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THE SEDIMENTOLOGY AND PALAEOENVIRONMENTAL SIGNIFICANCE
OF VLEI SEDIMENTS ON THE WINTERBERG RANGE,
SOUTH AFRICA

BY

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"One never notices what has been done; one can only see what remains to be done."

Marie Curie
(1867 - 1934)
French scientist.

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'There are three things that make the world go round: support, education and communication'.

Anon.

1948.

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ABSTRACT

Palaeoenvironmental reconstruction of the late Pleistocene and Holocene geological periods for central and southern Africa has been hampered by the erratic distribution of suitable sites, incomplete and inaccurately dated sequences and the limited nature of published data. One geomorphological feature which has supplied valuable evidence for fluctuations in past environmental conditions, is the vlei or dambo. The type-site of these waterlogged features is in south central Africa, but similar features have been described on other continents. The clastic and organic sediments contained within these features are affected by, and therefore reflect to some degree, the environment under which they were formed. The characteristics of the sediments supply information as to their origin, transport and mechanisms of deposition. From these processes, the environmental conditions at the time of vlei formation can be inferred. The environmental history of the Eastern Cape region has been considerably neglected, and is far less well understood than other countries such as Malawi, Zambia and Zimbabwe. A study site in the Winterberg Range (Eastern Cape) was selected which permitted the comparison of two vleis, the objective being to establish an accurate late Pleistocene sediment chronology for the entire plateau area. Radiocarbon dates from organic layers indicate that these sediments span the last 12 000 years BP, suggesting that organic accumulation at this site began at roughly the same time as at sites further afield. The vlei sediments are analysed in terms of their morphology, particle size distribution, and other physical and chemical characteristics. These data facilitate the construction of detailed stratigraphic diagrams and a chronological summary of sediment accumulation, from which the period and governing processes of vlei development under changing environments may be described. It is found that the

Winterberg vleis contain sediments which respond to changes in the prevailing environment. This makes these sediments useful indices from which to trace such changes during the late Pleistocene and Holocene times. These features are found to be similar in many respects to those described elsewhere in Southern Africa. The study attempts to provide greater understanding of contemporary vlei processes and emphasises the necessity of their preservation, as finite and valuable resources, by future generations.

CHAPTER 1

INTRODUCTION

That all aspects of the environment have fluctuated markedly in past geological time was only acknowledged towards the end of the 18th Century, some 2 to 3 million years after human populations inhabited the Earth! The evolution of ideas and growing interest in recent years in global environmental change throughout the late Quaternary, has generated research in several fields. This study is concerned with this most recent and well documented geological time period, the period when human populations disseminated across the face of the Earth. At this point, the terminology used to describe these changes through time should be clarified. The Pleistocene and Holocene series together constitute the Quaternary period, although it is difficult to determine when exactly one time period ended and another began. These terms are used more for classification than to express the reality (Goudie, 1983). There is much confusion about the use of such terms, and a considerable range of views exist. However, in this study, the term 'late Pleistocene' is used to describe the last 50 000 or so years, with the 'Holocene' beginning around 10 000 years ago.

As information and evidence accumulated from widespread studies, it was appreciated that the last 20 000 years have encompassed significant environmental changes. These fluctuations have included the glacial maximum around 18 000 years BP when two-thirds of the Earth's surface was covered in ice, and isostatic movements of continents when unburdened of ice in warmer periods. Smaller fluctuations during the Holocene seem to have occurred randomly and without pattern, and it is this confusion of evidence which needs to be substantiated by further evidence. The second chapter of this study is entitled "Environmental change", and is primarily a review of literature pertaining to this aspect of

study. This review explains the concept of environmental change and outlines the existing data available for southern Africa. These data, particularly on the magnitude and frequency of past changes, are valuable in understanding our contemporary environment, as well as defining a clear history of human evolution.

A number of studies on geomorphological features in Central and Southern Africa have resulted in valuable information on environmental change. One feature which has aroused much interest in palynological and sedimentological study is the vlei or dambo, and is described in a review chapter (3). Vleis can be defined using a combination of geomorphological, hydrological and climatological terms, which are interrelated to form a complex system enabling the formation and development of dambos (Mackel, 1974). Being products of their prevailing environmental, the clastic and organic sediments contained within vleis are affected by and therefore reflect to some degree, any change in this environment (Bond, 1964, 1965, 1967; Meadows, 1985). These features appear to be fairly sensitive in their response to environmental change (Meadows, 1988). However, the distribution of available sites and, therefore, existing data is erratic and fairly limited, and has failed to produce a distinct pattern of change for the late Quaternary throughout Southern Africa. Due to the scarcity of published data for South Africa, and particularly the Cape Province, a site in the Winterberg (Eastern Cape Province) was chosen for this study, the environmental history of this area being considerably less well understood or documented than some areas in Malawi, Zambia, Zimbabwe, Botswana and Namibia.

This study attempts to accurately and chronologically describe sediments within waterlogged vlei areas, from which inferences of past environments may be made. The study is a comparative assessment of two Winterberg vleis, the objective being to establish a sedimentary

history for the entire catchment area. Radiocarbon dating of these organic sediments indicates that they span the Holocene and later Pleistocene of around 12 000 years BP. It has been suggested that radiocarbon dating of similar sediments elsewhere can reveal the sensitivity of these features to late Quaternary environmental change.

1.1 STUDY AIMS

Analysis of the vleis of the Winterberg, in terms of morphology, chemical and physical sediment characteristics, is made in an attempt to expound the problem of environmental variability throughout the Holocene for southern Africa. The Winterberg vleis offer a suitable opportunity to trace changes in the environment by interpreting the complex sediment stratigraphy within them. In this stratigraphy, it is hoped that a record of climatic and environmental variations and associated catchment erosion and deposition processes for the last 12 000 years can be inferred. A chronological summary of the depositional environment throughout the period of vlei development might enable comparison with similar trends from other parts of Southern Africa, particularly the dambos of the Nyika Plateau, Malawi.

1.2 SPECIFIC OBJECTIVES

A number of specific objectives for this study can be outlined.

- (a) To describe the characteristics of the Winterberg vleis and their sediments, in terms of similar features elsewhere.
- (b) To characterise some of the chemical and physical properties of the sediments found in the two Winterberg vleis, Dunedin and Salisbury.
- (c) To establish a sediment chronology by radiocarbon dating, so providing accurate ages of the deposits
- (d) To construct comparative stratigraphic diagrams to facilitate interpretation of sedimentary changes.

- (e) To infer from stratigraphic changes, fluctuations in environmental conditions at the time of deposition.
- (f) To outline the significance and contