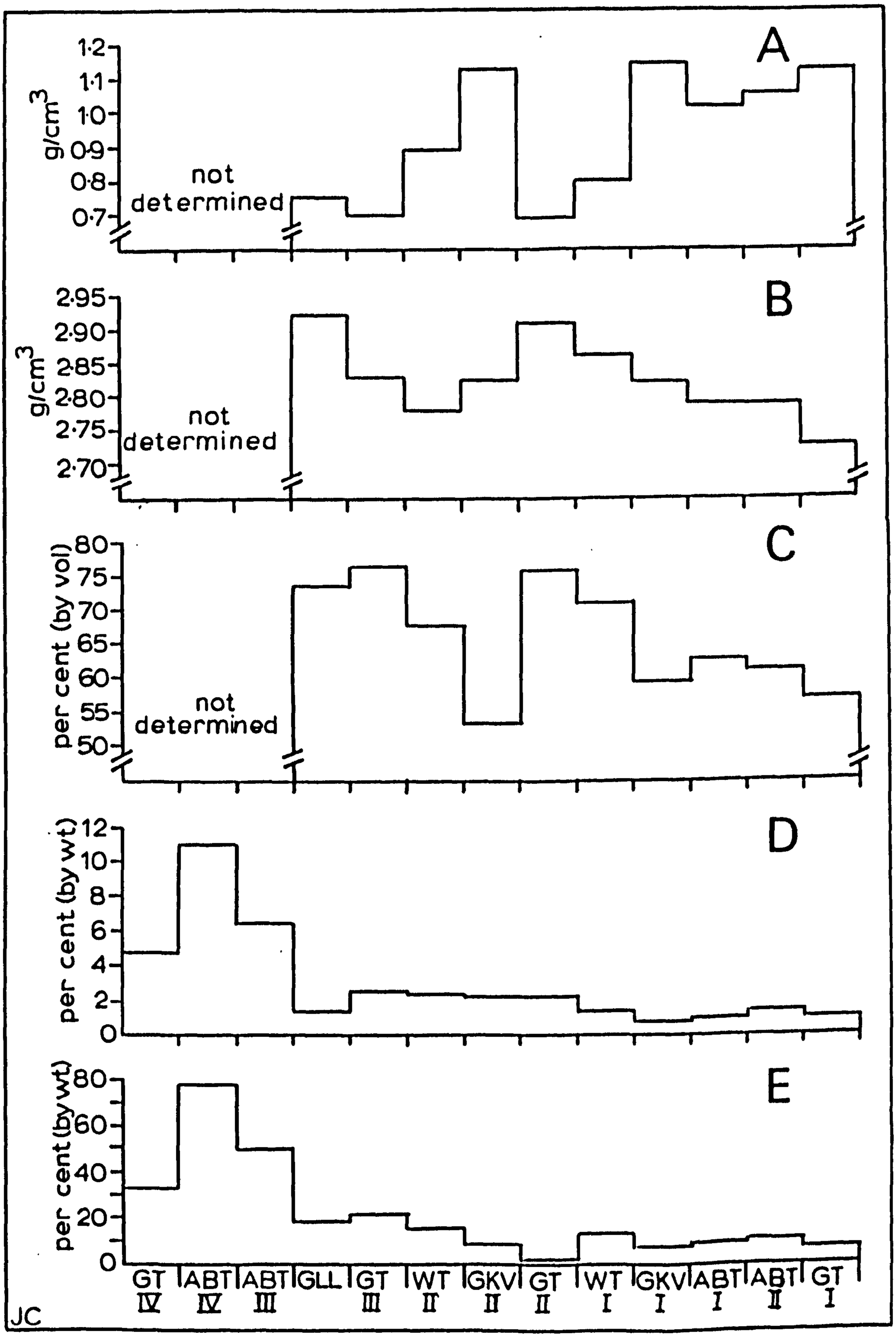


Figure 24 Physical properties and organic content of soils in the 13 plots:

A. Bulk density at 20 cm,
B. Particle density (maximum),
C. Pore space at 20 cm,
D. Organic carbon at 10 cm,
E, Loss on ignition at 10 cm.

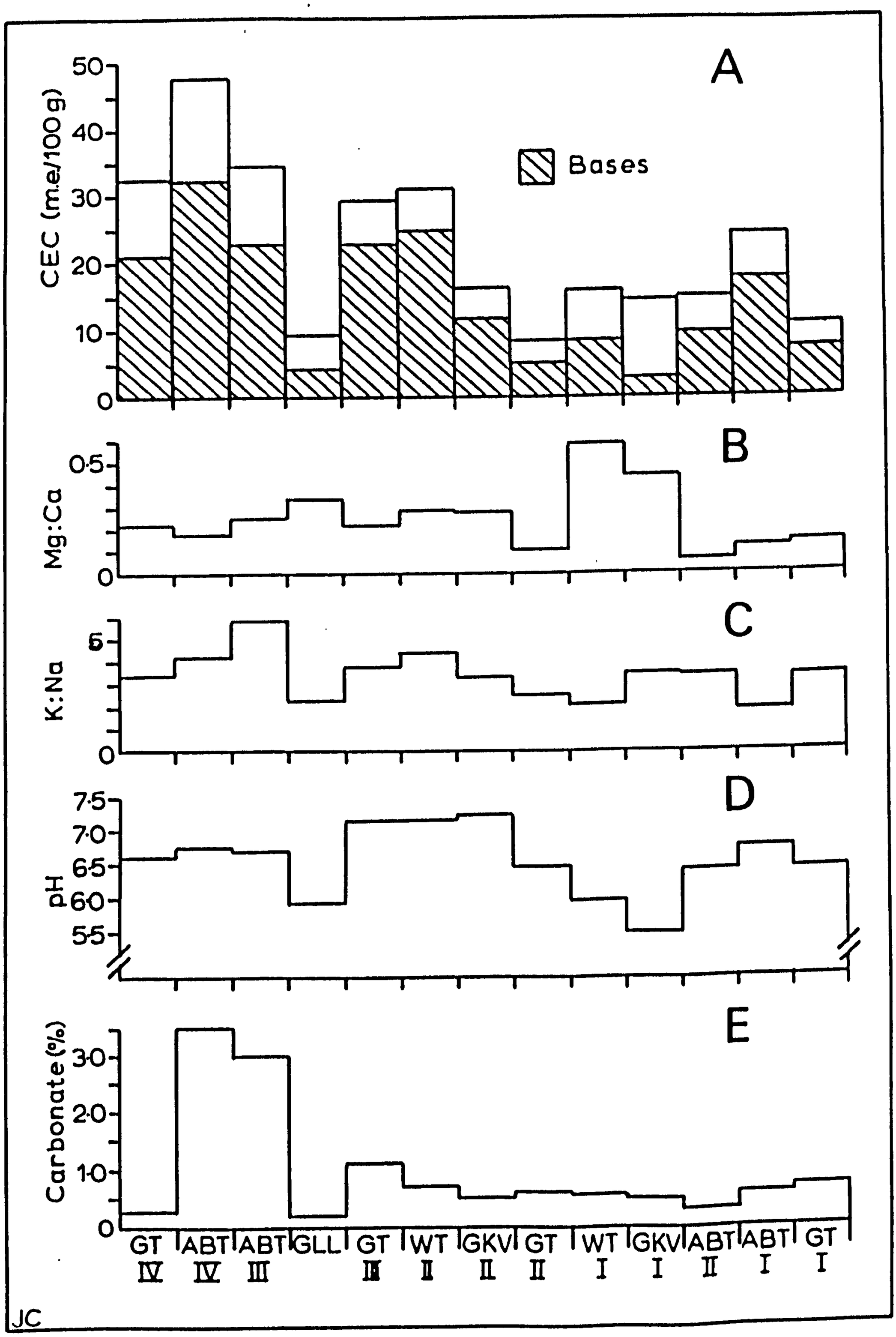


Figure 25 Average chemical properties of soils in the 13 plots:
A. C.E.C.,
B. Mg:Ca ratio of exchangeable cations,
C. K:Na ratio of exchangeable cations,
D. pH,
E. Carbonate content.

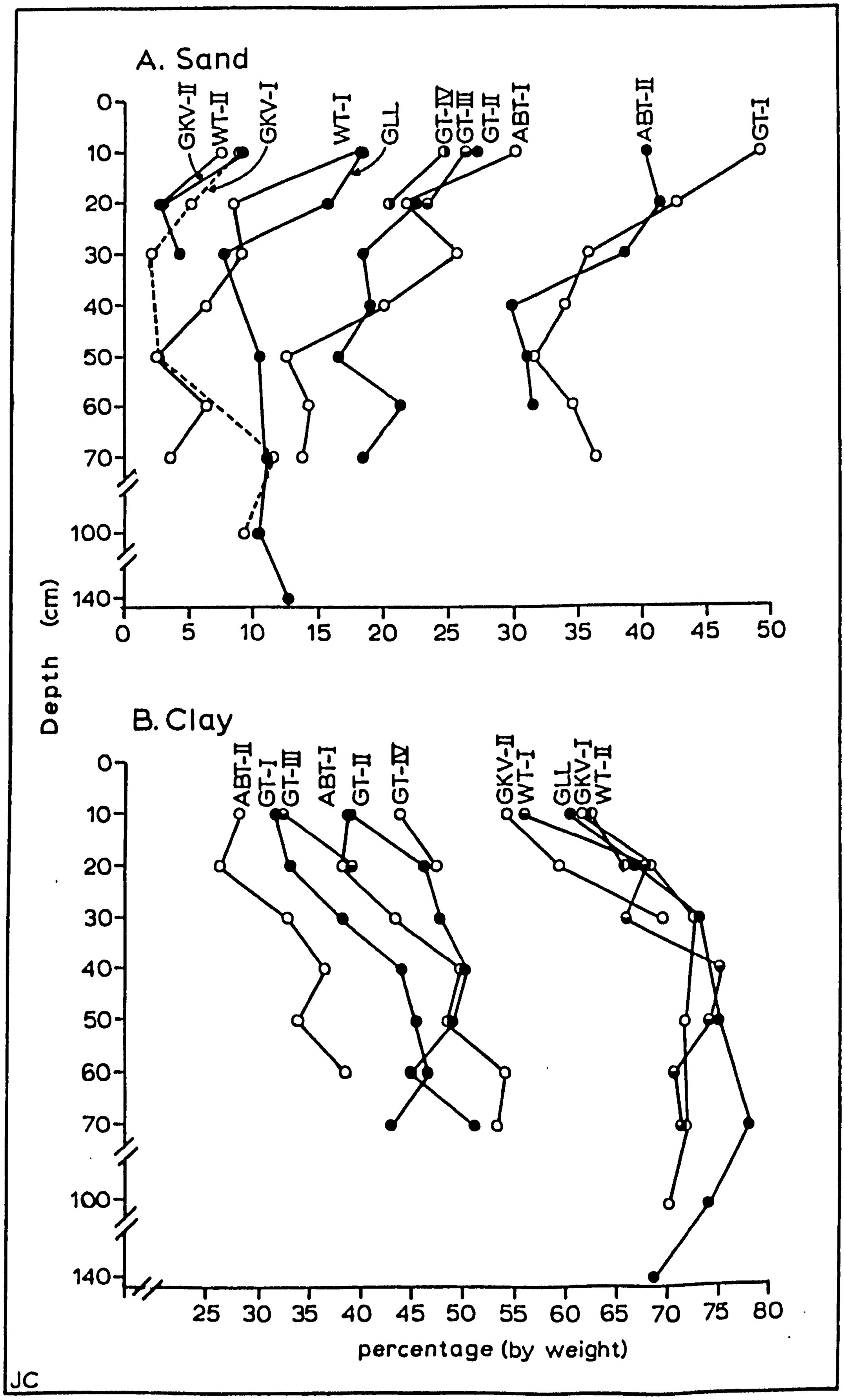


Figure 26 Variations in particle size with depth in the study plots.

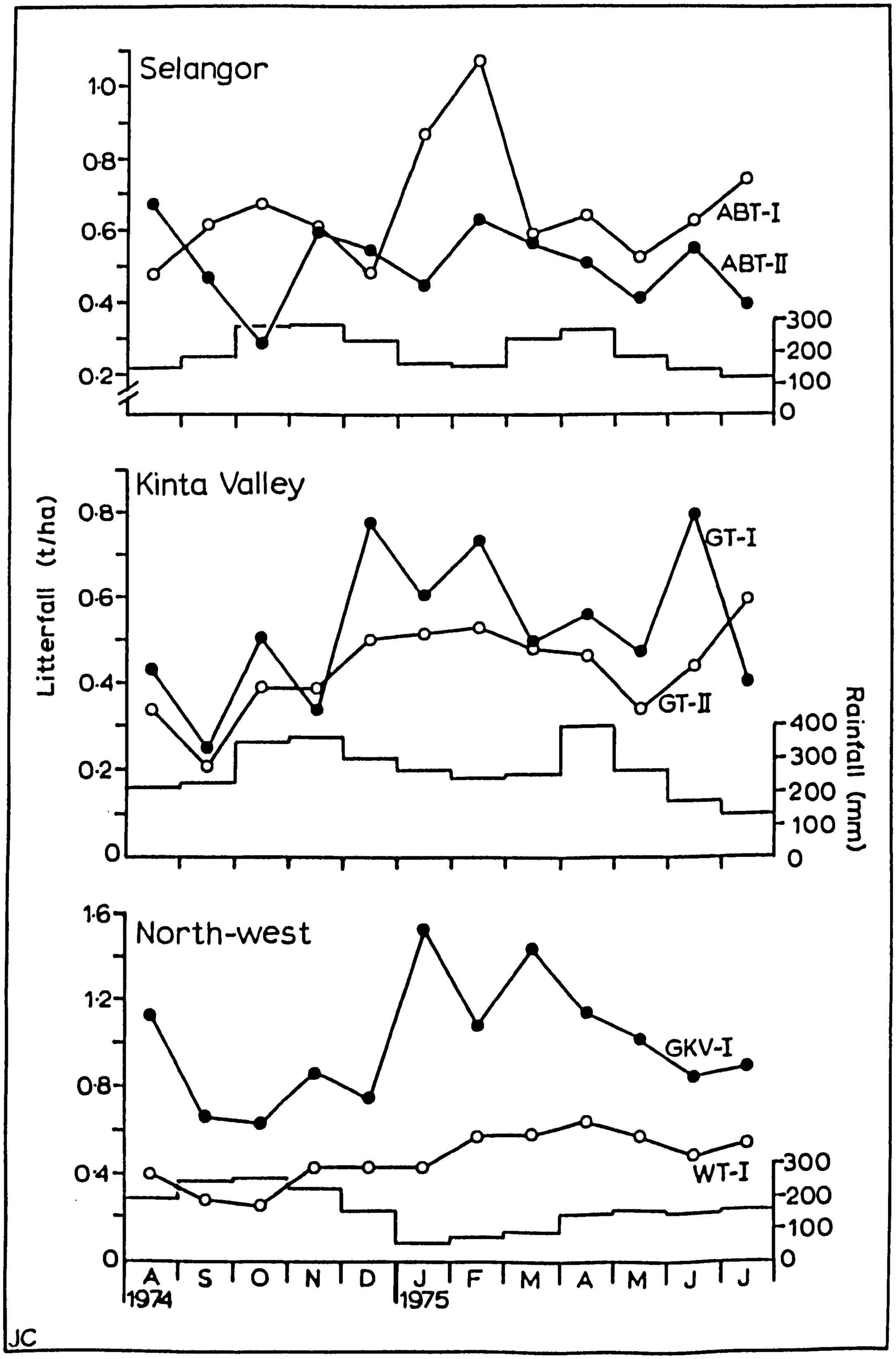


Figure 27 Monthly variations in litterfall in the three field areas. Rainfall shown is that of the long-term average at the following stations: 1602 (Selangor); 3601 (based on 3606, Kinta Valley); 4601 (North-west region).

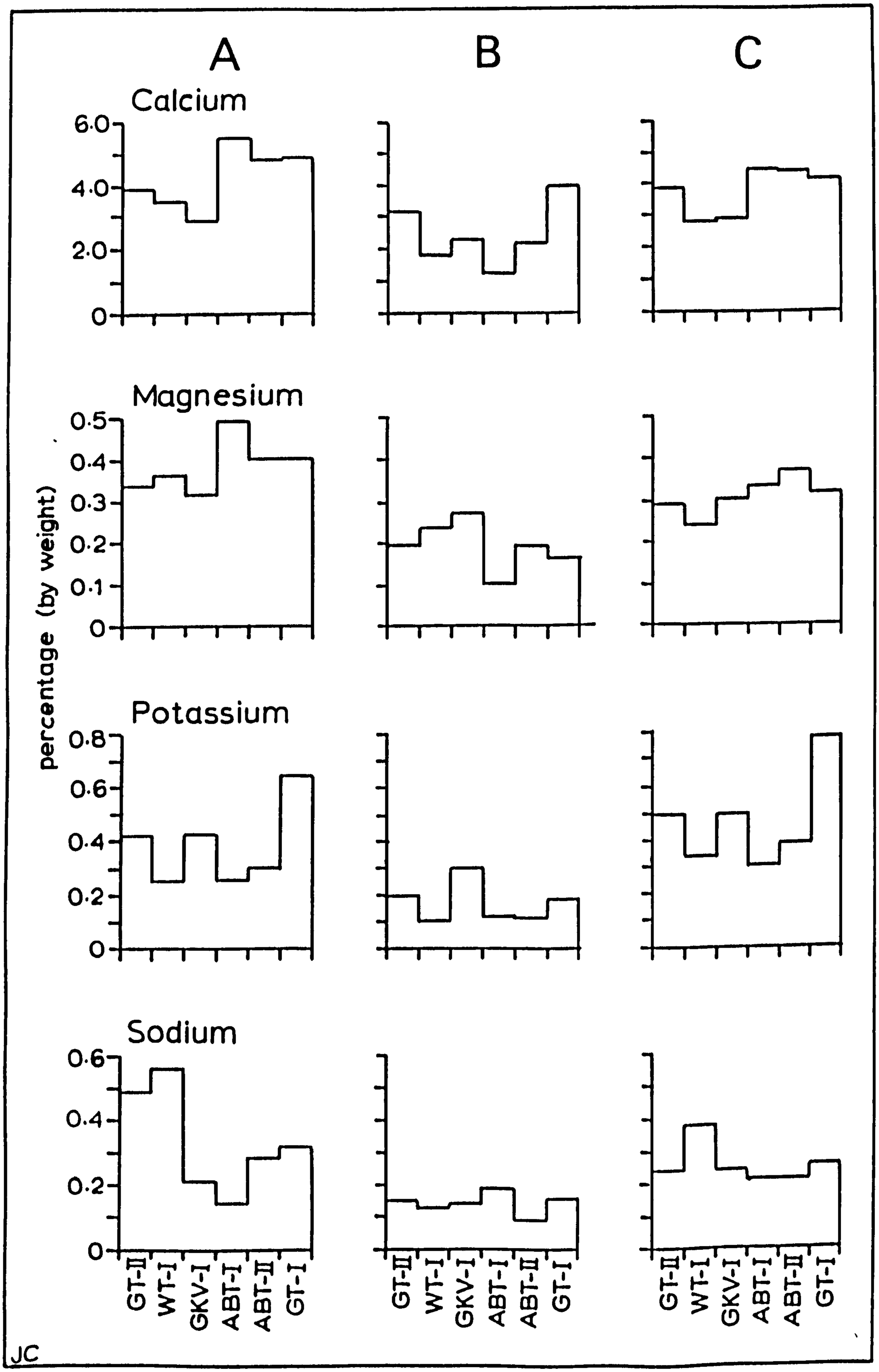


Figure 28 Nutrient content of fine litter in the six main study plots: A. Leaves; B. Twigs; C. Miscellaneous fraction.

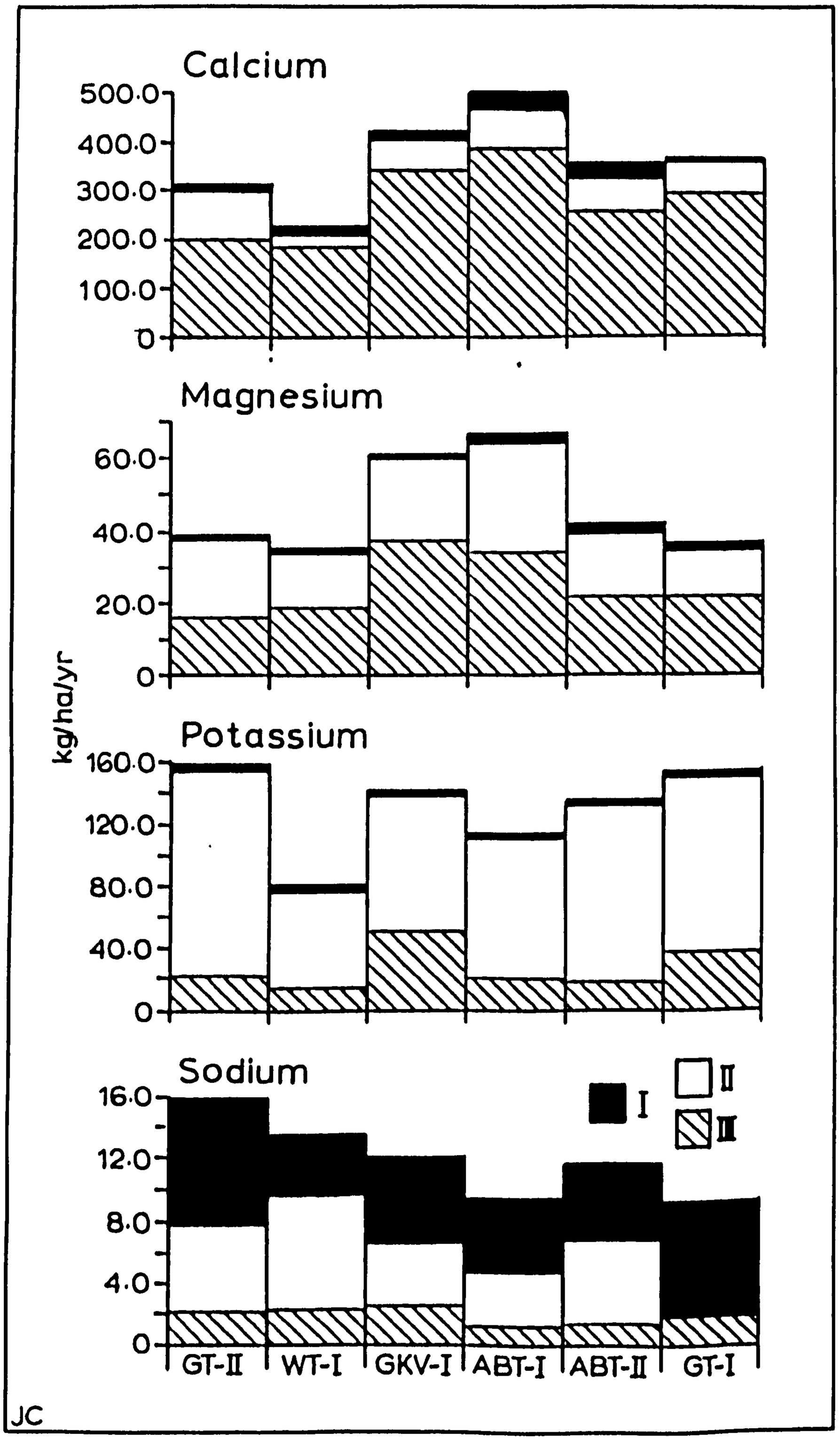


Figure 29 Total input of major elements at the soil surface in the six main study plots during the one-year sampling period. Shading:
I. Rainfall,
II. Canopy leaching,
III. Litterfall.

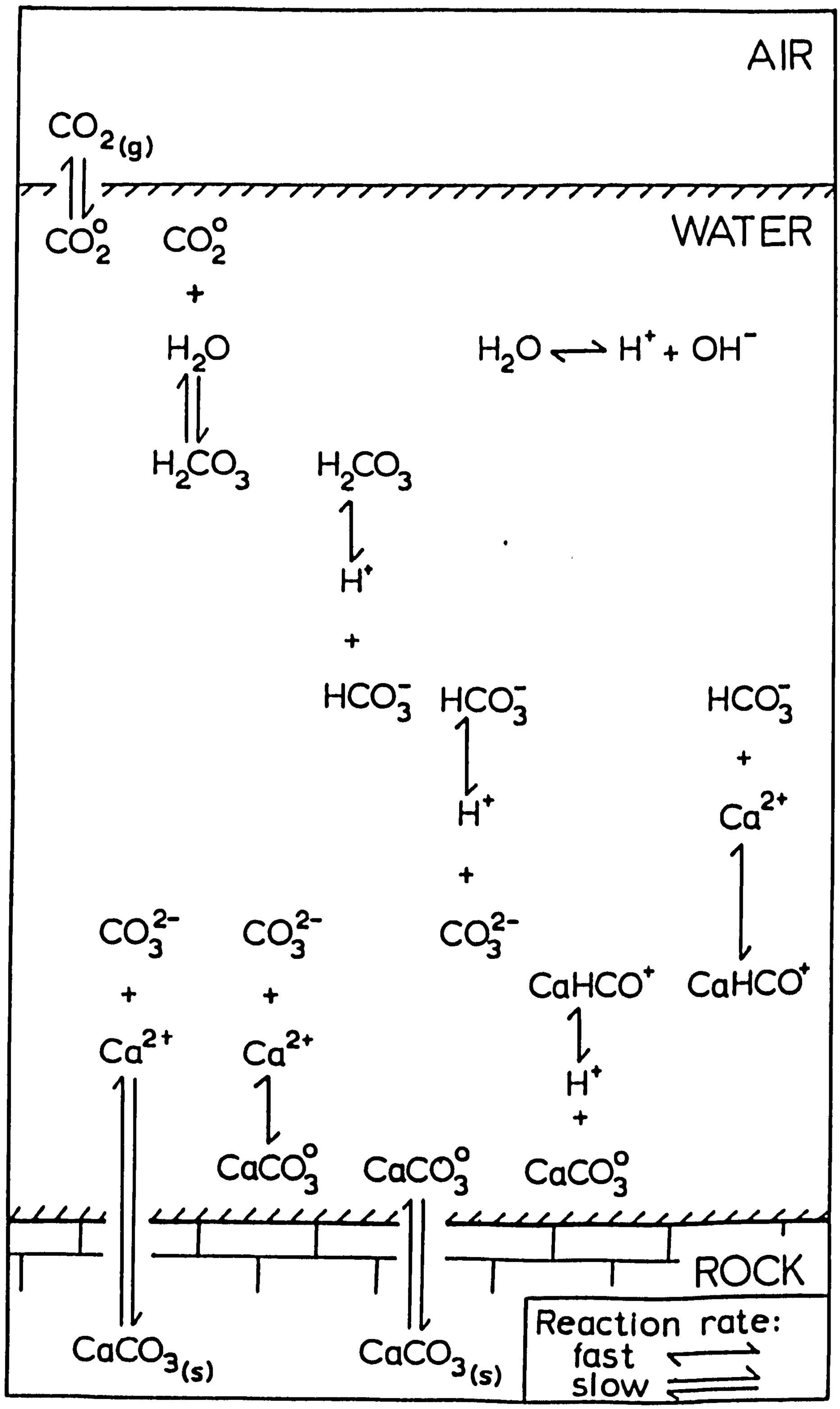


Figure 30 Principal reactions in the $CaCO_3$-CO_2-H_2O system.

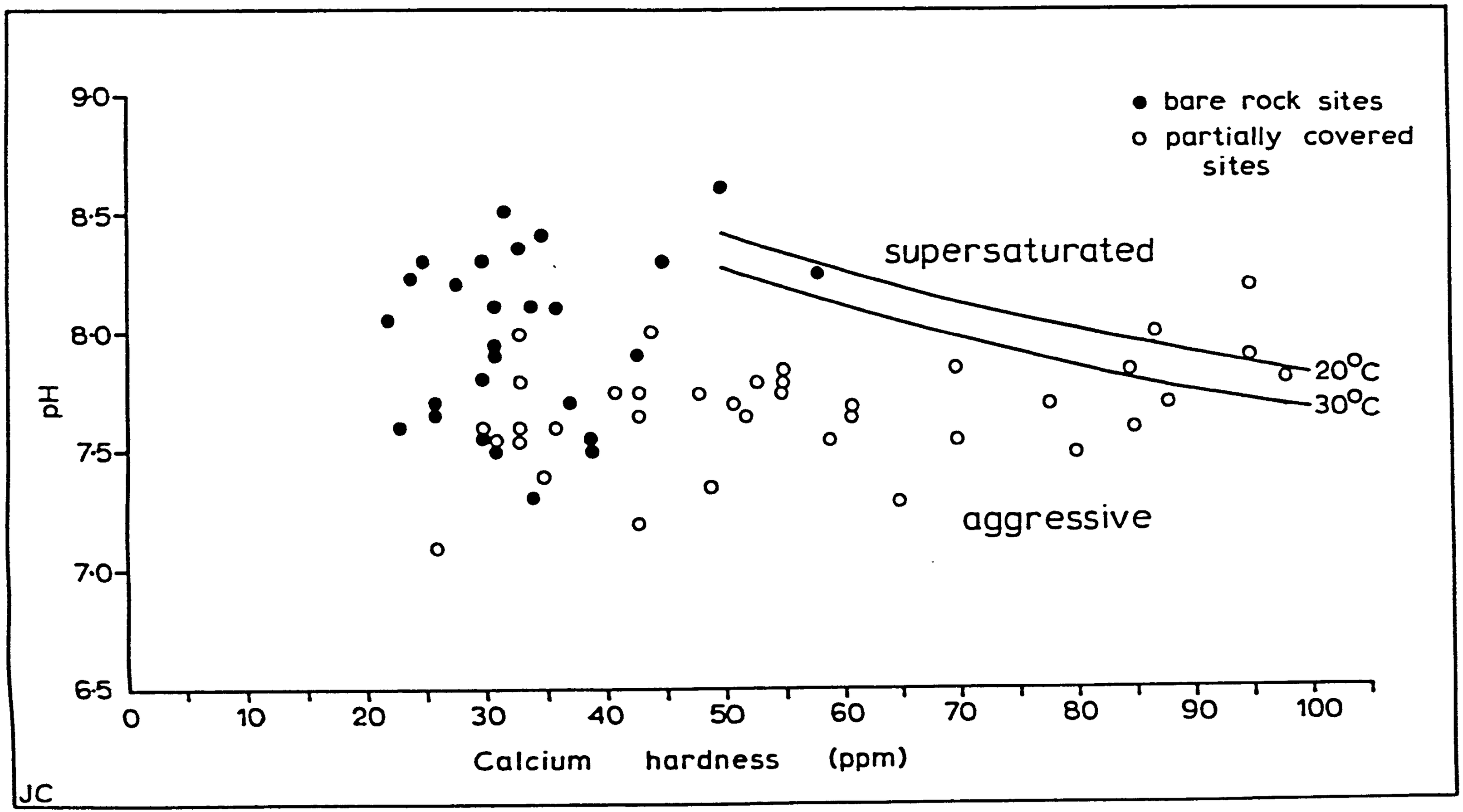


Figure 31 Plot of calcium hardness against pH for the water sample with the median hardness concentration in each of the 65 micro-catchments. Equilibrium curves are for saturated calcite solutions (after Picknett, 1972).

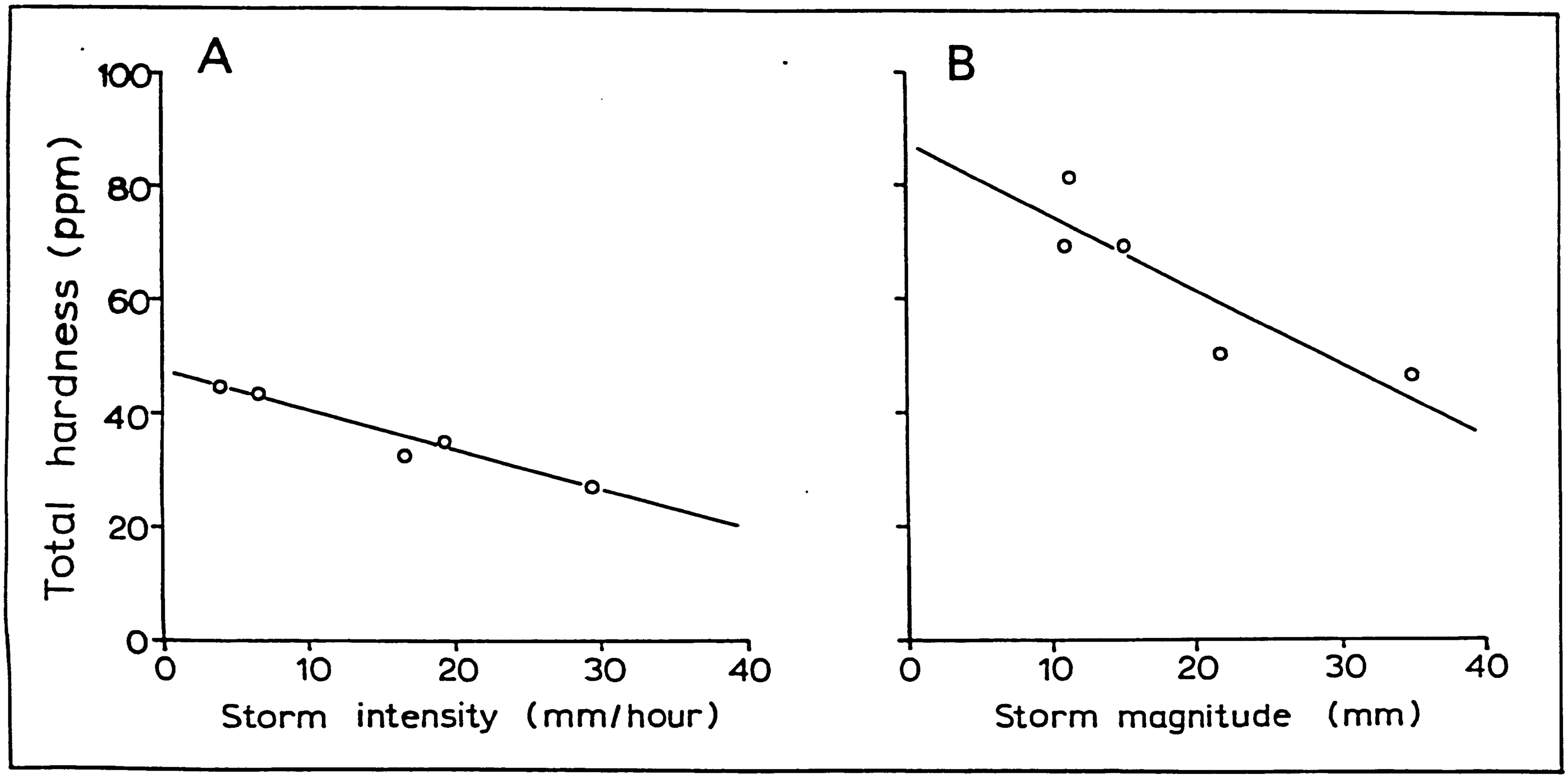


Figure 32 Relationship between total hardness and storm characteristics: A. Bare rock micro-catchments; B. Micro-catchments with a partial soil/litter cover.

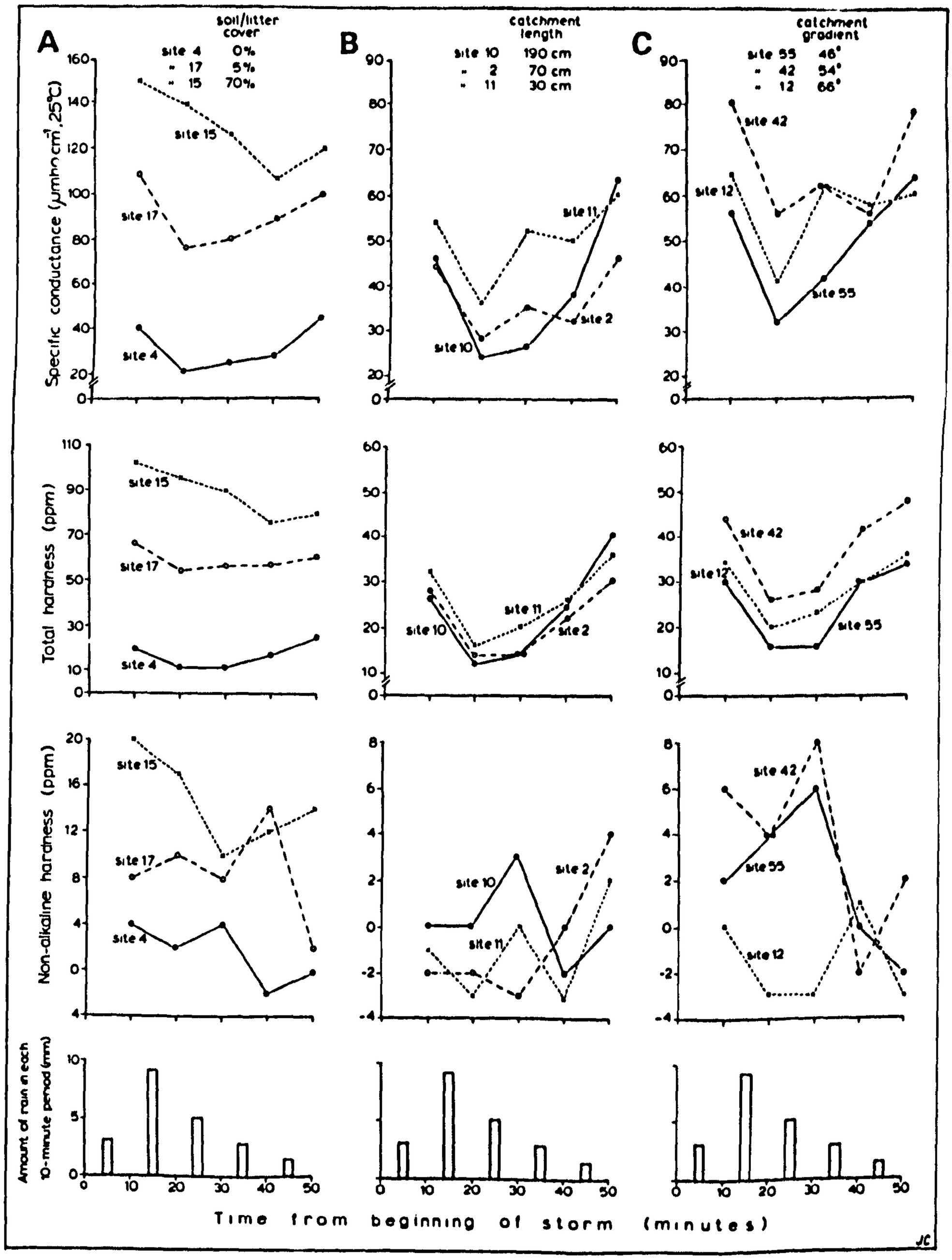


Figure 33 Variations in specific conductance and water hardness properties in selected micro-catchments during a storm of 21.8 mm in 45 minutes:

A. Sites with different amounts of surface cover,
B. Sites of different catchment size,
C. Sites of different gradient.

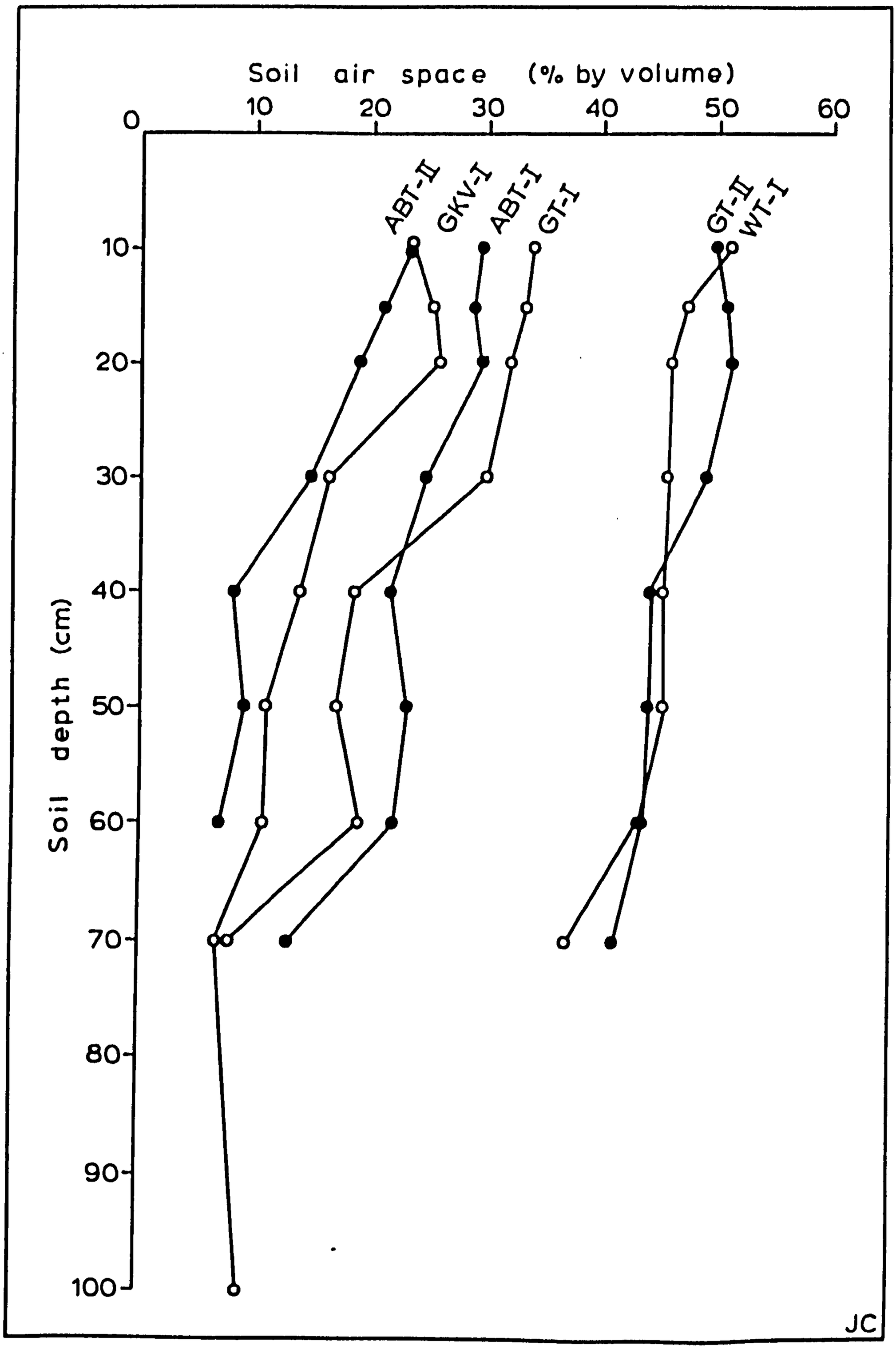


Figure 34 Variations in soil-air space with depth in the six main sampling plots.

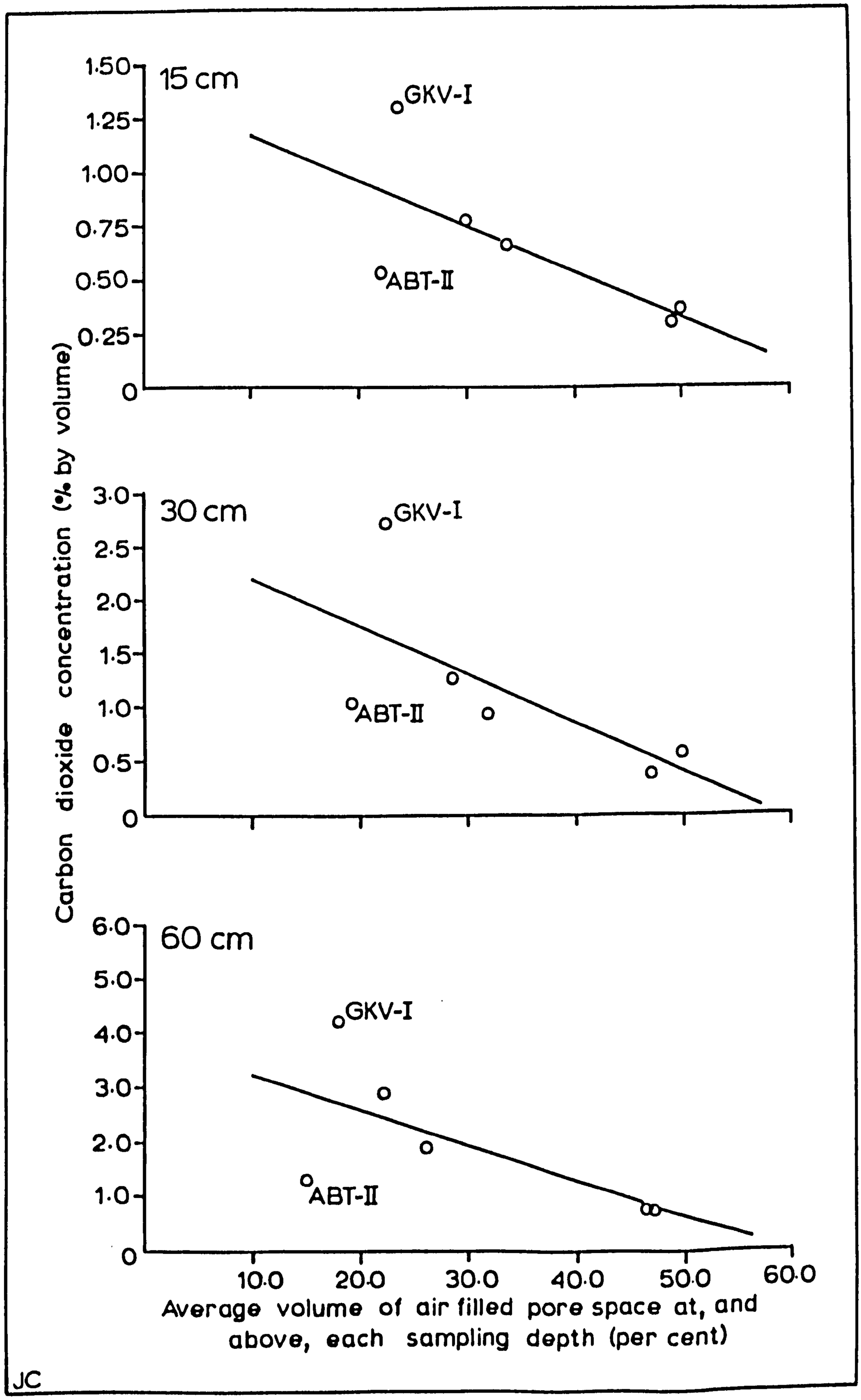


Figure 35 Relationship between carbon dioxide concentration and soil porosity at different depths in the six main sampling plots.

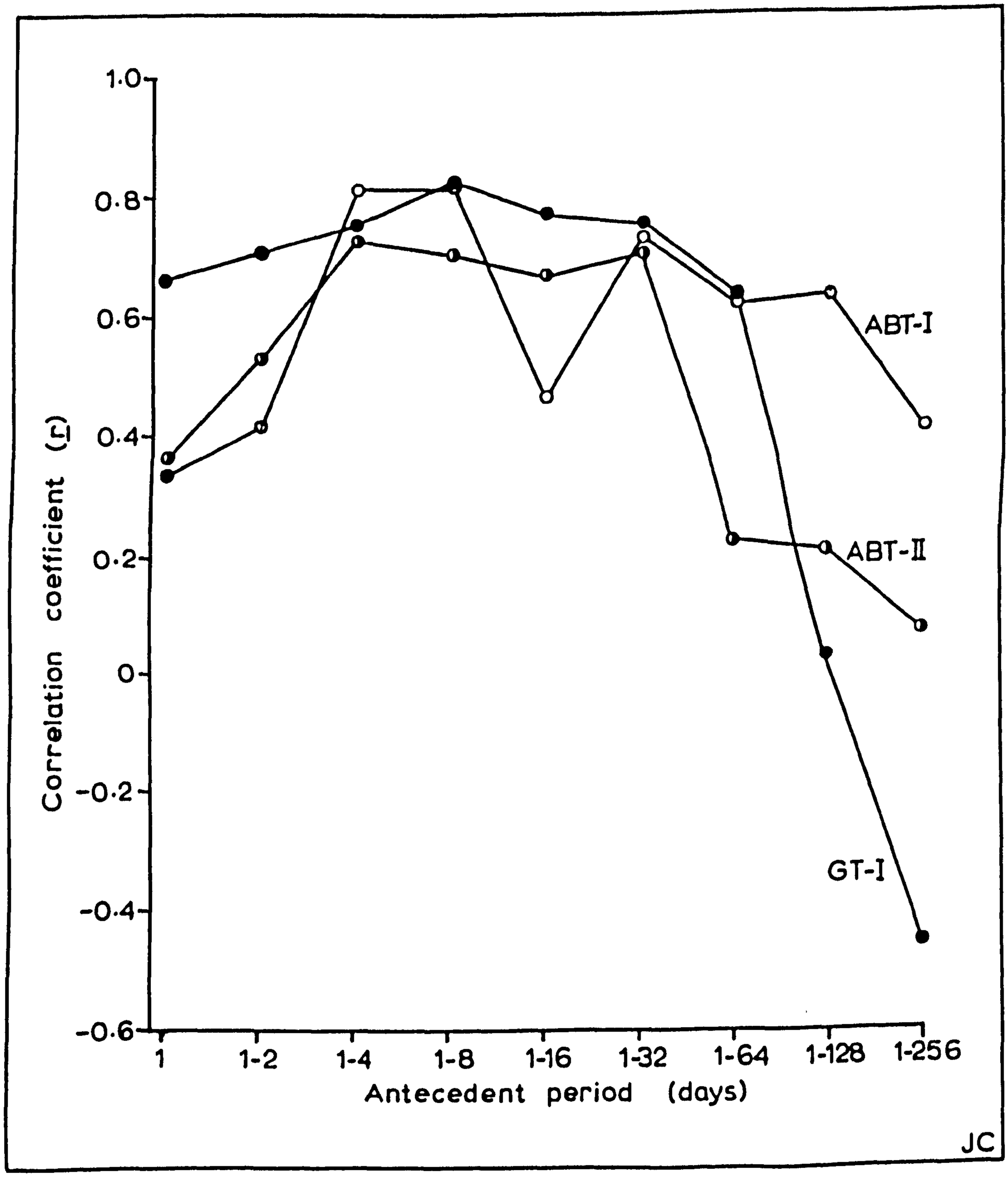


Figure 36 Correlation coefficients(r) for the relationships between soil moisture in the main profile of the three foothill plots and rainfall during pre-set antecedent periods.

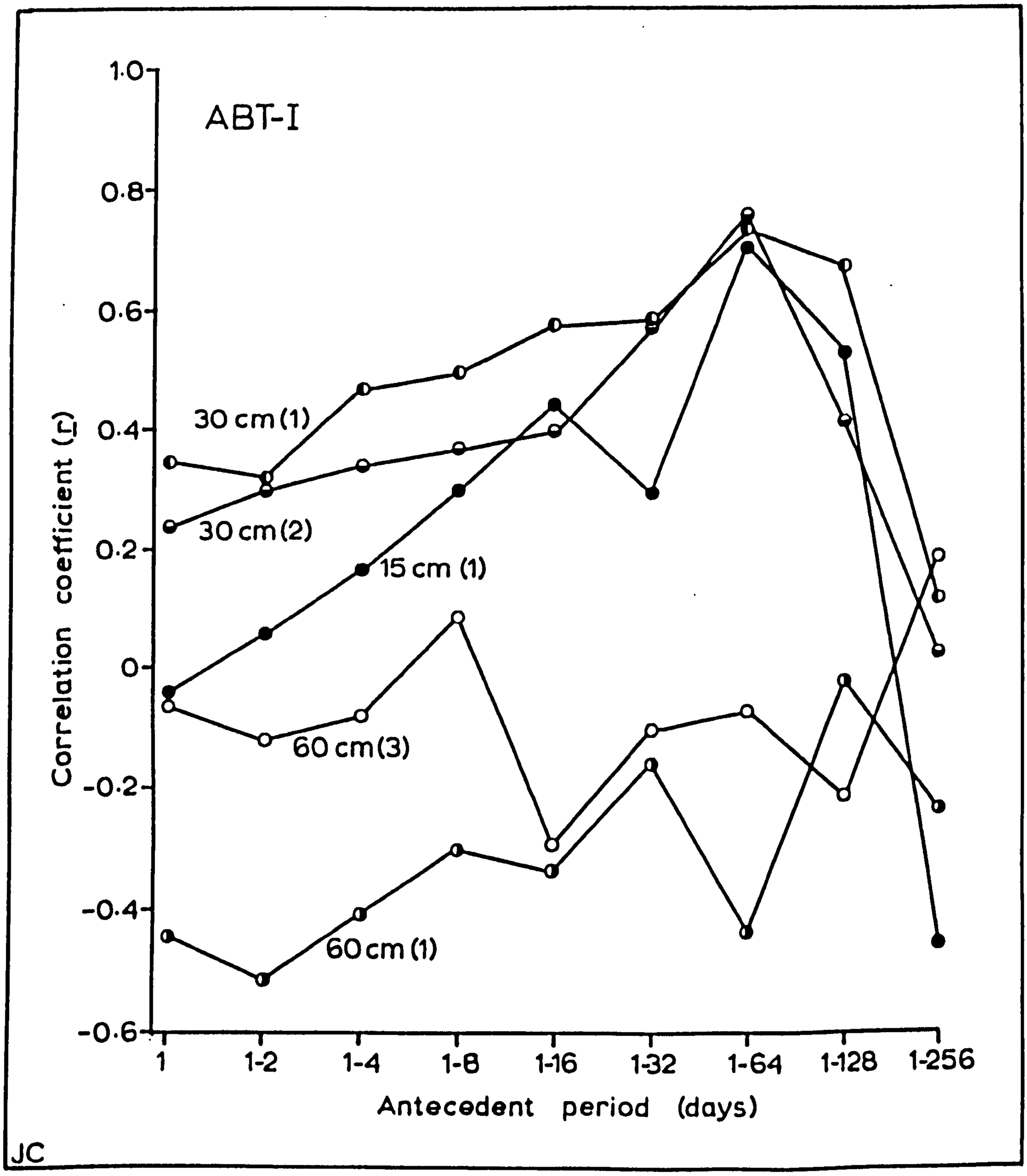


Figure 37 Correlation coefficients (r) for the relationships between soil carbon dioxide at various depths in the profiles studied in ABT-I and rainfall during pre-set antecedent periods.

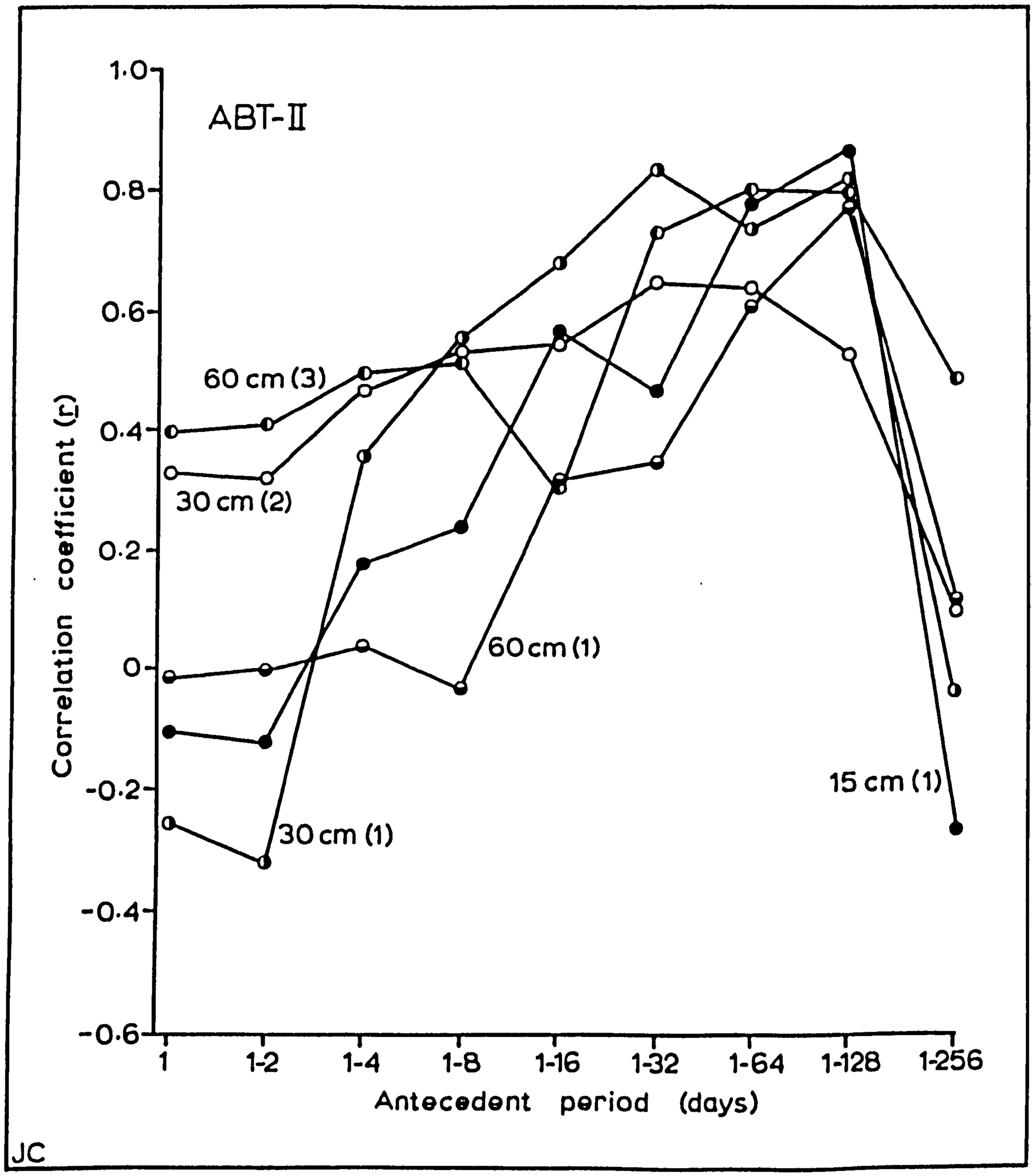


Figure 38 Correlation coefficients (r) for the relationships between soil carbon dioxide at various depths in the profiles studied in ABT-II and rainfall during pre-set antecedent periods.

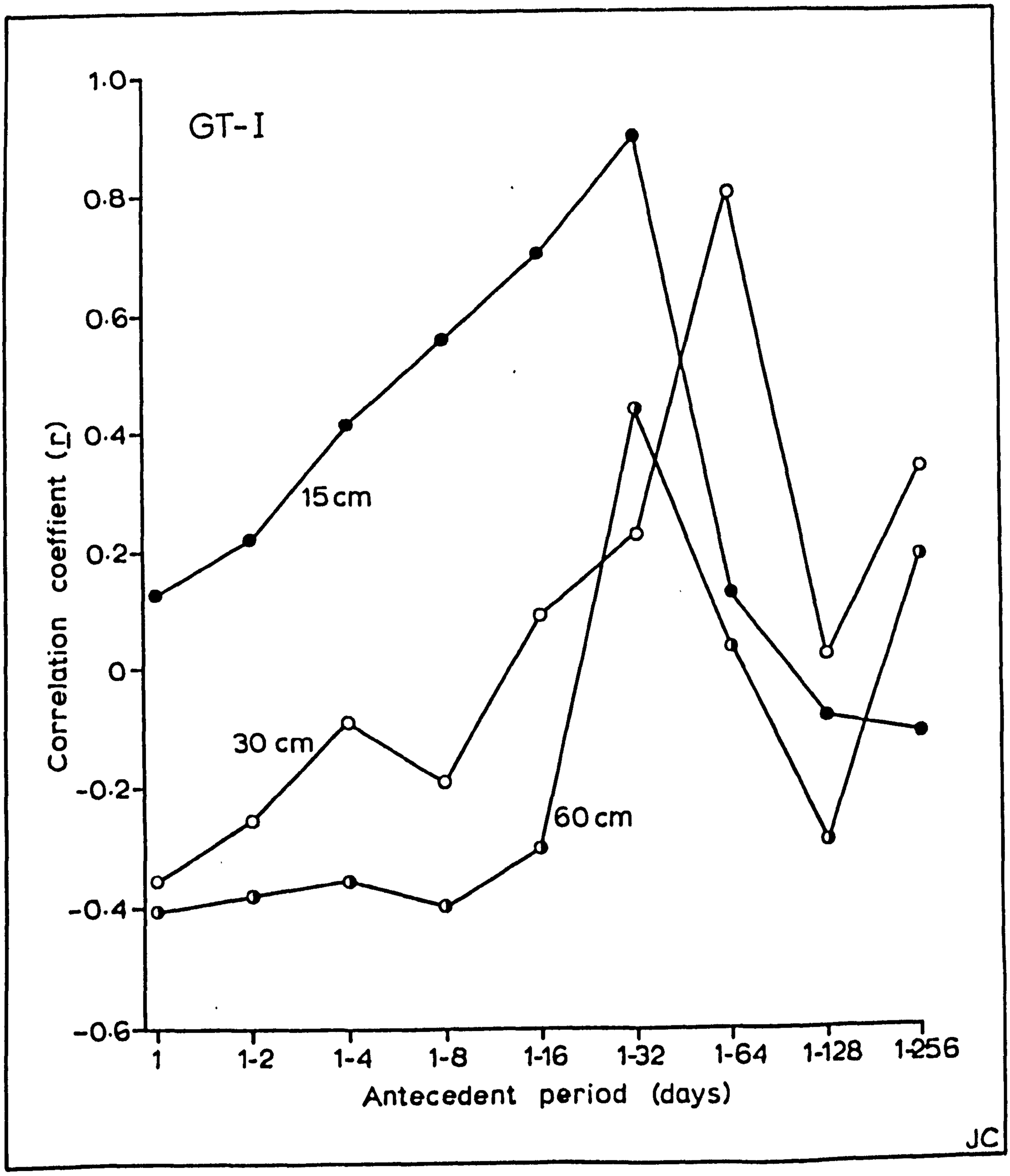


Figure 39 Correlation coefficients (r) for the relationship between soil carbon dioxide at various depths in the profiles studied in GT-I and rainfall during pre-set antecedent periods.

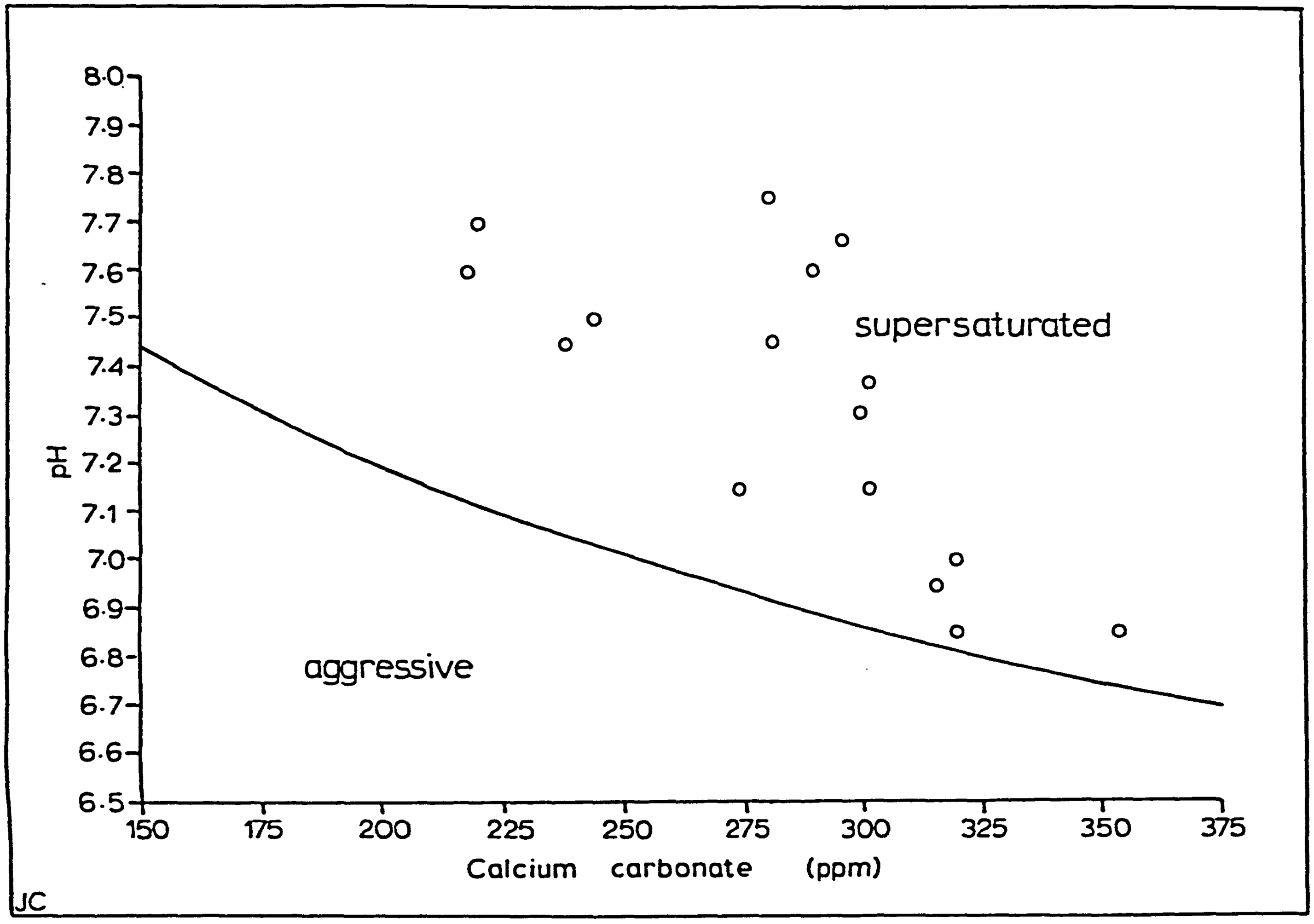


Figure 40 Plot of calcium hardness against pH for the water sample with the median hardness concentration for each of the subcutaneous-flow seepages in Wang Tangga. Equilbirium curves are for saturated calcite solutions at 25°C (after Picknett, 1972).

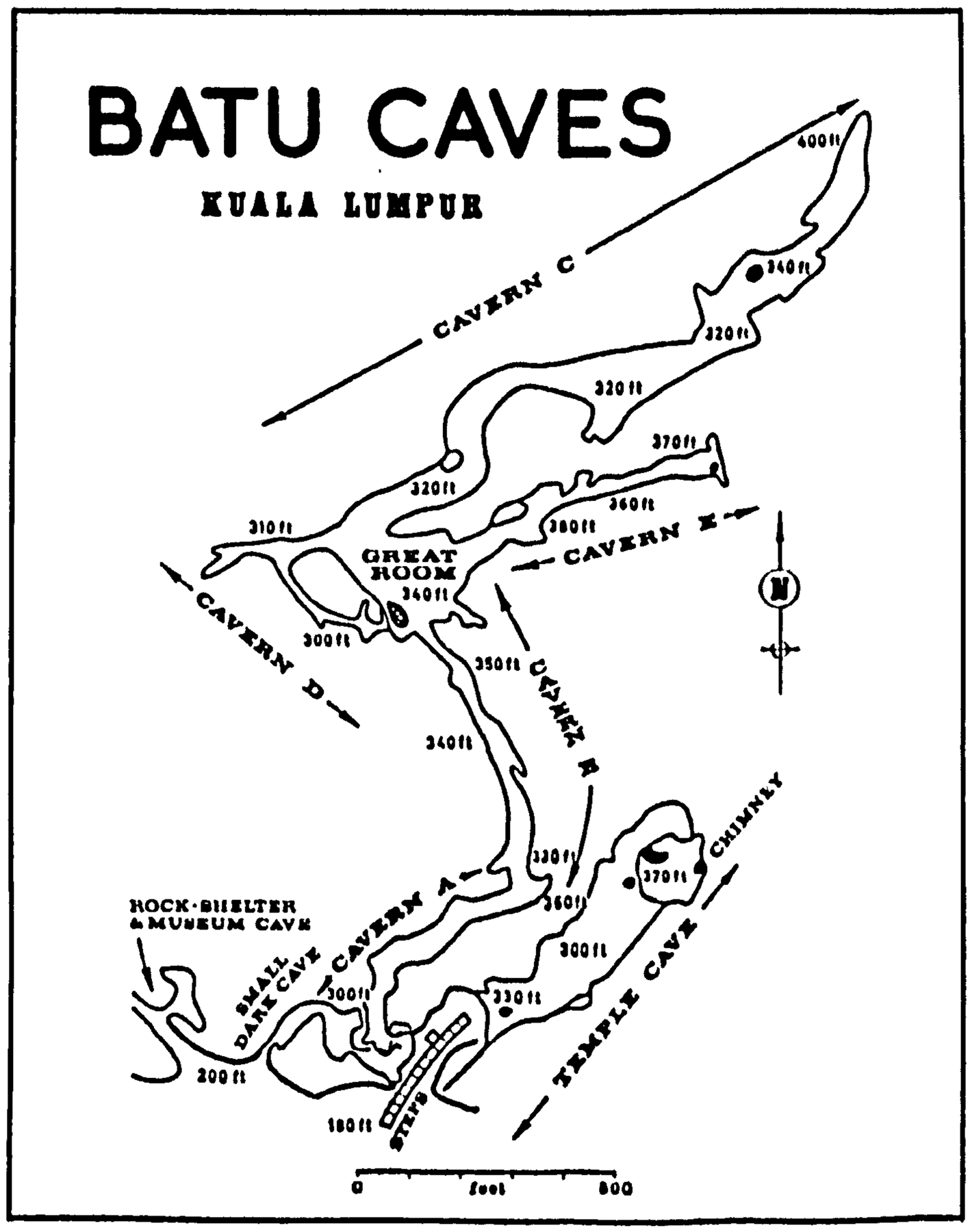


Figure 41 Survey of Gua Batu (from Wycherley, 1973, p.3).

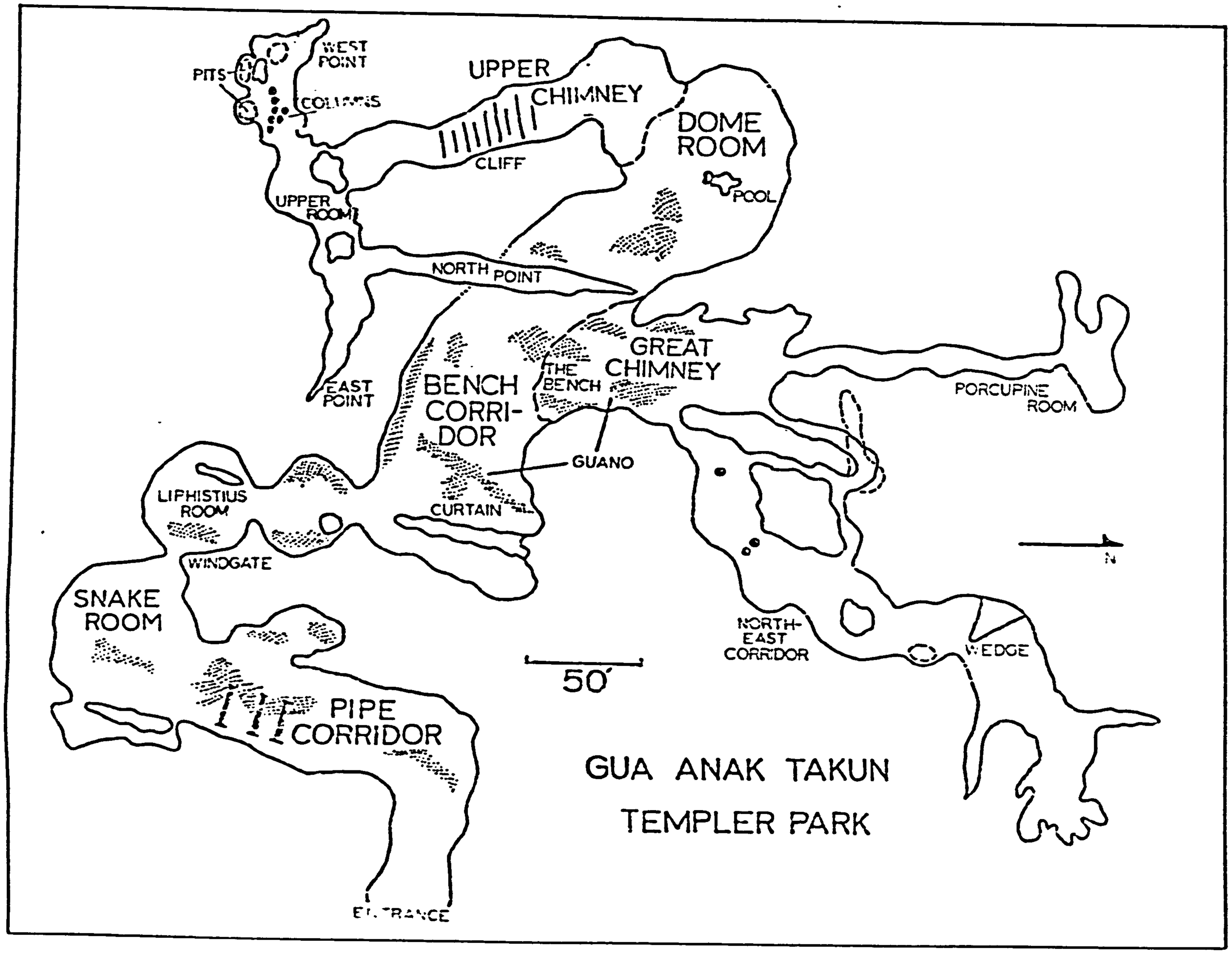


Figure 42

Survey of Gua Anak Takun (from Dunn, 1965).

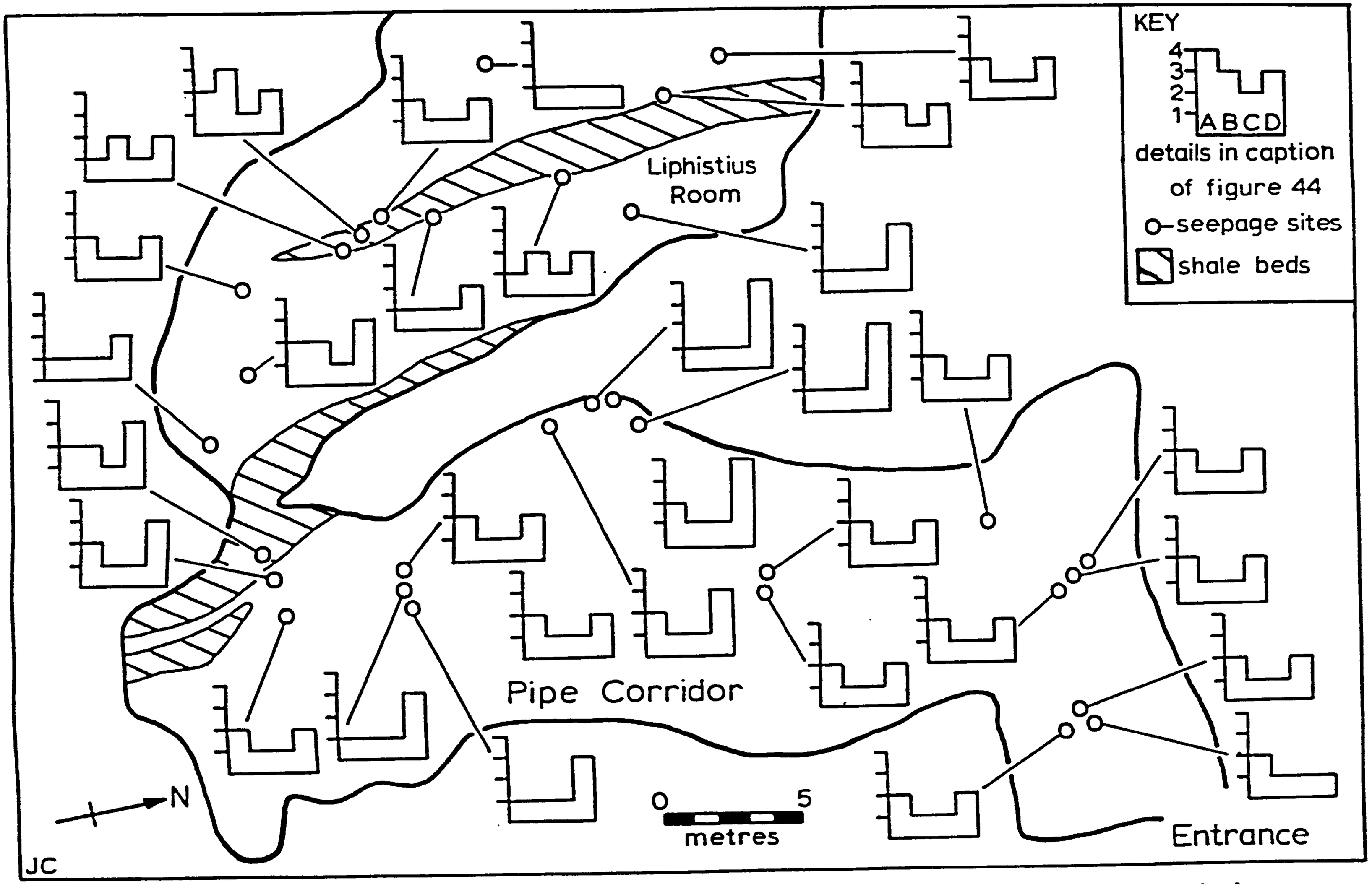


Figure 43 Variations in the chemical properties of diffuse-flow seepages in Pipe Corridor and Liphistius Room, Gua Anak Takun.

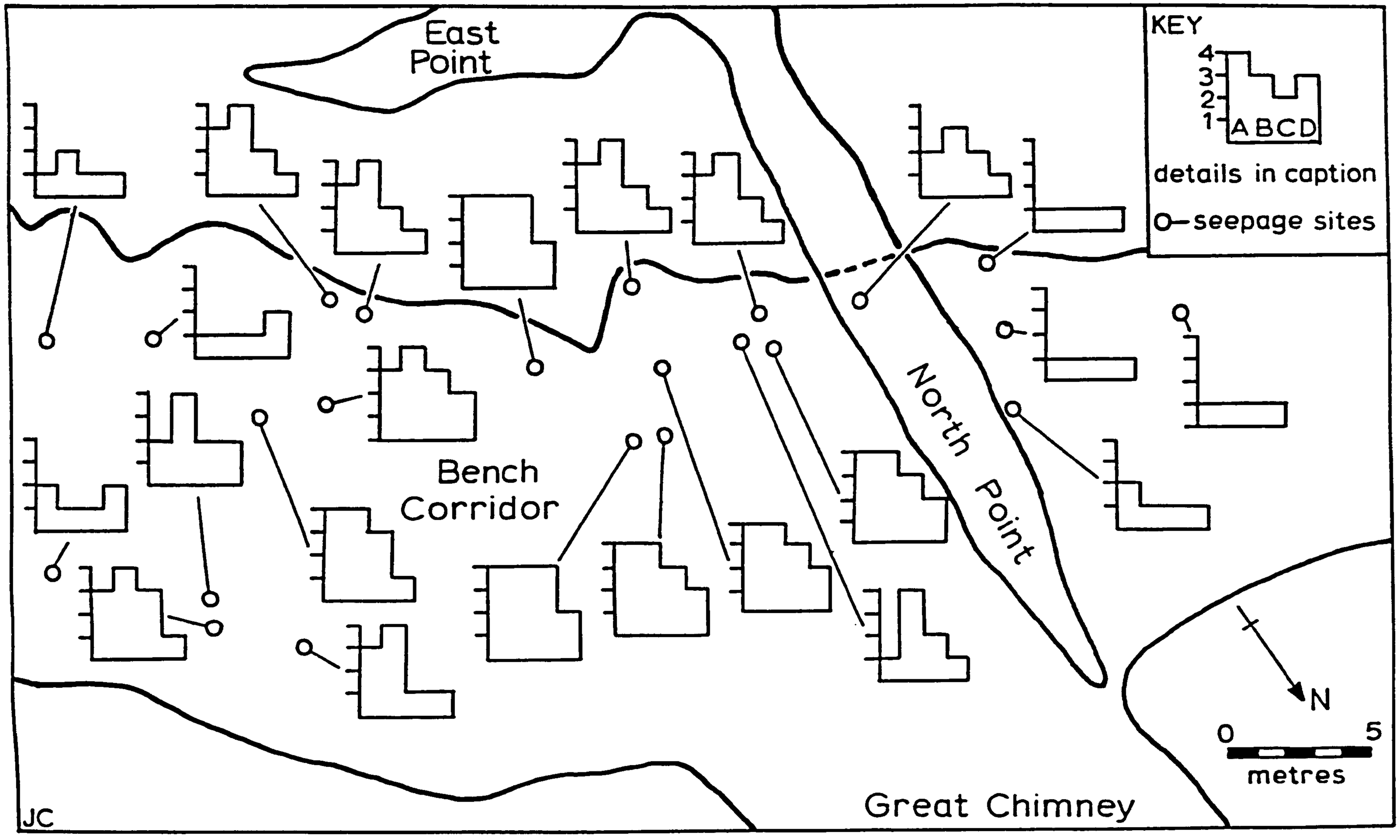


Figure 44 Variations in the chemical properties of diffuse-flow seepages in Bench Corridor, Gua Anak Takun: A. Total hardness (ppm) - 1. 0-99.9, 2. 100-199.9, 3. 200-399.9, 4. >400; B. Non-alkaline:Total hardness (%) - 1. 0-9.99, 2. 10-19.9, 3. 20-39.9, 4. >40; C. Potassium (ppm) - 1. 0-1.99, 2. 2-9.99, 3. 10-19.99, 4. >20; D. Mg:Ca (molar) - 1. 0-0.049, 2. 0.05-0.099, 3. 0.1-0.199, 4. > 0.2.

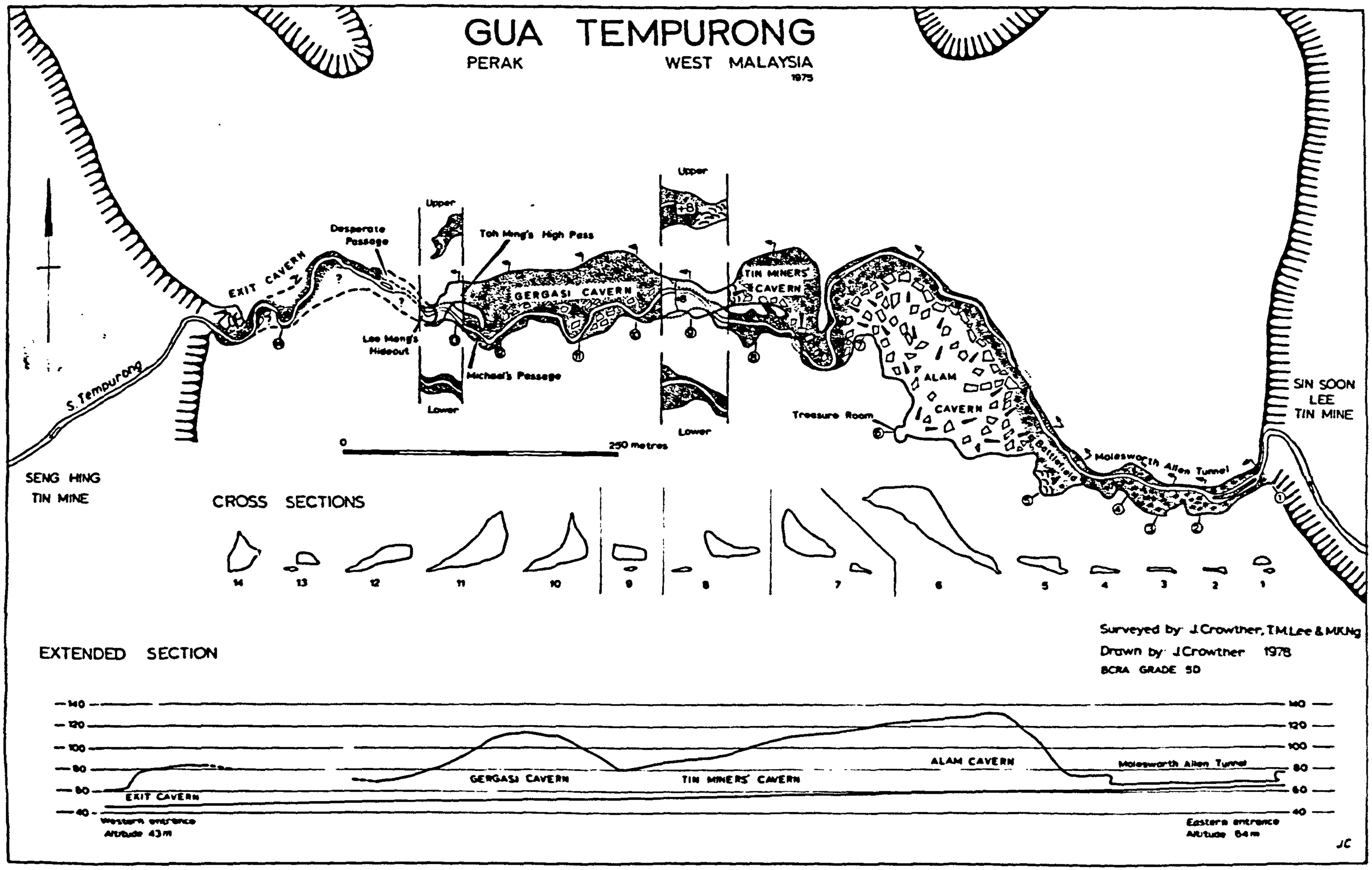


Figure 45 Survey of Gua Tempurong.

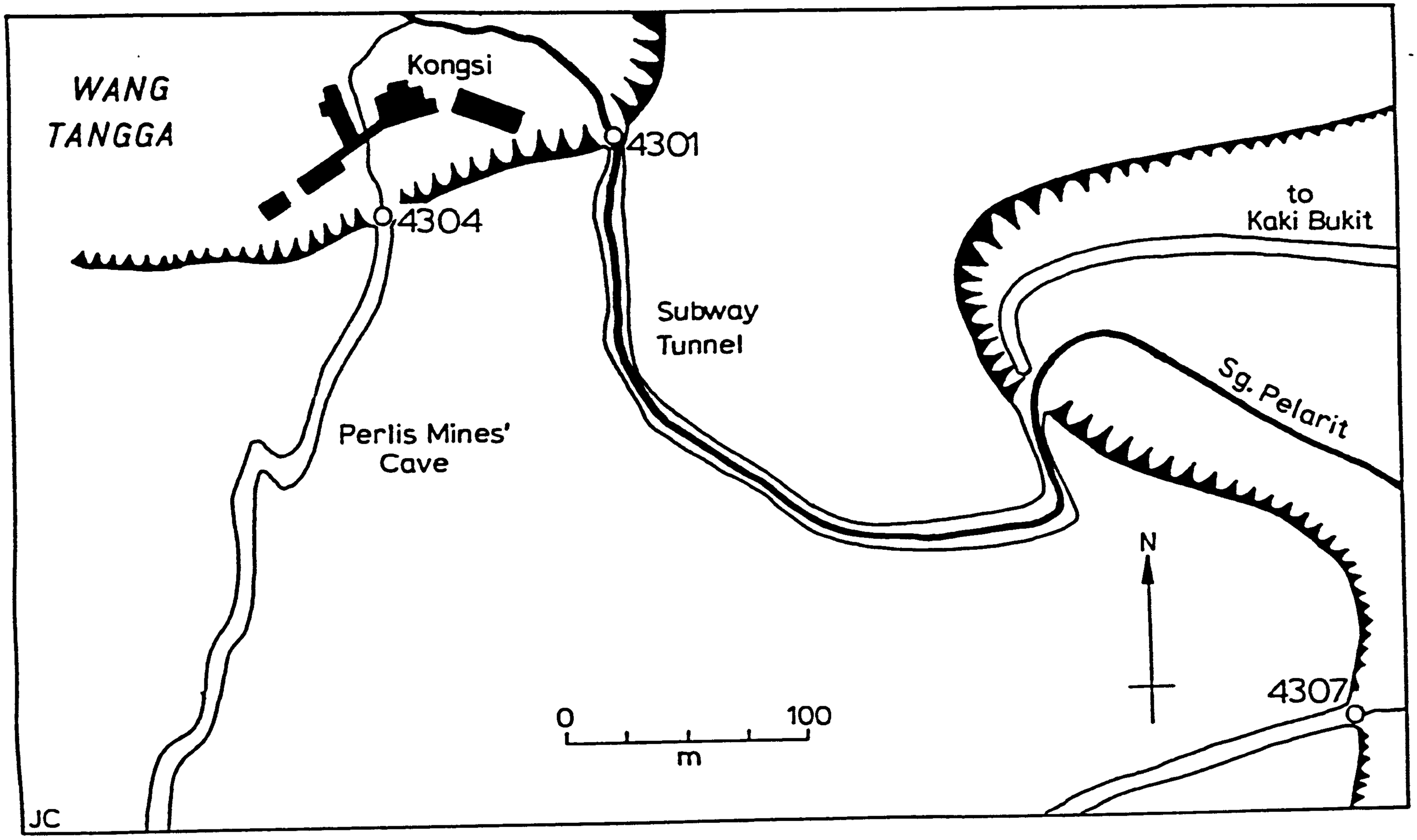


Figure 46 Survey of Perlis Mines' Cave/Subway Tunnel (after mine survey).

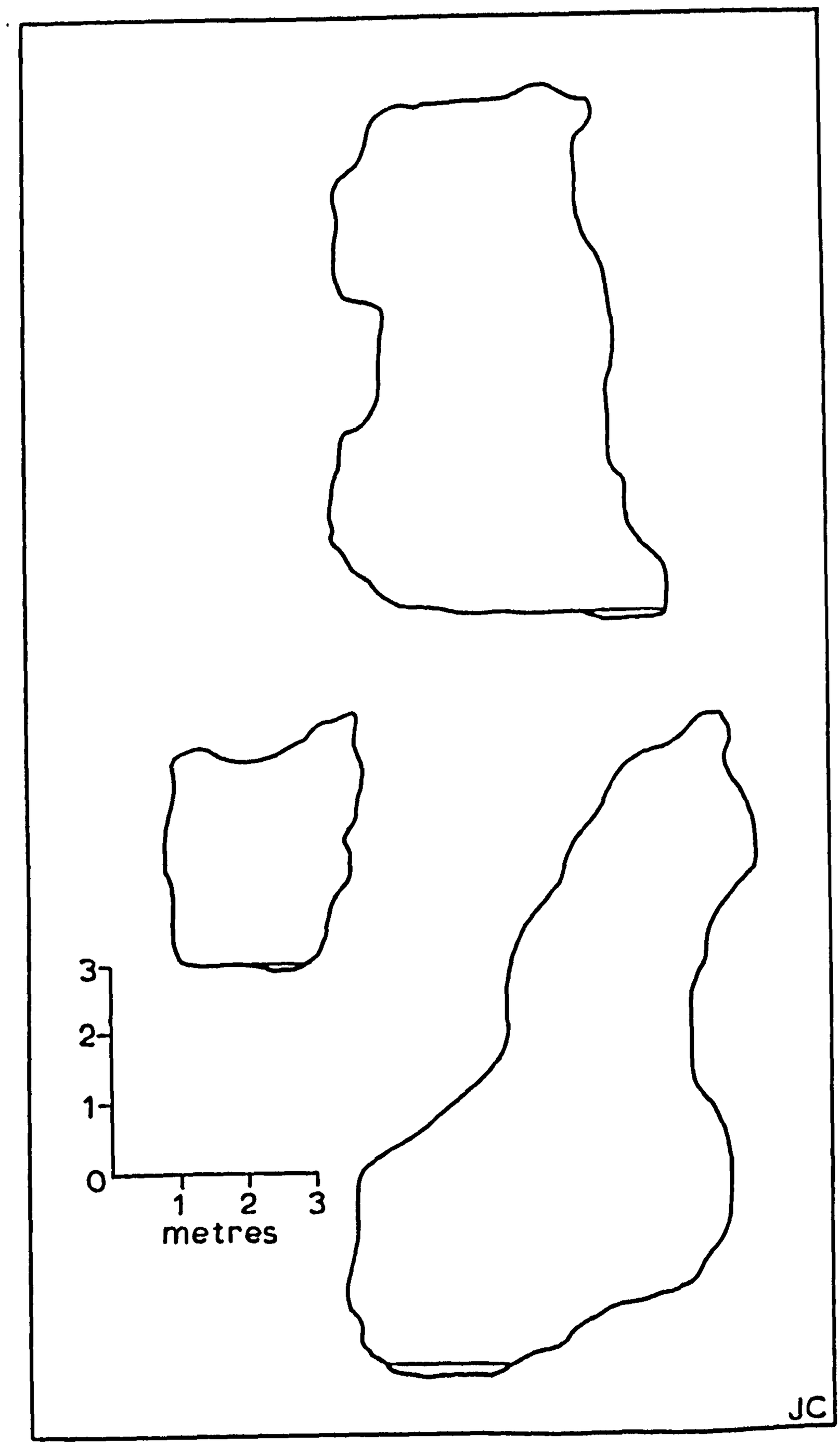


Figure 47 Selected cross-sections in Perlis Mines' Cave.

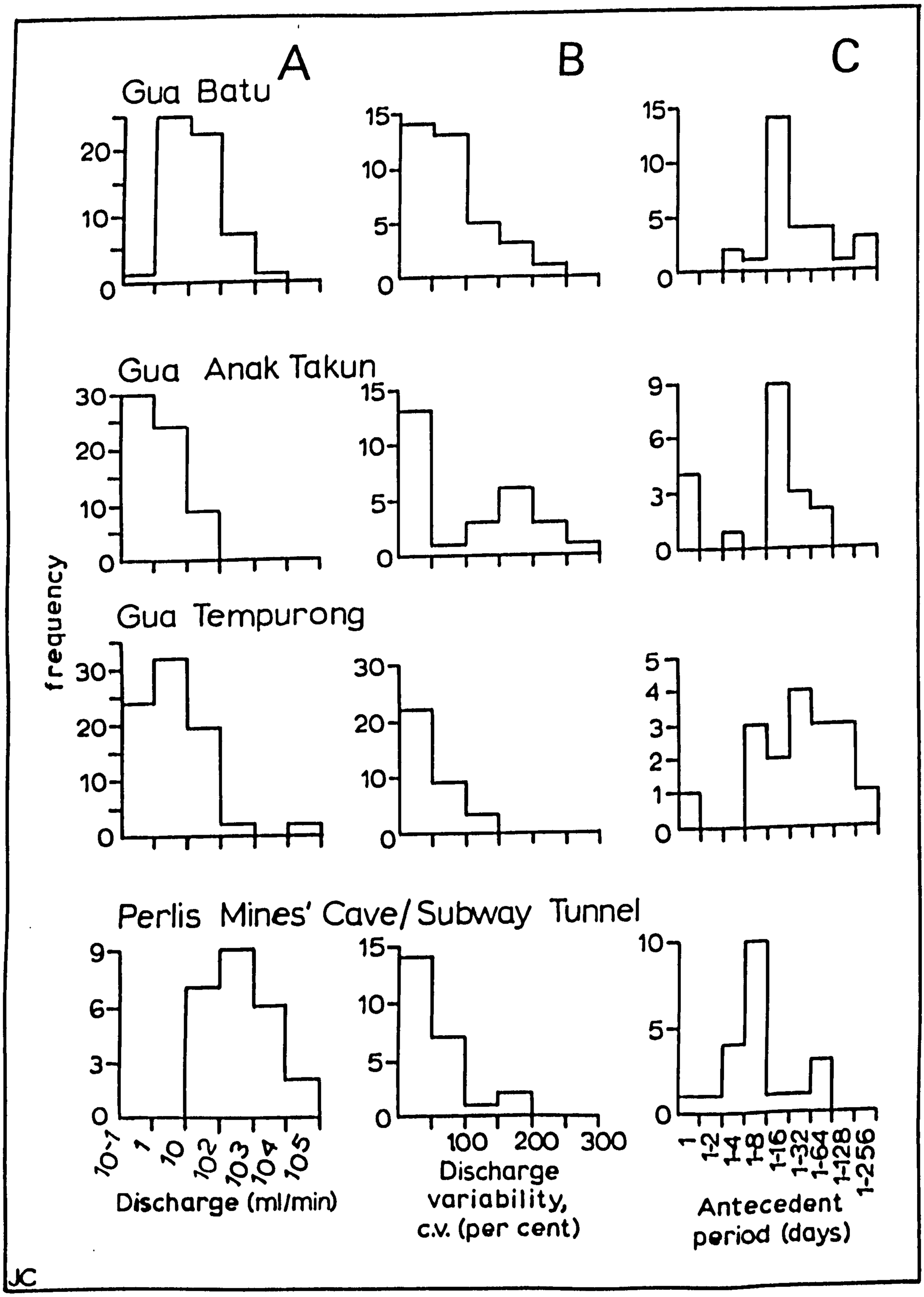


Figure 48 Hydrological characteristics of the four cave systems:
A. Discharge,
B. Discharge variability,
C. Antecedent rainfall period with which seepage discharge correlates most closely.

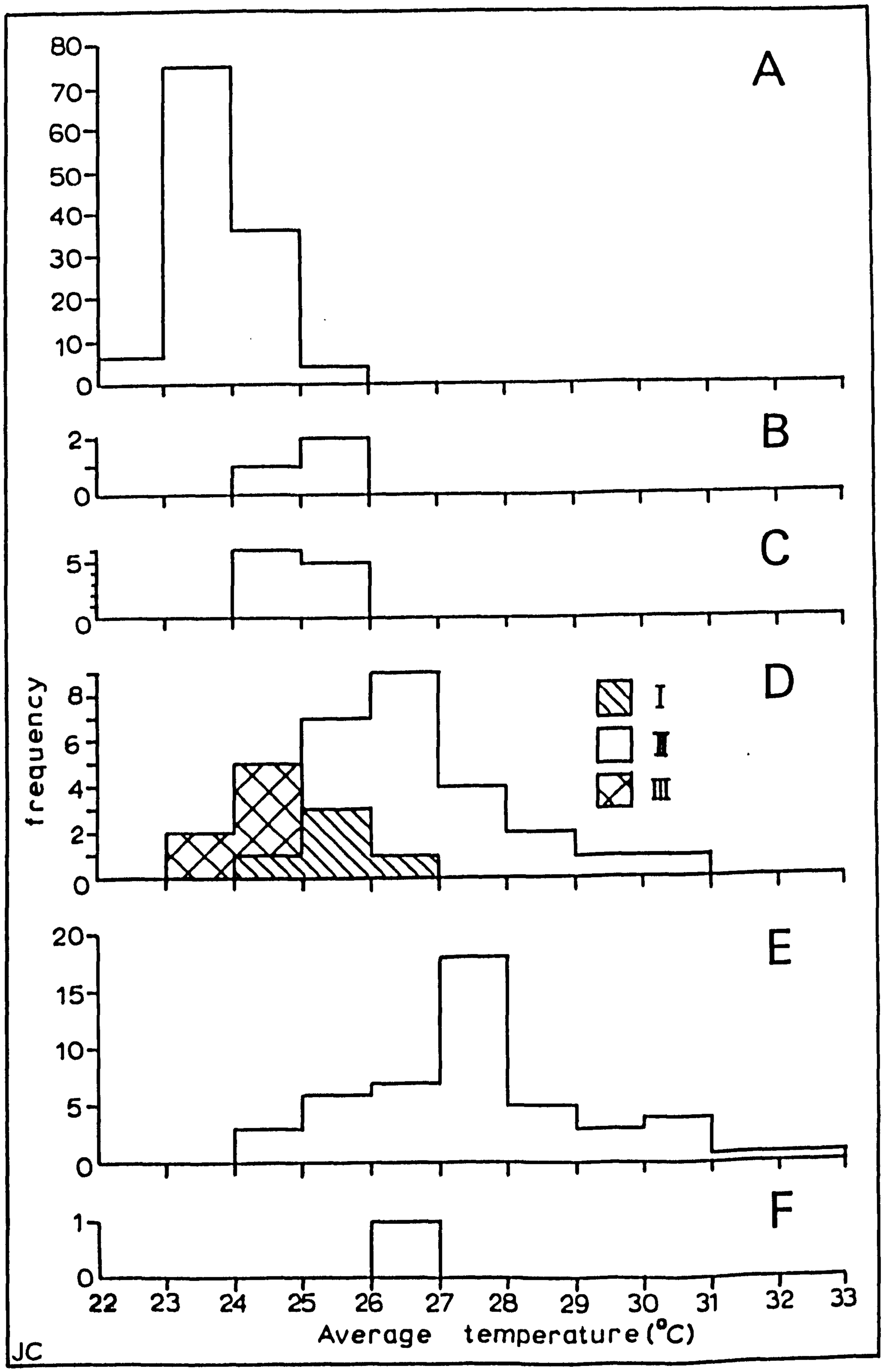


Figure 49 Average temperatures of waters in the three karst regions: A. Underground diffuse-flow seepages; B. Karst springs; C. Underground streams; D. Surface streams - I. Streams with significant groundwater contribution, II. Streams which derive much of their flow from the karst alluvial plains; III. Streams draining non-carbonate terrain. E. Seepages in alluvium of the karst plains; F. Deep groundwater.

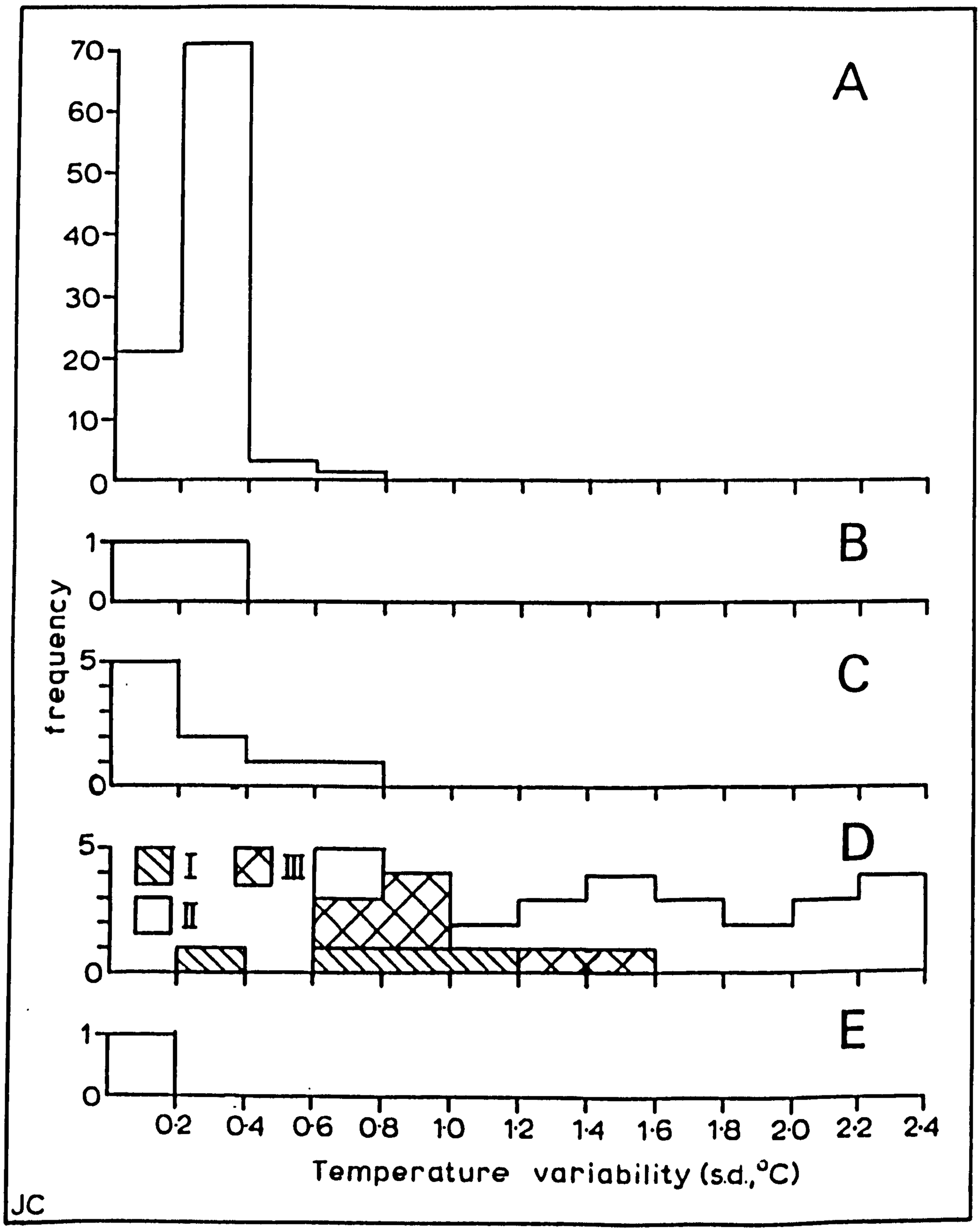


Figure 50 Water temperature variability in the three karst regions: A. Underground diffuse-flow seepages; B. Karst springs; C. Underground streams; D. Surface streams - I. Streams with significant groundwater contribution, II. Streams which derive much of their flow from the karst alluvial plains; III. Streams draining non-carbonate terrain. E. Deep groundwater.

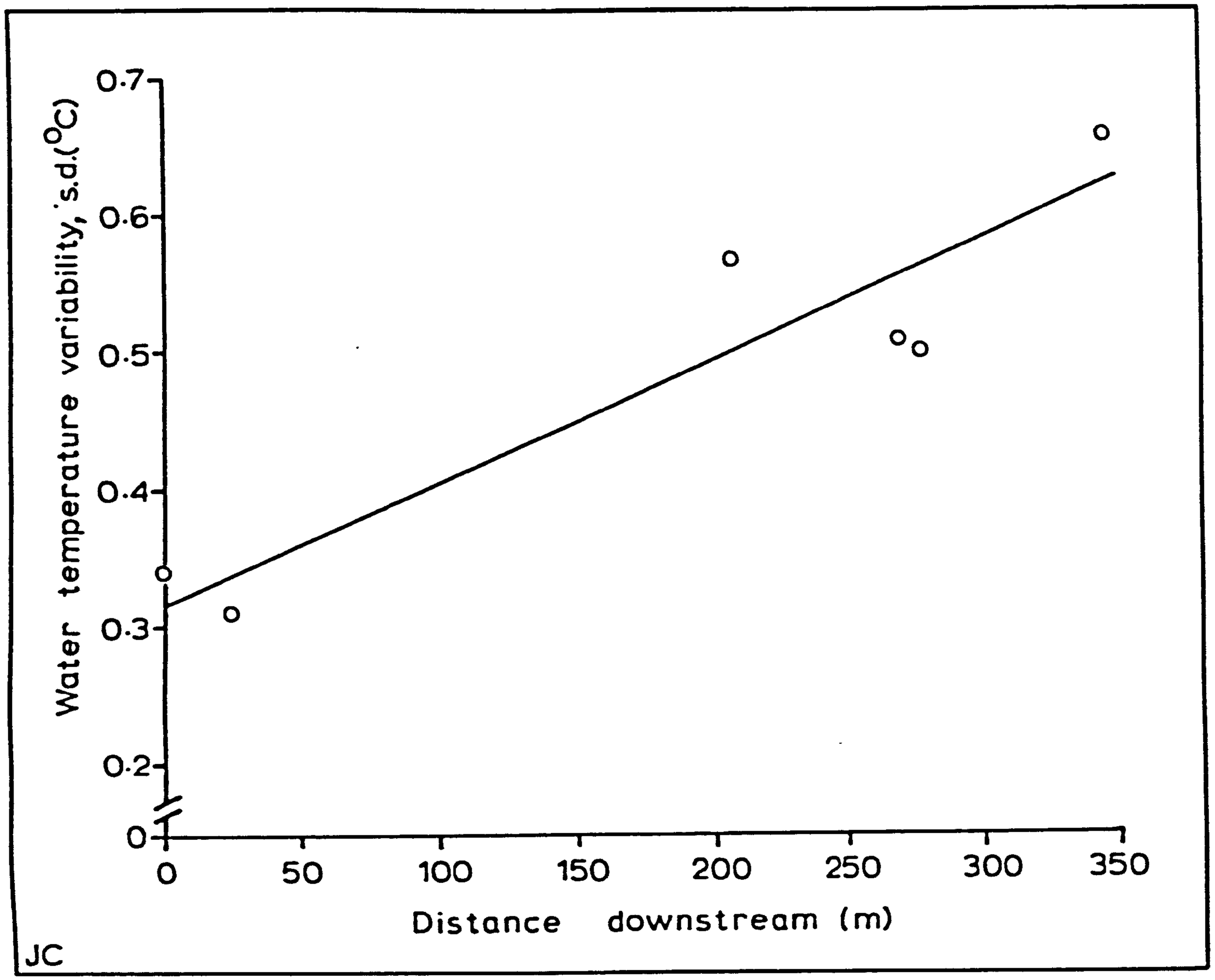


Figure 51 Changes in water temperature variability downstream from the main headwater rising of the Sg. Gelang Kerbau.

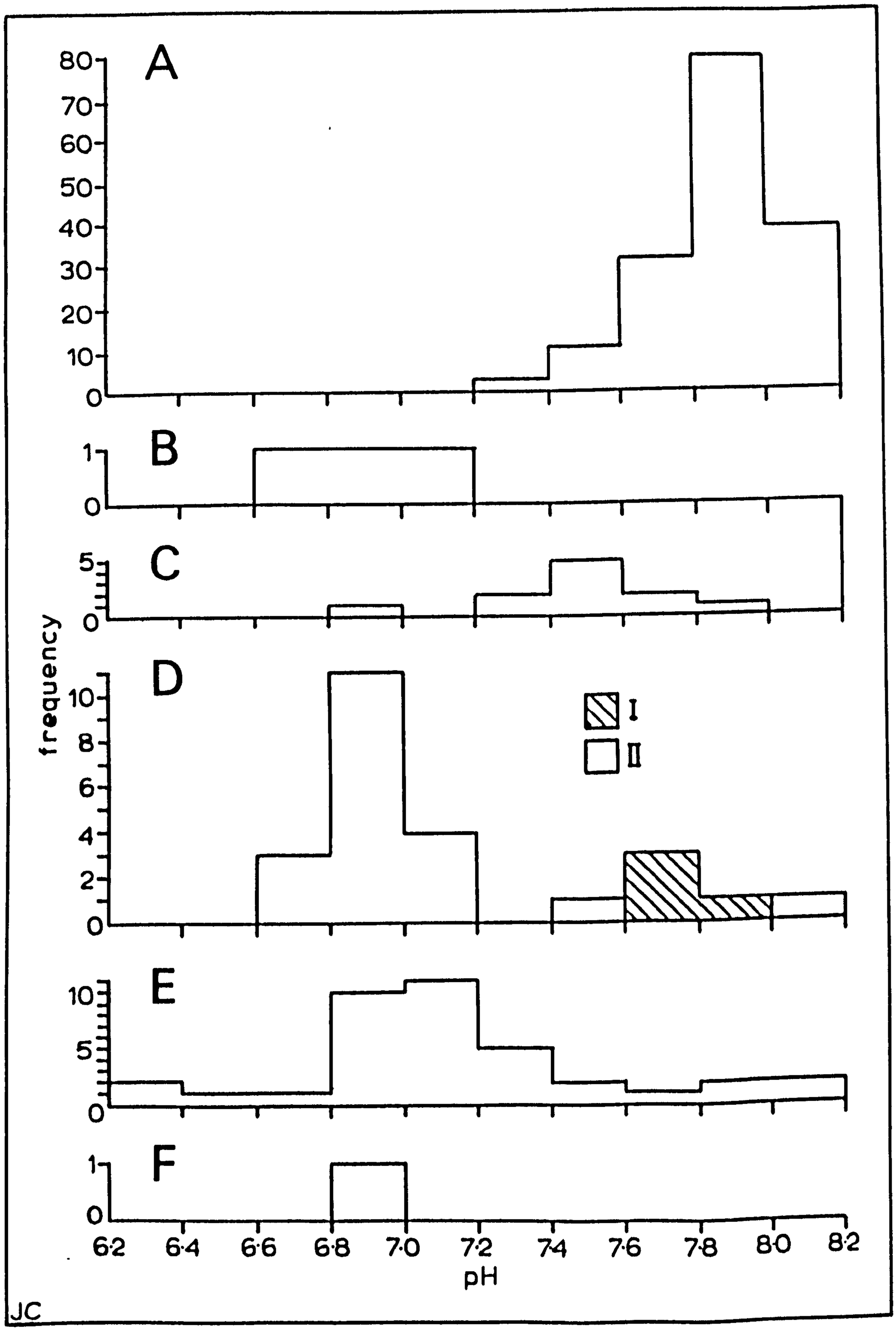


Figure 52 Average pH of karst waters in the three karst regions: A. Underground diffuse-flow seepages; B. Karst springs; C. Underground streams; D. Surface streams - I. Streams with significant groundwater contribution, II. Streams which derive much of their flow from the karst alluvial plains; E. Seepages in alluvium of the karst plains; F. Deep groundwater.

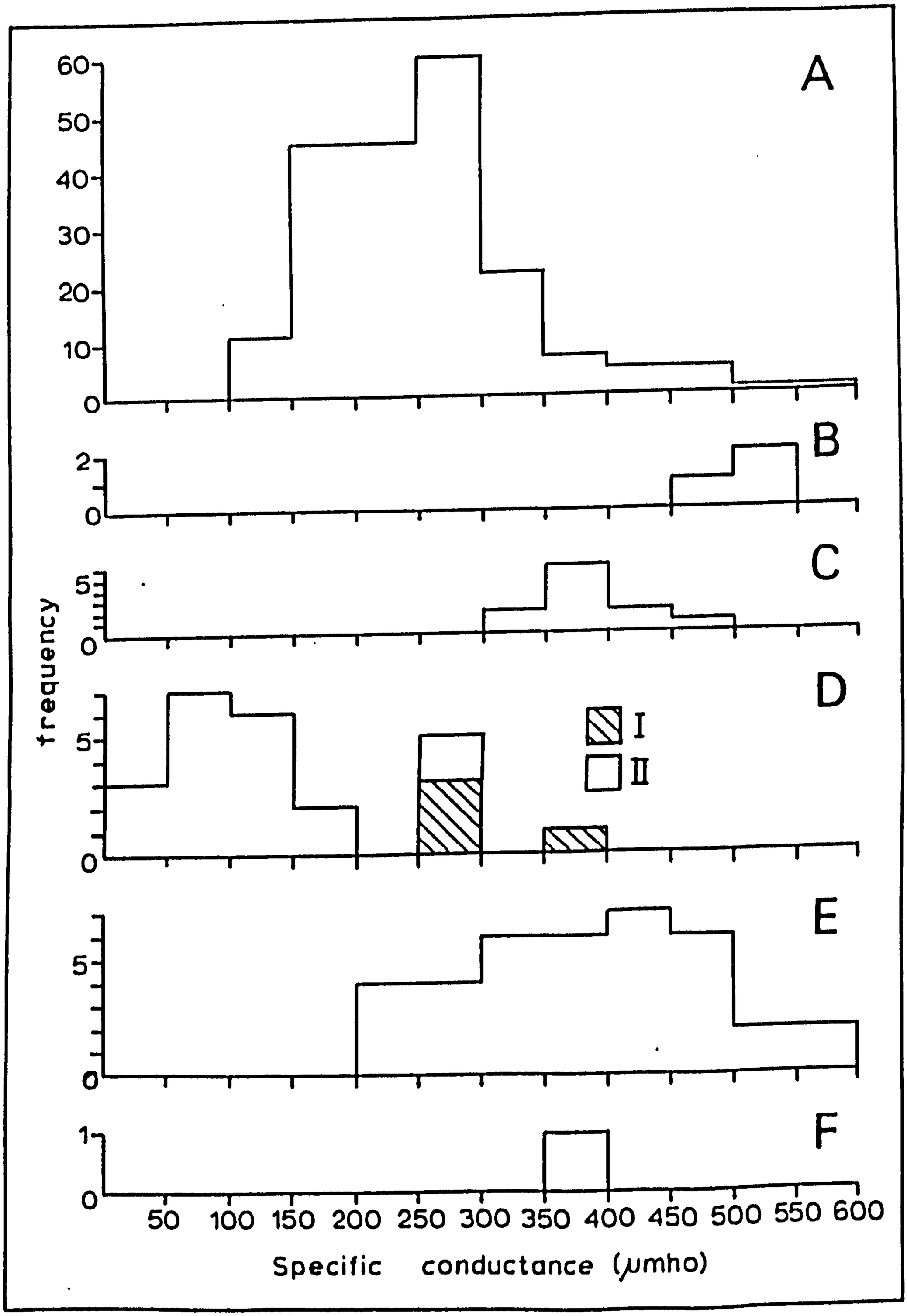


Figure 53 Average specific conductance of karst waters in the three field areas: A.Underground diffuse-flow seepages; B. Karst springs; C. Underground streams; D. Surface streams - I. Streams with a significant groundwater contribution, II. Streams which derive much of their flow from the karst alluvial plains; E. Seepages in alluvium of the karst plains; F. Deep groundwater.

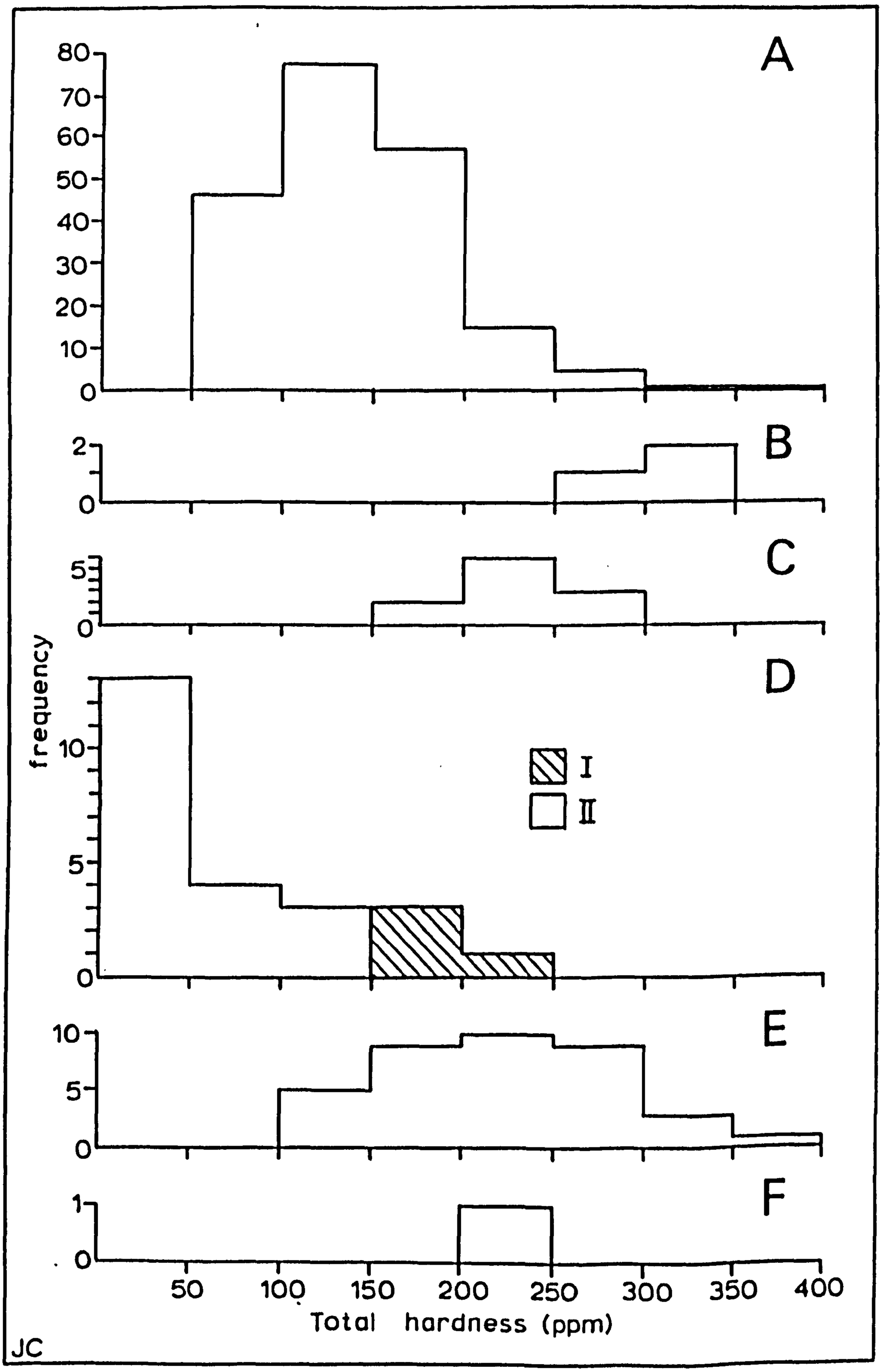


Figure 54 Average total hardness of karst waters in the three field areas: A. Underground diffuse-flow seepages; B. Karst springs; C. Underground streams; D. Surface streams - I. Streams with a significant groundwater contribution, II. Streams which derive much of their flow from the karst alluvial plains; E. Seepages in alluvium of the karst plains; F. Deep groundwater.

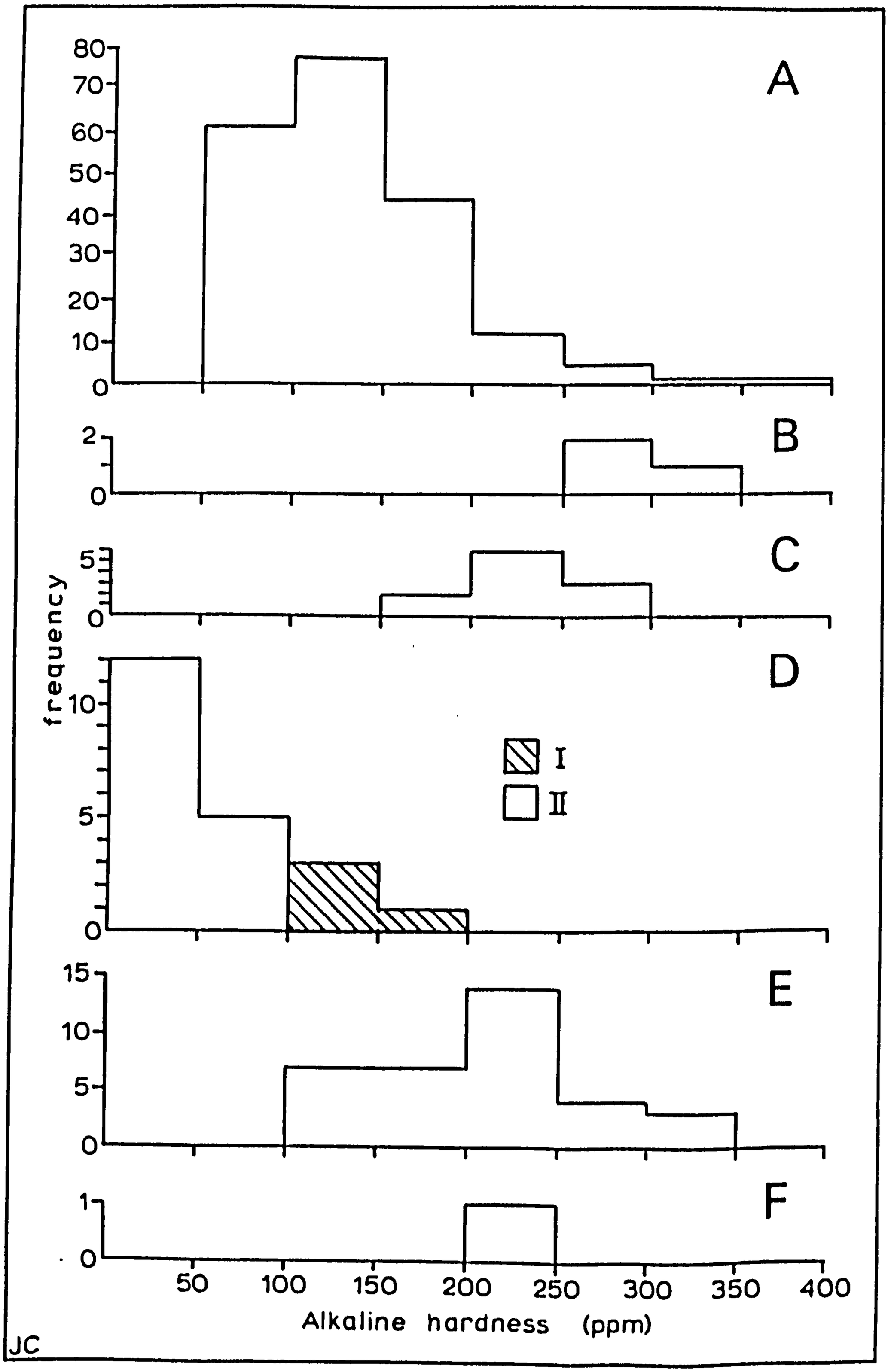


Figure 55 Average alkaline hardness of karst waters in the three field areas: A. Underground diffuse-flow seepages; B. Karst springs; C. Underground streams; D. Surface streams :- I. Streams with a significant groundwater contribution, II. Streams which derive much of their flow from the karst alluvial plains; E. Seepages in alluvium of the karst plains; F. Deep groundwater.

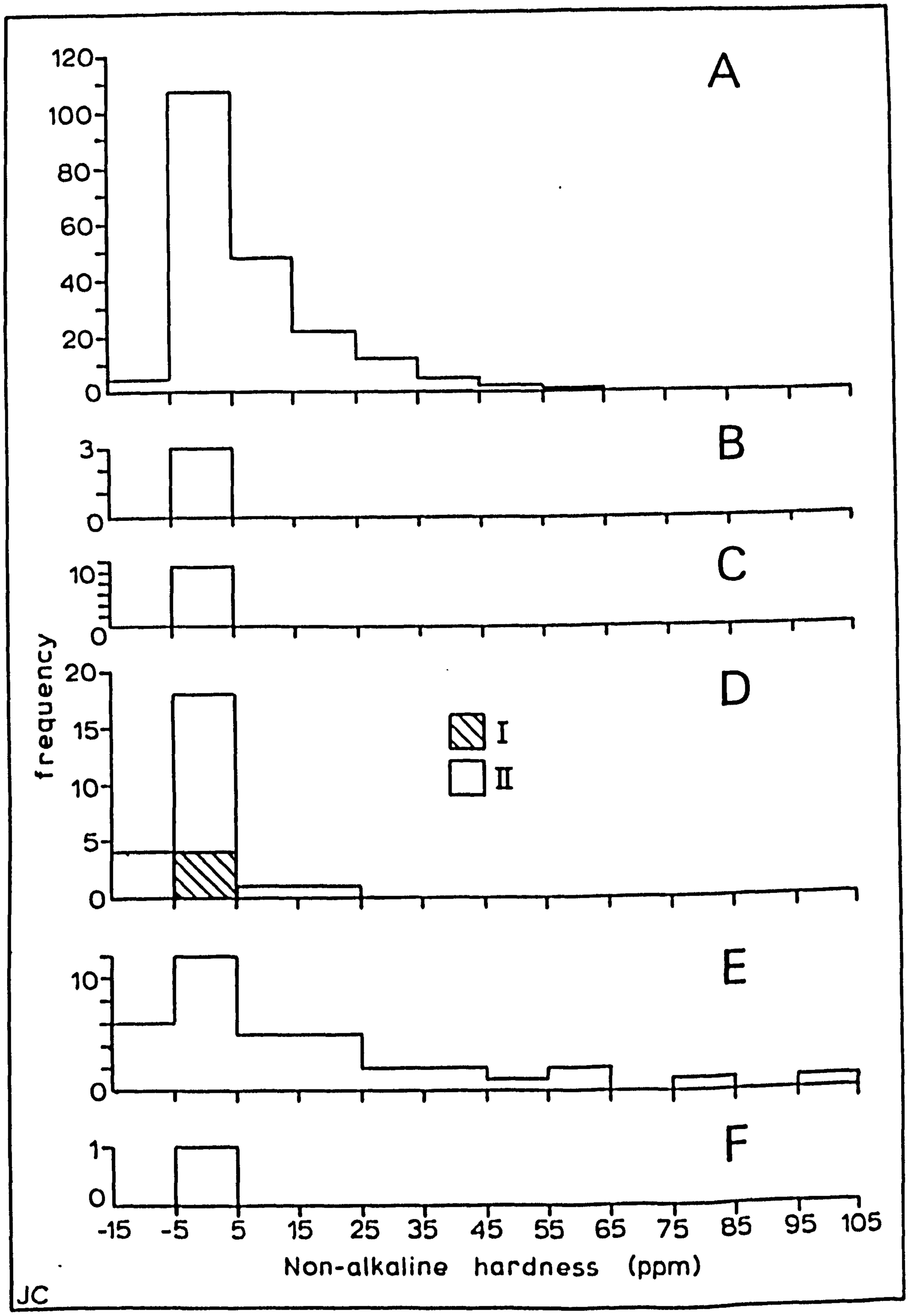


Figure 56 Average non-alkaline hardness of karst waters in the three field areas: A. Underground diffuse-flow seepages; B. Karst springs; C. Underground streams; D. Surface streams - I. Streams with a significant groundwater contribution, II. Streams which derive much of their flow from the karst alluvial plains; E. Seepages in alluvium of the karst plains; F. Deep groundwater.

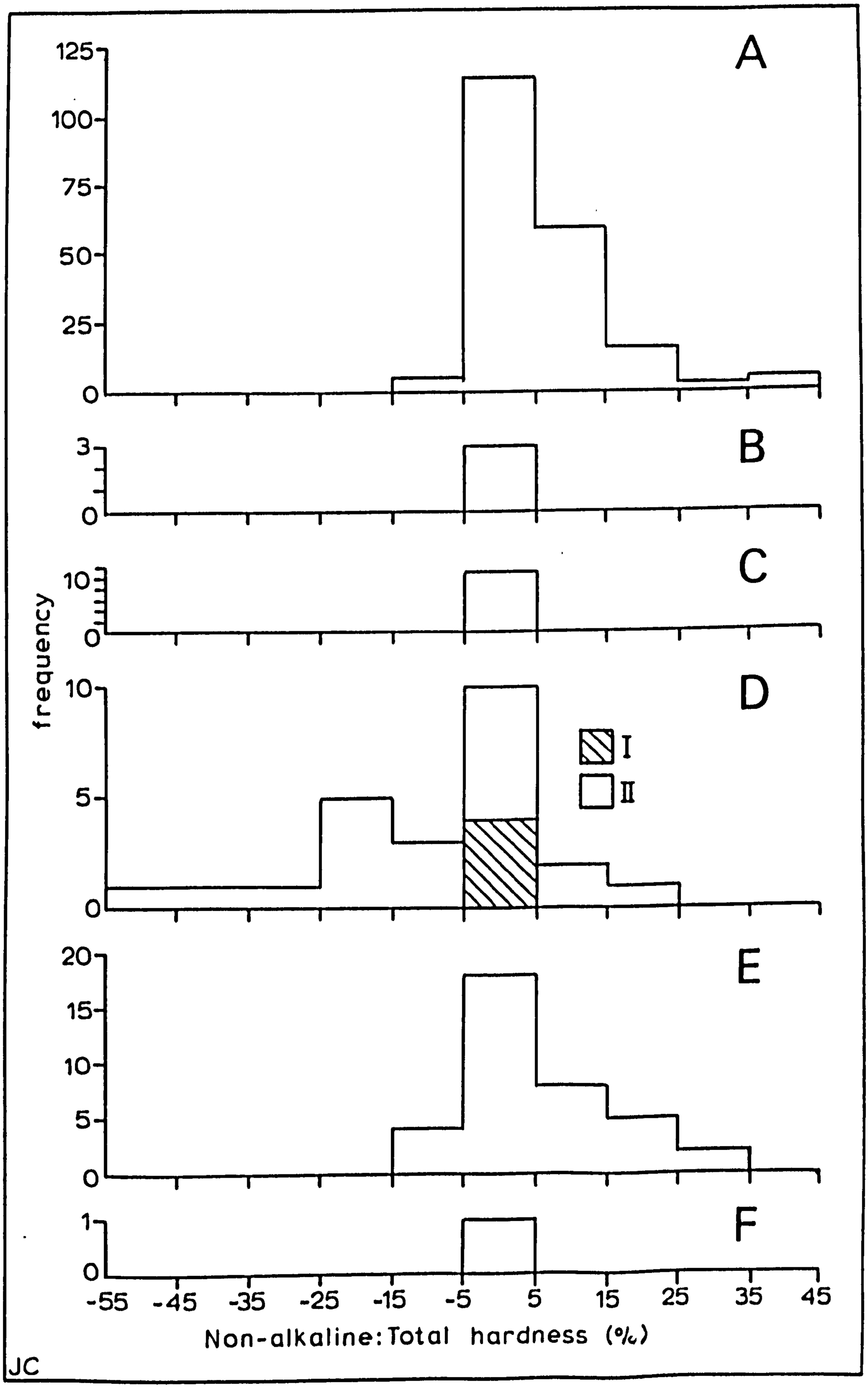


Figure 57 Ratio of non-alkaline:total hardness (%) of karst waters in the three field areas: A. Underground diffuse-flow seepages; B. Karst springs; C. Underground streams; D. Surface streams - I. Streams with a significant groundwater contribution, II. Streams which derive much of their flow from the karst alluvial plains; E. Seepages in alluvium of the karst plains; F. Deep groundwater.

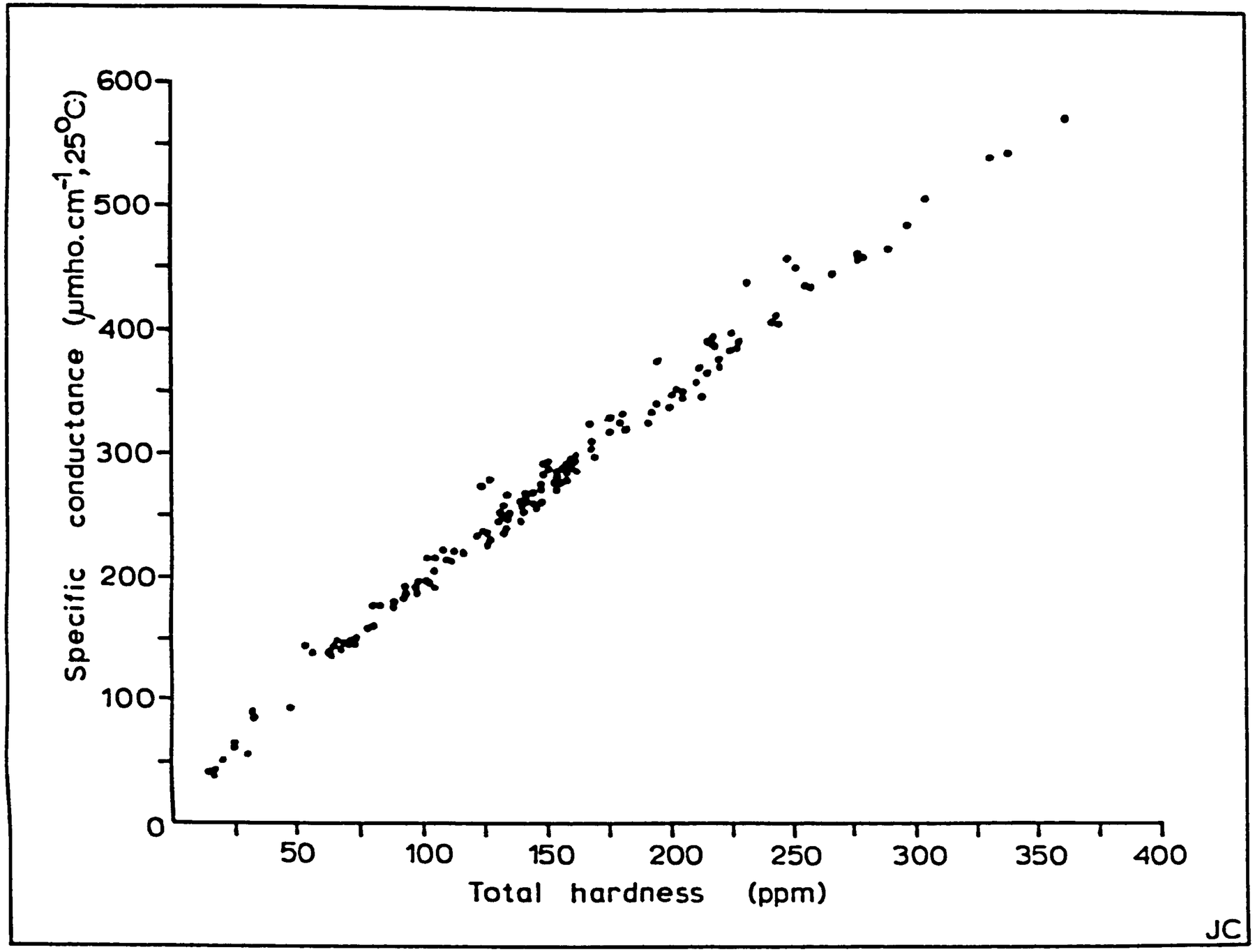


Figure 58 Relationship between total hardness and specific conductance for the karst waters sampled by the author in the three field areas.

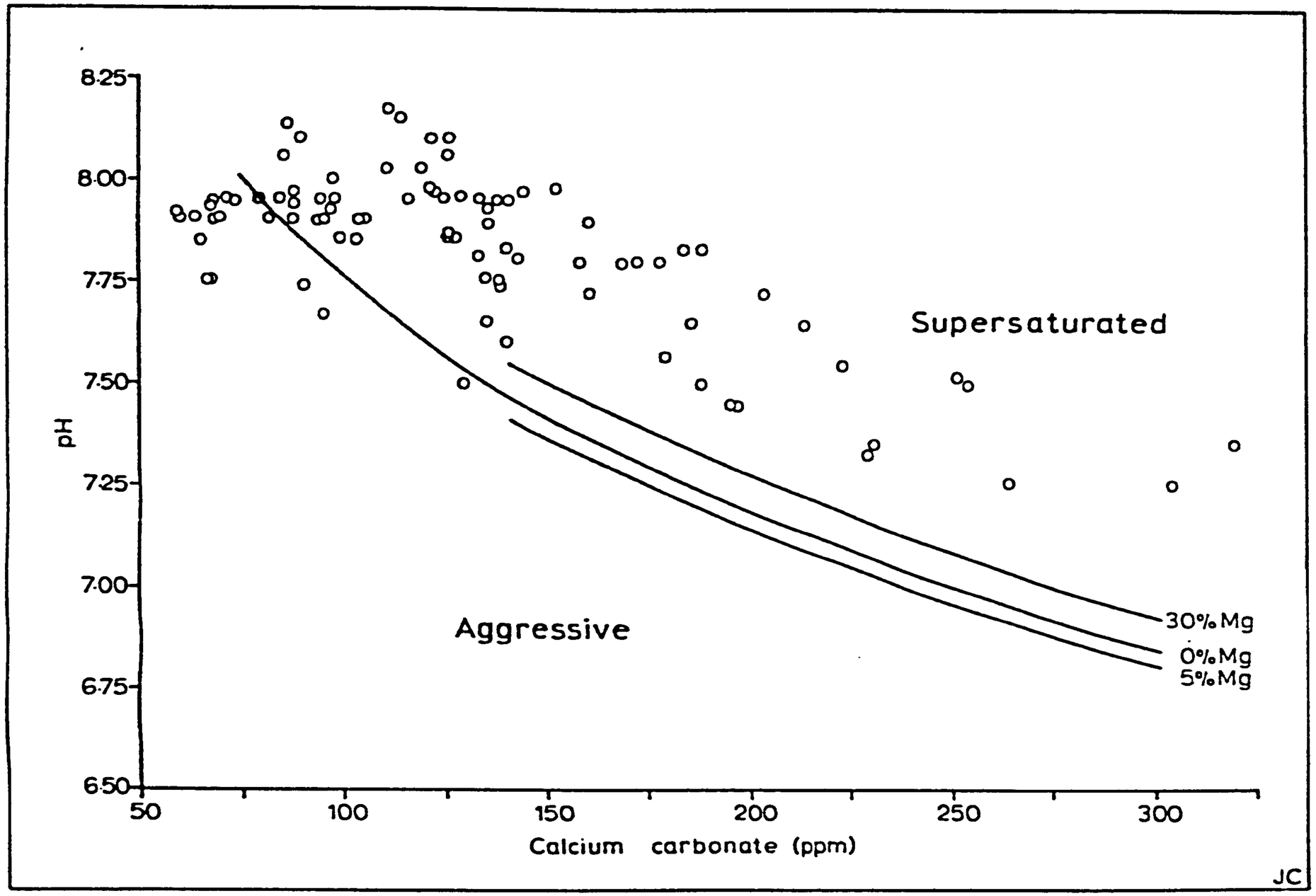


Figure 59 Plot of calcium hardness against pH for the water sample with the median hardness concentration for each of the uncontaminated, underground, diffuse-flow seepages with Mg:Ca ratios of less than 30 per cent. Equilibrium curves are for saturated calcite solutions at 25°C (after Picknett, 1972).

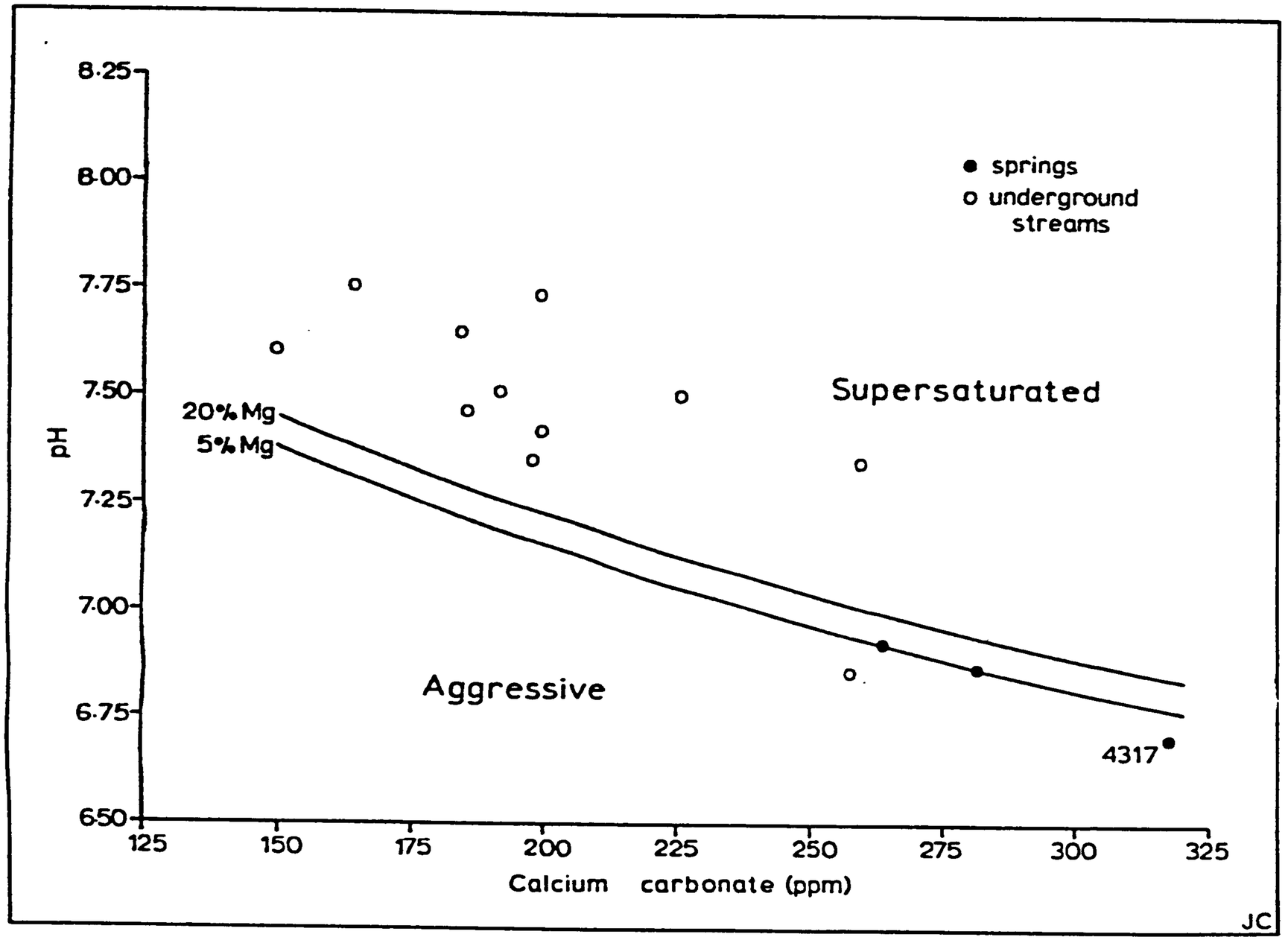


Figure 60 Plot of calcium hardness against pH for the water sample with the median hardness concentration for each of the karst springs and underground streams. Equilibrium curves are for saturated calcite solutions at 25°C (after Picknett, 1972).

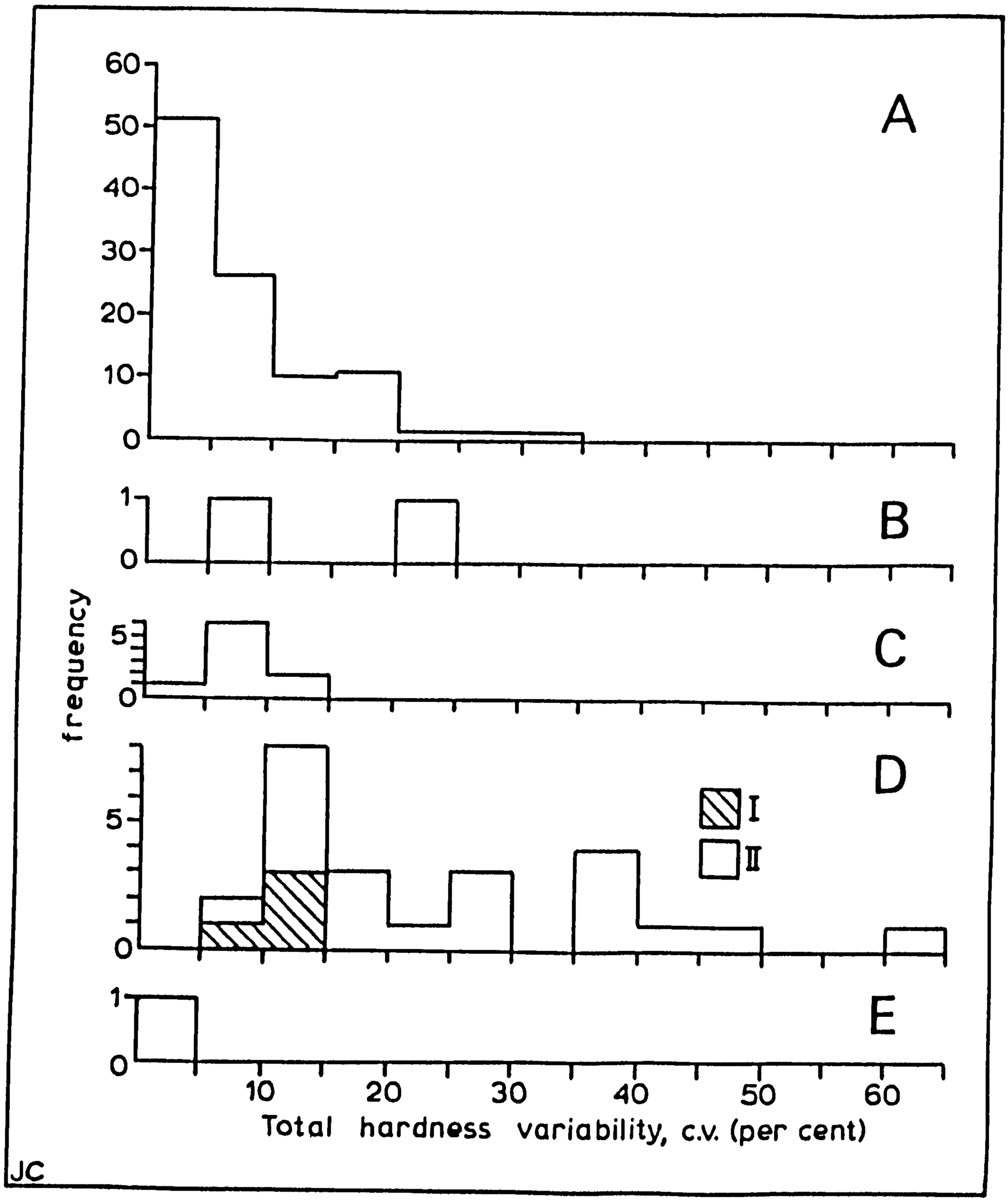


Figure 61 Total hardness variability of karst waters in the three field areas: A. Underground diffuse-flow seepages; B. Karst springs; C. Underground streams; D. Surface streams - I. Streams with a significant groundwater contribution, II. Streams which derive much of their flow from the karst alluvial plains; E. Seepages in alluvium of the karst plains; F. Deep groundwater.

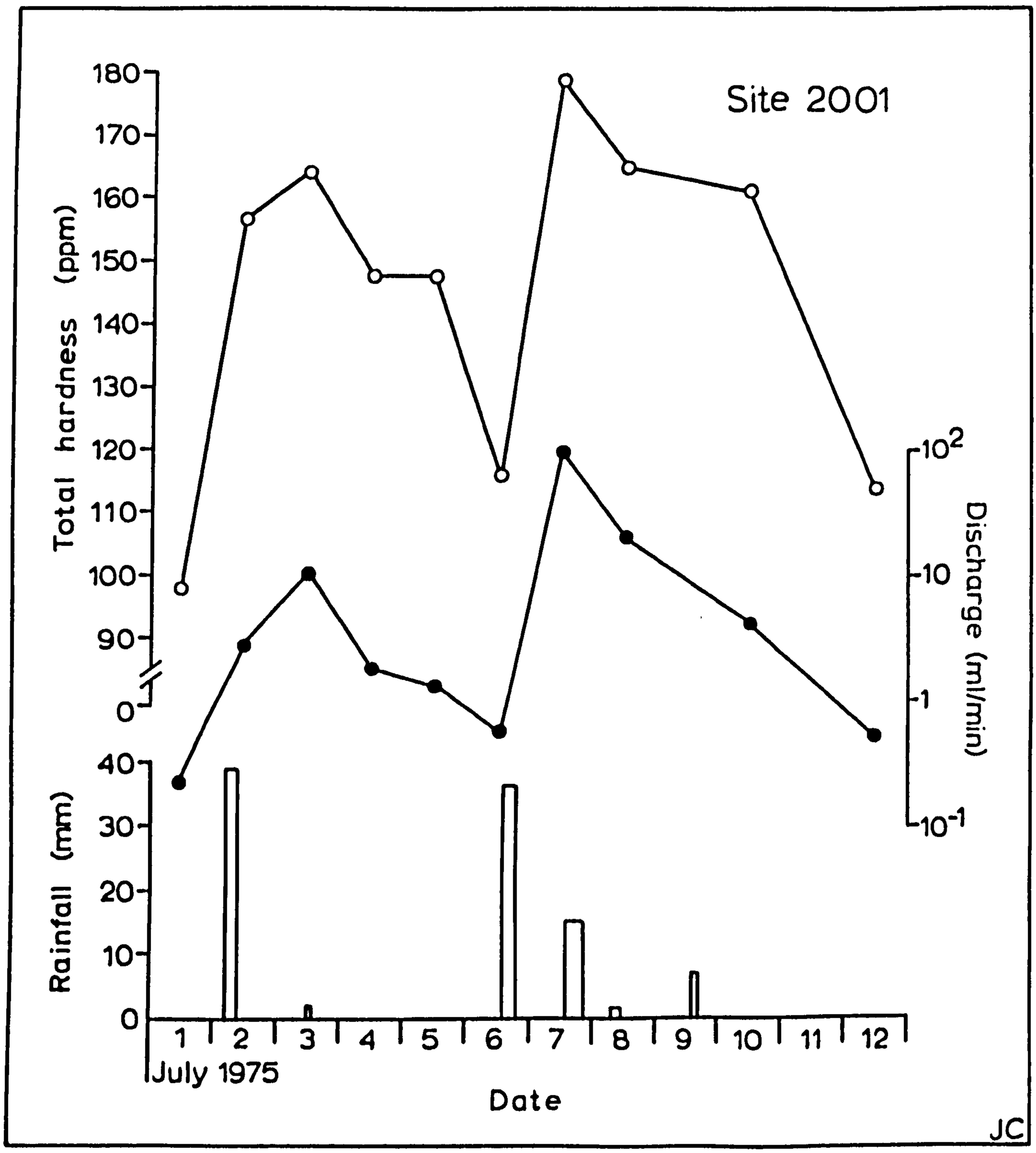


Figure 62 Variations in discharge and total hardness at site 2001 over a 12-day period.

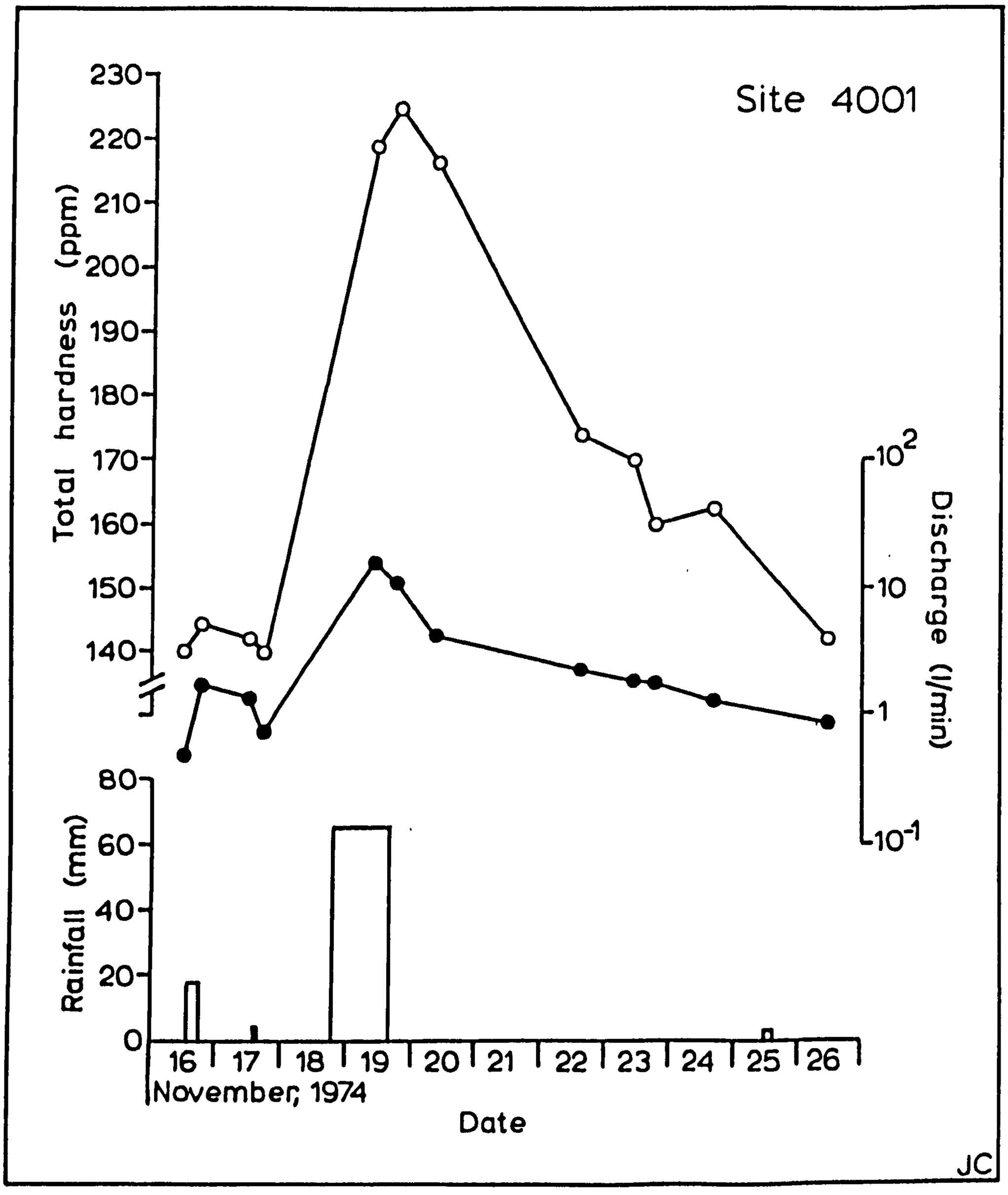


Figure 63 Variations in discharge and total hardness at site 4001 over an 11-day period.

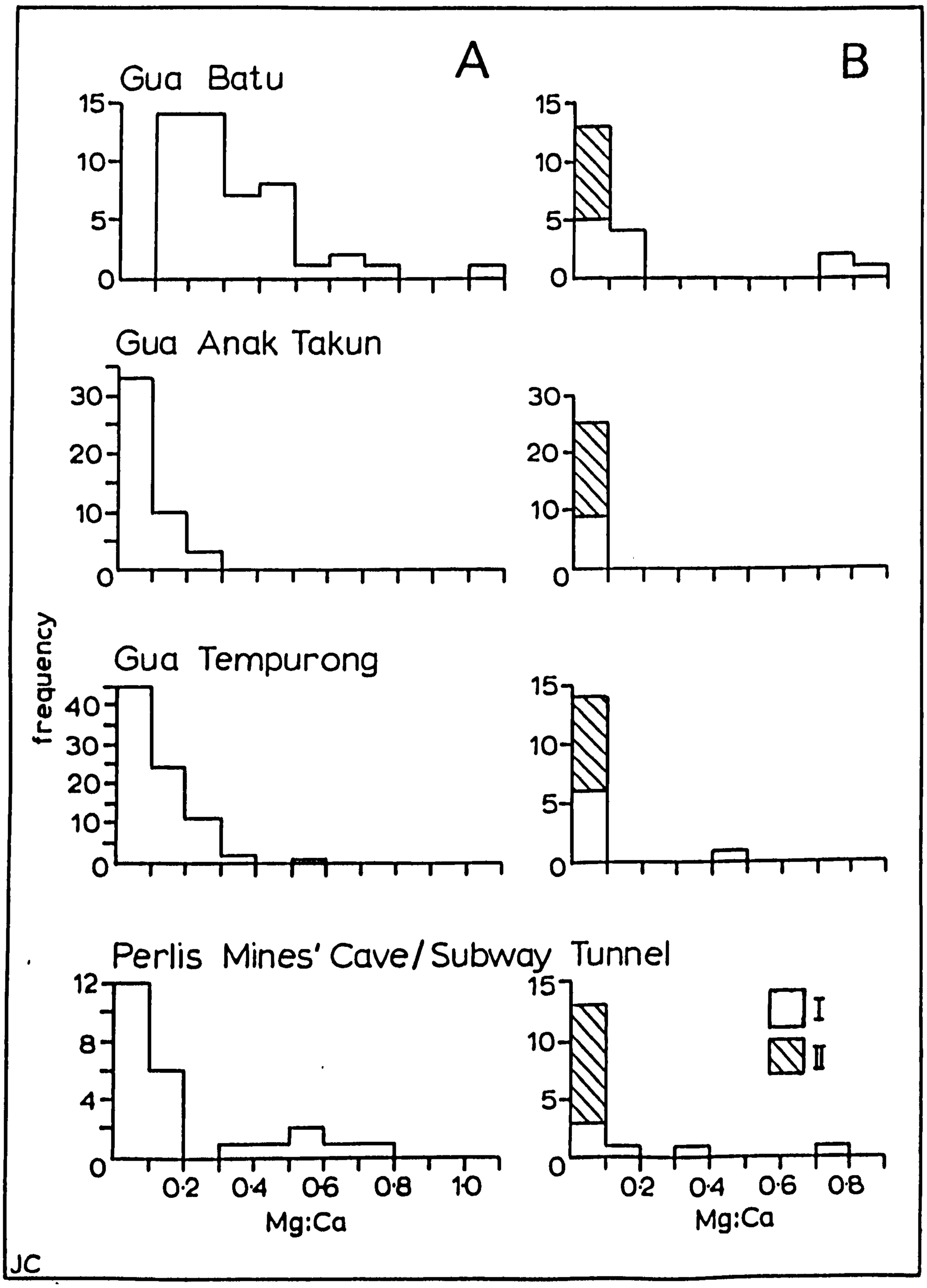


Figure 64 Mg:Ca ratios of karst waters and limestones in the three field areas:
A. Diffuse-flow seepages,
B. Limestones - I. Bedrock, II. Secondary deposits.

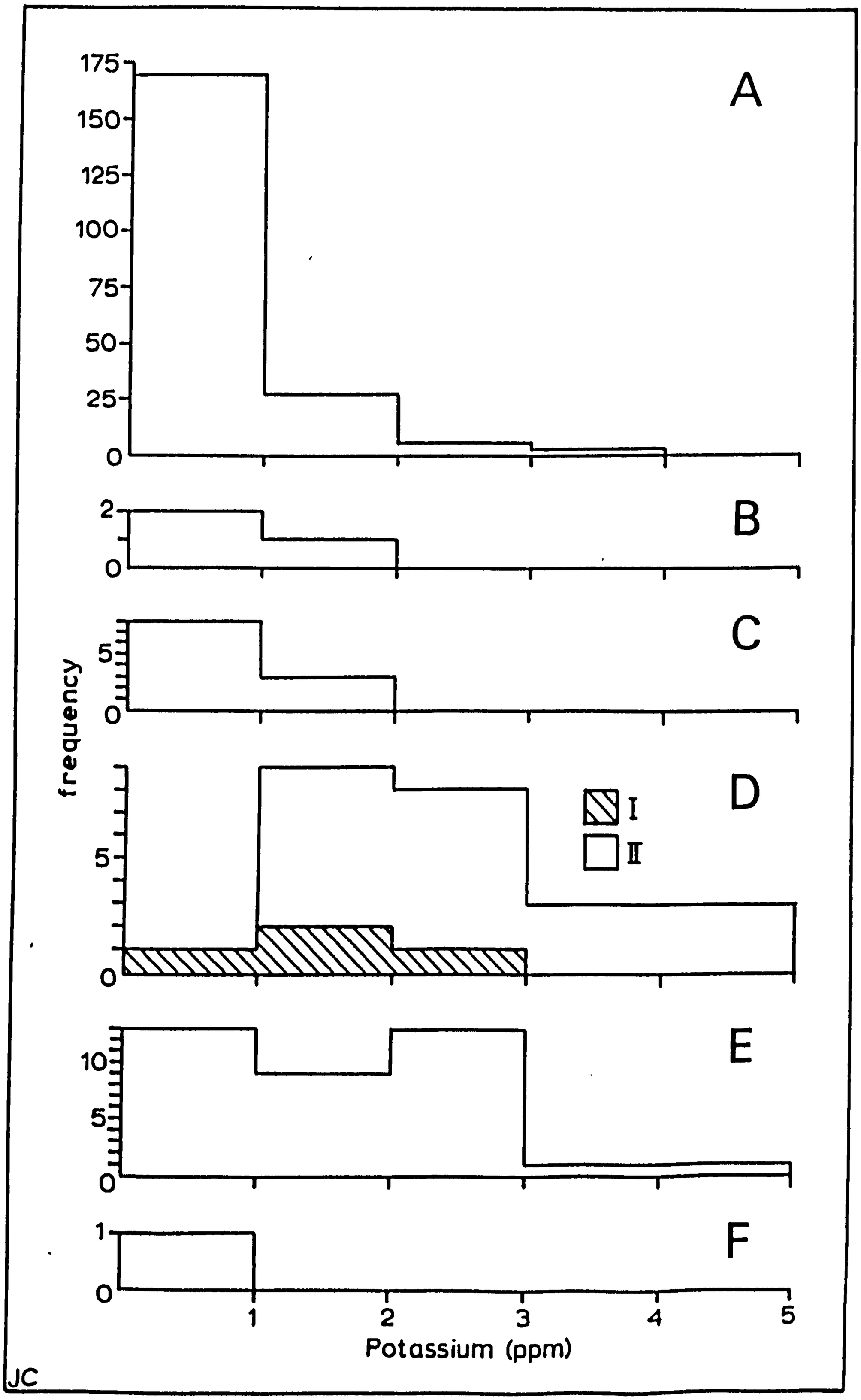


Figure 65 Average potassium content of karst waters in the three field areas: A. Diffuse-flow underground seepages; B. Karst springs; C. Underground seepages; D. Surface streams - I. Streams with a significant groundwater contribution, II. Streams which derive much of their flow from the karst alluvial plains; E. Seepages in alluvium of the karst plains; F. Deep groundwater.

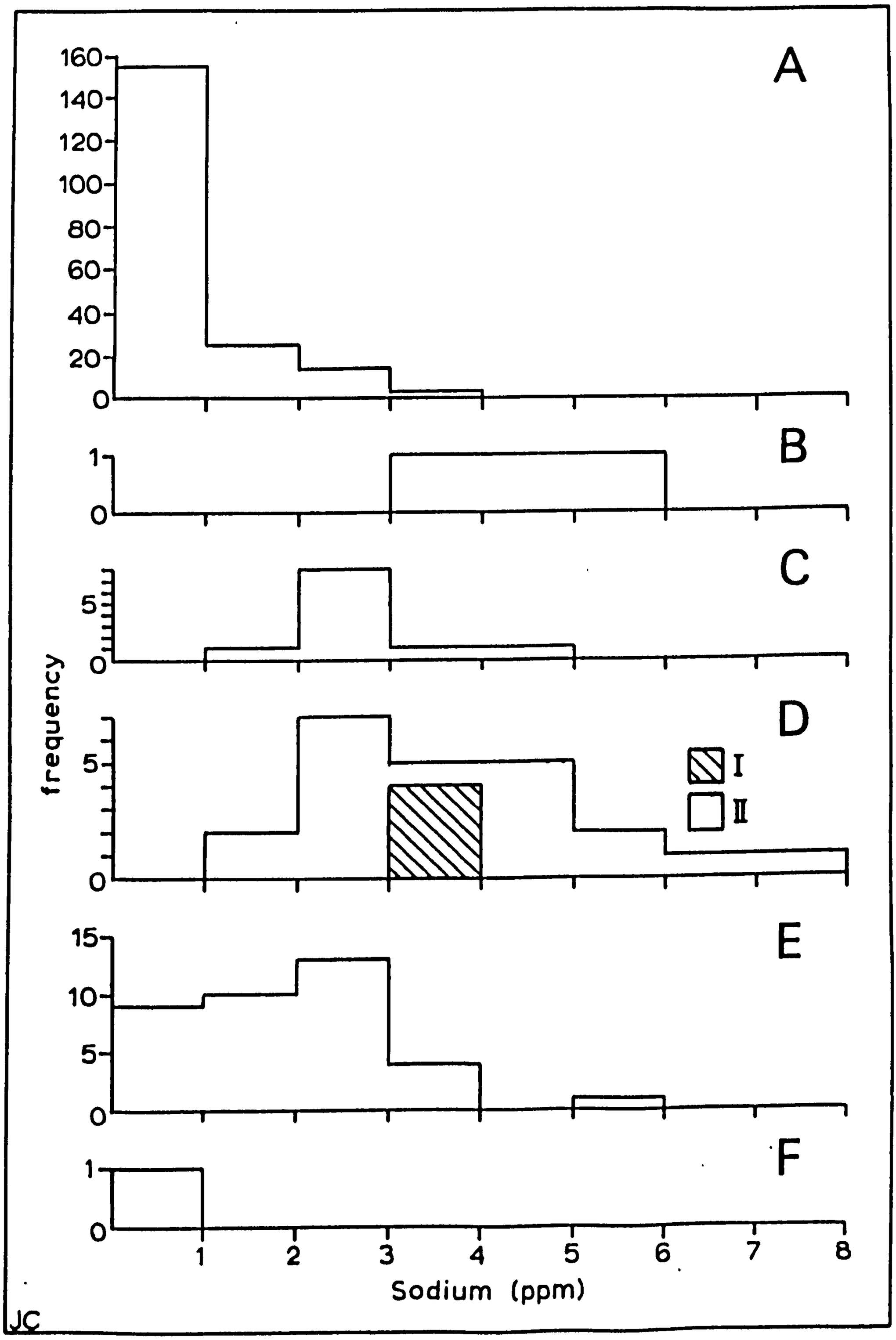


Figure 66 Average sodium content of karst waters in the three field areas: A. Underground diffuse-flow seepages; B. Karst springs; C. Underground streams; D. Surface streams - I. Streams with a significant groundwater contribution, II. Streams which derive much of their flow from the karst alluvial plains; E. Seepages in alluvium of the karst plains; F. Deep groundwater.

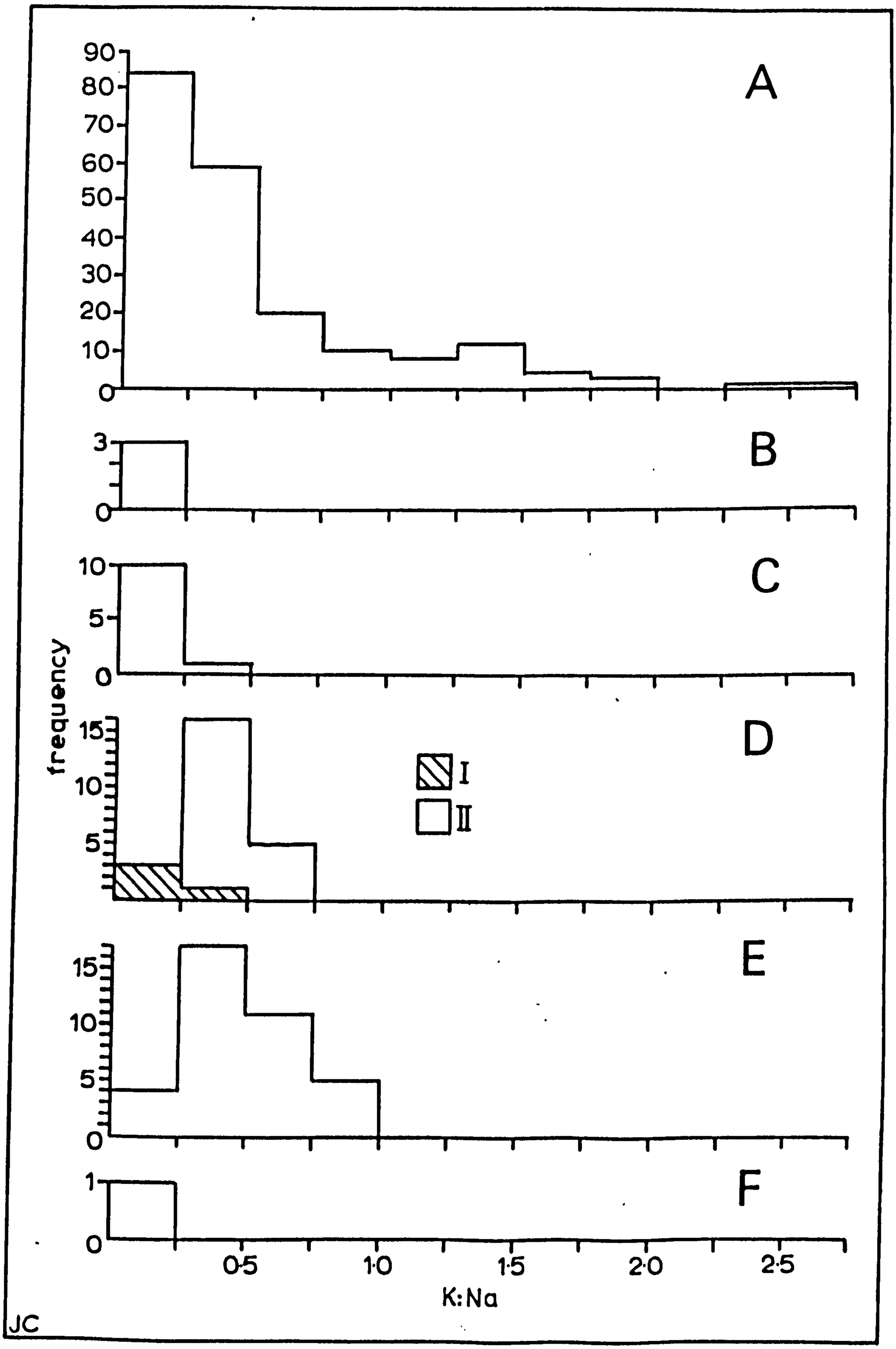


Figure 67 Average K:Na ratio of karst waters in the three field areas: A. Underground diffuse-flow seepages; B. Karst springs; C. Underground streams; D. Surface streams - I. Streams with a significant groundwater contribution, II. Streams which derive much of their flow from the karst alluvial plains; E. Seepages in alluvium of the karst plains; F, Deep groundwater.

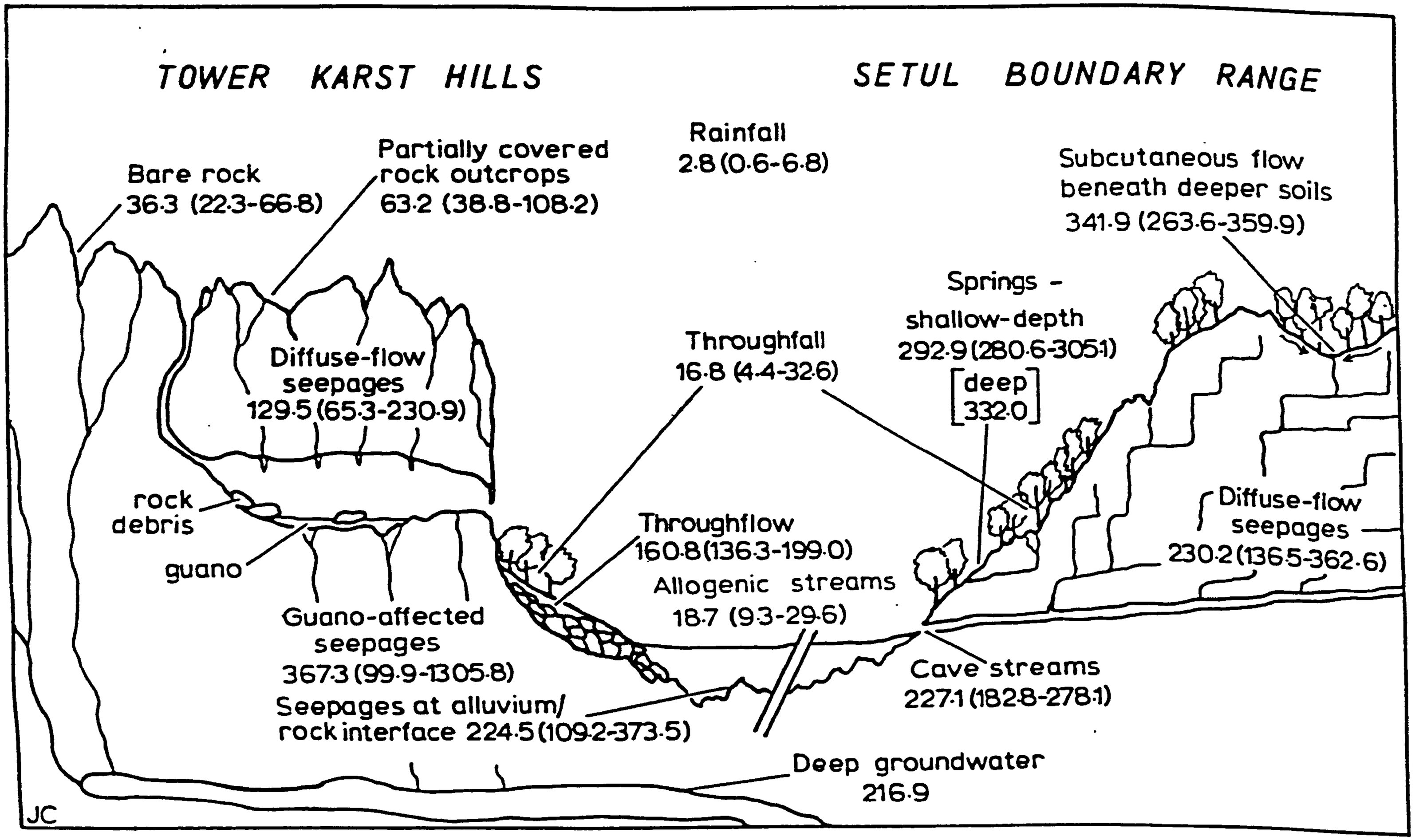


Figure 68 Summary of systematic variations in total hardness in the karst regions of West Malaysia, west of the Main Range.

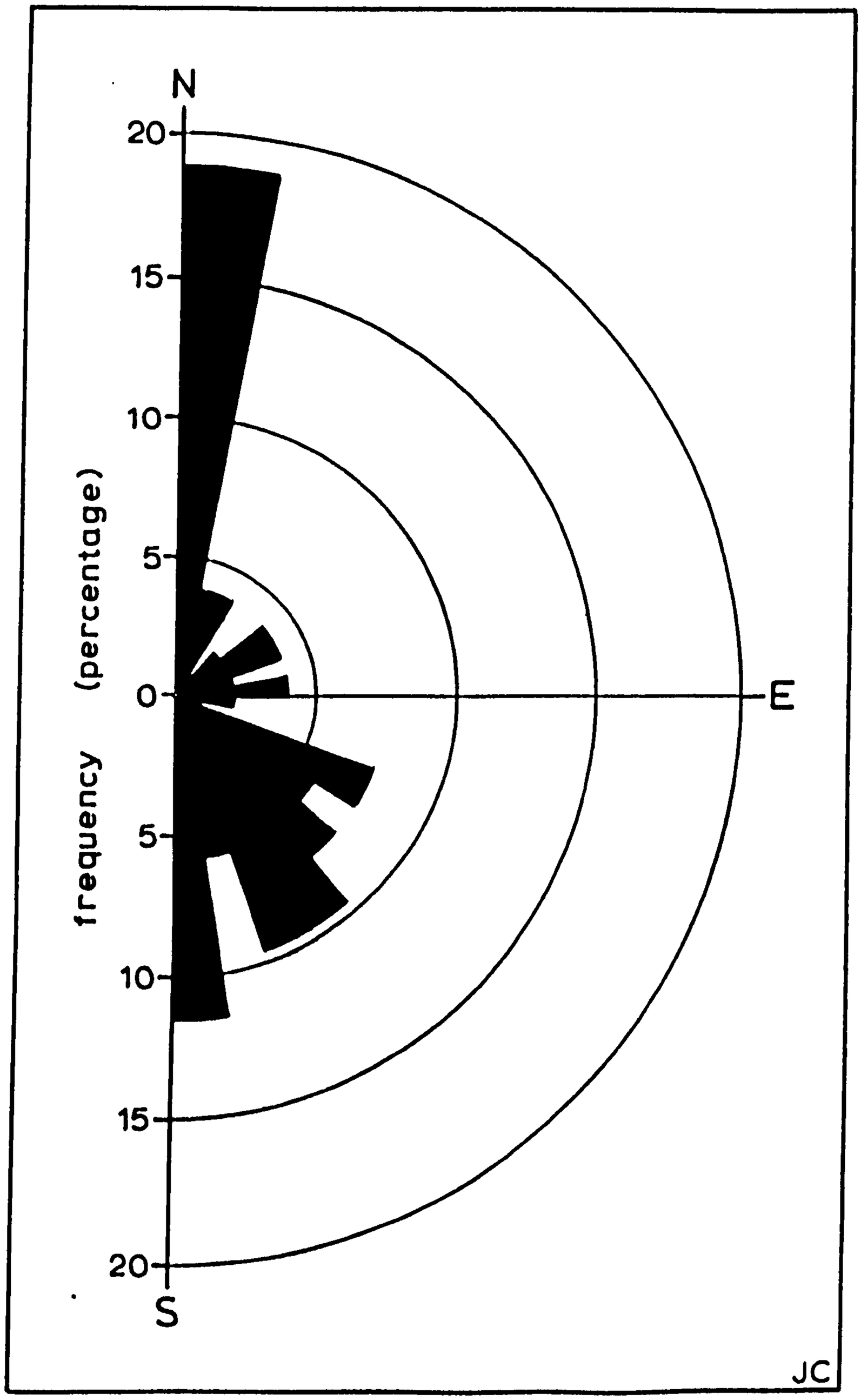


Figure 69 Frequency distribution of the orientations of the major axes of the tower karst hills in the Kinta Valley.

TABLES

Table 1 Meteorological stations used in assessment of daily rainfall during and immediately prior to study period.

CODE NUMBER[1]	STATION	SOURCE[2]	RELIAB-ILITY CLASS[3]	YEARS OF RECORD USED	COMMENTS
A. Selangor					
0601	Universiti Malaya	B	-	1973-75	
1601	Batu Caves Estate	A	A	1970-74	Closed: 31.10.74
1602	Kepong, F.R.I.	A	A	1970-75	
1604	Kg. Sg. Tua	A	-	1974-75	Opened: 1.6.74.
1605	Rawang Reservoir	A	B	1973-75	NR: 1.12.74 to 31.12.74
1607	St. John's School, Kuala Lumpur.	B	-	1974-75	
1609	Subang Airport	B	-	1970-75	
2601	Templer Park	C	-	1974-75	Opened: 10.9.74 NR: 12.11.74 to 11.12.74 and 12.1.75 to 22.1.75.
B. Kinta Valley					
3601	Sg. Siput (South)	C	-	1974-75	Opened: 11.9.74
3603	New Kinta Waterworks, Tg. Rambutan	A	-	1970-75	NR: 1.1.74 to 31.12.74
3604	Ipoh Hospital	A	-	1970-75	
3605	Batu Gajah Hospital	A	-	1970-75	
3606	Kampar Hospital	A	-	1970-75	
C. North-west					
4601	Kaki Bukit	A	-	1970-75	NR: 1.1.70 tO 30.6.70
4602	Padang Besar	A	-	1972-75	NR: 1.1.72 to 30.6.72
4603	Tasoh	A	-	1970-75	NR: 1.1.70 to 30.6.70
4604	Abi Head Works	A	-	1974-75	NR: 1.1.74 to 31.5.74
4605	Abi Kg. Bahru	A	-	1974-75	NR: 1.1.74 to 31.5.74
4606	Kangar	A	-	1970-75	NR: 1.12.72 to 31.12.72.

Notes: 1. Identification code adopted in present study. 2. Sources: A. D.I.D Malaysia; B. M.M.S.; C. Weekly autographic chart recorder installed and maintained by author. 3. After Leigh and Low (1973), details in table 4 4. NR, no records available.

Table 2 Meteorological stations of D.I.D., Malaysia, used in assessment of long-term monthly and annual rainfall in the field areas (after Drainage and Irrigation Dept., Malaysia, 1961).

CODE NUMBER[1]	STATION	RELIABILITY CLASS[2]	FIRST YEAR OF RECORD	LENGTH OF RECORD (yrs)
A. Selangor				
1601	Batu Caves Estate	A	1909	50
1605	Rawang Reservoir	B	1951	8
1611	Edinburgh Est., Kepong	B	1913	33
1612	General Hosp., Kuala Lumpur	B	1905	43
B. Kinta Valley				
3604	Ipoh Hospital	-	1891	59
3605	Batu Gajah Hospital	-	1888	65
3606	Kampar Hospital	-	1897	57
3607	Tg. Rambutan Asylum	-	1916	38
3608	Gopeng Hospital	-	1888	66
C. North-west				
4601	Kaki Bukit	-	1946	13
4602	Padang Besar	-	1928	26
4604	Abi Head Works	-	1946	13
4607	Kangar Hospital	-	1909	50
4608	Arau	-	1946	12
4609	Pulau Langkawi Hospital[3]	-	1917	37

Notes: 1. Identification code adopted in present study.
2. After Leigh and Low (1973), details in table.
3. Station used in determination of 24-hour point precipitation after Lockwood (1967).

Table 3 Meteorological stations of the Malaysian Meteorological Service used in assessment of monthly temperatures, sunshine hours (after Dale, 1964) and relative humidities in the field areas.

CODE NUMBER	STATION	ALTITUDE (m)	TYPE OF DATA[1]
A. Selangor			
1609	Subang Airport	16	A,C
1613	Port Klang	3	B
B. Kinta Valley			
3609	Ipoh Airport	38	A,B,C
C. North-west			
4609	Alor Star	5	A,B,C

Type of data: A. Temperature; B. Sunshine hours; C. Relative humidity.

Table 4 Reliability of rainfall stations in Selangor, based on double-mass curves (after Leigh and Low, 1973)

RELIABILITY CLASS	PERCENTAGE OF STATIONS
A. Straight lines with few or no deviations. Such stations are considered extremely reliable.	17.9
B. Minor deviations from a straight line. Such stations are considered to be generally reliable and adjustments might not be necessary.	44.2
C. Several distinct departures from a generally straight line. Such stations will require adjustments before they can be considered to be reliable.	29.5
D. Numerous deviations, often of considerable magnitude, from a straight line. Such stations are considered to be extremely unreliable and should be used only with caution.	8.4

Table 5 Correlation coefficients (r) of daily rainfall for pairs of stations in Selangor, based on data for 1973-1975

STATION	0601	1601	1602	1604	1605	1607	1609
A. Rainfall during approximate period of N.E. Monsoon (Nov. to April)							
1601	0.435						
1602	0.502	0.744					
1604	(0.124)	n.d.	(0.282)				
1605	0.283	0.296	0.362	(0.168)			
1607	0.739	0.590	0.590	(0.220)	0.294		
1609	0.446	0.454	0.454	(0.286)	0.278	0.414	
2601	(0.191)	n.d.	(0.551)	(0.292)	(0.459)	(0.359)	(0.127)
B. Rainfall during approximate period of S.W. Monsoon (May to Oct.)							
1601	0.420						
1602	0.423	0.676					
1604	(0.051)	n.d.	(0.634)				
1605	0.229	0.387	0.429	(0.127)			
1607	0.701	0.506	0.485	(0.293)	0.281		
1609	0.543	0.422	0.403	(0.212)	0.239	0.431	
2601	(0.073)	n.d.	(0.524)	(0.714)	(0.094)	(0.511)	(0.301)

n=270, except those cases in parentheses where n=89

n=274, except those cases in parentheses where n=81

Table 6 Correlation coefficients (r) of daily rainfall for pairs of stations in the Kinta Valley, based on data for 1973-75.

STATION	3601	3603	3604	3605
A. Rainfall during approximate period of N.E. Monsoon (Nov. to April)				
3603	n.d.			n=544, except those cases in parentheses where n=150
3604	(0.320)	0.427		
3605	(0.159)	0.360	0.439	
3606	(0.391)	0.321	0.315	0.383
B. Rainfall during approximate period of S.W. Monsoon (May to Oct.)				
3603	n.d.			n=552, except those cases in parentheses where n=142
3604	(0.292)	0.361		
3605	(0.166)	0.296	0.465	
3606	(0.461)	0.278	0.373	0.427

Table 7 Correlation coefficients (r) of daily rainfall for pairs of stations in the North-west region, based on data for 1973-75.

STATION	4601	4602	4603	4604	4605
A. Rainfall during approximate period of N.E. Monsoon (Nov. to April).					
4602	0.497			n=181	
4603	0.593	0.469			
4604	0.206	0.280	0.321		
4605	0.361	0.439	0.453	0.623	
4606	0.184	0.255	0.319	0.969	0.610
B. Rainfall during approximate period of S.W. Monsoon (May to Oct.)					
4602	0.385			n=184	
4603	0.536	0.324			
4604	0.422	0.203	0.363		
4605	0.432	0.289	0.529	0.552	
4606	0.258	0.109	0.271	0.796	0.399

Table 8 Classification of climate in the three field areas.

	KÖPPEN SYSTEM[1]	SCHMIDT AND FERGUSON SYSTEM[2]	AFTER DALE (1959)[3]	AFTER MORGAN (1971)[4]
Selangor	Af	A	West	IIc
Kinta Valley	Af	A	West	IIc
North-west	Am	B	North-west	IIa

Notes: 1. For explanation, see test p.
2. From Whitmore (1975), after Schmidt and Ferguson (1951), based on calculation of Q, where:

$$Q = \frac{\text{number of dry months}}{\text{number of wet months}} \times 100$$

A dry month is defined as any month with rainfall <60mm, and a wet month as one having >100mm rainfall. This quotient is calculated for each month of the climatic record, not for monthly averages. Values of Q between 0 and 14.3 are of climatic type 'A', whilst values between 14.3 and 33.3 are of type 'B'.
3. For explanation, see (Chapter II, Section A2).
4. IIc = West coast climate, Kuala Lumpur type; IIa = West coast climate, Langkawi type.

Table 9 Mean monthly and annual rainfall (mm) in field areas, based on the arithmetic means of long-term records of selected stations (see table 2).

	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	ANNUAL TOTAL
Selangor[1]	151	154	261	287	202	142	122	153	204	301	285	233	2495
Kinta Valley[2]	215	180	268	318	244	153	133	186	227	326	322	275	2847
North-west[3]	52	55	105	166	178	158	181	191	256	266	242	104	1960

Stations: 1. 1601, 1605, 1611, 1612; 2. 3604, 3605, 3606, 3607, 3608; 3. 4601, 4602, 4604, 4607, 4608.

Table 10 Frequency of rainfall events of given magnitude in field areas (1970-1974), after Crowther (1978a)

	AVERAGE NO. DAYS PER ANNUM WITH RAINFALL EXCEEDING:			
	0.99mm	25.0mm	50.0mm	100.0mm
Selangor[1]	159	34	10	0.5
Kinta Valley[2]	157	33	9	0.7
North-west[3]	126	24	7	0.4

Stations: 1. 1601, 1602, 1605, 1607; 2. 3602, 3603, 3604, 3605; 3. 4601, 4602, 4603, 4606.

Table 11 Typical maximum 24 hour point precipitation (mm) in field areas, after Lockwood (1967).

	MAXIMUM 24-HOUR RAINFALL FOR RETURN PERIODS IN YEARS OF:				PROBABLE MAXIMUM	MAXIMUM RECORDED
	10	100	1,000	10,000		
Selangor[1]	122	168	211	254	422	163
Kinta Valley[2]	147	203	262	319	536	244
North-west[3]	158	239	318	396	681	206

Stations: 1. 1612; 2.3608; 3.4609

Table 12 Typical mean monthly and annual relative humidities (%) in field areas.

	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	ANNUAL MEAN
Selangor[1]	82.9	82.5	82.4	86.1	85.4	84.3	84.9	83.3	86.0	85.3	86.5	86.3	84.6
Kinta Valley[2]	80.0	78.4	78.7	82.8	83.5	80.9	82.4	81.2	84.1	83.7	86.3	83.8	82.2
North-west[3]	76.9	73.2	75.1	81.7	84.7	84.6	84.4	83.6	85.6	86.3	86.7	81.6	82.0

Stations: 1. 1609; 2. 3609; 3. 4609.

Table 13 Correlations (r_s) between rainfall and other climatic parameters in the three field areas (n=12)

	RELATIVE HUMIDITY	SUNSHINE HOURS	TEMPERATURE
Selangor	0.462	-0.330	-0.057
Kinta Valley	0.580[1]	-0.523	-0.258
North-west region	0.860	-0.860	-0.064

[1]Values underlined are significant at 95 per cent confidence level.

Table 14 Estimates of potential evapo-transpirational loss (mm) in field areas, based on Nieuwolt (1965, fig.8).

	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	ANNUAL TOTAL
Selangor	121	133	124	124	122	124	124	121	112	109	107	109	1430
Kinta Valley	130	130	133	130	124	130	130	122	117	117	114	122	1499
North-west	150	156	156	150	135	135	133	130	122	122	122	135	1646

Table 15 Typical mean monthly and annual temperatures (°C) in field areas

	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	ANNUAL MEAN
Selangor[1]	26.0	26.1	26.5	26.6	26.7	26.7	26.3	26.5	26.1	26.3	26.1	25.6	26.3
Kinta Valley[2]	26.2	26.8	27.1	27.1	26.8	27.1	26.3	26.9	26.1	26.3	25.8	25.8	26.5
North-west[3]	26.2	27.2	27.6	27.8	27.2	27.2	26.9	27.2	26.6	27.1	26.2	25.9	26.9

Stations: 1. 1609; 2. 3609; 3. 4609.

Table 16 Typical average number of bright sunshine hours per day in field areas, after Dale (1964)

	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	ANNUAL MEAN
Selangor[1]	5.9	6.5	7.0	6.7	6.3	6.8	6.4	6.5	5.6	5.5	5.5	5.6	6.2
Kinta Valley[2]	6.9	7.5	7.4	7.0	6.6	7.1	6.9	6.3	5.9	5.7	5.6	6.3	6.6
North-west[3]	8.0	8.5	8.5	8.2	6.8	6.5	6.7	6.4	5.7	5.6	5.7	7.0	7.0

Stations: 1. 1613; 2.3609; 3.4609.

Table 17 Typical diurnal temperature fluctuations (°C) in field areas, after Dale (1963)

	ANNUAL MEAN DIURNAL RANGE	ABSOLUTE EXTREMES RECORDED	
		HIGHEST MAXIMUM	LOWEST MINIMUM
Selangor[1]	9.9	36.6	17.8
Kinta Valley[2]	10.4	37.7	17.2
North-west[3]	8.8	37.2	16.7

Stations: 1. Kuala Lumpur; 2. Ipoh; 3. Alor Star.

Table 18 Soil temperature (^{o}C) at 13.00 to 14.00 hrs., Ak.Bt. Takun, Selangor (plot ABT-I, altitude 122 m).

DEPTH (cm)	n	MEAN	MINIMUM	MAXIMUM	S.DEV.
7.6	10	24.4	23.7	25.1	0.54
15.2	10	24.2	23.8	24.8	0.36
30.1	10	24.3	23.9	24.9	0.35

Table 19 Comparison of rainfall at Kepong (1602), Selangor, during study period with previous years' records.

	AUG	SEP	OCT	NOV	DEC	JAN	FEB	MAR	APR	MAY	JUN	JUL	TOTAL
Monthly totals (mm)													
1974/75	71	250	51	236	340	137	268	172	323	83	105	282	2318
1913-58	151	191	282	288	231	166	164	240	276	184	140	127	2441
Number of dry days (rainfall <1.0 mm)													
1974/75	26	8	25	14	13	16	10	18	11	21	17	10	189
1970-73	18	15	15	12	13	22	20	20	15	15	17	23	205
Maximum number of successive dry days													
1974/75	23	13	12	5	3	4	4	5	9	5	8	4	-
1970-73	8	10	7	4	5	23	21	13	7	10	11	14	-
Number of days with more than 25.0 mm													
1974/75	1	4	0	3	5	2	3	2	3	0	1	3	27
1970-73	2	3	4	6	3	3	2	2	5	4	2	1	37
Maximum 24-hour rainfall (mm)													
1974/75	33	45	22	42	72	36	80	48	81	23	46	36	-
1970-73	90	80	89	87	94	148	82	59	115	71	80	69	-
Maximum 96-hour rainfall (mm)													
1974/75	79	75	22	96	154	58	130	81	109	32	61	75	-
1970-73	189	165	164	163	193	270	171	125	165	165	186	130	-

Table 20. Comparison of rainfall at Kampar (3606), Kinta Valley, during study period with previous years' records.

	AUG	SEP	OCT	NOV	DEC	JAN	FEB	MAR	APR	MAY	JUN	JUL	TOTAL
Monthly totals (mm)													
1974/75	117	557	103	46	408	280	177	410	383	348	114	255	3198
1897-1958	236	256	395	411	339	298	268	390	452	294	183	143	3663
Number of dry days (rainfall <1.0 mm)													
1974/75	22	7	19	11	12	9	10	11	10	17	23	16	167
1970-73	14	9	11	10	10	15	13	14	10	12	19	21	158
Maximum number of successive dry days													
1974/7	12	3	10	4	5	4	4	3	7	10	8	10	-
1970-73	6	6	6	8	5	5	9	11	4	6	11	12	-
Number of days with more than 25.0 mm.													
1974/75	2	6	1	0	5	3	2	8	5	4	1	6	43
1970-73	3	3	5	6	4	4	1	4	5	5	3	1	44
Maximum 24-hour rainfall (mm)													
1974/75	49	91	44	10	104	98	38	90	97	118	67	47	-
1970-73	67	89	117	87	79	58	14	68	128	134	87	65	-
Maximum 96-hour rainfall (mm)													
1974/5	203	216	54	15	198	145	94	155	176	133	68	105	-
1970-73	120	243	254	192	141	207	77	181	146	240	142	87	-

Table 21 Comparison of rainfall at Kaki Bukit (4601), North-west region, during study period with previous years' records.

	AUG	SEP	OCT	NOV	DEC	JAN	FEB	MAR	APR	MAY	JUN	JUL	TOTAL
Monthly totals (mm)													
1974/75	178	202	163	228	84	314	34	139	159	213	171	197	2082
1946-58	192	240	247	225	151	55	70	88	142	154	143	152	1859
Number of dry days (rainfall<1.0 mm)													
1974/75	21	13	18	15	23	22	26	24	19	13	18	16	228
1970-73	20	12	14	17	20	28	25	26	22	20	17	20	241
Maximum number of successive dry days													
1974/75	9	4	7	3	19	34	25	17	9	3	5	8	-
1970-73	11	5	6	13	15	51	25	24	15	12	10	10	-
Number of days with more than 25.0 mm													
1974/75	3	1	1	3	0	5	1	2	2	2	2	3	25
1970-73	2	5	3	3	3	1	1	1	3	1	3	2	28
Maximum 24-hour rainfall (mm)													
1974/75	33	35	64	56	17	76	28	46	51	31	42	49	-
1970-73	108	98	58	65	123	60	40	31	60	76	72	86	-
Maximum 96-hour rainfall (mm)													
1974/75	64	99	78	90	49	138	34	111	61	66	57	92	-
1970-73	160	228	149	119	197	139	82	73	114	89	118	123	-

Table 22 Chemical properties of rainwater in the three field areas.

FIELD AREA[2]	pH	SPECIFIC CONDUCTANCE ($\mu mho.cm^{-1}$ 25°C)	Ca^{2+} (ppm)	Mg^{2+} (ppm)	K^+ (ppm)	Na^+ (ppm)
A. Selangor (n=15)						
mean	5.65	11.4	1.48	0.14	0.15	0.21
minimum	4.80	8.0	0.40	0.08	0.04	0.10
maximum	6.65	19.0	3.40	0.20	0.30	0.46
B. Kinta Valley (n=5)						
mean	5.31	13.6	0.40	0.05	0.12	0.28
minimum	4.65	8.0	0.24	0.02	0.03	0.08
maximum	6.15	20.0	0.60	0.10	0.22	0.55
C. North-west (n=5)						
mean	6.02	15.2	1.02	0.09	0.13	0.26
minimum	5.47	6.0	0.68	0.05	0.06	0.18
maximum	6.72	31.0	2.45	0.15	0.21	0.32

Notes: 1. Since the rain gauges remained uncovered between storms, these figures include the water-soluble components of atmospheric dust.
2. Stations used: A. 2601 and Ak. Bt. Takun (Chapter V, Section C1.2); B. 3601; C. 4601.

Table 23 Chemical properties of rainwater in other parts of the Malay Peninsula

	Ca^{2+} (ppm)	Mg^{2+} (ppm)	K^{+} (ppm)	Na^{+} (ppm)
Pasoh Forest Reserve[1], Negri Sembilan	0.20	0.04	0.35	1.09
Gombak Field Studies Centre[2], Selangor	0.56	0.13	0.50	n.d.

Sources: 1. Manokaran (1978), 2. Kenworthy (1971).

Table 24 Input of major elements from rainfall during the one-year study period (1.8.74 to 31.7.75).

	RAINFALL (mm)	Ca (kg/ha/yr)	Mg (kg/ha/yr)	K (kg/ha/yr)	Na (kg/ha/yr)
Selangor[1]	2234	33.06	3.13	3.35	4.69
Kinta Valley[2]	2846	11.38	1.42	3.42	7.97
North-west region[3]	2052	20.94	1.85	2.67	5.34
Arithmetic mean		21.79	2.13	3.15	6.00

Rainfall based on the following stations: 1.2601; 2. 3601; 3. 4601 and 4604.

Table 25 Summary of characteristics of the major limestone formations, west of the Main Range.

	A. SELANGOR	B. KINTA VALLEY	C. NORTH-WEST		
Limestone	Kuala Lumpur Limestone	Kinta Limestone	Setul Formation	Chuping Formation	Kodiang limestone
Age	Silurian	Silurian to Permian	Ordovician to Silurian	Permian	Triassic
Estimated thickness(m)	1830[1]	>3000[2]	1547[3]	600[4]	?
Number of samples	52[5]	72[6]	37[5]	25[7]	11[8]
CaO (%)	49.21	46.44	46.88	49.95	48.78
MgO (%)	5.40	7.70	3.10	4.10	5.33
Acid insoluble residue(%)	1.80	1.55	9.89	3.44	2.81

Sources:
1. Gobbett (1964)
2. Ingham and Bradford (1960)
3. Jones (1973)
4. Jones (in press)
5. Author; Hutchison (1966)
6. Author; Hutchison (1966); Ingham and Bradford (1960)
7. Hutchison (1966)
8. Alexander et al. (1964); Hutchison (1966)

Table 26 Summary of chemical analyses of limestones from the three field areas[1] (n=197)

	MEAN	MINIMUM	MAXIMUM	S.DEV.
Acid insoluble residue (%)	3.50	0.01	42.29	5.86
CaO (%)	47.83	26.43	56.70	30.27
MgO (%)	5.64	0.02	24.44	7.87
Mg:Ca (molar ratio)	0.230	0.001	1.285	0.355

[1]Data from present study and from Alexander et al. (1964), Hutchison (1966), and Ingham and Bradford (1960)

Table 27 Summary of chemical analyses of secondary deposits from the three field areas[1] (n=44)

	MEAN	MINIMUM	MAXIMUM	S.DEV.
Acid insoluble residue (%)	0.89	0.01	8.67	1.57
CaO (%)	54.94	51.69	56.88	1.11
MgO (%)	0.32	0.02	0.96	0.26
Mg:Ca (molar ratio)	0.008	0.0003	0.025	0.0066

[1]Data from present study and Hutchison (1966)

Table 28 Summary of chemical analyses of the major limestone lithologies[1]

	MEAN	MINIMUM	MAXIMUM	S.DEV.
Kuala Lumpur Limestone (n = 52)				
Acid insoluble residue (%)	1.84	0.01	14.20	2.39
CaO (%)	49.21	27.16	56.70	9.52
MgO (%)	5.40	0.07	22.05	7.83
Mg:Ca (molar ratio)	0.220	0.002	1.023	0.354
Kinta Limestone (n = 72)				
Acid insoluble reside (%)	1.55	0.05	12.00	1.98
CaO (%)	46.44	26.43	56.28	10.78
MgO (%)	7.70	0.12	24.44	9.08
Mg:Ca (molar ratio)	0.325	0.003	1.285	0.413
Setul Formation (n = 37)				
Acid insoluble residue (%)	9.89	1.90	42.29	9.75
CaO (%)	46.88	28.45	53.06	6.39
MgO (%)	3.10	0.02	21.36	4.45
Mg:Ca (molar ratio)	0.108	0.001	1.043	0.200
Chuping Formation (n = 25)				
Acid insoluble residue (%)	3.44	0.10	26.20	5.20
CaO (%)	49.95	31.64	56.56	8.93
MgO (%)	4.10	0.17	20.39	6.97
Mg:Ca (molar ratio)	0.164	0.004	0.896	0.299
Kodiang limestone (n = 11)				
Acid insoluble residue (%)	2.81	0.10	8.20	2.18
CaO (%)	48.78	31.92	55.72	9.34
MgO (%)	5.33	0.17	18.57	6.95
Mg:Ca (molar ratio)	0.208	0.004	0.809	0.295

[1]Sources as given in table 25.

Table 29 Plots used in assessment of slope, soil and vegetation characteristics.

PLOT	REGION[1]	CODE[1]	ALTITUDE (m)	SIZE (m)	TYPES OF STUDY UNDERTAKEN[3]
1. Rocky hilltops					
G. Tempurong, IV	KV	GT-IV	400	30x10	A,D,E
Ak.Bt.Takun, IV	Sel	ABT-IV	162	30x10	A,D,E,H
Ak.Bt.Takun, III	Sel	ABT-III	153	30x10	A,D,E,H
2. Hilltop depressions					
G. Layang Layang	KV	GLL	290	30x20	A,D,E
3. Steep hill slopes					
G. Tempurong, III	KV	GT-III	287	30x20	A,D,E
Wang Tangga, II	NW	WT-II	244	30x10	A,D,E
Gelang Kerbau V., II	NW	GKV-II	327	30x10	A,D,E
4. Moderately-inclined hill slopes					
G. Tempurong, II	KV	GT-II	196	30x30	A,B,C,D,E,F,G,H
Wang Tangga, I	NW	WT-I	180	30x10	A,B,C,D,E,F,G,H,I
Gelang Kerbau V., I	NW	GKV-I	189	30x10	A,B,C,D,E,F,G,H
5. Moderately-inclined foothill slopes					
Ak.Bt.Takun, I	Sel	ABT-I	122	30x30	A,B,C,D,E,F,G,H,I
Ak.Bt.Takun, II	Sel	ABT-II	113	30x30	A,B,C,D,E,F,G,H,I
G. Tempurong, I	KV	GT-I	124	30x30	A,B,C,D,E,F,G,H,I

Notes: 1. Sel = Selangor, KV = Kinta Valley, NW = North-west.
2. Abbreviations used in text.
3. A. Soil analysis, B. Soil moisture, C. Soil carbon dioxide, D. Micro-relief, soil depths and surface cover, E. Tree density, F. Throughfall and interception loss, G. Litterfall, H. Erosional weight-loss tablet studies, I. Throughflow.

Table 30 Relief and surface cover of the sampling plots.

PLOT	n[1]	AVERAGE GRADIENT[2] (degrees)	NATURE OF SURFACE COVER (% of area)			SOIL DEPTH(cm)	
			BARE ROCK[3]	ROCK/ORGANIC LITTER	SOIL	MEAN[4]	MAXIMUM
1. Rocky hilltops							
GT-IV	60	49.2	54.9	8.6	36.5	6.4	60
ABT-IV	30	67.3	83.3	7.4	9.3	0.8	15
ABT-III	90	45.4	60.6	8.1	31.3	2.6	27
2. Hilltop depressions							
GLL	20	15.1	0.3	3.7	96.0	162.1 ~~13.4~~	240
3. Steep hill slopes							
GT-III	90	46.7	44.1	5.4	50.5	11.7	65
WT-II	60	45.3	40.7	7.7	51.6	14.7	145
GKV-II	20	46.9	43.7	3.2	53.1	11.9	60
4. Moderately-inclined hill slopes							
GT-II	90	40.1	34.0	6.9	59.1	26.5	155
WT-I	90	35.2	29.4	12.0	58.6	21.1	135
GKV-I	90	32.1	5.8	0.3	93.9	43.6	138
5. Moderately-inclined hill slopes							
ABT-I	90	32.8	15.1	1.0	83.0	38.2	165
ABT-II	60	33.9	15.8	3.2	81.0	23.3	90
GT-I	90	36.6	26.8	11.9	61.3	24.0	130

Notes:
1. Number of slope segments measured.
2. In case of rocky hilltop plots, the gradients of segments which are negative with respect to the overall gradient of the transect line are treated as positive.
3. Refers to surfaces free of soil and litter, but includes areas with lichen cover.
4. Average depth along the transect lines, including those points with no soil cover.

Table 31 Composition of limestone (by weight) in the thirteen study plots

	$CaCO_3$ (%)	$MgCO_3$ (%)	Mg:Ca (molar)	A.I.R. (%)	Fe_2O_3 (%)	Fe_2O_3/A.I.R. (%)
GT-IV	99.25	0.66	0.008	0.05	0.02	40.0
ABT-IV	98.53	1.16	0.014	0.31	0.09	29.0
ABT-III	98.53	1.16	0.014	0.31	0.09	29.0
GLL	93.00	6.94	0.088	0.40	0.12	30.0
GT-III	99.25	1.04	0.012	1.40	0.23	16.4
WT-II	93.50	4.16	0.053	3.00	0.84	28.0
GKV-II	88.50	4.51	0.060	6.40	0.37	5.8
GT-II	99.68	1.14	0.014	0.38	0.24	63.2
WT-I	72.65	22.50	0.367	2.35	0.57	24.3
GKV-I	88.63	5.55	0.074	9.65	0.44	4.6
ABT-I	98.25	1.39	0.017	0.50	0.21	42.0
ABT-II	99.20	0.69	0.008	0.20	0.08	40.0
GT-I	97.97	1.53	0.018	0.33	0.06	18.2

Table 32 Frequency of trees and cross-sectional areas of tree trunks in the thirteen plots

	TREE[1] DENSITY (no./ha)	TREES EXCEEDING 100 cm girth (no./ha)	BASAL AREA OF TREE TRUNKS[2] (m^2/ha)
GT-IV	200	0	4.9
ABT-IV	233	0	2.2
ABT-III	478	89	26.0
GLL	1783	0	17.5
GT-III	800	33	16.4
WT-II	1033	100	33.7
GKV-II	1167	100	28.2
GT-II	656	56	27.5
WT-I	1050	100	36.8
GKV-I	1167	233	70.2
ABT-I	556	56	22.2
ABT-II	600	17	9.5
GT-I	456	111	63.2

Notes: 1. Trees greater than 15 cm girth.
2. Circular cross-sections are assumed.

Table 33 Average composition (percentage by weight) of the fine-earth fraction of soils in the thirteen study plots

	Ca	Mg	K	Na	Si	Fe	Al
GT-IV	0.9	0.6	0.1	0.09	3.4	10.6	33.6
ABT-IV	2.0	0.5	3.2	2.88	31.0	2.6	3.4
ABT-III	0.8	0.5	3.7	1.38	33.0	2.8	6.1
GLL	0.2	1.2	0.1	0.05	13.4	14.8	20.4
GT-III	0.7	2.1	0.2	0.07	8.1	11.1	29.5
WT-II	0.5	1.4	4.0	0.17	18.7	9.8	15.0
GKV-II	0.3	1.2	3.7	0.40	23.0	4.8	9.5
GT-II	0.5	0.2	0.8	1.07	14.8	10.8	25.3
WT-I	0.1	1.2	0.6	0.16	23.0	13.4	12.0
GKV-I	0.1	0.9	3.8	0.10	25.8	6.4	12.6
ABT-I	1.2	0.4	2.1	1.54	33.7	6.1	10.9
ABT-II	1.5	0.9	3.6	2.92	39.5	2.2	4.3
GT-I	0.7	0.6	2.8	1.60	36.1	2.2	11.3

Table 34 Soil analytical data - GT-IV

profile depth = 28 cm

SAMPLE DEPTH/SHALLOW SOIL NO. ( ):	10 cm	20 cm	SHALLOW SOILS (1)	(2)	(3)
1. Particle-size distribution* (% by weight)					
Gravel (6.0-2.0 mm)	0.0	0.8	0.2	0.0	0.0
Sand (0.06-2.0 mm)	21.9	17.3	26.0	15.7	13.2
Silt (0.002-0.06 mm)	33.9	37.2	43.1	28.9	48.4
Clay (<0.002 mm)	44.2	45.5	30.9	55.4	38.4
2. Other physical properties					
Colour (moist soil)	5YR3/1	5YR3/2	5YR3/2	5YR3/2	5YR3/1
Bulk density ($g.cm^{-3}$)	n.d.	n.d	n.d	n.d.	n.d.
Particle density ($g.cm^{-3}$)	n.d.	n.d.	n.d.	n.d.	n.d.
Pore space (% by volume)	n.d.	n.d.	n.d.	n.d.	n.d.
3. Organic matter (% by weight)					
Carbon	4.73	2.20	6.47	3.51	8.52
Loss on ignition	32.9	18.3	47.1	26.6	59.4
4. Cation exchange properties					
C.E.C. (m.e./100 g)	34.76	19.61	38.48	29.28	40.29
Exchangeable Ca (m.e./100 g)	13.78	12.67	16.57	14.23	25.73
" Mg (")	3.04	2.68	4.01	3.61	4.18
" K (")	0.30	0.10	0.30	0.26	0.26
" Na (")	0.10	0.02	0.10	0.08	0.09
Molar ratio Mg:Ca	0.221	0.212	0.242	0.254	0.162
" " K:Na	3.0	3.3	3.0	5.0	2.9
Base saturation (%)	49.5	78.9	54.5	62.1	75.1
pH (water, 1:2.5)	6.35	6.9	6.15	6.6	7.05
5. Carbonate content (%)	0.1	0.5	0.1	0.1	0.5

*Gravel expressed as percentage of whole soil; sand, silt, and clay as percentage of <2.0 mm fraction

Table 35 Soil analytical data - ABT-IV and ABT-III

	ABT-IV		ABT-III		
SHALLOW SOIL NO. ( ):	(1)	(2)	(1)	(2)	(3)
1. Particle-size distribution* (% by weight)					
Gravel (6.0-2.0 mm)	0.0	0.0	0.0	0.0	0.0
Sand (0.06-2.0 mm)	20.8	6.2	18.3	16.0	9.8
Silt (0.002-0.06 mm)	30.9	55.5	40.3	46.7	34.0
Clay (<0.002 mm)	48.3	38.3	41.5	37.3	56.2
2. Other physical properties					
Colour (moist soil)	10YR2.5/1	10YR2.5/1	10YR2.5/2	10YR2.5/1	10YR2.5/1
Bulk density ($g.cm^{-3}$)	n.d.	n.d.	n.d.	n.d.	n.d.
Particle density ($g.cm^{-3}$)	n.d.	n.d.	n.d.	n.d.	n.d.
Pore space (% by volume)	n.d.	n.d.	n.d.	n.d.	n.d.
3. Organic matter (% by weight)					
Carbon	11.37	10.69	4.81	9.18	5.16
Loss on ignition	76.4	78.1	42.4	65.5	40.9
4. Cation exchange properties					
C.E.C. (m.e./100 g)	46.21	48.90	32.33	42.79	27.85
Exchangeable Ca (m.e./100 g)	31.97	21.40	19.10	16.86	15.30
" Mg (")	4.80	4.36	4.32	3.96	4.38
" K (")	0.38	0.45	0.52	0.43	0.35
" Na (")	0.10	0.10	0.08	0.10	0.05
Molar ratio Mg:Ca	0.150	0.204	0.226	0.235	0.286
" " K:Na	3.8	4.5	6.5	4.3	7.0
Base saturation (%)	80.6	53.8	74.3	49.9	72.1
pH (water, 1:2.5)	7.2	6.25	6.8	6.1	7.15
5. Carbonate content (%)	5.0	2.0	2.0	2.0	5.0

*Gravel expressed as percentage of whole soil; sand, silt, and clay as percentage of <2.0 mm fraction

Table 36 Soil analytical data - GLL

profile depth = 145 cm

SAMPLE DEPTH (cm):	10	20	30	50	70	100	140
1. Particle-size distribution* (% by weight)							
Gravel (6.0-2.0 mm)	1.2	0.8	3.9	5.3	8.7	5.8	1.3
Sand (0.06-2.0 mm)	18.3	15.7	7.7	10.6	11.4	10.7	12.9
Silt (0.002-0.06 mm)	21.0	17.4	18.8	14.2	10.4	15.0	17.8
Clay (<0.002 mm)	60.7	66.9	73.5	75.2	78.2	74.3	69.3
2. Other physical properties							
Colour (moist soil)	2.5YR3/4	2.5YR3/4	2.5YR3/4	2.5YR3/4	2.5YR3/4	2.5YR3/4	2.5YR3/4
Bulk density ($g.cm^{-3}$)	0.71	0.75	0.82	0.86	0.94	1.03	1.14
Particle density ($g.cm^{-3}$)	2.80	2.83	2.88	2.92	2.89	2.92	2.92
Pore space (% by volume)	74.6	73.5	71.6	70.6	67.5	64.7	61.0
3. Organic matter (% by weight)							
Carbon	1.34	0.90	0.72	0.40	0.18	0.06	0.08
Loss on ignition	18.5	17.4	16.9	19.5	16.1	16.3	15.3
4. Cation exchange properties							
C.E.C. (m.e./100 g)	9.98	8.63	9.60	9.44	8.81	8.09	9.28
Exchangeable Ca (m.e./100 g)	1.39	1.26	2.90	3.34	3.89	4.09	4.58
" Mg (")	0.82	0.54	0.86	1.24	0.87	0.93	0.87
" K (")	0.05	0.05	0.04	0.02	<0.01	<0.01	<0.01
" Na (")	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Molar ratio Mg:Ca	0.590	0.429	0.297	0.371	0.224	0.227	0.190
" " K:Na	5.0	5.0	4.0	2.0	<1.0	n.d.	<1.0
Base saturation (%)	22.8	21.6	39.7	48.8	54.1	62.1	58.8
pH (water, 1:2.5)	5.2	5.35	5.7	6.1	6.2	6.5	6.5
5. Carbonate content (%)	0.1	0.5	0.5	0.1	0.1	0.1	0.1

*Gravel expressed as percentage of whole soil; sand, silt, and clay as percentage of <2.0 mm fraction

Table 37 Soil analytical data - GT-III

profile depth = 23 cm

SAMPLE DEPTH/SHALLOW SOIL No. ( ):	10 cm	20 cm	SHALLOW SOILS (1)	(2)	(3)
1. Particle-size distribution* (% by weight)					
Gravel (6.0-2.0 mm)	0.0	0.4	0.0	0.0	0.0
Sand (0.06-2.0 mm)	26.4	23.0	18.4	22.1	7.2
Silt (0.002-0.06 mm)	40.7	37.3	42.9	45.1	34.2
Clay (<0.002 mm)	32.9	39.7	38.7	32.8	58.6
2. Other physical properties					
Colour (moist soil)	2.5YR3/4	2.5YR3/4	2.5YR2.5/2	2.5YR3/2	2.5YR3/4
Bulk density ($g.cm^{-3}$)	0.65	0.70	n.d.	n.d.	n.d.
Particle density ($g.cm^{-3}$)	2.78	2.83	n.d.	n.d.	n.d.
Pore space (% by volume)	76.6	75.3	n.d.	n.d.	n.d.
3. Organic matter (% by weight)					
Carbon	2.60	1.63	7.48	5.64	7.91
Loss on ignition	21.3	18.4	55.4	49.6	60.3
4. Cation exchange properties					
C.E.C. (m.e./100 g)	25.86	14.43	34.42	31.89	39.17
Exchangeable Ca (m.e./100 g)	14.62	8.61	21.78	20.14	26.31
" Mg (")	2.71	1.64	5.28	5.97	4.25
" K (")	0.32	0.19	0.32	0.27	0.40
" Na (")	0.09	0.05	0.18	0.06	0.08
Molar ratio Mg:Ca	0.185	0.190	0.242	0.296	0.162
" " K:Na	3.7	3.8	1.8	4.5	5.0
Base saturation (%)	68.6	72.7	80.1	82.9	79.2
pH (water, 1:2.5)	6.7	6.7	7.35	7.5	7.55
5. Carbonate content (%)	0.5	0.1	2.0	1.0	2.0

*Gravel expressed as percentage of whole soil; sand, silt, and clay as percentage of <2.0 mm fraction

Table 38 Soil analytical data - WT-II

profile depth = 28 cm

SAMPLE DEPTH/SHALLOW SOIL NO. ( ):	10 cm	20 cm	SHALLOW SOILS (1)	(2)	(3)
1. Particle-size distribution* (% by weight)					
Gravel (6.0-2.0 mm)	1.2	0.5	0.0	0.0	0.0
Sand (0.06-2.0 mm)	7.4	2.8	8.4	1.7	9.6
Silt (0.002-0.06 mm)	29.9	31.2	24.2	25.5	26.7
Clay (<0.002 mm)	62.7	66.0	67.4	72.8	63.7
2. Other physical properties					
Colour (moist soil)	5YR3/3	5YR3/3	5YR3/1	5YR3/1	5YR2.5/1
Bulk density ($g.cm^{-3}$)	0.86	0.93	n.d.	n.d.	n.d.
Particle density ($g.cm^{-3}$)	2.72	2.78	n.d.	n.d.	n.d.
Pore space (% by volume)	68.4	66.5	n.d.	n.d.	n.d.
3. Organic matter (% by weight)					
Carbon	2.29	2.27	3.61	4.54	4.27
Loss on ignition	15.0	13.2	26.7	29.0	36.1
4. Cation exchange properties					
C.E.C. (m.e./100 g)	33.40	28.54	27.26	31.09	33.71
Exchangeable Ca (m.e./100 g)	17.84	18.10	17.40	17.88	23.55
" Mg (")	5.08	7.17	4.66	5.32	4.90
" K (")	0.29	0.34	0.32	0.42	0.38
" Na (")	0.09	0.08	0.08	0.10	0.06
Molar ratio Mg:Ca	0.285	0.396	0.268	0.298	0.208
" " K:Na	3.2	4.3	4.0	4.2	6.3
Base saturation (%)	69.8	90.0	82.4	76.3	85.7
pH (water, 1:2.5)	7.15	7.6	7.55	7.25	7.5
5. Carbonate content (%)	0.5	0.5	1.0	0.5	1.0

*Gravel expressed as percentage of whole soil; sand, silt, and clay as percentage of <2.0 mm fraction

Table 39 Soil analytical data - GKV-II

profile depth = 32 cm

SAMPLE DEPTH (cm)	10	20	30
1. Particle-size distribution* (% by weight)			
Gravel (6.0-2.0 mm)	0.8	1.4	2.1
Sand (0.06-2.0 mm)	9.2	2.6	4.3
Silt (0.002-0.06 mm)	36.4	37.8	25.9
Clay (<0.002 mm)	54.4	59.6	69.8
2. Other physical properties			
Colour (moist soil)	7.5YR5/4	7.5YR5/6	7.5YR5/6
Bulk density ($g.cm^{-3}$)	1.10	1.13	1.36
Particle density ($g.cm^{-3}$)	2.71	2.82	2.84
Pore space (% by volume)	59.4	53.5	52.1
3. Organic matter (% by weight)			
Carbon	2.17	1.16	0.81
Loss on ignition	8.9	7.4	7.6
4. Cation exchange properties			
C.E.C. (m.e./100 g)	21.34	15.38	10.64
Exchangeable Ca (m.e./100 g)	10.18	9.50	6.21
" Mg (")	3.48	2.52	1.56
" K (")	0.38	0.20	0.16
" Na (")	0.10	0.06	0.06
Molar ratio Mg:Ca	0.342	0.265	0.251
" " K:Na	3.8	3.3	2.7
Base saturation (%)	66.3	79.8	75.5
pH (water, 1:2.5)	7.0	7.35	7.35
5. Carbonate content (%)	0.5	0.5	0.5

*Gravel expressed as percentage of whole soil; sand, silt, and clay as percentage of <2.0 mm fraction

Table 40 Soil analytical data - GT-II

profile depth = 72 cm

SAMPLE DEPTH (cm)	10	20	30	40	50	60	70
1. Particle-size distribution* (% by weight)							
Gravel (6.0-2.0 mm)	1.4	0.0	0.8	1.1	2.7	1.0	0.2
Sand (0.06-2.0 mm)	27.2	22.8	18.3	19.2	16.7	21.4	18.5
Silt (0.002-0.06 mm)	33.5	30.8	33.7	30.5	34.2	33.4	30.3
Clay (<0.002 mm)	39.3	46.4	48.0	50.3	49.1	45.2	51.2
2. Other physical properties							
Colour (moist soil)	2.5YR3/2	2.5YR3/4	2.5YR3/4	2.5YR3/4	2.5YR3/4	2.5YR3/4	2.5YR3/4
Bulk density ($g.cm^{-3}$)	0.67	0.68	0.72	0.80	0.81	0.82	0.87
Particle density ($g.cm^{-3}$)	2.82	2.87	2.89	2.91	2.90	2.91	2.90
Pore space (% by volume)	76.2	76.3	75.1	72.5	72.1	71.8	70.0
3. Organic matter (% by weight)							
Carbon	2.16	0.82	0.72	0.63	0.66	0.69	0.55
Loss on ignition	0.5	0.1	0.5	0.5	1.0	0.5	1.0
4. Cation exchange properties							
C.E.C. (m.e./100 g)	17.50	8.32	6.09	6.72	6.90	6.37	5.84
Exchangeable Ca (m.e./100 g)	12.03	4.39	2.95	3.42	3.34	3.45	3.58
" Mg (")	1.72	0.52	0.38	0.40	0.36	0.21	0.31
" K (")	0.11	0.03	0.03	0.01	0.02	0.01	0.01
" Na (")	0.02	0.01	0.01	0.01	0.01	0.01	0.01
Molar ratio Mg:Ca	0.143	0.118	0.129	0.117	0.108	0.081	0.087
" " K:Na	5.5	3.0	3.0	1.0	2.0	1.0	1.0
Base saturation (%)	79.3	59.5	55.3	57.1	54.1	54.9	64.6
pH (water, 1:2.5)	6.75	6.25	6.3	6.5	6.35	6.5	6.55
5. Carbonate content (%)	0.5	0.1	0.5	0.5	1.0	0.5	1.0

*Gravel expressed as percentage of whole soil; sand, silt, and clay as percentage of <2.0 mm fraction

Table 41 Soil analytical data - WT-I

profile depth = 79 cm

SAMPLE DEPTH (cm)	10	20	30	40	50	60	70
1. Particle-size distribution* (% by weight)							
Gravel (6.0-2.0 mm)	0.6	0.0	1.8	1.3	0.0	0.0	0.4
Sand (0.06-2.0 mm)	18.1	8.3	9.2	6.3	2.3	6.5	2.8
Silt (0.002-0.06 mm)	25.7	23.6	24.8	18.2	23.4	22.7	25.3
Clay (<0.002 mm)	56.2	68.1	66.0	75.5	74.3	70.8	71.9
2. Other physical properties							
Colour (moist soil)	2.5YR3/4	2.5YR3/4	2.5YR3/4	2.5YR3/4	2.5YR3/4	2.5YR3/4	2.5YR3/4
Bulk density ($g.cm^{-3}$)	0.72	0.80	0.80	0.75	0.91	0.85	0.94
Particle density ($g.cm^{-3}$)	2.75	2.80	2.81	2.86	2.84	2.82	2.81
Pore space (% by volume)	73.8	71.4	71.5	73.8	68.0	69.9	66.5
3. Organic matter (% by weight)							
Carbon	1.31	1.18	1.22	0.98	0.94	0.91	0.84
Loss on ignition	13.4	11.4	12.8	12.3	12.1	11.8	12.1
4. Cation exchange properties							
C.E.C. (m.e./100 g)	14.44	15.49	14.65	16.87	15.58	14.76	15.82
Exchangeable Ca (m.e./100 g)	6.36	4.86	4.23	5.01	4.51	5.18	5.45
" Mg (")	4.25	2.94	3.54	2.60	2.72	2.54	2.60
" K (")	0.18	0.08	0.10	0.09	0.07	0.06	0.07
" Na (")	0.05	0.05	0.05	0.04	0.05	0.04	0.04
Molar ratio Mg:Ca	0.668	0.605	0.837	0.519	0.470	0.490	0.447
" " K:Na	3.6	1.6	2.0	2.3	1.4	1.5	1.8
Base saturation (%)	75.1	51.2	54.1	45.9	47.2	53.0	51.6
pH (water, 1:2.5)	6.2	6.0	5.95	5.75	5.7	6.1	6.0
5. Carbonate content (%)	0.5	0.5	0.5	0.5	0.5	0.5	1.0

*Gravel expressed as percentage of whole soil; sand, silt, and clay as percentage of <2.0 mm fraction

Table 42 Soil analytical data - GKV-I

profile depth = 102 cm

SAMPLE DEPTH (cm):	10	20	30	50	70	100
1. Particle-size distribution* (% by weight)						
Gravel (6.0-2.0 mm)	3.7	2.3	3.4	8.0	12.2	6.5
Sand (0.06-2.0 mm)	9.1	5.2	2.1	2.9	11.7	9.4
Silt (0.002-0.06 mm)	29.6	26.4	24.8	25.1	16.4	20.3
Clay (<0.002 mm)	61.3	68.4	73.1	72.0	71.9	70.3
2. Other physical properties						
Colour (moist soil)	7.5YR5/6	7.5YR5/8	7.5YR5/8	7.5YR5/6	7.5YR5/8	7.5YR5/6
Bulk density ($g.cm^{-3}$)	1.22	1.14	1.26	1.30	1.30	1.33
Particle density ($g.cm^{-3}$)	2.82	2.80	2.82	2.77	2.81	2.81
Pore space (% by volume)	56.7	59.3	55.3	53.1	53.7	52.7
3. Organic matter (% by weight)						
Carbon	0.80	0.45	0.46	0.36	0.29	0.07
Loss on ignition	7.4	8.1	7.9	8.1	8.1	6.9
4. Cation exchange properties						
C.E.C. (m.e./100 g)	14.53	13.98	13.72	14.51	14.81	14.42
Exchangeable Ca (m.e./100 g)	1.84	1.52	1.96	1.49	1.66	2.48
" Mg (")	1.04	0.75	0.92	0.57	0.64	0.87
" K (")	0.37	0.28	0.26	0.13	0.10	0.10
" Na (")	0.07	0.06	0.07	0.04	0.06	0.05
Molar ratio Mg:Ca	0.565	0.493	0.469	0.383	0.386	0.351
" " K:Na	5.3	4.7	3.7	3.3	1.8	2.0
Base saturation (%)	22.9	18.7	23.4	15.4	16.7	24.3
pH (water, 1:2.5)	5.4	5.5	5.4	5.5	5.6	5.6
5. Carbonate content (%)	0.5	0.5	0.5	0.5	0.5	0.5

*Gravel expressed as percentage of whole soil; sand, silt, and clay as percentage of <2.0 mm fraction

Table 43 Soil analytical data - ABT-I

profile depth = 75 cm

SAMPLE DEPTH (cm):	10	20	30	40	50	60	70
1. Particle-size distribution* (% by weight)							
Gravel (6.0-2.0 mm)	0.0	0.0	0.0	0.0	0.6	0.0	0.0
Sand (0.06-2.0 mm)	30.2	21.6	25.9	20.1	12.7	14.3	13.8
Silt (0.002-0.06 mm)	30.5	40.0	30.3	29.7	38.5	31.5	32.6
Clay (<0.002 mm)	39.3	38.4	43.8	50.2	48.8	54.2	53.6
2. Other physical properties							
Colour (moist soil)	10YR/3	10YR3/3	10YR3/4	10YR3/4	10YR3/4	10YR3/4	10YR3/4
Bulk density ($g.cm^{-3}$)	0.97	1.02	1.10	1.16	1.14	1.16	1.26
Particle density ($g.cm^{-3}$)	2.71	2.75	2.75	2.77	2.79	2.74	2.75
Pore space (% by volume)	64.2	62.9	60.0	58.1	59.1	57.7	54.2
3. Organic matter (% by weight)							
Carbon	0.85	0.62	0.42	0.32	0.28	0.31	0.31
Loss on ignition	9.7	9.1	8.5	8.4	8.6	8.6	8.9
4. Cation exchange properties							
C.E.C. (m.e./100 g)	14.76	17.96	15.43	12.36	11.27	12.96	17.16
Exchangeable Ca (m.e./100 g)	8.86	12.07	9.68	8.01	5.59	7.24	11.25
" Mg (")	1.16	0.84	0.55	0.35	0.26	0.28	0.29
" K (")	0.04	0.05	0.04	0.04	0.03	0.03	0.05
" Na (")	0.01	0.01	0.01	0.01	0.02	0.01	0.02
Molar ratio Mg:Ca	0.131	0.070	0.057	0.044	0.047	0.039	0.026
" " K:Na	4.0	5.0	4.0	4.0	1.5	3.0	2.5
Base saturation (%)	68.2	72.2	66.6	68.0	52.35	58.33	67.66
pH (water, 1:2.5)	6.65	6.7	6.5	6.2	6.1	6.2	6.5
5. Carbonate content (%)	0.5	0.5	0.5	0.1	0.1	0.5	0.1

*Gravel expressed as percentage of whole soil; sand, silt, and clay as percentage of <2.0 mm fraction

Table 44 Soil analytical data - ABT-II

profile depth = 68 cm

SAMPLE DEPTH (cm):	10	20	30	40	50	60
1. Particle-size distribution* (% by weight)						
Gravel (6.0-2.0 mm)	1.5	0.9	0.0	0.8	0.0	0.0
Sand (0.06-2.0 mm)	40.2	41.4	38.8	30.0	31.2	31.6
Silt (0.002-0.06 mm)	31.2	32.1	28.0	33.1	34.7	29.5
Clay (<0.002 mm)	28.6	26.5	33.2	36.9	34.1	38.9
2. Other physical properties						
Colour (moist soil)	10YR3/1	10YR3/1	10YR3/1	10YR3/1	10YR3/1	10YR3/1
Bulk density ($g.cm^{-3}$)	0.93	1.06	1.14	1.27	1.28	1.27
Particle density ($g.cm^{-3}$)	2.64	2.74	2.72	2.76	2.79	2.78
Pore space (% by volume)	64.8	61.3	58.1	54.0	54.1	54.3
3. Organic matter (% by weight)						
Carbon	1.40	1.02	0.92	0.81	0.76	0.75
Loss on ignition	10.2	8.4	8.0	8.5	8.3	8.1
4. Cation exchange properties						
C.E.C. (m.e./100 g)	27.82	22.60	23.39	21.36	22.84	22.47
Exchangeable Ca (m.e./100 g)	17.92	15.09	13.34	13.65	15.72	16.21
" Mg (")	2.46	2.08	1.52	1.78	1.95	1.89
" K (")	0.15	0.11	0.10	0.10	0.06	0.05
" Na (")	0.05	0.06	0.05	0.06	0.06	0.04
Molar ratio Mg:Ca	0.137	0.138	0.114	0.130	0.124	0.117
" " K:Na	3.0	1.8	2.0	1.7	1.0	1.3
Base saturation (%)	74.0	76.7	64.2	73.0	77.9	81.0
pH (water, 1:2.5)	7.0	6.8	6.6	6.7	6.7	6.8
5. Carbonate content (%)	0.5	0.5	0.5	1.0	0.5	0.5

*Gravel expressed as percentage of whole soil; sand, silt, and clay as percentage of <2.0 mm fraction

Table 45 Soil analytical data - GT-I

profile depth = 77 cm

SAMPLE DEPTH (cm):	10	20	30	40	50	60	70
1. Particle-size distribution* (% by weight)							
Gravel (6.0-2.0 mm)	1.9	0.2	0.0	0.0	0.4	0.0	0.0
Sand (0.06-2.0 mm)	49.2	42.6	36.0	34.2	31.7	34.6	36.5
Silt (0.002-0.06 mm)	18.7	23.9	25.6	21.4	22.4	18.5	20.3
Clay (<0.002 mm)	32.1	33.5	38.4	44.4	45.9	46.9	43.2
2. Other physical properties							
Colour (moist soil)	10YR3/3	10YR4/4	10YR4/4	10YR4/4	10YR4/4	10YR4/4	10YR4/4
Bulk density ($g.cm^{-3}$)	1.06	1.13	1.22	1.42	1.46	1.46	1.65
Particle density ($g.cm^{-3}$)	2.64	2.62	2.67	2.71	2.72	2.71	2.73
Pore space (% by volume)	59.8	56.9	54.3	47.6	46.3	40.6	39.6
3. Organic matter (% by weight)							
Carbon	0.84	0.55	0.38	0.34	0.33	0.17	0.16
Loss on ignition	7.1	7.1	7.2	7.3	6.7	5.8	6.0
4. Cation exchange properties							
C.E.C. (m.e./100 g)	18.88	13.05	9.10	8.92	7.38	8.55	7.68
Exchangeable Ca (m.e./100 g)	10.35	7.80	5.73	4.85	3.81	4.86	5.03
" Mg (")	2.09	1.47	0.84	0.58	0.45	0.55	0.73
" K (")	0.26	0.19	0.13	0.10	0.05	0.07	0.05
" Na (")	0.06	0.05	0.03	0.03	0.02	0.02	0.02
Molar ratio Mg:Ca	0.202	0.188	0.147	0.120	0.118	0.113	0.145
" " K:Na	4.3	3.8	4.3	3.3	2.5	3.5	2.5
Base saturation (%)	67.6	72.9	74.0	62.3	58.7	64.3	75.9
pH (water, 1:2.5)	6.85	6.75	6.5	6.15	6.1	6.25	6.5
5. Carbonate content (%)	0.1	0.1	0.5	1.0	1.0	1.0	1.0

*Gravel expressed as percentage of whole soil; sand, silt, and clay as percentage of <2.0 mm fraction

Table 46 Basal area of tree trunks weighted to provide a measure of the canopy cover in the central sub-plots of the six main sampling plots (details of calculation, Chapter IV, Section E).

PLOT	WEIGHTED BASAL AREA INDEX (m^2/ha)
GT-II	22.0
WT-I	25.4
GKV-I	69.7
ABT-I	37.9
ABT-II	10.6
GT-I	53.8

Table 47 Percentage of rainfall intercepted by vegetation canopy in the six plots.

	n[1]	MEAN	MINIMUM[2]	MAXIMUM[2]	S.DEV.
GT-II	5	27.6	15.2	34.6	8.13
WT-I	5	18.9	8.1	30.2	7.05
GKV-I	5	34.3	26.8	43.6	6.50
ABT-I	12	26.9	8.2	47.4	5.98
ABT-I	12	24.6	13.2	35.8	7.04
GT-I	9	32.5	22.1	45.3	7.33

Notes: 1. Number of occasions on which measurements were made.
2. Average values for four collecting funnels.

Table 48 Chemical properties of throughfall in the six plots.

	pH	SPECIFIC CONDUCTANCE µmho.cm^{-1},25°C	Ca^{2+} ppm	Mg^{2+} ppm	K^{+} ppm	Na^{+} ppm
GT-II (n=14)						
mean	6.69	77.6	5.36	1.13	6.71	0.57
minimum	5.82	43.0	2.16	0.36	1.75	0.12
maximum	7.43	139.0	8.98	1.90	13.35	0.93
s. dev.	0.49	25.1	1.76	0.42	3.42	0.25
WT-I (n=16)						
mean	6.76	35.5	2.76	0.97	3.85	0.66
minimum	5.78	20.0	0.85	0.55	1.55	0.28
maximum	7.60	49.0	5.80	1.40	7.70	0.92
s. dev.	0.57	9.6	1.19	0.23	1.70	0.18
GKV-I (n=16)						
mean	6.72	85.1	6.29	1.76	6.91	0.71
minimum	5.45	40.0	3.08	0.90	0.88	0.10
maximum	7.51	147.0	12.6	3.61	15.2	1.37
s. dev.	0.64	31.7	2.41	0.71	3.80	0.33
ABT-I (n=28)						
mean	6.52	80.2	7.04	1.99	5.71	0.50
minimum	5.80	34.0	1.50	0.85	1.55	0.08
maximum	7.32	162.0	14.0	3.50	9.0	1.41
s. dev.	0.41	30.4	3.5	0.77	2.34	0.40
ABT-II (n=26)						
mean	6.57	70.5	5.74	1.21	6.98	0.60
minimum	6.05	26.0	2.10	0.42	2.25	0.09
maximum	7.13	132.0	10.2	1.86	12.2	1.67
s. dev.	0.27	25.7	2.3	0.39	3.08	0.43
GT-I (n=28)						
mean	6.01	44.61	3.91	0.73	6.09	0.39
minimum	5.22	15.0	1.30	0.15	1.15	0.07
maximum	6.63	116.0	9.80	1.96	19.0	1.80
s. dev.	0.38	23.96	2.89	0.47	5.18	0.41
ABT-III/IV (n=20)						
mean	6.14	17.79	2.43	0.30	0.64	0.20
minimum	5.70	9.0	1.20	0.20	0.21	0.15
maximum	6.60	33.0	4.00	0.50	1.80	0.30
s. dev.	0.27	5.82	0.95	0.09	0.42	0.05

Table 49 Matrix of the closer correlations (r_s) between the chemical properties of throughfall waters (n=7).

	Specific conductance	Ca	Mg	K	Na
pH	0.500				0.893[1]
Specific conductance		0.893	0.929	0.571	0.643
Ca			0.964	0.607	
Mg				0.536	0.536
K					0.536

[1] Coefficients underlined are significant at the 95 per cent confidence level.

Table 50 Weight of fine litterfall (t/ha) in the six plots during the one-year study period (1.8.74 to 31.7.75).

	LEAVES	TWIGS	MISCELLANEOUS FRACTION[1]	TOTAL
GT-II	3.769	0.749	0.671	5.189
WT-I	4.044	0.529	1.161	5.734
GKV-I	8.940	1.857	1.251	12.048
ABT-I	6.257	0.889	0.721	7.867
ABT-II	4.376	0.960	0.708	6.044
GT-I	4.494	1.029	0.888	6.411

[1]Comprises mostly bracts, scales, flowers and fruits.

Table 51 Mean nutrient contents (percent by weight) and the more important molar ratios of litter in the six plots.

	LEAVES	TWIGS	MISCELLANEOUS FRACTION
Ash	12.68	6.38	6.60
Ca	4.23	2.43	3.72
Mg	0.39	0.20	0.31
K	0.38	0.17	0.45
Na	0.025	0.014	0.026
Si	1.41	0.38	0.69
Fe	0.012	0.053	0.015
Al	0.025	0.154	0.075
Molar(Ca+Mg:K+Na)	12.30	13.95	8.58
Molar Mg:Ca	0.153	0.145	0.140
Molar K:Na	8.12	7.08	11.06

Table 52 Components of nutrient input at the soil surface (kg/ha) averaged over the six plots (percentage contribution in brackets) during the one-year study period (1.8.74 to 31.7.75).

	Ca	Mg	K	Na
Rainfall	21.79 (5.98)	2.13 (4.55)	3.15 (2.42)	6.00 (50.38)
Canopy leaching	66.03 (18.13)	19.51 (41.64)	100.54 (77.20)	4.01 (33.67)
Litterfall	276.46 (75.89)	25.21 (53.81)	26.54 (20.38)	1.90 (15.95)
	364.28	46.85	130.23	11.91

Table 53 Ca+Mg:K+Na ratios of the various components of nutrient flow in the six plots.

	GT-II	WT-I	GKV-I	ABT-I	ABT-II	GT-I	MEAN
Rainfall	0.79	1.99	1.99	3.30	3.30	0.79	2.03
Canopy leaching	0.91	0.67	1.01	1.33	0.73	0.73	0.90
Litterfall	9.05	11.18	7.24	20.32	14.83	7.91	11.76
Total from vegetation	2.08	2.82	3.25	4.78	2.66	2.66	3.04
Total surface input	1.96	2.72	3.15	4.64	2.71	2.47	2.94
Exchangeable cations	62.8	29.2	4.9	89.6	58.0	22.8	44.6

Table 54 Mg:Ca ratios of nutrient stores and components of nutrient flow in the six plots.

	GT-II	WT-I	GKV-I	ABT-I	ABT-II	GT-I	MEAN
Rainfall	0.205	0.145	0.145	0.156	0.156	0.205	0.169
Canopy leaching	0.363	0.942	0.564	0.590	0.446	0.326	0.539
Litterfall	0.135	0.173	0.180	0.146	0.140	0.127	0.150
Total from vegetation	0.211	0.267	0.240	0.223	0.200	0.162	0.217
Total surface input	0.211	0.256	0.229	0.219	0.196	0.163	0.212
Exchangeable cations in soil:							
10 cm	0.143	0.668	0.565	0.131	0.137	0.202	0.308
Base of profile	0.087	0.477	0.351	0.026	0.117	0.145	0.201
Mean	0.112	0.581	0.441	0.059	0.127	0.148	0.245
Limestone bedrock	0.014	0.367	0.074	0.017	0.008	0.018	0.083

Table 55 K:Na ratios of the various components of nutrient flow in the six plots.

	GT-II	WT-I	GKV-I	ABT-I	ABT-II	GT-I	MEAN
Rainfall	0.259	0.295	0.295	0.421	0.421	0.295	0.325
Canopy leaching	13.79	6.40	12.57	15.21	12.40	n.d.[1]	12.07
Litterfall	5.75	3.60	12.06	9.34	6.92	12.19	8.31
Total from vegetation	11.63	5.58	12.39	23.26	11.24	68.22	22.05
Total surface input	5.90	3.46	7.03	7.06	6.86	9.78	6.68
Exchangeable cations	2.36	2.01	3.46	3.43	1.79	3.47	2.75

[1] No sodium leached from canopy

Table 56 Equilibrium equations and constants for the main reactions in the $CaCO_3$-CO_2-H_2O system, based on Picknett *et al.*, 1976.

EQUATION[1]	EQUILIBRIUM CONSTANT 10°C	EQUILIBRIUM CONSTANT 25°C	SOURCE[2]
$K_o = (CO_2^o)/(H_2CO_3)$	493	381	A,B
$K_1 = (H^+).(HCO_3^-)/(H_2CO_3)$	1.7×10^{-4}	1.7×10^{-4}	B
$K_2 = (H^+).(CO_3^{2-})/(HCO_3^-)$	3.24×10^{-11}	4.69×10^{-11}	C
$K_3 = (Ca^{2+}).(HCO_3^-)/(CaHCO_3^+)$	0.13	0.04	D
$K_4 = (Ca^{2+}).(CO_3^{2-})/(CaCO_3^o)$	1×10^{-4}	4×10^{-4}	D
K_s(calcite) $= (Ca^{2+}).(CO_3^{2-})$	4.0×10^{-9}	3.7×10^{-9}	D
$Kc = [CO_2^o]/P_{CO_2}$	0.0526	0.0329	E
$K_\omega = (H^+).(OH^-)$	2.92×10^{-15}	1.01×10^{-14}	F

Notes: 1. Square brackets represent the molar concentration of the ion or molecule, while round brackets represent the activity of a particular ion with respect to a standard state at infinite dilution (Garrels and Christ, 1965).
2. Sources: A. Harned and Owen (1958); B. Wissbrun *et al.* (1954); C. Harned and Scholes (1941): D. Picknett (1973); E. Siedell (1958); F. Harned and Robinson (1941).

Table 57 Percentage soil and litter cover on summit of Ak. Bt. Takun.

	40m x 30m PLOT [1]	64 MICRO-CATCHMENTS
Bare limestone)	66.3	24.1
Rock covered by lichen)		54.2
Rock covered by organic litter	7.9	7.4
Shallow soil, depth <10cm	19.3	14.3
Deeper soil, depth >10cm	6.5	-

[1] Based on four transect lines in ABT-III (3) and ABT-IV (1).

Table 58 Characteristics of the 64 micro-catchments.

	MEAN	MINIMUM	MAXIMUM	S. DEV.
Length[1] (cm)	98.0	10.0	380.0	68.1
Width[2] (cm)	21.9	7.0	84.0	15.0
Depth[3] (cm)	10.8	2.0	32.5	7.7
Surface area[4] (m^2)	0.21	0.01	2.52	0.47
Gradient[5] (degrees)	49.4	23.0	85.0	14.7
Soil/litter cover (%)	21.7	0.0	95.0	26.1

Notes:
1. Distance between collection point and uppermost point in micro-catchment, measured along the line of the main channel.
2. Average width, based on measurements made at points ¼, ½, and ¾ of the way along the main channel.
3. Average distance, measured at right-angles to the gradient, from the base of the flute to the straight line joining the width extremities measured in (2).
4. Length x width.
5. Average gradient of the line measured in (1).

Table 59 Details of the five storms sampled.

AMOUNT OF RAINFALL (mm)	DURATION (min)	INTENSITY ($mm.hr^{-1}$)
35.0	110	19.1
15.0	134	6.7
21.8	45	29.1
11.4	170	4.0
11.2	40	16.8

Table 60 Properties of runoff from micro-catchments, Ak. Bt. Takun (n=64). Values for each site are based on the arithmetic means of 5 storms.

	MEAN	MINIMUM	MAXIMUM	S.DEV.
pH	7.84	7.39	8.63	0.27
Specific conductance (μmho.cm^{-1}, 25°C)	85.24	4.40	173.4	0.88
Total hardness (ppm)	51.83	22.3	108.2	21.18
Calcium hardness (ppm)	48.01	21.4	102.2	18.74
Magnesium hardness (ppm)	3.82	0.5	18.7	3.16
Alkaline hardness (ppm)	46.79	22.0	96.2	17.67
Non-alkaline hardness (ppm)	5.04	-3.1	22.3	6.85
Potassium (ppm)	1.50	0.1	5.3	1.36
Sodium (ppm)	0.36	0.14	2.28	0.36

Table 61 Matrix of correlation coefficients (r), significant at 95% confidence level, for relationships between chemical properties of runoff from bare and partially-covered micro-catchments, Ak.Bt. Takun (n=64).

	pH	SPEC. COND.	TOTAL HARD.	ALK. HARD.	NON-ALK. HARD.	K
Specific conductance	-0.294					
Total hardness	-0.239	0.955				
Alkaline hardness		0.865	0.941			
Non-alkaline hardness	-0.383[1]	0.689	0.651	0.369		
K	-0.682	0.592	0.575	0.490	0.481	
Na	-0.266	0.477	0.424	0.234	0.660	0.401

1. Coefficients underlined are significant at 99.9 per cent confidence level.

Table 62 Matrix of correlation coefficients (r), significant at 95% confidence level, showing relationships between the chemical properties of runoff and selected characteristics of the micro-catchments (n=64)

MICRO-CATCHMENT CHARACTERISTICS	pH	SPEC. COND.	TOTAL HARD.	ALK. HARD	NON-ALK. HARD.	K^+	Na^+
Length							
Width							
Surface area							
Depth		0.241					0.212
Width / Depth							
Gradient	0.506[1]	-0.460	-0.418	-0.264	-0.539	-0.472	-0.572
Soil/litter cover (%)	-0.490	0.679	0.622	0.405	0.792	0.516	0.552
Soil/litter cover (g/m^2)	-0.355	0.660	0.544	0.317	0.772	0.362	0.477

[1]Coefficients underlined are significant at 99.9 per cent confidence level.

Table 63 Properties of runoff from bare rock micro-catchments, Ak.Bt.Takun (n=27). Values for each site are based on the arithemetic means of 5 storms.

	MEAN	MINIMUM	MAXIMUM	S.DEV.
pH	8.02	7.61	8.63	0.26
Specific conductance ($\mu mho.cm^{-1}$, $25^{o}C$)	61.10	44.0	98.8	12.60
Total hardness (ppm)	36.30	22.3	66.8	9.31
Calcium hardness (ppm)	34.89	21.4	65.0	9.43
Magnesium hardness (ppm)	1.41	0.5	2.6	0.51
Alkaline hardness (ppm)	36.64	22.0	65.2	9.43
Non-alkaline hardness (ppm)	-0.34	-3.1	5.2	2.13
Potassium (ppm)	0.68	0.1	2.4	0.63
Sodium (ppm)	0.23	0.14	0.42	0.06

Table 64 Properties of runoff from micro-catchments with partial soil/litter cover, Ak. Bt. Takun (n=37). Values for each site are based on the arithmetic means of 5 storms.

	MEAN	MINIMUM	MAXIMUM	S.DEV.
pH	7.71	7.39	8.25	0.19
Specific conductance (μmho.cm^{-1}, 25°C)	102.86	54.2	137.4	27.09
Total hardness (ppm)	63.17	38.8	108.2	20.22
Calcium hardness (ppm)	57.58	36.6	102.2	18.19
Magnesium hardness (ppm)	5.59	1.7	18.7	3.11
Alkaline hardness (ppm)	54.20	27.6	96.2	18.65
Non-alkaline hardness (ppm)	8.97	-2.0	22.3	6.41
Potassium (ppm)	2.10	0.2	5.3	1.44
Sodium (ppm)	0.46	0.20	2.28	0.45

Table 65 Details of sites used in investigating the effects of surface cover, micro-catchment size and channel gradient upon the properties of runoff.

	A. SOIL/LITTER COVER			B. CHANNEL GRADIENT			C. CATCHMENT SIZE		
Site number:	4	17	15	55	42	12	10	2	11
Morphometry:									
length (cm)	90	80	90	80	80	90	190	70	30
width (cm)	25	22	23	9	10	9	35	11	9
gradient (degrees)	53	50	55	46	54	66	60	64	64
Surface cover:									
bare rock (%)	87	90	30	40	30	40	80	60	75
lichen (%)	13	5	0	60	70	60	20	40	25
soil/litter (%)	0	5	70	0	0	0	0	0	0
soil/litter, dry weight (g)	0	5	190	0	0	0	0	0	0
snails (number/100 cm^2)	0	0	0	3	5	3	0	0	0
Karst water properties (average of 5 storms):									
specific conductance (µmho)	50.6	89.3	141.2	73.2	72.0	56.6	58.4	50.2	59.0
total hardness (ppm)	28.6	54.0	95.0	37.6	44.6	32.2	36.7	29.9	33.9
non-alkaline hardness[1] (ppm)	0.4	8.8	14.2	1.55	<0.0	<0.0	<0.0	<0.0	<0.0

[1]Values <0.0 are all within range of analytical error.

Table 66 Depths of surface lowering (microns) in bare and partially-covered micro-catchments during the 5 storms, calculated on assumption that no water is lost by evaporation or evapotranspiration[1].

STORM MAGNITUDE (mm)	A. BARE ROCK SITES (microns)	B. SITES WITH PARTIAL SOIL/LITTER COVER (microns)
35.0	0.37	0.56
15.0	0.23	0.37
21.8	0.19	0.38
11.4	0.18	0.34
11.2	0.13	0.29

[1] See Chapter II, Section F1.1.

Table 67 Concentrations of carbon dioxide (% by volume) in the atmosphere of the six karst soils.

PROFILE NO.	DEPTH (cm)	n	MEAN	MINIMUM	MAXIMUM	S. DEV.
GT-II						
1	15	4	0.35	0.22	0.53	0.14
1	30	4	0.57	0.39	0.79	0.18
1	60	4	0.73	0.49	1.04	0.24
WT-I						
1	15	5	0.28	0.16	0.45	0.11
1	30	5	0.37	0.27	0.51	0.10
1	60	4	0.78	0.53	1.21	0.31
GKV-I						
1	15	4	1.32	1.08	1.58	0.21
1	30	4	2.84	2.35	3.26	0.37
1	60	4	4.21	2.66	5.83	1.34
ABT-I						
1	15	9	0.77	0.15	1.38	0.43
1	30	11	1.23	0.69	1.68	0.35
2	30	15	1.13	0.52	1.88	0.37
1	60	11	2.96	1.55	4.19	0.90
3	60	11	1.34	0.98	1.83	0.30
ABT-II						
1	15	7	0.66	0.26	0.96	0.25
1	30	8	1.04	0.52	1.87	0.45
2	30	14	1.08	0.37	1.50	0.35
1	60	12	1.26	0.65	2.18	0.46
3	60	8	1.28	0.97	1.60	0.24
GT-I						
1	15	7	0.65	0.38	0.96	0.21
2	15	4	0.35	0.18	0.55	0.16
1	30	7	0.94	0.71	1.36	0.26
2	30	4	0.84	0.61	1.08	0.19
1	60	7	1.94	1.03	3.67	0.98
2	60	4	1.78	0.78	2.85	0.86
3	60	5	1.81	1.23	2.52	0.46

Table 68 Significance of differences in mean carbon dioxide concentration between different sampling depths.

PLOT	SAMPLE DEPTHS 15 and 30	15 and 60	30 and 60
GT-II	*[1]	*	-
WT-I	-	*	*
GKV-I	*	*	*
ABT-I	*	*	*
ABT-II	*	*	-
GT-I	*	*	*

[1]Asterisks denote differences which are significant at 95% confidence level (randomization test).

Table 69 Significance of differences in mean carbon dioxide concentrations between the six plots at different sampling depths.

	GT-II	WT-I	GKV-I	ABT-I	ABT-II
15 cm					
WT-I	-				
GKV-I	*[1]	*			
ABT-I	*	*	*		
ABT-II	-	*	*	-	
GT-I	-	*	*	-	-
30 cm					
WT-I	-				
GKV-I	*	*			
ABT-I	*	*	*		
ABT-II	*	*	*	-	
GT-I	*	*	*	-	-
60 cm					
WT-I	-				
GKV-I	*	*			
ABT-I	*	*	*		
ABT-II	*	*	*	*	
GT-I	*	*	*	-	*

[1]Asterisks denote differences which are significant at 95% confidence level (randomization test).

Table 70 Soil moisture contents (% by volume) at 15 cm depth in the six karst soils.

Profile No.	n	MEAN	MINIMUM	MAXIMUM	S.DEV.
GT-II					
1	6	25.7	22.3	29.9	2.85
WT-I					
1	6	24.7	21.4	29.2	3.44
GKV-I					
1	6	34.0	28.9	39.4	4.37
ABT-I					
1	12	34.0	28.9	41.1	3.97
2	15	33.7	28.1	45.4	5.18
3	15	38.7	29.6	51.2	5.65
ABT-II					
1	16	42.2	34.2	69.1	8.51
2	16	38.0	31.0	51.0	5.17
3	15	40.4	32.8	50.2	5.57
GKV-I					
1	8	25.4	19.4	30.8	4.55
2	5	28.4	21.0	33.8	5.47
3	5	24.5	20.5	27.8	2.94

Table 71 Significance of differences in mean soil moisture contents (15 cm) between the six plots.

	GT-II	WT-I	GKV-I	ABT-I	ABT-II
WT-I	-				
GKV-I	*[1]	*			
ABT-I	*	*	-		
ABT-II	*	*	*	*	
GT-I	-	-	*	*	*

[1]Asterisks denote differences which are significant at 95% confidence level (randomization test).

Table 72 Average soil moisture contents (% by volume) at different depths in the six plots. Values are based on samples taken in October, January, April and July.

DEPTH (cm)	GT-II	WT-I	GKV-I	ABT-I profile(1)	ABT-II profile(1)	GT-I profile (1)
10	28.8	22.9	34.6	33.9	38.4	29.3
15	27.9	24.1	34.1	33.3	39.1	28.6
20	27.0	25.2	33.6	32.5	39.4	27.9
30	28.5	25.3	39.2	34.4	40.3	27.2
40	30.5	23.5	41.1	35.5	42.4	32.5
50	30.6	22.5	42.0	35.2	42.6	32.7
60	31.1	26.0	42.6	35.0	43.5	30.9
70	31.2	28.4	46.7	40.1	-	35.7
100	-	-	45.5	-	-	-

Table 73 Matrix of correlation coefficients (r_s) for the relationships between the carbon dioxide content of soil air at different depths in profile 1 of the three foothill plots.

PLOT	SAMPLE DEPTHS (cm)		
	15 and 30	15 and 60	30 and 60
ABT-I	0.849[1]	-0.647	-0.600
ABT-II	0.786	0.900	0.943
GT-I	0.500	0.714	0.500

[1]Coefficients underlined are significant at 95% confidence level.

Table 74 Matrix of correlation coefficients (r_s) for the relationships between soil moisture(15cm) and soil carbon dioxide in the three foothill plots.

PROFILE No.	SAMPLE DEPTH (cm) 15	30	60
ABT-I			
1	0.227	0.436	-0.281
2		0.290	
3			-0.022
ABT-II			
1	0.107	0.333	0.301
2		0.495[1]	
3			0.000
GT-I			
1	0.857	-0.143	0.429

[1]Coefficients underlined are significant at 95% confidence level.

Table 75 Matrix of correlation coefficients (r_s) for the relationships between the carbon dioxide content of soil air and litterfall in the antecedent period 1-60 days before sampling.

PROFILE No.	SAMPLE DEPTH (cm)		
	15	30	60
ABT-I			
1	0.387	0.300	0.218
2		-0.167	
3			0.027
ABT-II			
1	0.857[1]	0.477	0.490
2		0.503	
3			0.143
GT-I			
1	-0.429	-0.476	-0.429

[1]Coefficients underlined are significant at 95% confidence level.

Table 76 List of antecedent periods of rainfall with which soil carbon dioxide is most closely correlated. Correlation coefficients (r_s) for each period shown are positive and are statistically significant at the 95% confidence level.

PROFILE No.	SAMPLE DEPTH (cm) 15	30	60
ABT-I			
1	1-64	1-64	—
2		1-64	
3			—
ABT-II			
1	1-128	1-32	1-128
2		1-32	
3			1-128
GT-I			
1	1-32	1-64	1-32

Table 77 Average chemical characteristics of throughflow waters in the three foothill plots.

	ABT-I	ABT-II	CT-I
Temperature (°C)	n.d.	n.d.	n.d.
pH	7.05	7.28	7.12
Specific conductance (μmho)	273.4	362.0	278.6
Total hardness (ppm)	136.3	199.0	147.0
Alkaline hardness (ppm)	130.9	170.2	131.2
Non-alkaline hardness (ppm)	5.4	28.8	15.8
Calcium hardness (ppm)	129.5	175.1	124.1
Magnesium hardness (ppm)	6.8	23.9	23.9
K (ppm)	1.26	1.95	4.72
Na (ppm)	0.81	0.74	0.81
Non-alk:Total hardness (%)	4.0	16.9	10.8
Mg:Ca	0.053	0.136	0.193
K:Na	0.92	1.55	3.43

Table 78 Properties of seepage waters in Wang Tangga (n=16)

	MEAN	MINIMUM	MAXIMUM	S.DEV
Temperature (°C)	24.40	24.07	24.85	0.206
pH	7.36	6.87	7.72	0.249
Specific conductance (μmho)	543.1	437.9	621.0	58.0
Total hardness (ppm)	341.9	263.6	359.9	38.6
Alkaline hardness (ppm)	340.7	263.3	395.3	37.7
Non-alkaline hardness (ppm)	1.24	-2.54	5.87	2.24
Calcium hardness (ppm)	280.8	211.0	326.0	31.1
Magnesium hardness (ppm)	61.1	49.6	80.2	10.2
K (ppm)	0.51	0.23	0.85	0.154
Na (ppm)	4.25	3.68	4.83	0.302
Non-alk:Total hardness (%)	0.26	-1.04	1.50	0.65
Mg:Ca	0.219	0.180	0.260	0.026
K:Na	0.074	0.030	0.130	0.026

[1]Values based on 3 to 8 samples from each seepage point

Table 79 Estimated rates of erosion beneath mineral soils based on the total hardness of throughflow waters.

	AMOUNT OF THROUGHFLOW[1] (mm/yr)	TOTAL HARDNESS (ppm)	EROSION RATE[2] (mm/1000 yrs)
ABT-I	1479	136.3	65.2
ABT-II	1584	199.0	106.3
GT-I	3819	147.0	182.9
WT-I	1526	341.9	173.7

Notes: 1. Details of calculation given in Chapter VI, Section F1.

2. After correction for cation input from rainfall and runoff from rock outcrops (Chapter VI, Section F2), and assuming a rock density of 2.65 g/cm^3.

Table 80 Rates of limestone erosion (mm/1000 yrs) at different depths within mineral soils in the six plots, estimated by the erosional weight-loss tablet technique.

	DEPTH cm	MEAN[1]	MINIMUM	MAXIMUM
GT-II				
	15	1.45	1.08	1.92
	30	2.73	2.24	3.03
	60	4.09	3.36	4.96
WT-I				
	15	1.94	1.25	2.53
	30	1.79	1.19	2.48
	60	2.44	1.68	3.41
GKV-I				
	15	3.29	2.76	3.95
	30	3.92	1.84	5.31
	60	5.03	4.29	6.36
ABT-I				
	15	1.36	0.96	1.53
	30	1.41	1.26	1.50
	60	1.58	1.14	1.87
ABT-II				
	15	1.76	1.38	2.41
	30	2.27	1.44	3.15
	60	2.01	0.98	2.83
GT-I				
	15	0.95	0.89	1.04
	30	2.14	1.36	2.98
	60	2.26	1.74	2.45

[1]Values based on 3 tablets at each depth.

Table 81 Characteristics of the 11 cave streams sampled in the Setul Boundary Range. Locations shown in fig. 5.

CODE No[1]	SITE	FLOW TYPE[2]	CATCHMENT ZONE[3]	DISCHARGE (l/sec) MEAN	MINIMUM	MAXIMUM
4304	Perlis Mines' Cave - Main stream	A	I	282.5	70.0	2200.0
4019	Perlis Mines' Cave-tributary I	A	-	3.8	0.2	8.4
4025	Perlis Mines' Cave-tributary II	A	-	0.8	0.2	1.3
4305	Wang Tangga cave stream	B	I	4.1	0.6	10.8
4307	Kaki Bukit cave stream	A	I	14.6	0,8	52.6
4309	Rubber factory, m.s. 15, Kangar-Kaki Bukit Rd.	A	III	117.2	4.4	900.0
4310	Bt. Wang Mu - cave stream I	B	III	8.5	<0.1	75.0
4311	Bt. Wang Mu - cave stream II	A	III	48.5	<0.1	280.8
4315	Abi Head - main rising	B	IV	163.2	28.7	666.3
4316	Abi Head - subsidiary rising	B	IV	31.1	0.2	250.1
4318	Gua Ipoh	B	II	56.0	0.1	207.9

Notes: 1. Identification code adopted in present study.
2. Flow type: A. Vadose stream flow with free air circulation through open cave entrance;
B. Supposed vadose stream flow. Limited entry of air at rising, but air may enter the system at a higher level.
3. Catchment zone: I/II - areas of relatively pure limestone; III/IV - areas of comparatively impure limestone (see fig. 5 and Chapter VII, Section D3.3.ii).

Table 82 Characteristics of the three karst springs sampled. Locations shown in fig. 5.

CODE No[1]	SITE	SPRING TYPE[2]	DISCHARGE (l/sec) MEAN	MINIMUM	MAXIMUM
4306	Wang Tangga	B	3.1	<0.1	7.1
4317	Kg. Gtg. Kabu	A	5.4	0.7	9.6
4312	Sg. Gelang Kerbau, headwater rising	B	2.1	<0.1	5.8

Notes: 1. Identification code used in present study
2. Spring type: A. Deep-groundwater spring,
B. Shallow-groundwater spring.

Table 83 Details of river catchments (table continued on next page)

CODE NO[1]	SAMPLING STATION	AREA km^2	TOPOGRAPHY/ROCK TYPE (% OF CATCHMENT) LIMESTONE HILLS	KARST ALLUVIAL PLAIN	NON-CARBONATE LITHOLOGIES
Selangor					
1301	Sg. Batu[2] at Kg. Sg. Tua	51.0	0.0	0.0	100.0
1302	Sg. Batu at Bt. Batu	52.4	0.4	5.1	94.5
1303	Sg. Batu[2] at Sentul	145.0	0.9	30.6	88.5
1304	Sg. Gombak[3] at Jl. Pekeliling	140.0	1.0	34.0	65.0
1305	Sg. Klang[2] at Kg. Bahru	142.1	0.0	6.8	93.2
1306	Sg. Klang[2] at Jl. Brickfield	457.0	0.3	20.0	79.7
1314	Sg. Lui[2], D.I.D. catchment	69.0	0.0	0.0	100.0
1320	Sg. Rangkap	9.3	0.0	0.0	100.0
1321	Stream from granitic terrain, Templer Park	0.6	0.0	0.0	100.0
1322	Stream from granitic terrain, Templer Park	1.1	0.0	0.0	100.0
North-west region					
4301	Sg. Pelarit at Wang Tangga	17.4	45.7	11.2	43.1
4302	Sg. Pelarit at Kaki Bukit	17.7	46.3	11.1	42.6
4303	Sg. Pelarit at m.s. 12, Kaki Bukit-Kangar Rd.	61.4	56.1	3.2	40.7
4327	Headwaters of Sg. Gelang Kebau	0.5	100.0	0.0	0.0
4319	Sg. Tasoh at Kaki Bukit-Kangar Rd.	112.3	2.1	2.0	95.9[4]

Table 83 (ctd.) Details of river catchments

CODE NO[1]	SAMPLING STATION	AREA km^2	TOPOGRAPHY/ROCK TYPE (% OF CATCHMENT) LIMESTONE HILLS	KARST ALLUVIAL PLAIN	NON-CARBONATE LITHOLOGIES
Kinta Valley					
3301	Sg. Tempurong southern tributary	0.6	0.0	0.0	100.0
3302	Sg. Tempurong northern tributary	0.7	0.0	0.0	100.0
3303	Sg. Tempurong at eastern entrace of Gua Tempurong	1.9	7.2	15.3	77.5
3305	Sg. Kinta at Anderson Rd.	313.6	0.5	9.6	89.9
3306	Sg. Kinta at Lahat-Simpang Pulai Rd.	650.4	0.3	10.1	89.6
3307	Sg. Pinji at Kuala Lumpur-Ipoh Rd.	31.9	15.5	75.1	9.4
3308	Sg. Pinji at Lahat-Simpang Pulai Rd.	64.8	9.6	83.5	6.9
3309	East bank tributary of Sg. Sengat	2.2	16.3	83.7	0.0
3310	Sg. Sengat, 100 m upstream of Lahat-Simpang Pulai Rd.	6.5	10.3	77.5	12.2
3312	Sg. Johan at Ipoh-Batu Gajah Rd.	22.5	0.0	7.0	93.0
3313	Sg. Raia at Kuala Lumpur-Ipoh Rd.	210.3	3.6	5.1	91.3
3314	Sg. Tekah at Kuala Lumpur-Ipoh Rd.	1.5	8.5	64.7	26.8
3315	Sg. Gopeng at Kuala Lumpur-Ipoh Rd.	24.7	0.0	14.6	85.4
3316	Sg. Seribu at Kuala Lumpur-Ipoh Rd.	4.7	2.6	40.1	57.3
3317	Sg. Kampar at Kg. K. Dipang-Sg. Siput (S) Rd.	273.9	0.4	3.0	96.6
3318	Sg. Dipang at Kg. K. Dipang-Sg. Siput (S) Rd.	42.0	0.0	10.2	89.8

Notes: 1. Identification code adopted in present study.
2. Catchment area and water quality data from unpublished records, Ministry of Agriculture and Rural Development, Malaysia, 1974/75.
3. Catchment area and topography/rock type from Douglas (1968). Water quality data from unpublished records, Ministry of Agriculture and Rural Development, Malaysia, 1974/75.
4. Non-carbonate lithology mostly comprises alluvial plain.

Table 84 Hardness characteristics of all uncontaminated karst waters based on average figures for each site.

	n	MEAN	MINIMUM	MAXIMUM	S.DEV.
pH	241	7.64	6.25	8.19	0.424
Specific conductance (μmho)	278	271.7	40.3	580.0	104.8
Total hardness (ppm)	278	151.2	15.5	373.5	68.2
Alkaline hardness (ppm)	278	143.3	17.7	358.9	65.8
Non-alkaline hardness (ppm)	278	7.9	-13.1	103.9	14.8
Non-alk:Total hardness (%)	278	4.12	-46.9	40.2	10.5

Table 85 Hardness characteristics of uncontaminated, underground, diffuse-flow seepages, based on average figures for each site.

	n	MEAN	MINIMUM	MAXIMUM	S.DEV.
pH	165	7.87	7.28	8.19	0.179
Specific conductance (μmho)	202	255.4	135.0	574.1	78.8
Total hardness (ppm)	202	140.7	65.3	362.6	51.0
Alkaline hardness (ppm)	202	132.7	58.9	358.9	51.8
Non-alkaline hardness (ppm)	202	8.0	-10.4	62.9	11.8
Non-alk:Total hardness (%)	202	5.81	-8.47	40.2	8.77

Table 86 Matrix of correlation coefficients (r) for relationships between water hardness of uncontaminated, underground, diffuse-flow seepages.

	SPECIFIC CONDUCTANCE	TOTAL HARDNESS	ALKALINE HARDNESS	NON-ALKALINE HARDNESS
pH	(-0.567)[1]	(-0.532)	(-0.566)	(0.277)
Specific conductance		0.983[2]	0.954	0.025
Total hardness			0.976	0.002
Alkaline hardness				-0.217

Notes: 1. n=202, except for values in parentheses where n=165
2. Values underlined are significant at 99.9 per cent confidence level.

Table 87 Hardness characteristics of uncontaminated, diffuse-flow seepages in the four main cave systems investigated, based on average figures for each site.

	n	MEAN	MINIMUM	MAXIMUM	S.DEV.
Gua Batu					
pH	43	7.96	7.55	8.19	0.134
Specific conductance (μmho)	48	272.4	170.0	350.0	46.12
Total hardness (ppm)	48	153.4	89.0	217.6	31.75
Alkaline hardness (ppm)	48	131.6	78.0	190.0	33.23
Non-alkaline hardness (ppm)	48	21.8	3.9	62.9	11.72
Non-alk:Total hardness (%)	48	14.5	2.6	37.7	8.08
Gua Anak Takun					
pH	28	7.85	7.64	8.15	0.109
Specific conductance (μmho)	46	214.1	135.0	330.5	49.98
Total hardness (ppm)	46	111.0	65.3	181.3	29.07
Alkaline hardness (ppm)	46	105.1	58.9	180.1	28.99
Non-alkaline hardness (ppm)	46	5.9	-7.7	35.8	7.90
Non-alk:Total hardness (%)	46	5.3	-8.5	34.9	7.47
Gua Tempurong					
pH	70	7.91	7.55	8.15	0.130
Specific conductance (μmho)	84	229.1	136.2	463.7	58.74
Total hardness (ppm)	84	124.2	67.7	230.9	35.26
Alkaline hardness (ppm)	84	120.4	62.0	228.3	35.10
Non-alkaline hardness (ppm)	84	3.8	-5.9	41.7	8.58
Non-alk:Total hardness (%)	84	2.9	-6.5	40.2	7.44
Perlis Mines' Cave/Subway Tunnel					
pH	24	7.60	7.28	7.88	0.176
Specific conductance (μmho)	24	392.5	259.6	574.1	80.41
Total hardness (ppm)	24	230.2	136.5	362.6	55.86
Alkaline hardness (ppm)	24	230.7	136.5	358.9	55.40
Non-alkaline hardness (ppm)	24	-0.5	-10.4	5.7	3.02
Non-alk:Total hardness (%)	24	-0.3	-4.2	2.2	1.22

Table 88 Hardness characteristics of diffuse-flow seepages contaminated by bat guano, based on average figures for each site.

	n	MEAN	MINIMUM	MAXIMUM	S.DEV.
pH	19	7.76	7.46	8.01	0.152
Specific conductance (μmho)	28	767.4	238.3	2030.0	434.9
Total hardness (ppm)	28	367.3	99.9	1305.8	241.0
Alkaline hardness (ppm)	28	84.67	32.1	130.0	24.99
Non-alkaline hardness (ppm)	28	282.6	44.4	1175.8	239.7
Non-alk:Total hardness (%)	28	68.5	31.6	95.1	16.99

Table 89 Hardness characteristics of karst spring waters (n=3), based on average figures for each site.

	MEAN	MINIMUM	MAXIMUM	S.DEV.
pH	6.96	6.73	7.18	0.225
Specific conductance (μmho)	507.5	473.6	540.3	33.37
Total hardness (ppm)	305.9	280.6	332.0	25.75
Alkaline hardness (ppm)	307.3	283.9	331.1	23.61
Non-alkaline hardness (ppm)	-1.4	-3.3	0.9	2.17
Non-alk:Total hardness (%)	-0.5	-1.0	0.2	0.614

Table 90 Hardness characteristics of underground streams (n=11), based on average figures for each site.

	MEAN	MINIMUM	MAXIMUM	S.DEV.
pH	7.47	6.98	7.84	0.226
Specific conductance (μmho)	387.9	320.1	460.4	42.79
Total hardness (ppm)	227.7	180.2	280.8	30.53
Alkaline hardness (ppm)	227.1	182.8	278.1	30.26
Non-alkaline hardness (ppm)	-0.6	-2.9	2.6	1.79
Non-alk:Total hardness (%)	-0.4	-1.6	1.2	0.825

Table 91 Hardness characteristics of surface streams (subdivided according to catchment type), based on average figures for each site.

	MEAN	MINIMUM	MAXIMUM	S.DEV.
Streams with significant groundwater contribution (n=4)				
pH	7.74	7.62	7.90	0.126
Specific conductance (μmho)	305.2	284.5	351.1	31.3
Total hardness (ppm)	170.6	156.8	203.8	22.5
Alkaline hardness (ppm)	172.3	158.5	205.8	22.5
Non-alkaline hardness (ppm)	-1.7	-2.5	-0.5	0.83
Non-alk:Total hardness (%)	-1.0	-1.6	-0.4	0.499
Karst streams with significant contribution from lowland plains (n=20)				
pH	7.00	6.76	8.01	0.294
Specific conductance (μmho)	116.0	40.3	274.3	69.2
Total hardness (ppm)	48.6	15.5	127.9	40.1
Alkaline hardness (ppm)	49.2	17.7	123.5	29.6
Non-alkaline hardness (ppm)	-0.6	-7.7	21.4	6.41
Non-alk:Total hardness (%)	-11.1	-46.9	18.6	17.0
Streams draining non-carbonate terrain (n=7)				
pH	6.94	6.56	7.22	0.231
Specific conductance (μmho)	45.0	24.0	76.5	18.5
Total hardness (ppm)	18.7	9.3	29.6	8.57
Alkaline hardness (ppm)	24.5	10.8	38.8	11.3
Non-alkaline hardness (ppm)	-5.8	-10.3	-0.6	3.65
Non-alk:Total hardness (%)	-49.1	-75.5	-31.8	17.5

Table 92 Hardness characteristics of seepages at the alluvium/rock interface (n=37), based on average figures for each site.

	MEAN	MINIMUM	MAXIMUM	S.DEV.
pH	7.12	6.25	8.05	0.416
Specific conductance (μmho)	384.3	205.0	580.0	96.9
Total hardness (ppm)	224.5	109.2	373.5	66.1
Alkaline hardness (ppm)	208.8	118.0	314.0	54.3
Non-alkaline hardness (ppm)	15.7	-13.1	103.9	27.1
Non-alk:Tot hardness (%)	5.5	-9.9	32.1	10.0

Table 93 Hardness characteristics of deep-groundwater from borehole site in Kinta Valley (n=1).

pH	6.97
Specific conductance (μmho)	329.3
Total hardness (ppm)	216.9
Alkaline hardness (ppm)	219.2
Non-alkaline hardness (ppm)	-2.3
Non-alk:Total hardness (%)	-1.1

Table 94 List of antecedent rainfall periods with which total hardness correlates most closely at the 52 diffuse-flow seepages where statistically significant positive correlations were found (r_s;$p<0.05$).

ANTECEDENT PERIOD (days)	GUA BATU	GUA ANAK TAKUN	GUA TEMPURONG	PERLIS MINES' CAVE/SUBWAY TUNNEL
1				
1-2				1
1-4	1	2		3
1-8		2		2
1-16	2			3
1-32				1
1-64	3	2		3
1-128				
1-256	1			
2-3	1			
4-7				1
8-15	1			
15-28		1	1	
29-42	2			
43-56	2			1
57-70	3	2		1
64-127	1			
71-84				1
85-98			1	
99-112		1	4	
128-255	1	1		

Table 95 Nutrient loss (kg/ha) from the karst outcrops during the one-year study period (1.8.74 to 31.7.75).

	Ca	Mg	K	Na
Selangor (Gua Batu[1]/Gua Anak Takun[2])	377.4	44.9	3.06	6.81
Kinta Valley (Gua Tempurong[3])	595.4	44.5	9.02	7.14
North-west region (springs/cave streams[4])	356.5	24.0	2.76	12.59
Arithmetic mean	443.1	37.8	4.95	8.85

Rainfall based on the following stations: 1. 1602; 2. 2601; 3. 3601; 4. 4601 and 4604.

Table 96 Estimation of long-term rates of net chemical denudation based on the total hardness of karst waters.

FIELD AREA	EFFECTIVE RAINFALL[1] (mm)	MEAN TOTAL HARDNESS[2] (ppm)	DENUDATION RATE[3] (mm/1000yrs)
Bt. Batu	1011[a]	153.4[e]	54.6
Ak. Bt. Takun	1011[a]	111.0[e]	38.4
G. Tempurong	1761[b]	124.2[e]	81.0
Setul Boundary Range	443[c]	244.0[f]	38.5
Sub-alluvial karst surface, Kinta Valley	1348[d]	244.5[g] (257.1)	112.9 (129.4)

Notes: 1. Rainfall stations on which effective rainfall based: a. 1602; b. 3601; c. 4601 and 4604; d. 3604, 3605, 3606, 3607 and 3608.

2. Solute data based on the following types of site: e. underground diffuse-flow seepage; f. springs and cave streams; g. seepages from alluvium/rock interface (figures in parentheses have been augmented according to measured levels of chemical aggressiveness).

3. After correction for cation input from rainfall, and assuming a rock density of $2.65 g/cm^3$.

PLATES

Plates 3 to 58 are arranged by region.

All photographs taken by author.

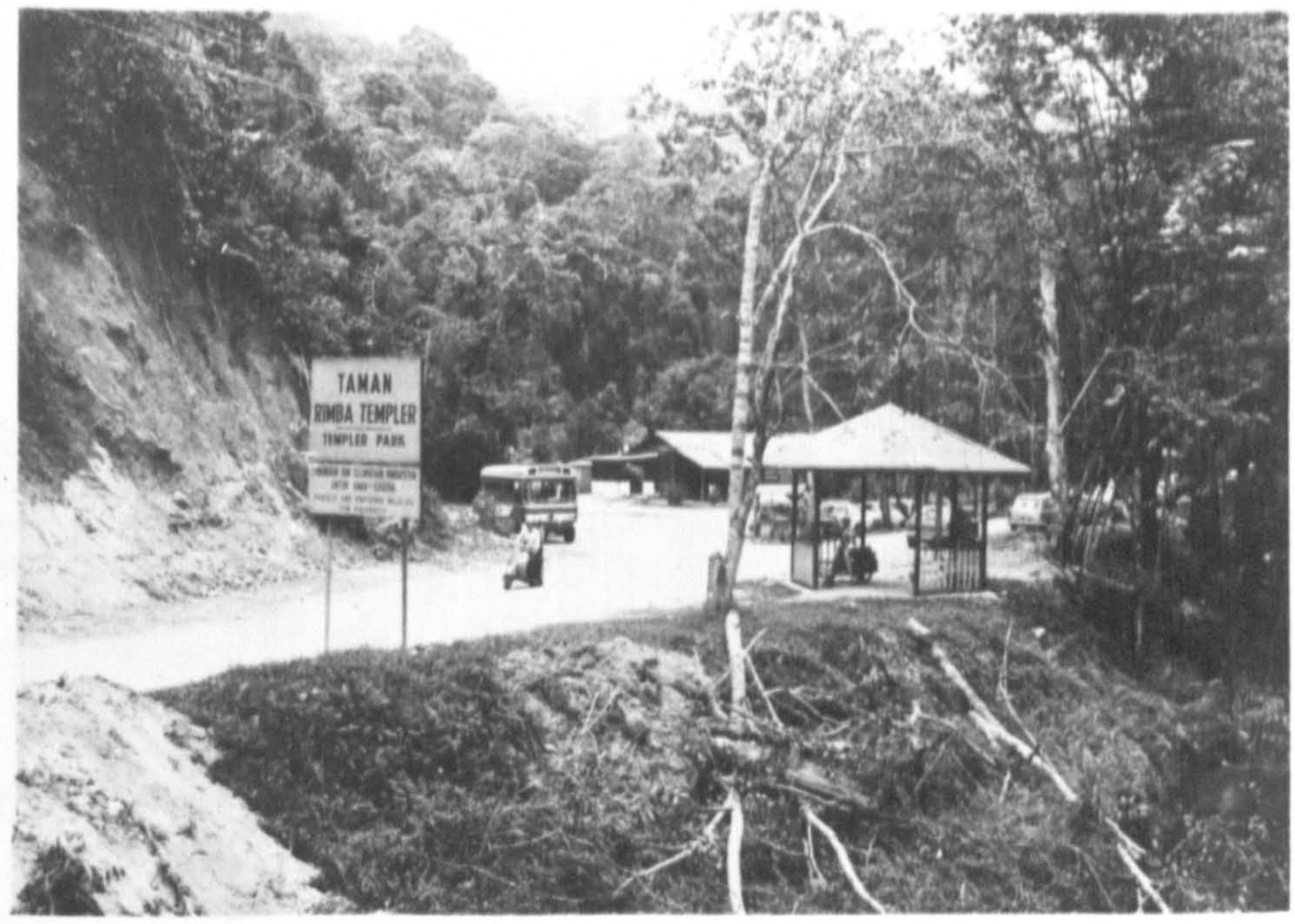


Plate 1 Entrance to Templer Park. The automatic rain gauge (station 2601) is located on the roof of the building immediately to the right of the coach.

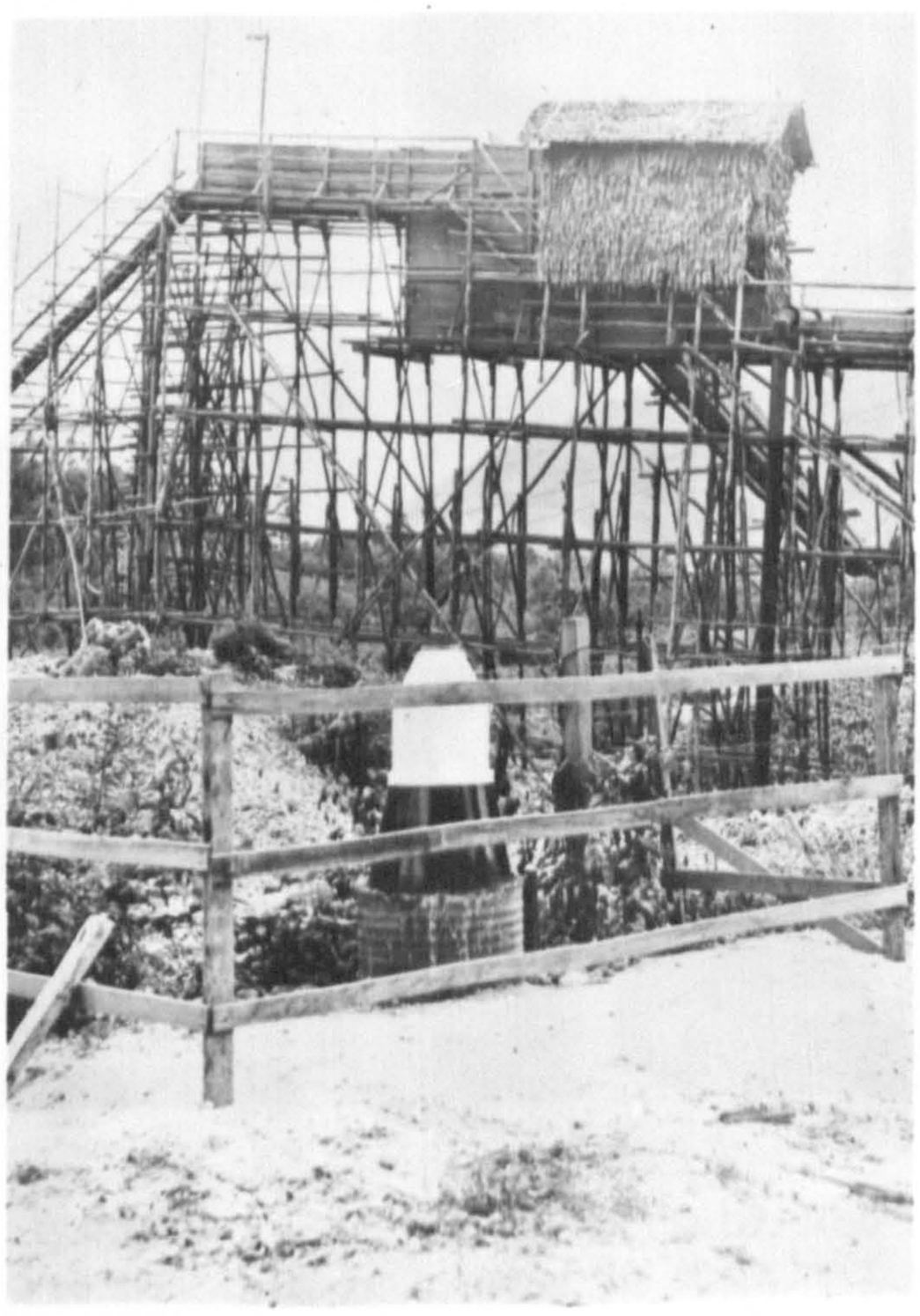


Plate 2 'Kent', siphon-type rain gauge at Siput Mines Ltd., Sg. Siput South (station 3601).

Plate 3 Bt. Batu, north of Kuala Lumpur, with extensively mined deposits of Old Alluvium in the foreground.

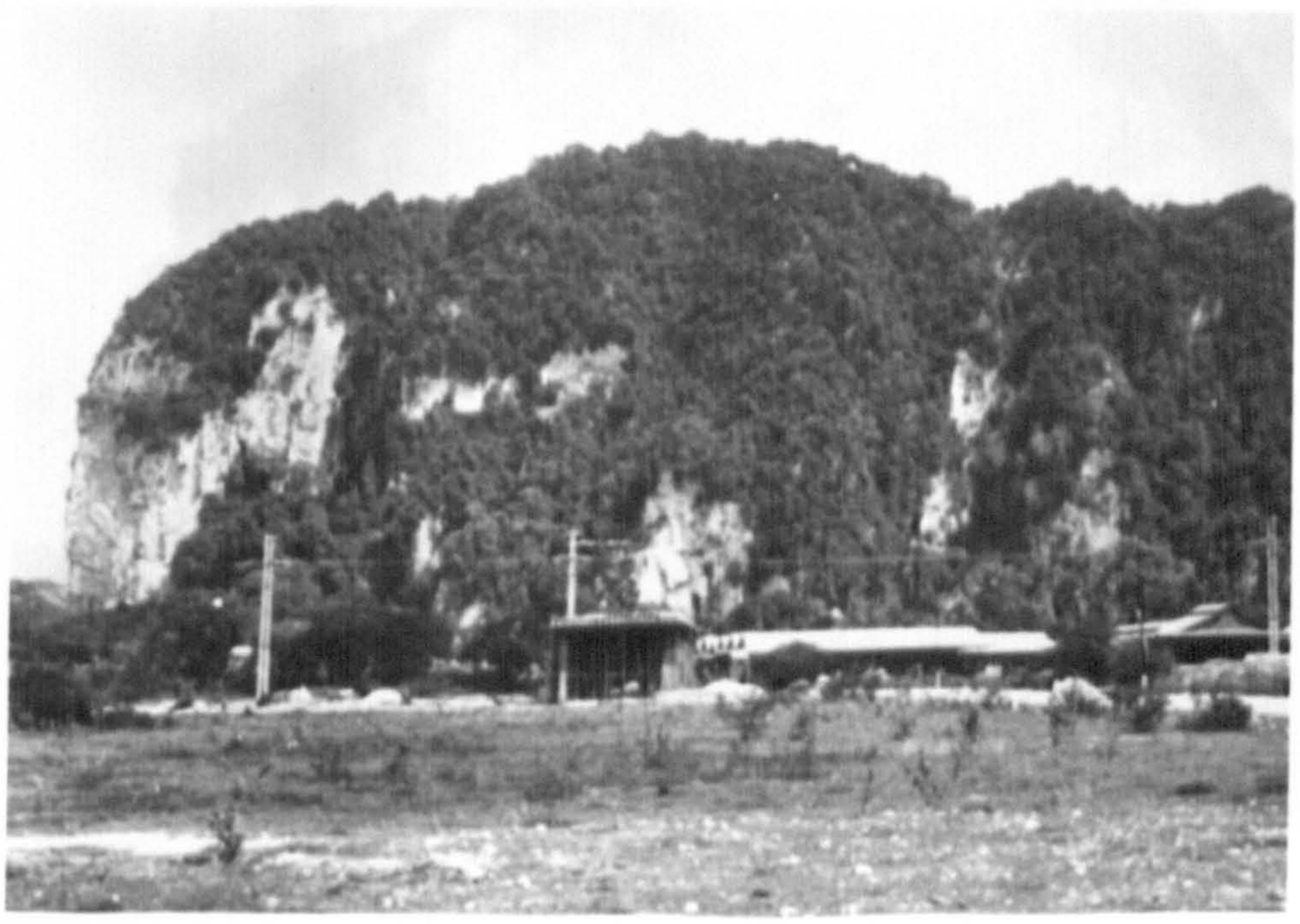


Plate 4 South-western face of Bt. Batu. The entrance to Gua Batu is located in the cliff at the right of the photograph.

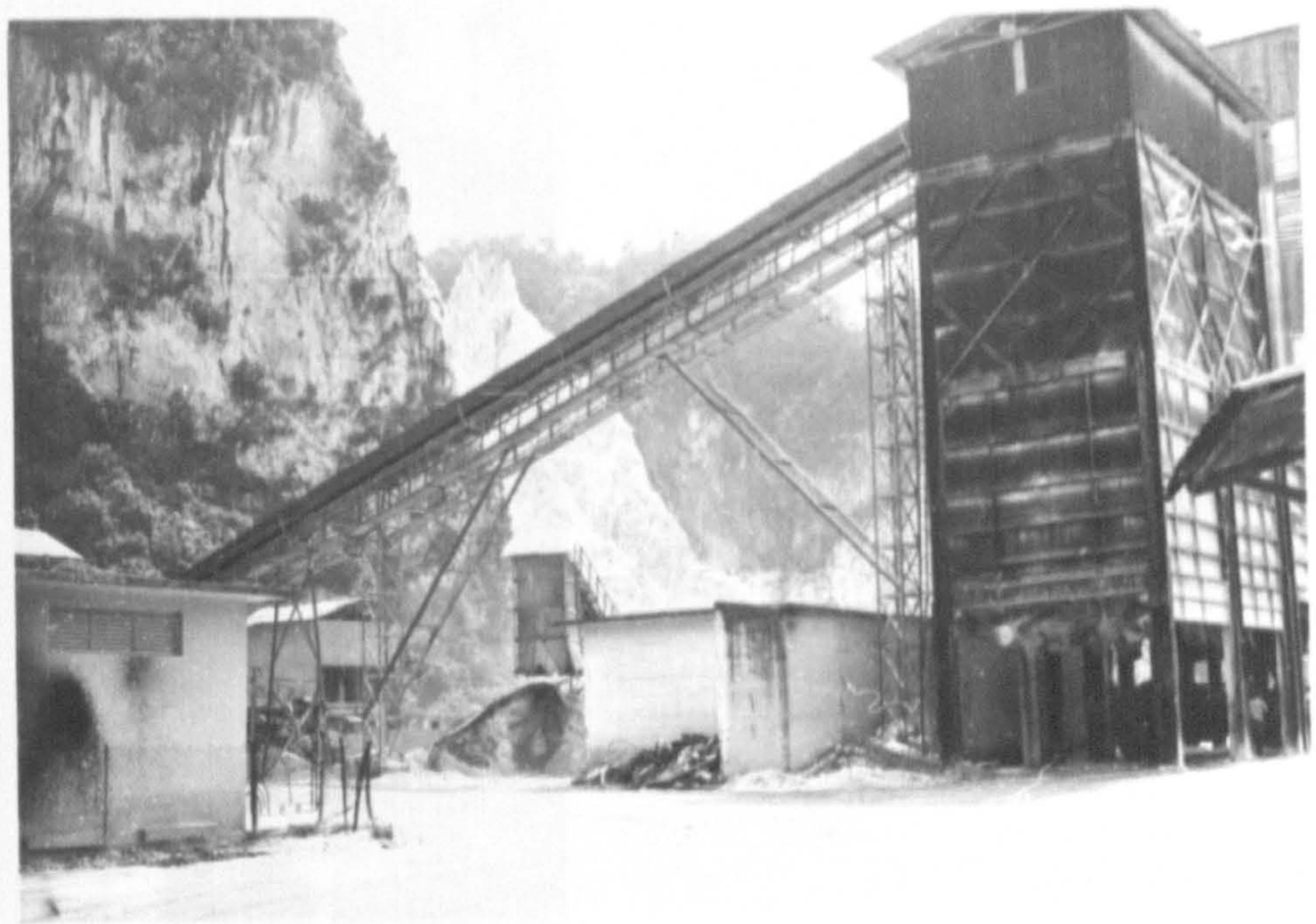


Plate 5 Quarry, 300 m east of Gua Batu.

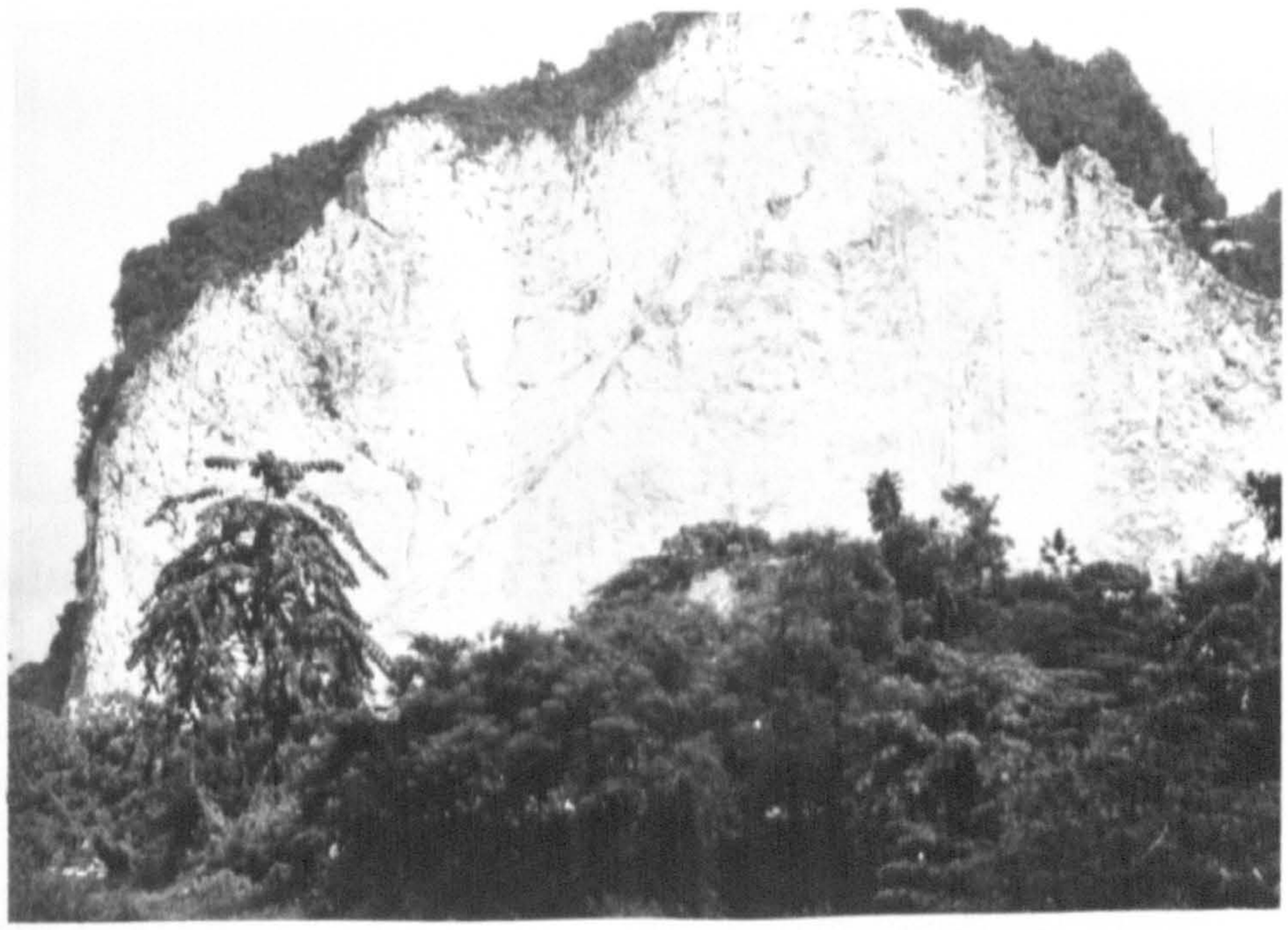


Plate 6 Fresh exposure of the Kuala Lumpur Limestone in the quarry face, showing the massiveness of the bedrock. Few traces of the original sedimentary structures remain.

Plate 7

Entrance to Gua Batu. Temple Cave lies beyond the gates at the top of the steps. Gua Gelap is to the left, immediately below these gates.

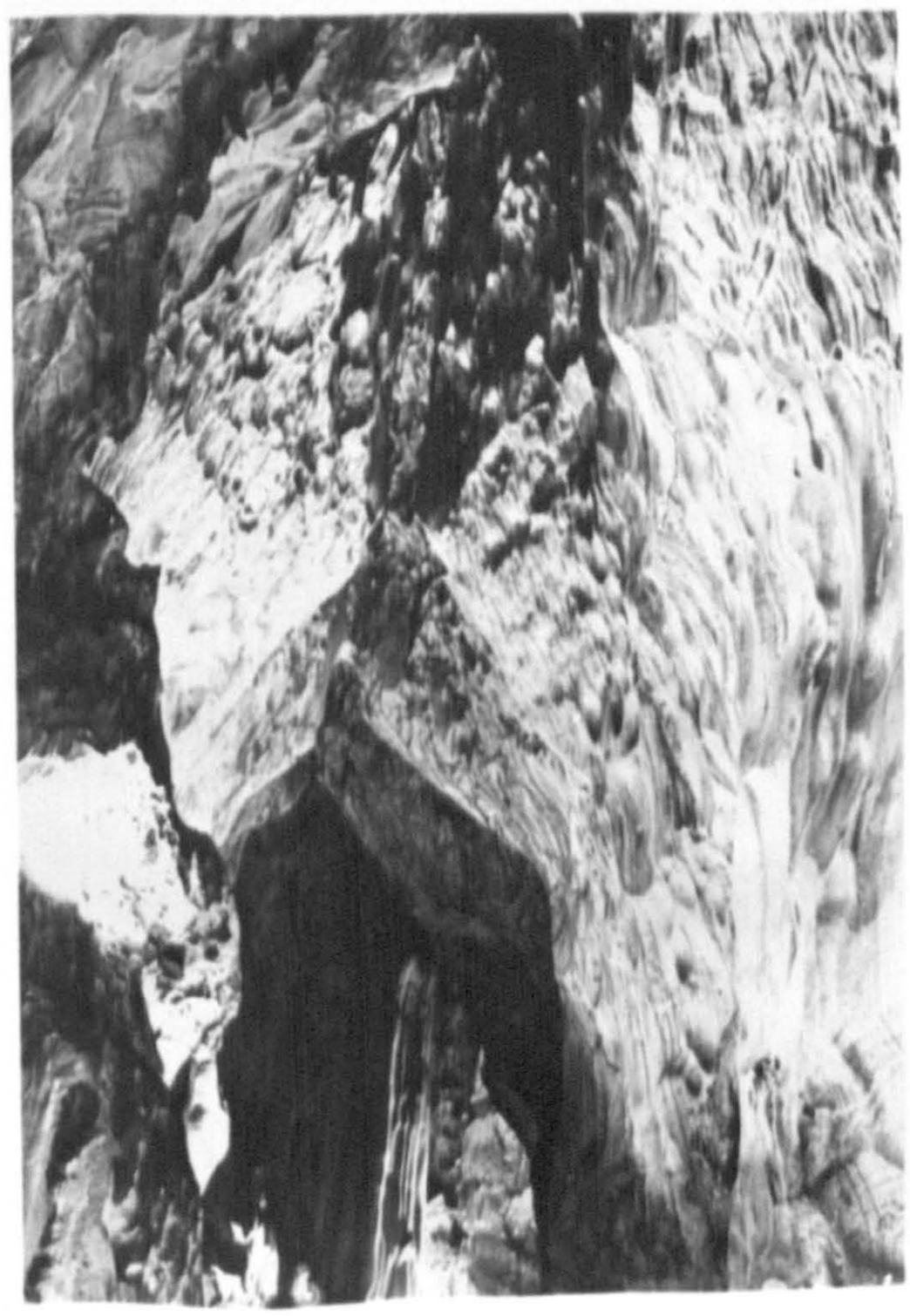


Plate 8

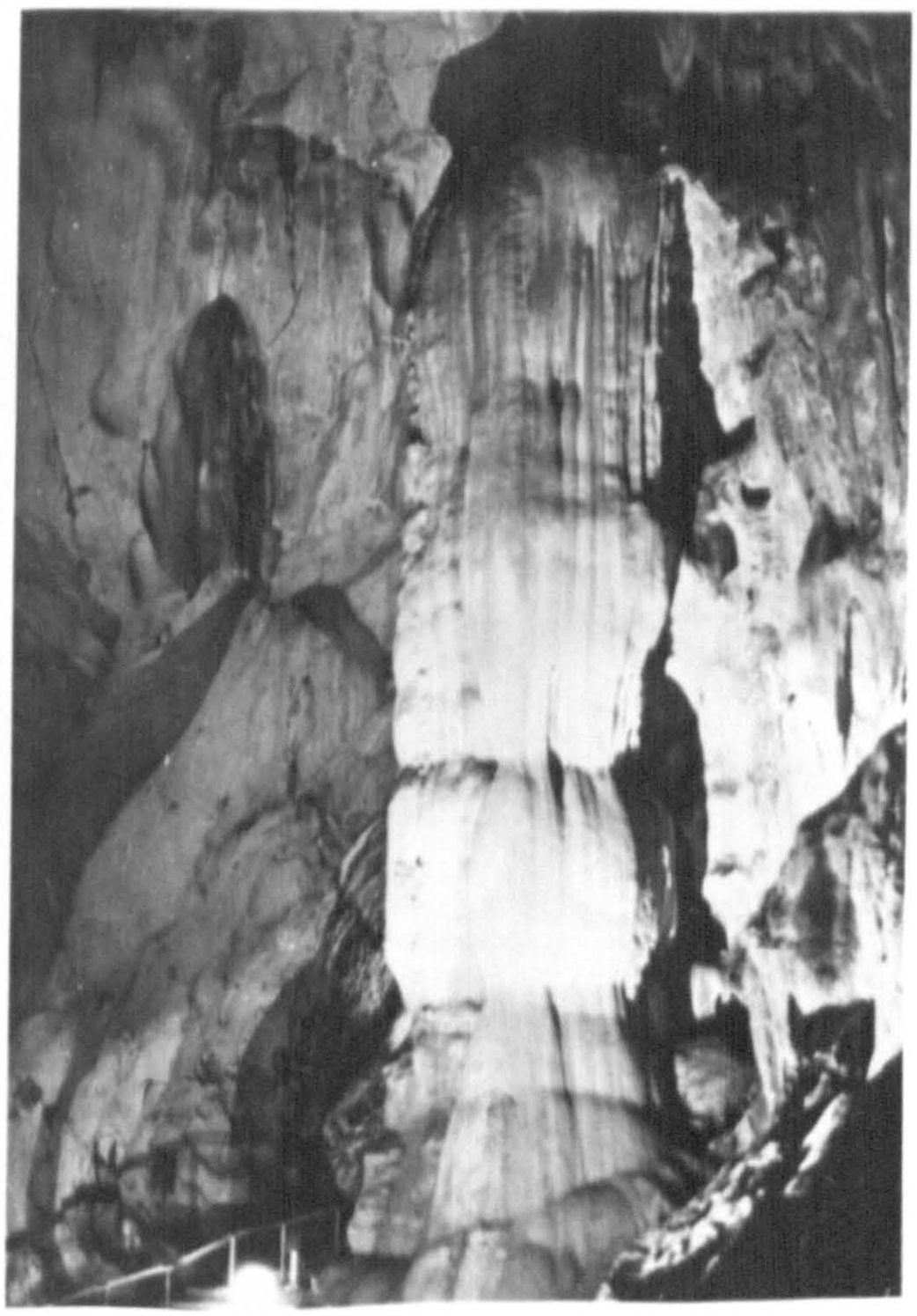
Cavern B, Gua Gelap (roof height approximately 26 m). The line of the cavern is structurally controlled. The passage is probably of phreatic origin, but the scalloped walls and the distinct notch in the right wall suggest subsequent vadose modification.

Plate 9

Open shaft in Temple Cave. The central column provides some indication of the enormous size of depositional features in the tower karst caves. Figure to left of column for scale.

Plate 10

Large depositional feature in Cavern B, Gua Gelap (seen also in plate 8). The rail at the base of the column is 0.9 m high.



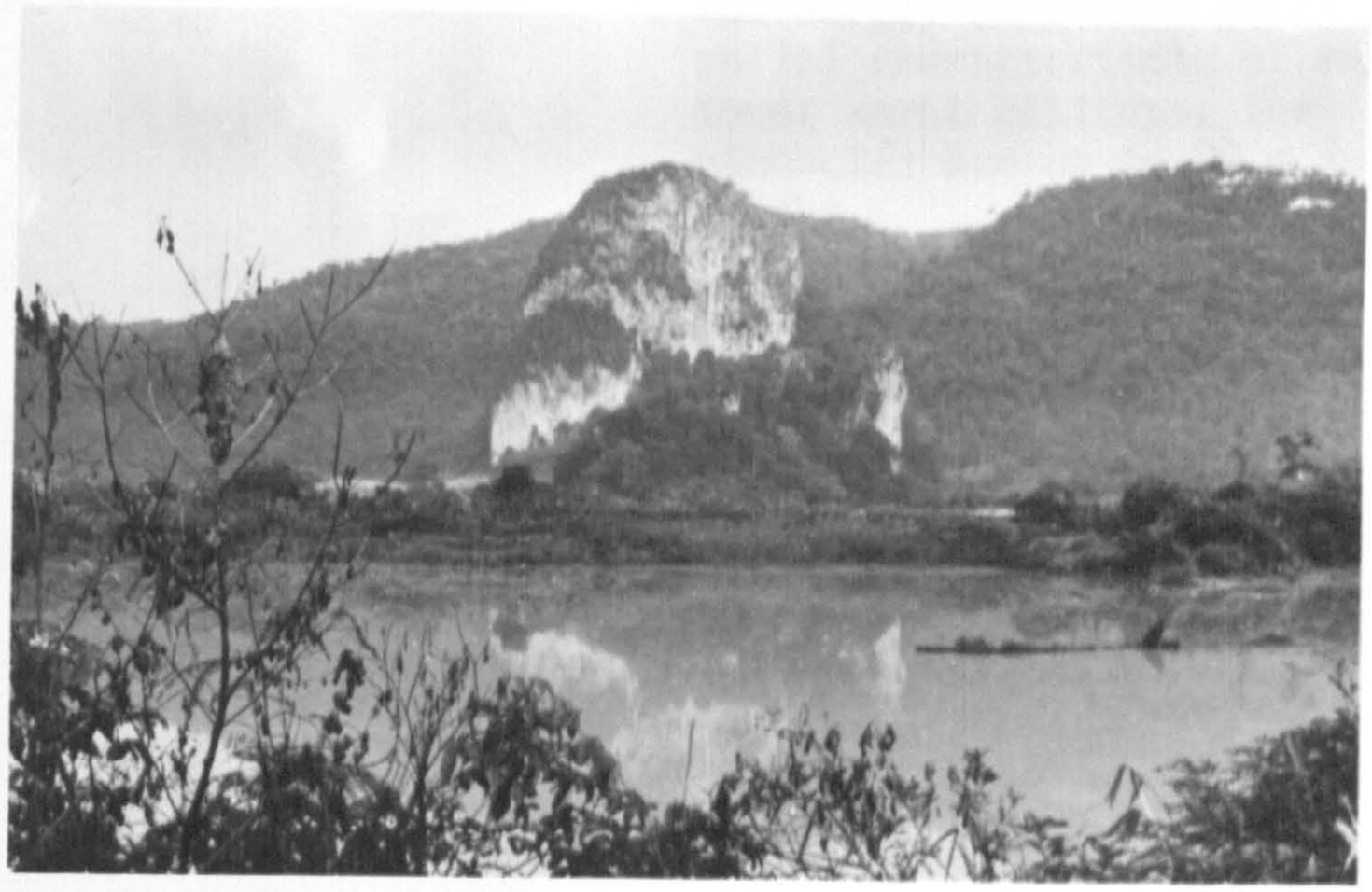


Plate 11 Ak. Bt. Takun, the low limestone hill in the middle distance, backed by the isolated tower of Bt. Takun.

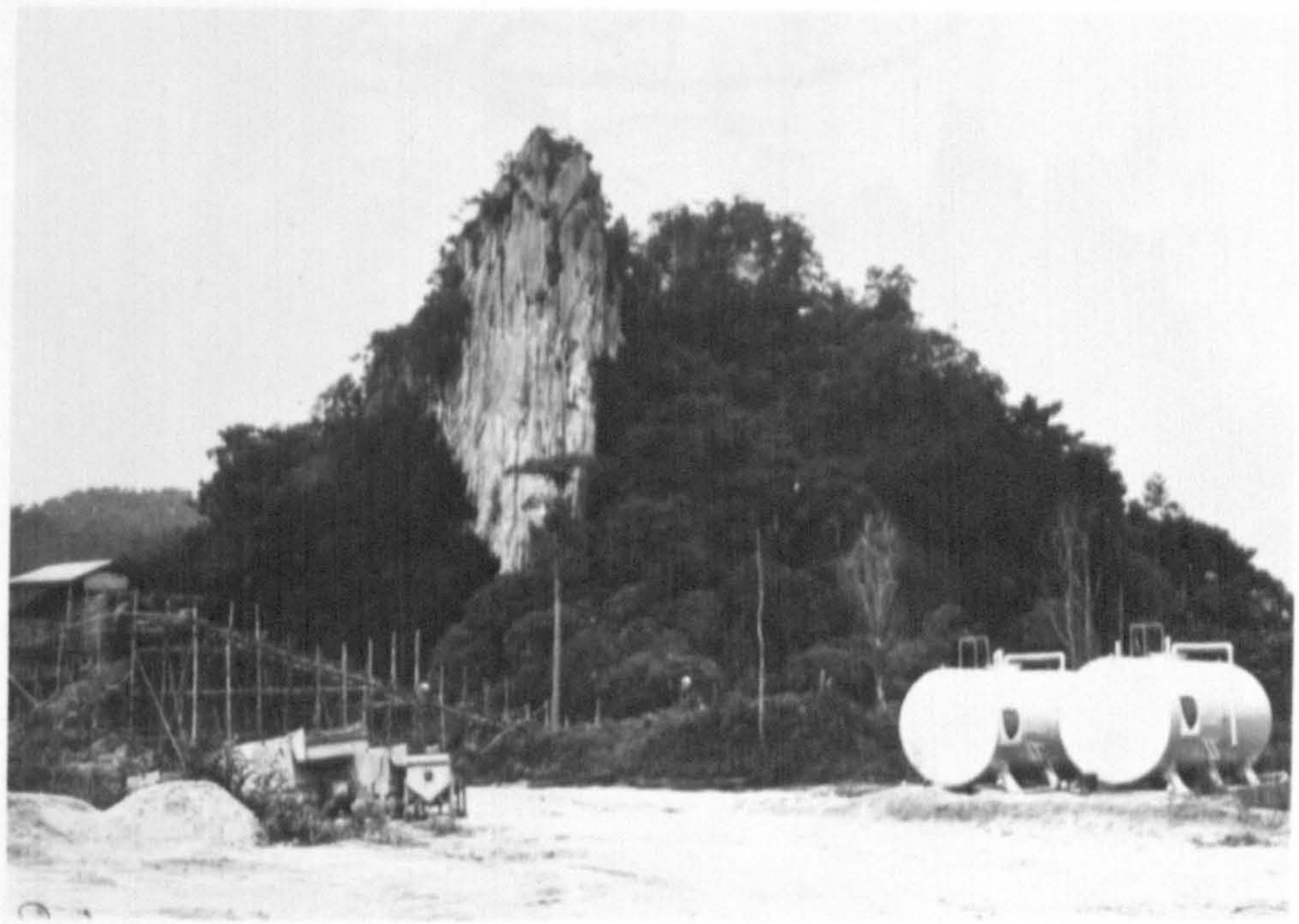


Plate 12 Precipitous cliff face of Ak. Bt. Takun which surmounts the entrance of Gua Anak Takun. Plots ABT-I and ABT-II are located, respectively, to the right and left of the cliff.

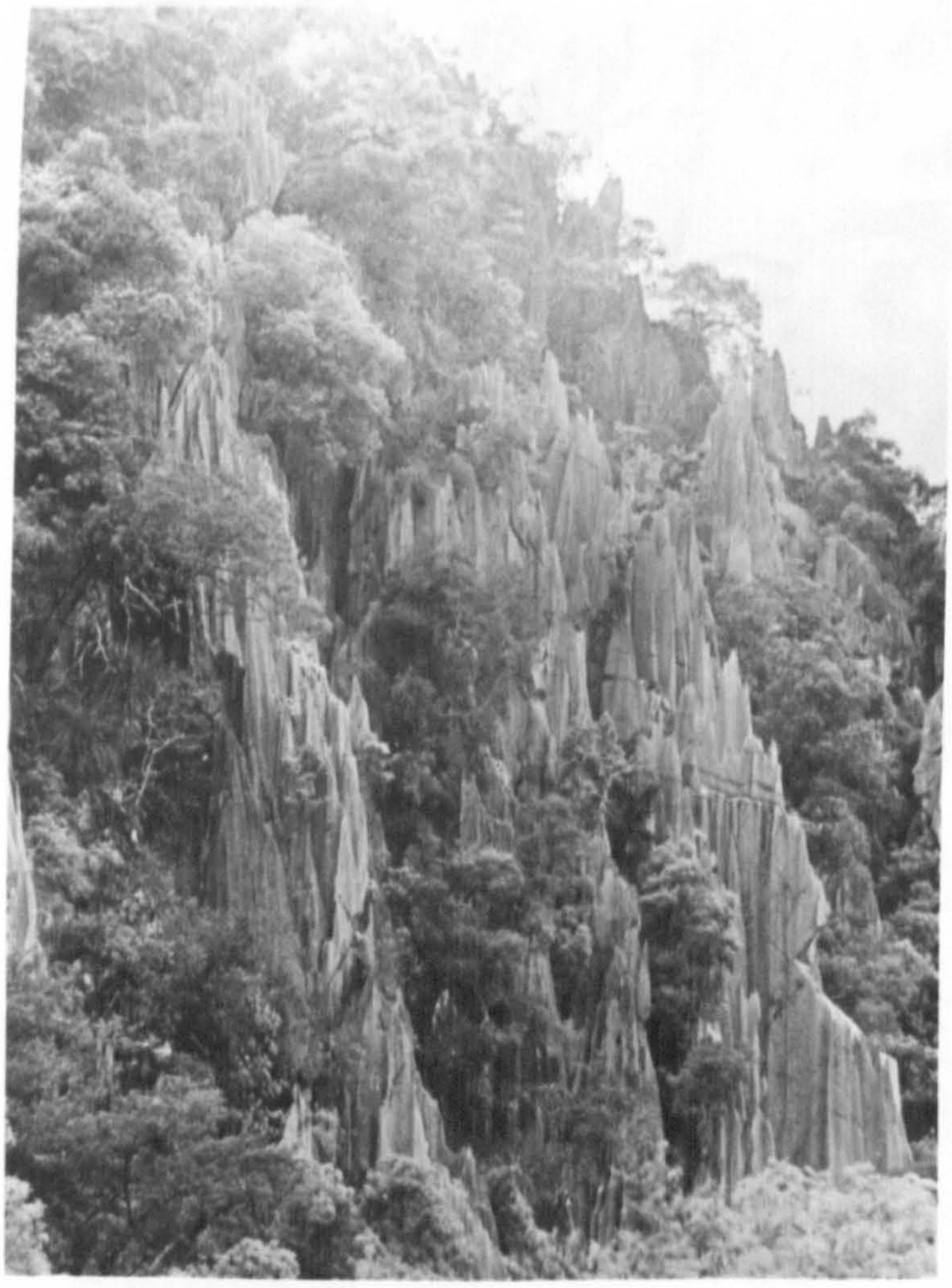


Plate 13

Summit of Bt. Takun, showing the deeply dissected meso-relief characteristic of many tower karst hilltops. The pinnacles are up to 20 m in height.

Plate 14

Summit of Ak. Bt. Takun, location of plots ABT-III and -IV.

Plate 15

Typical bare rock micro-catchment with sampling bottle in place.

Plate 16

Micro-catchment with a partial soil/litter cover.

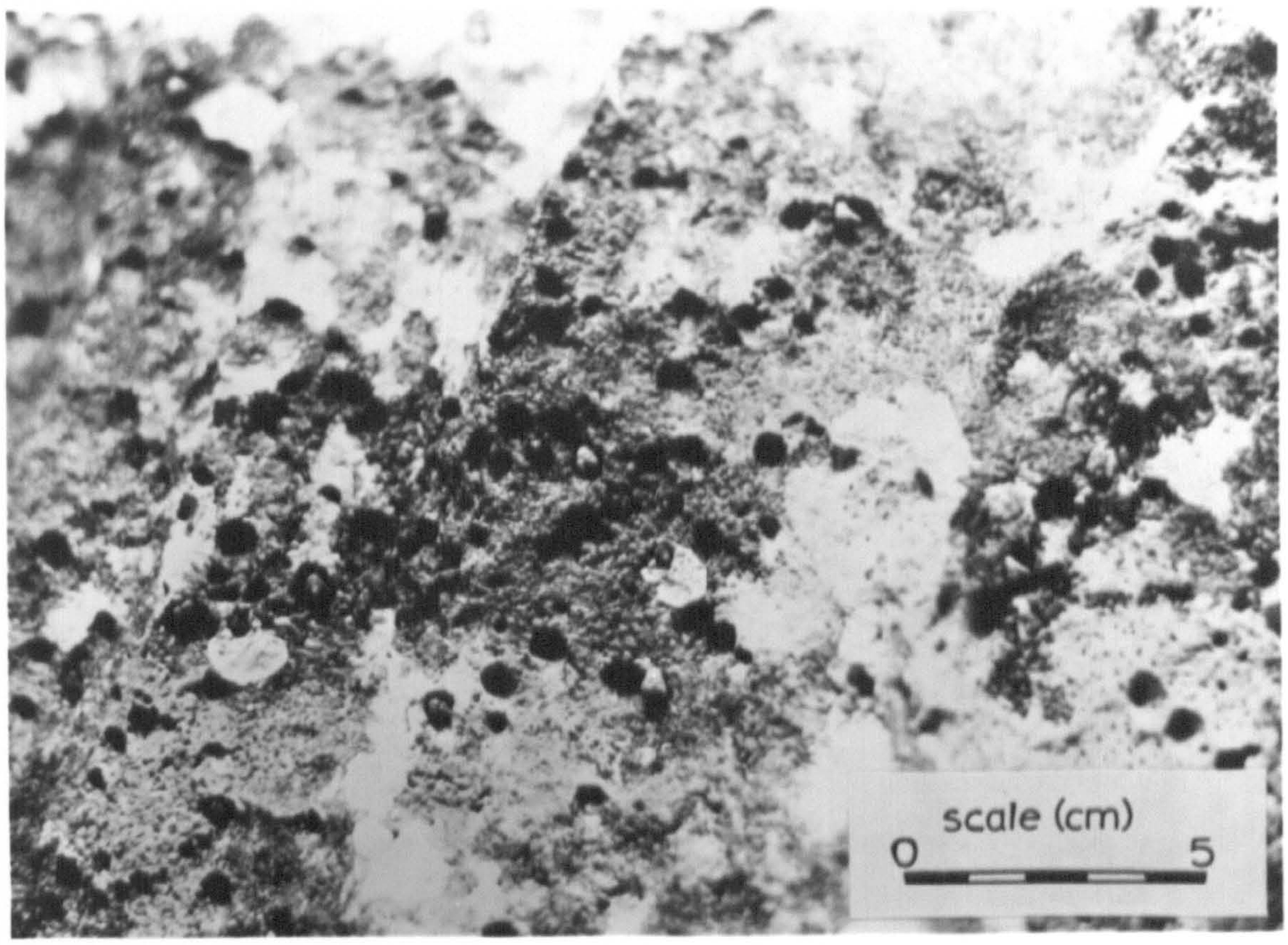


Plate 17 Small circular depressions on an exposed rock surface at the summit of Ak. Bt. Takun. Several contain small snails of the species Alycaeus.

Plate 18 Sharply-etched micro-relief of a bare rock outcrop.

Plate 19 Tower karst hills of the Kinta Valley, set amid broad alluvial plains. View is from G. Panjang, looking northwards. G.Meusah (226 m) is the hill to the left.

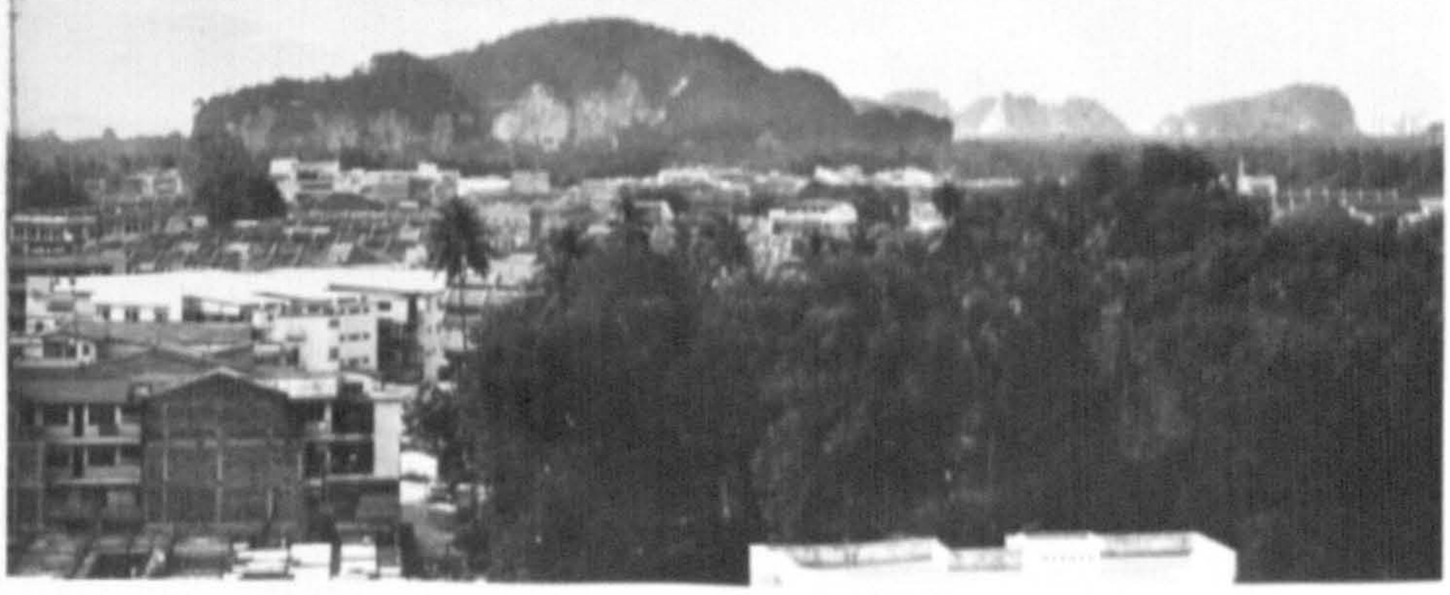


Plate 20 View northwards from Ipoh. G. Cheroh and G. Lang are the two outcrops in the middle distance. The granites of the Kledang Range are just visible to the left.

Plate 21 Cliff-foot notch exposed beneath Old Alluvium in tin workings near Tambun.

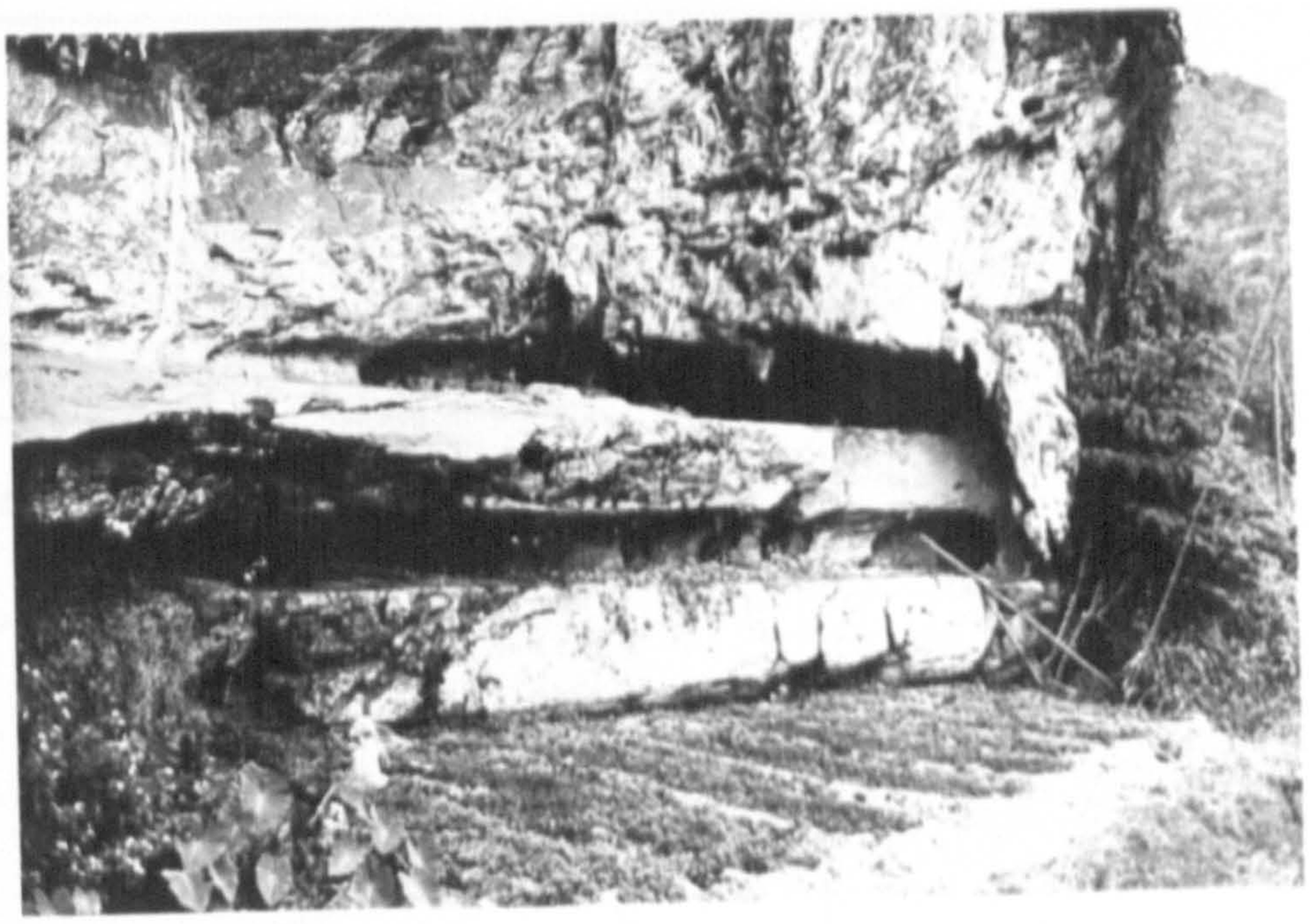
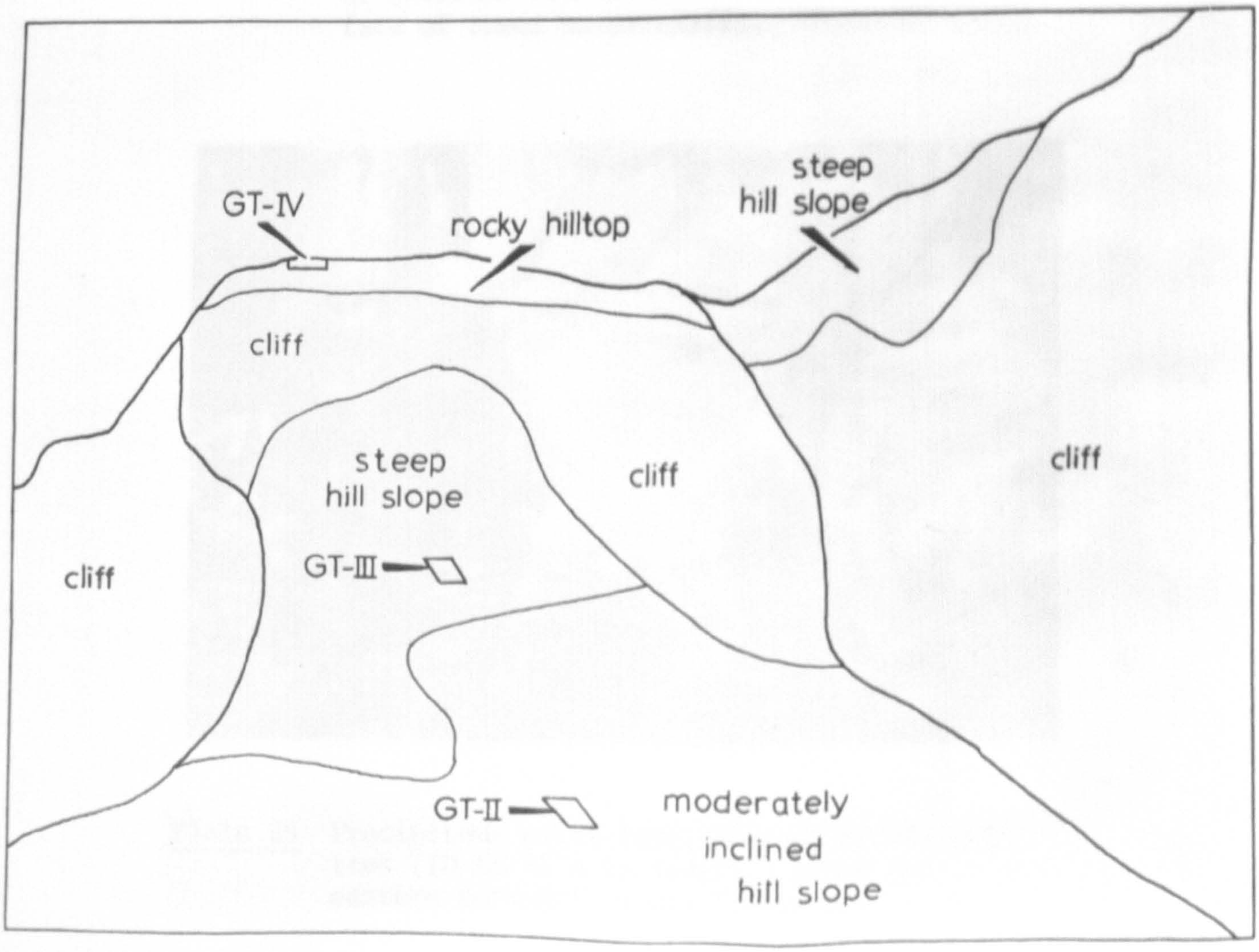


Plate 22 Horizontal slits, probably cut by swamp waters, at base of G. Rapat.Figure for scale, bottom left.

Plate 23 Southernmost spur of G. Tempurong. Annotated sketch (below) shows the main morphological units and the approximate locations of three of the study plots.



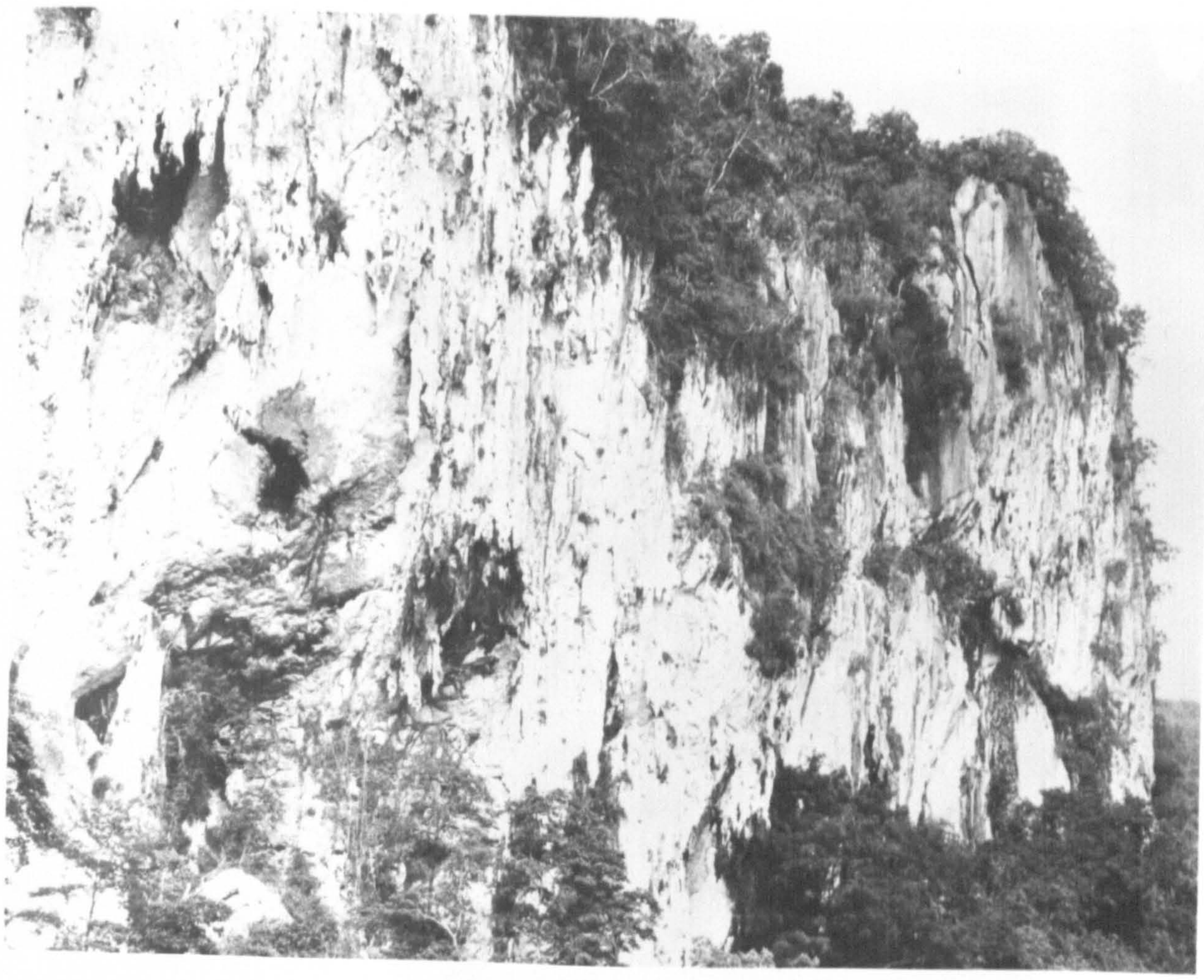


Plate 24 Part of the structurally-controlled, south-eastern wall of G. Datok, showing the scale of external stalactite development on the face of tower karst cliffs.

Plate 25 Precipitous cliff face, adorned by stalactites (10 to 15 m in length), above the eastern entrance of Gua Tempurong.

Plate 26 High-level cave in the G. Gajah-Tempurong massif.

Plate 27 One of the cave temples in the limestone cliffs of G. Rapat.

Plate 28 A joint-controlled corridor in G. Datok, with a mining pool in foreground.

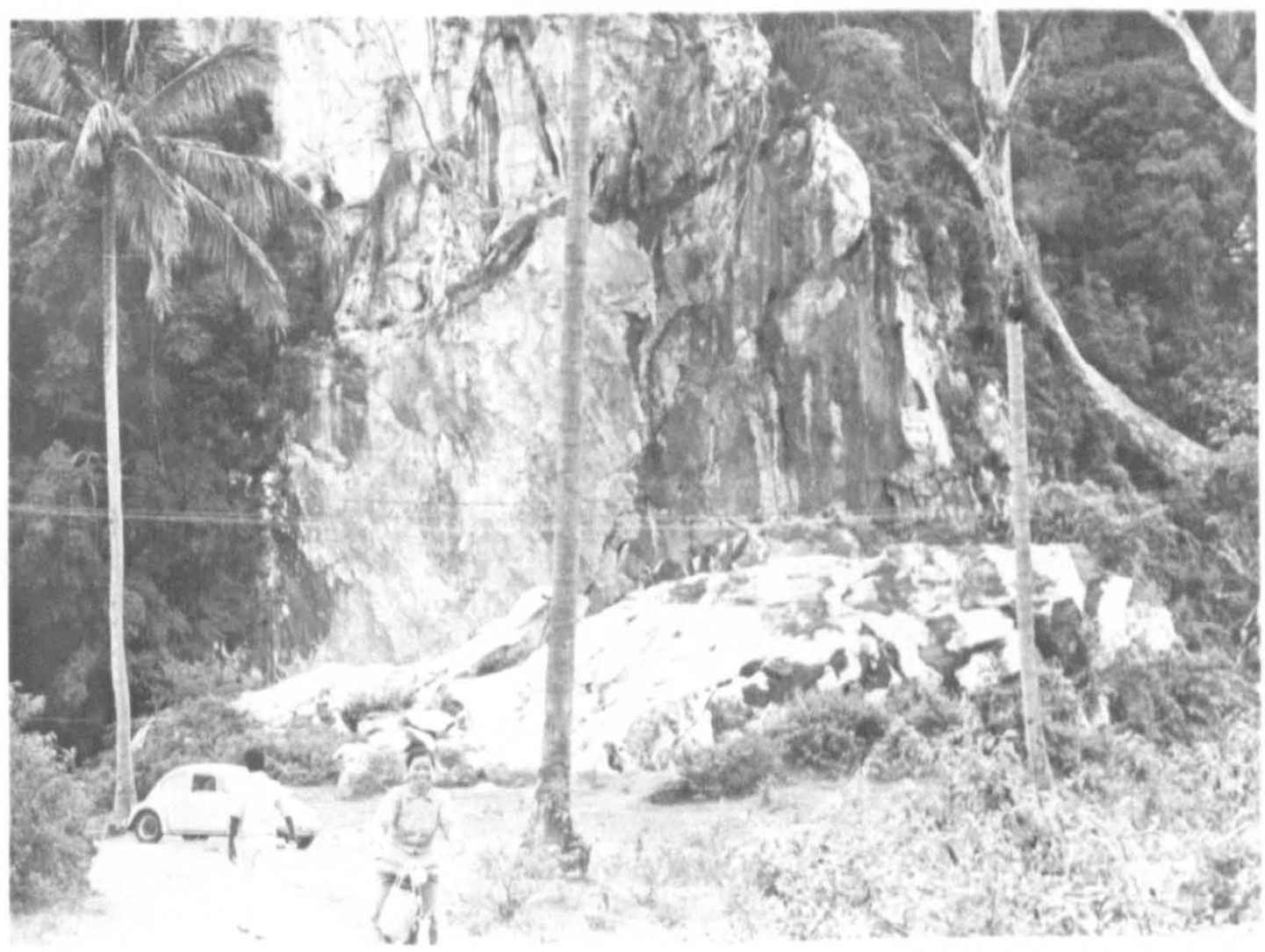


Plate 29 The slab of rock at G. Cheroh which fell on a longhouse.

Plate 30

Plot GT-IV (400 m), located on a ridge-crest of G. Tempurong (see plate 23).

Plate 31

Section of G. Cheroh exposed by rockfall (see plate 29), showing deep pockets of mineral soil, separated by thinly mantled rock outcrops.

Plate 32 Litter tray and throughfall collection funnel in plot GT-I.

Plate 33 Throughflow collection device installed to sample water at the soil/rock interface in GT-I.

Plate 34

Basal undercutting by stream action at the eastern entrance of Gua Tempurong.

Plate 35

Horizontal notches cut by stream action in Exit Cavern, Gua Tempurong.

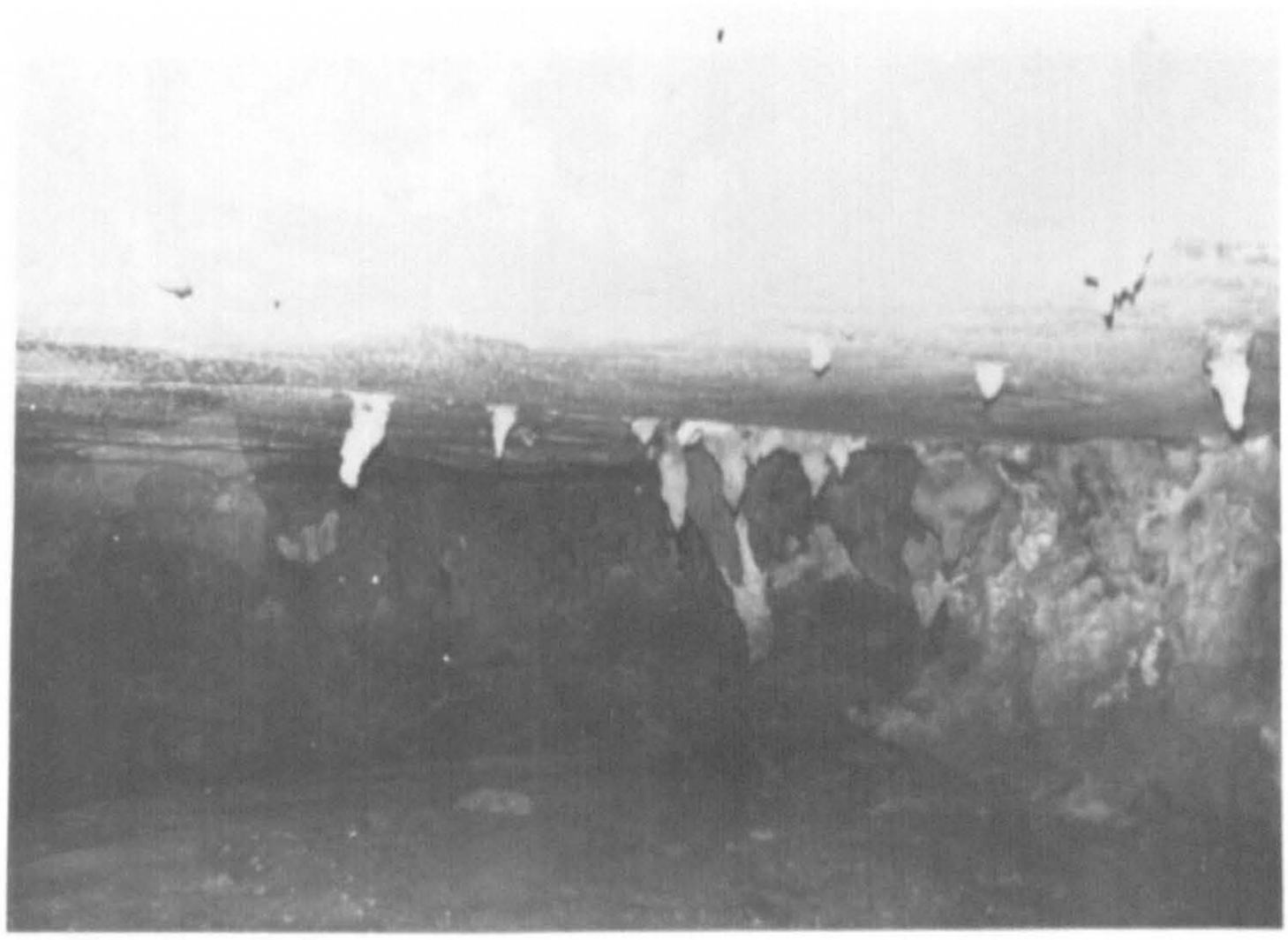


Plate 36 Isolated stalactites aligned along tight rock fractures in Molesworth Allen Tunnel, Gua Tempurong.

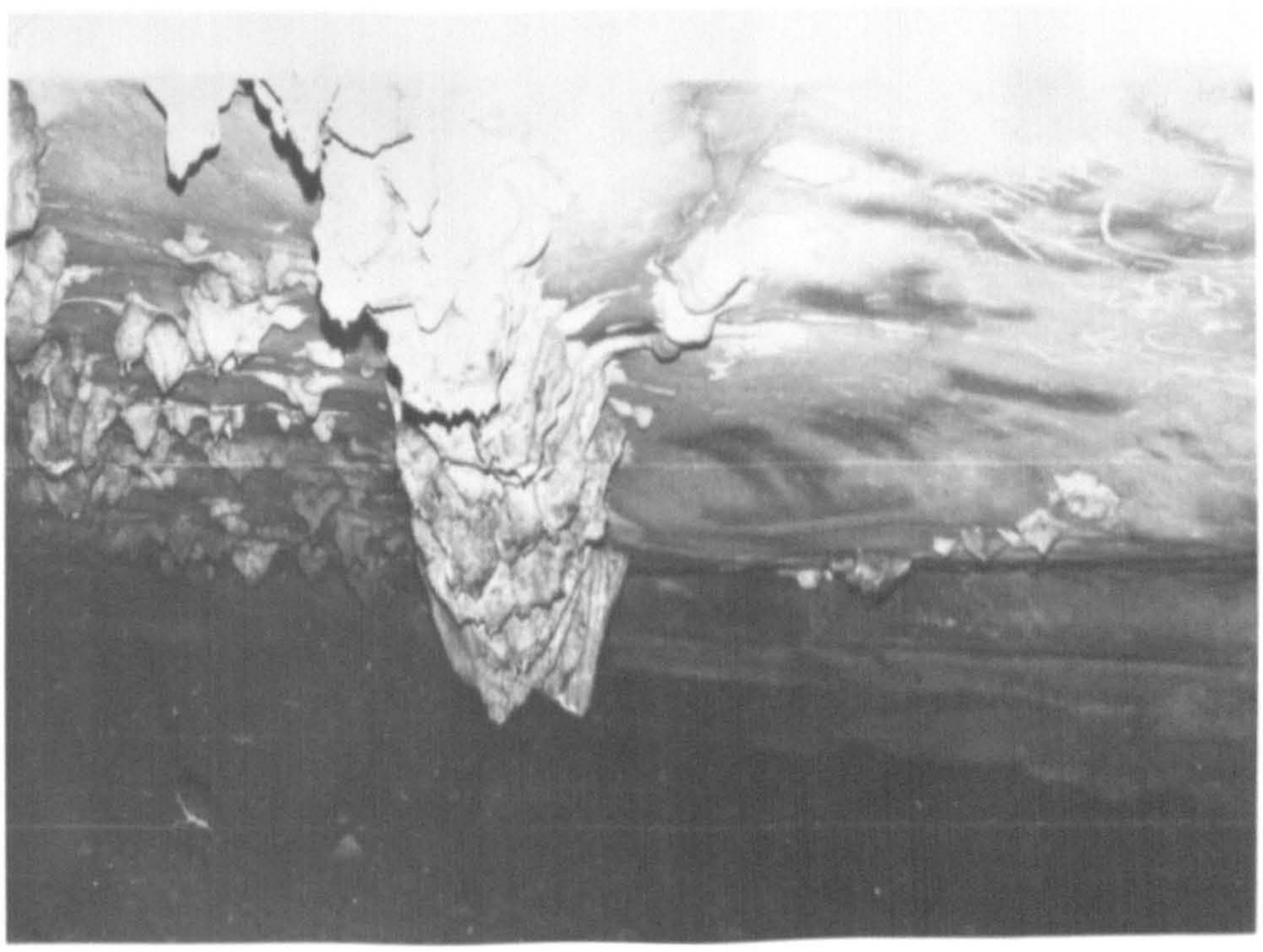


Plate 37 Small stalactite forms in the undercut section of Alam Cavern, Gua Tempurong.

Plate 38 East bank tributary of the Sg. Sengat, emerging from G. Kroh after a passage of about 150 m underground.

Plate 39 Limestone bedrock exposed at the surface of the flood plain of the Sg. Sengat, near G. Kroh.

Plate 40 Limestone bedrock exposed along the course of the Sg. Kampar.

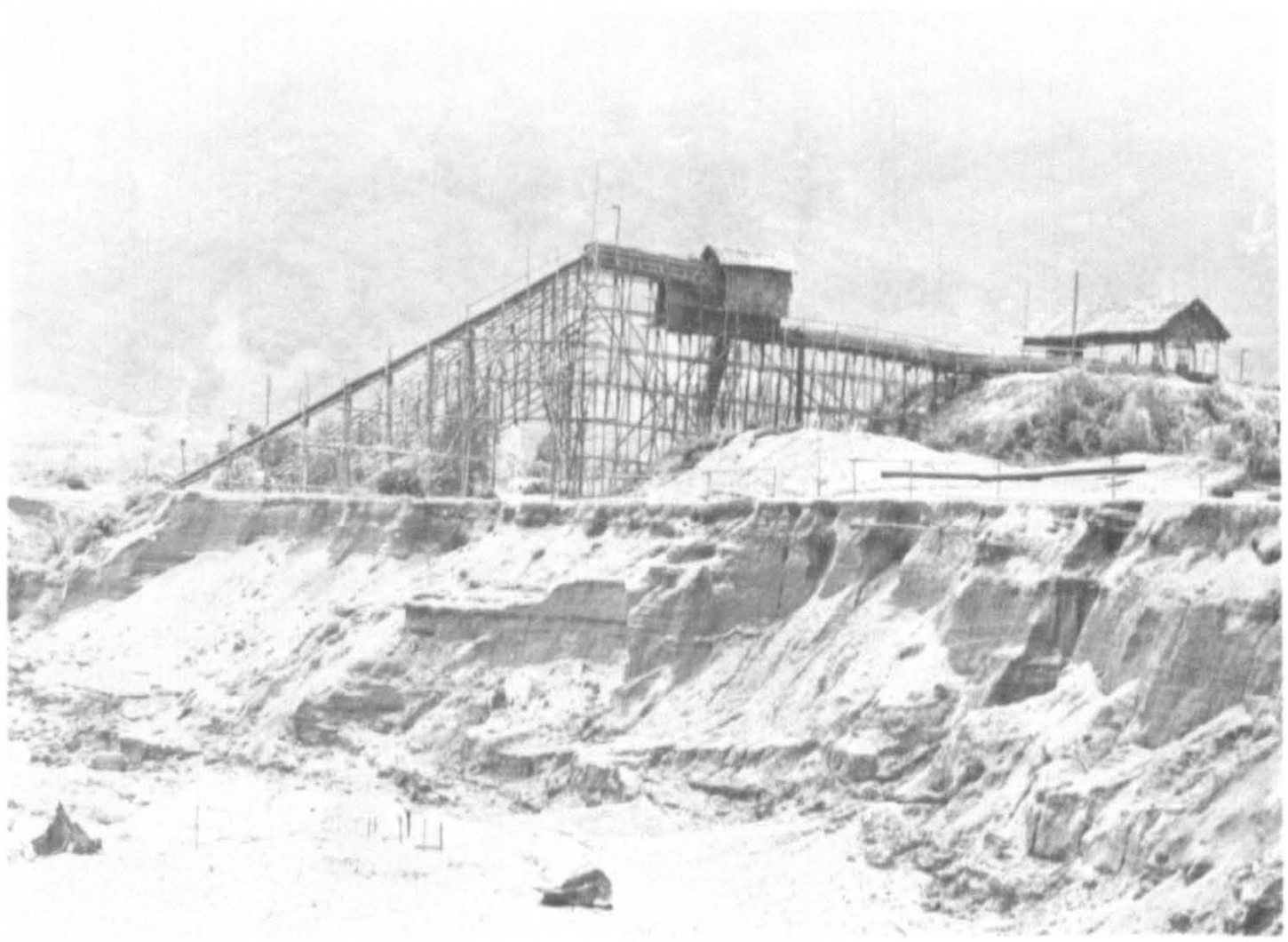


Plate 41 Mine workings in the Old Alluvium, south of G. Tempurong.

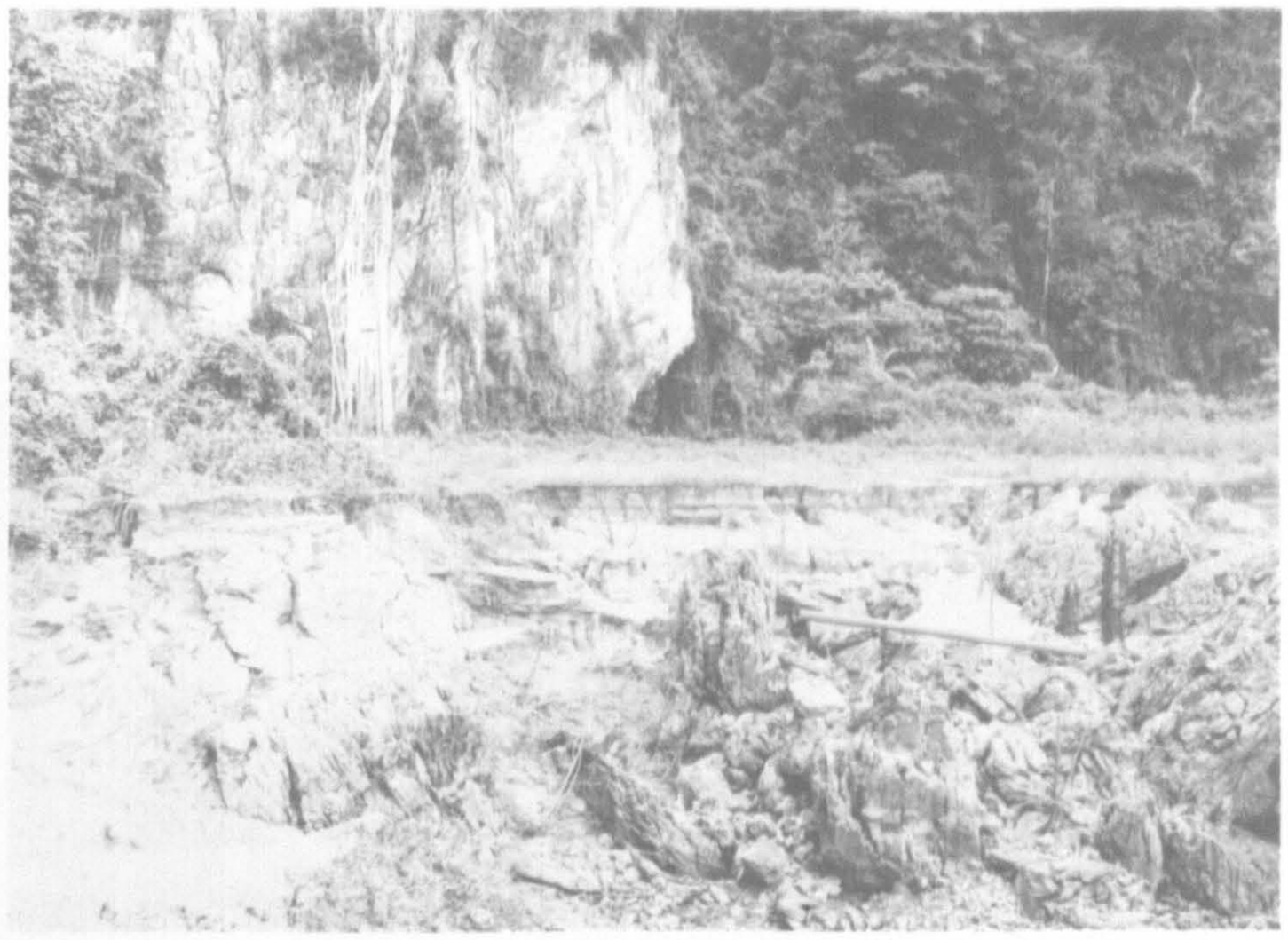


Plate 42 Sub-alluvial karst surface in densely jointed limestone, immediately north of G. Datok.

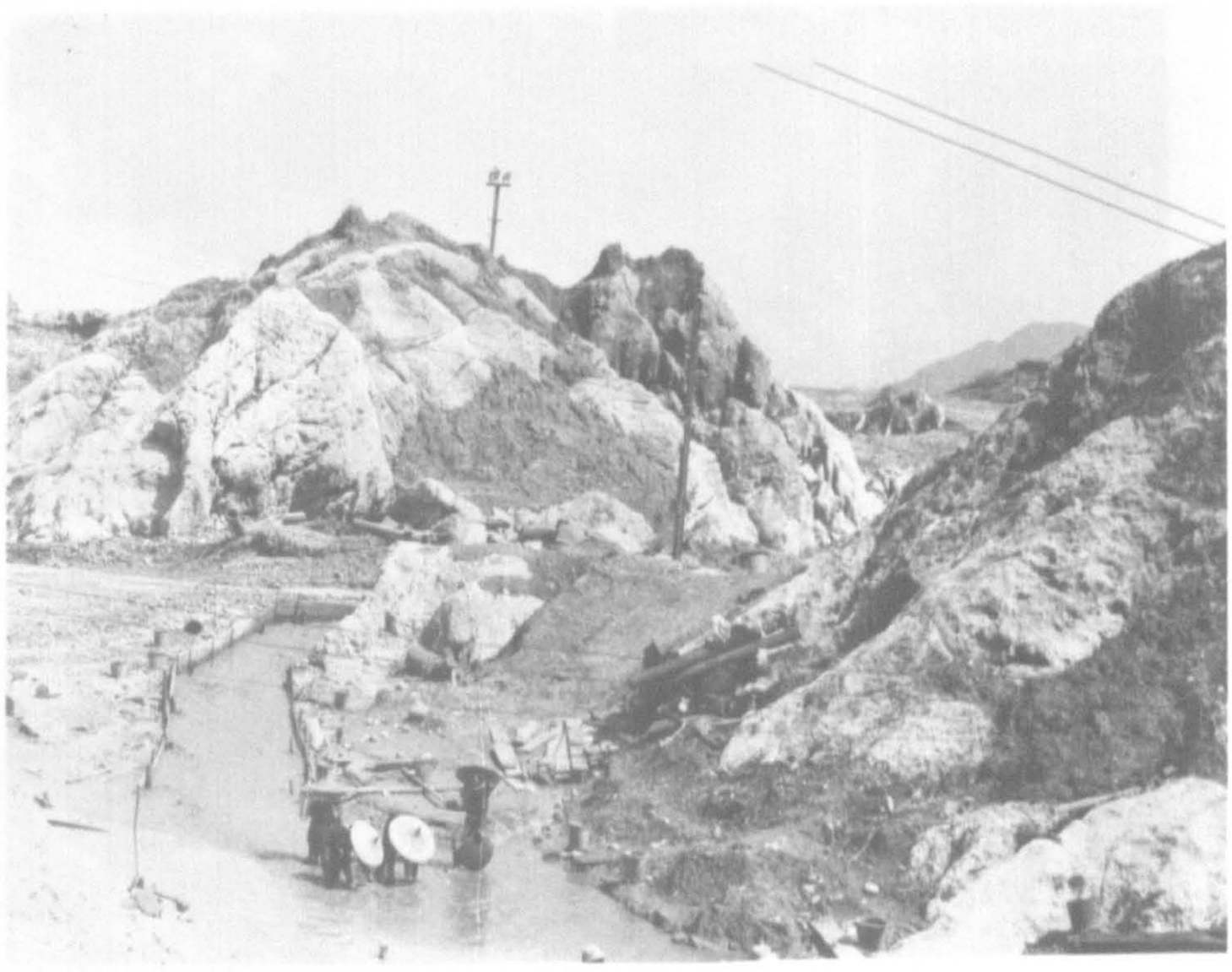


Plate 43 Residual blocks of more massive limestone exposed in mine workings near Gopeng.

Plate 44 Bt. Ketrii and other tower karst hills of the Chuping Limestone, viewed from the south-west.

Plate 45 Undercutting at the base of Bt. Kepala (Kodiang limestone). The upper, parabola-shaped notch is thought to be of marine origin (Jones, 1965, p. 27). The lower notch is probably attributable to more recent swamp action.

Plate 46 View of the eastern slopes of the Setul Boundary Range near Bt. Bintang. Photograph taken in February 1975 when many of the trees were bare.

Plate 47 Part of the narrow ridge of limestone which forms the south-eastern rim of Wang Tangga, viewed from Kaki Bukit.

Plate 48 Wang Tangga, viewed from south-east corner. Photograph taken in October 1974.

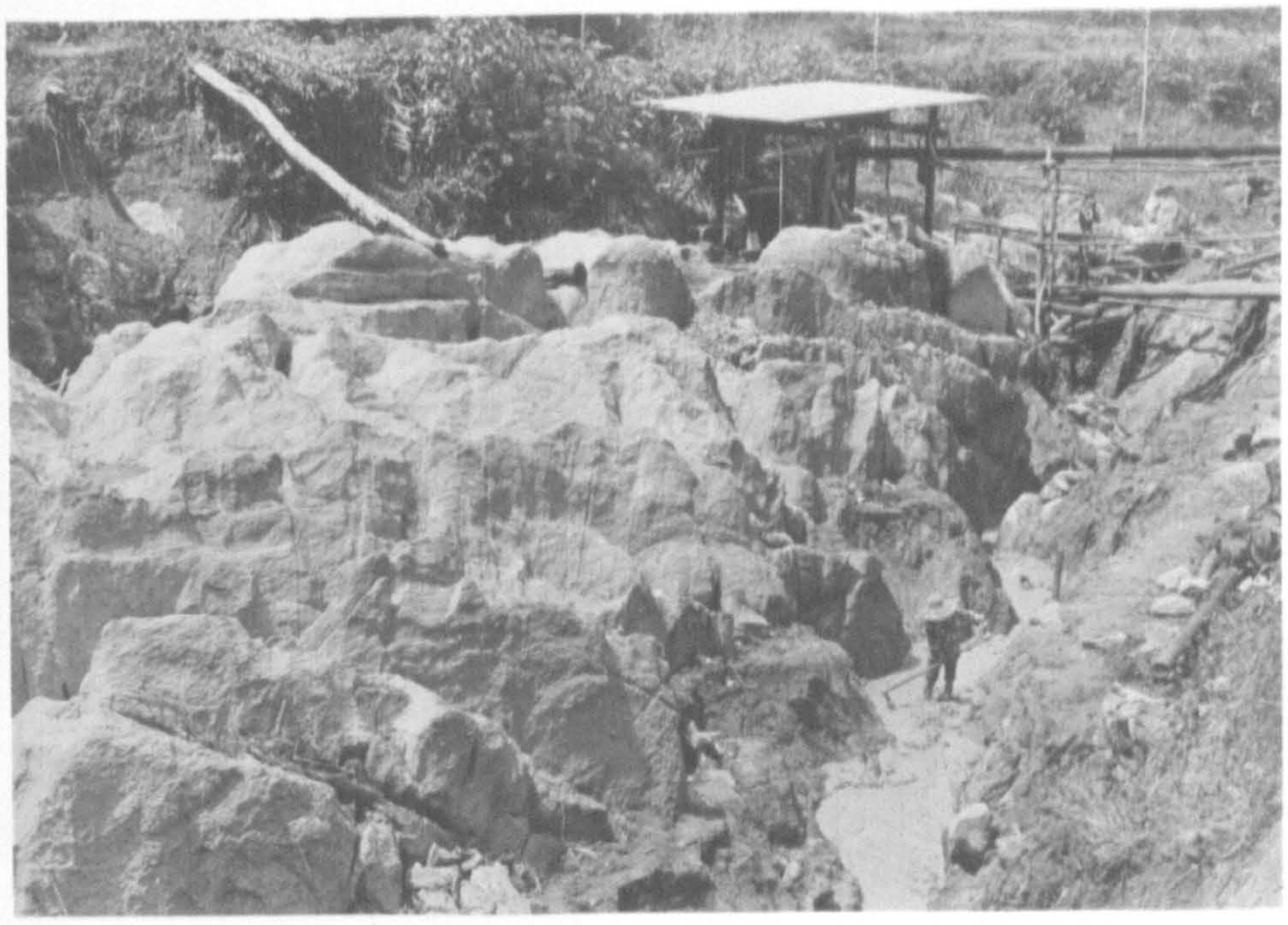


Plate 49 Sub-alluvial karst surface exposed in mine workings in Wang Tangga.

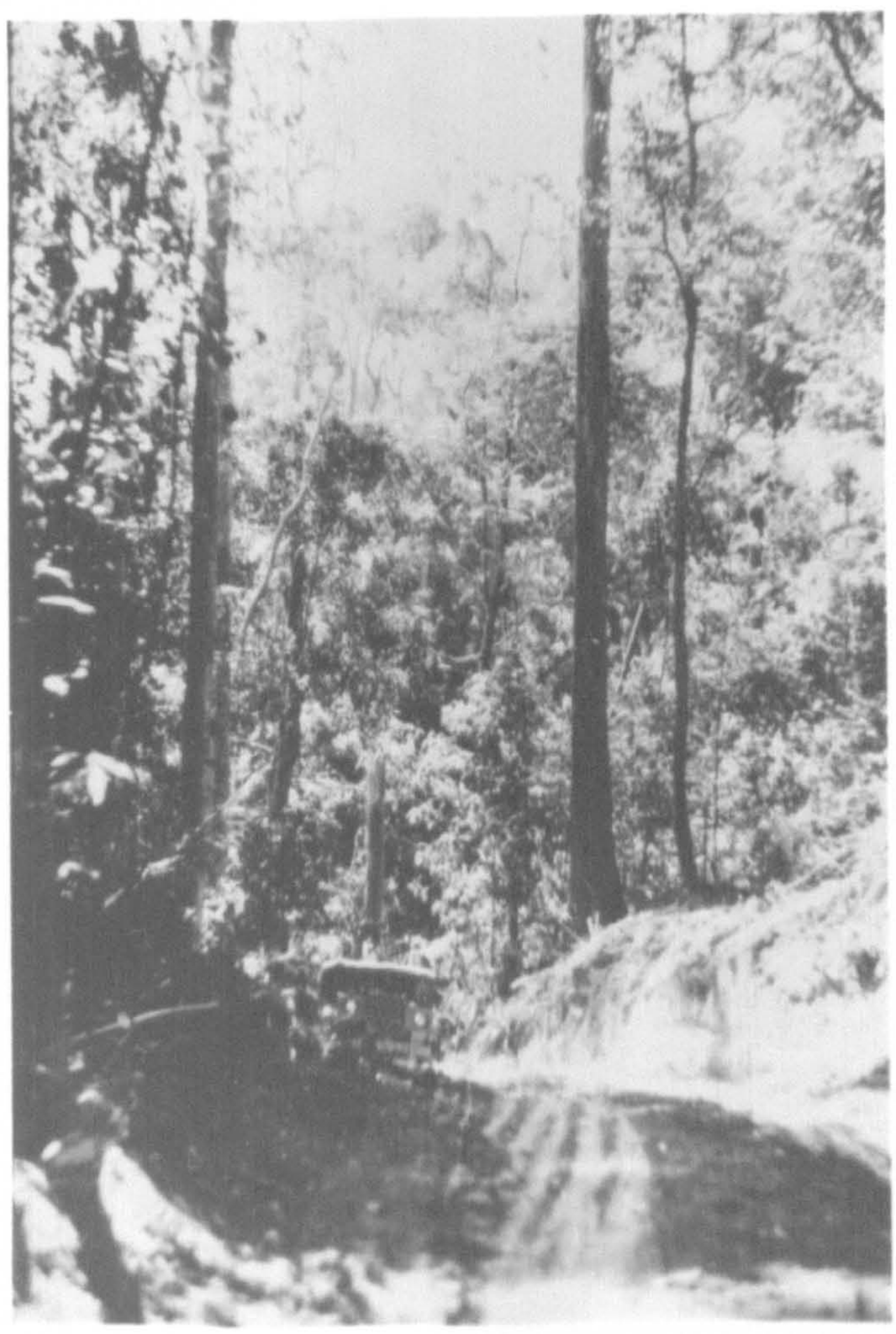


Plate 50 Deep soils and dense tree cover characteristic of slopes developed on impure limestones on the western side of the Gelang Kerbau Valley, Setul Boundary Range.

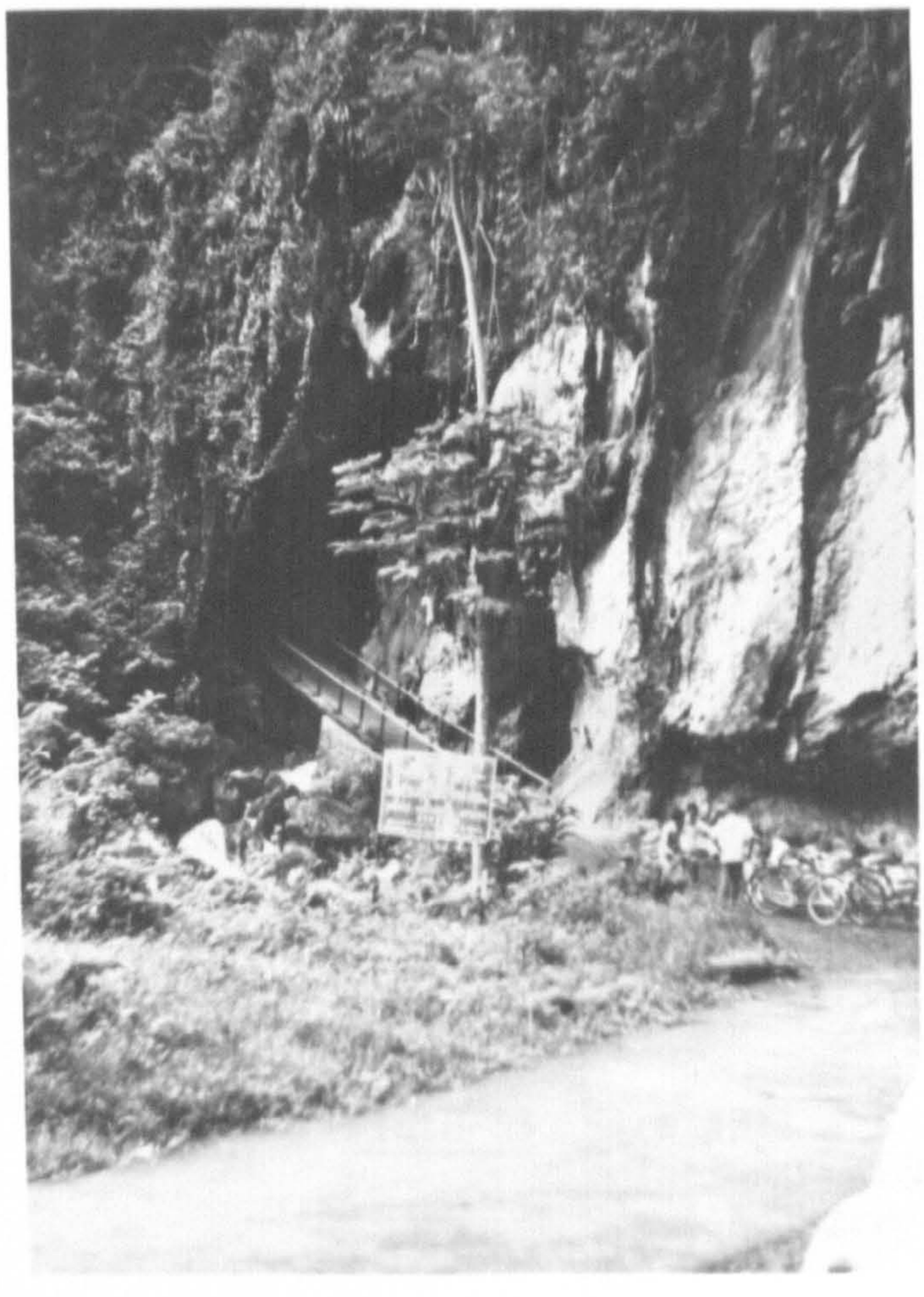


Plate 51

Eastern entrance of Subway Tunnel.

Plate 52

Entrance to Perlis Mines' Cave, Wang Tangga.

Plate 53 Site 4311 - a cave stream with free air circulation, which emerges at the foot of the Setul Boundary Range, south-east of Bt. Wang Mu.

Plate 54 Gua Ipoh (site 4318). The cave is penetrable for only several metres beyond the entrance.

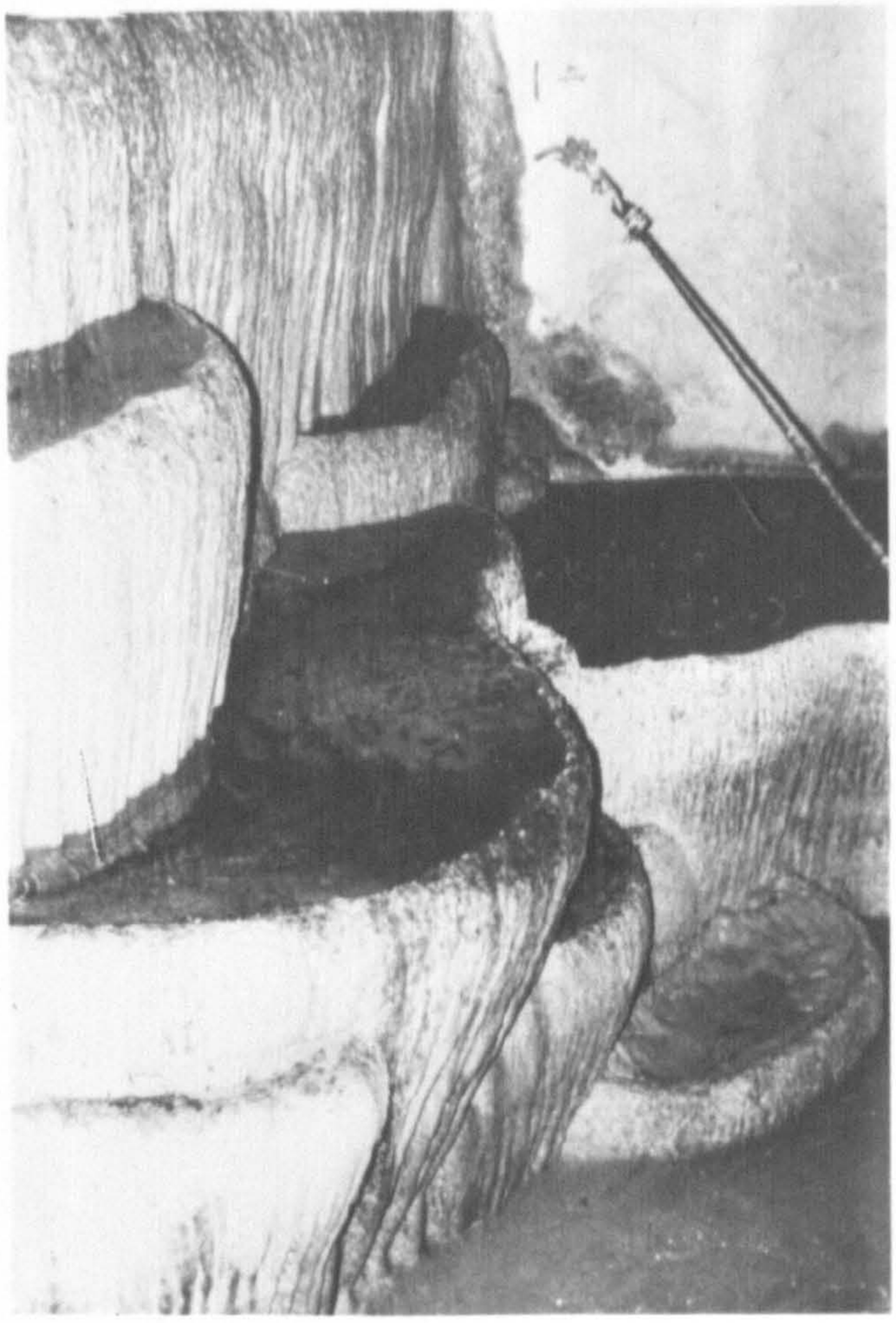


Plate 55

Active rimstone formations in Subway Tunnel. The walls of the individual dams are approximately 1 m high.

Plate 56

Surface deposition at the semi-natural seepage sites, Wang Tangga.

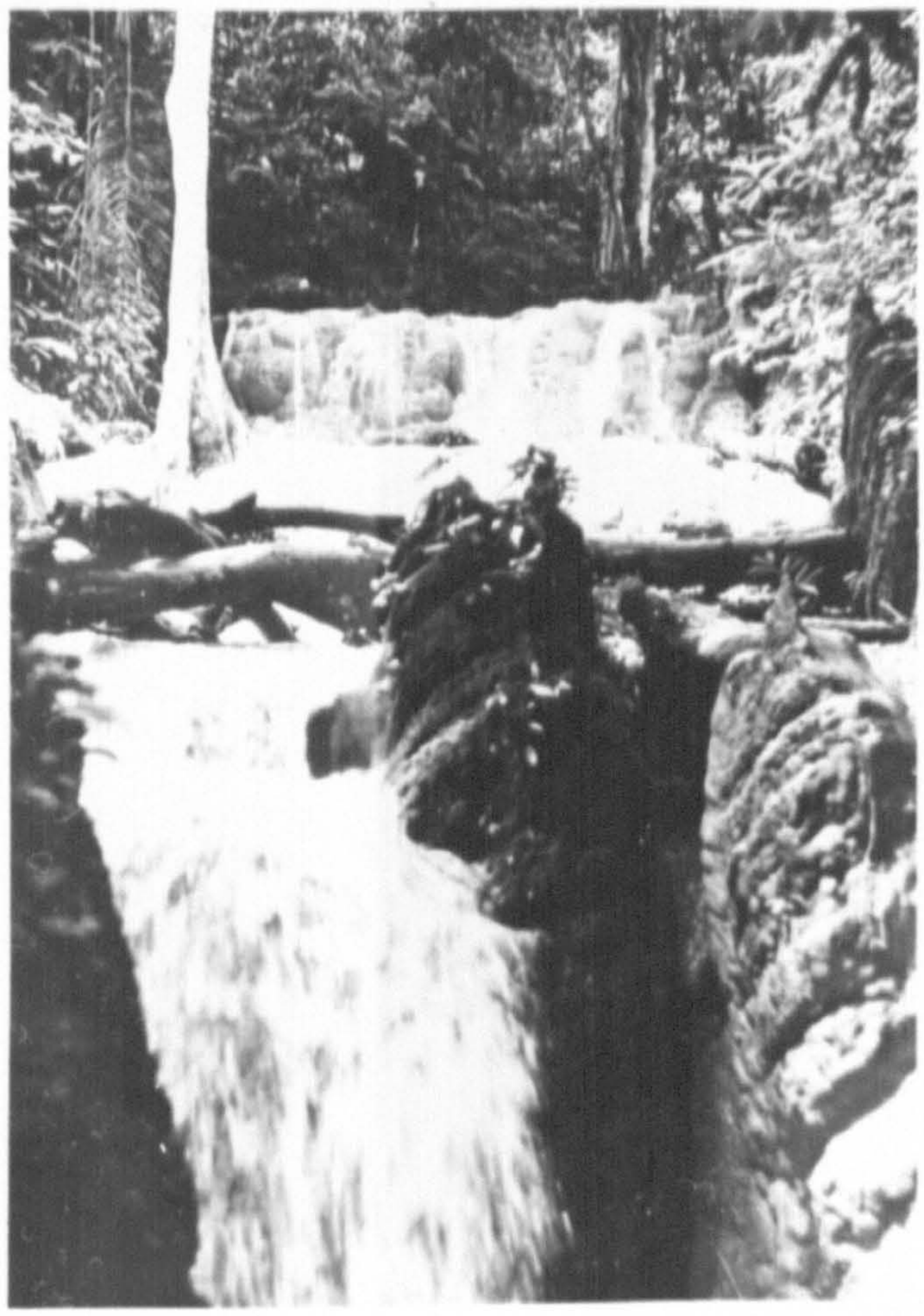


Plate 57 Tufa waterfall in the upper reaches of the Sg. Gelang Kerbau, approximately 200 m below the main rising (site 4315).

Plate 58 Small depositional dams, located about 100 m downstream from plate 57.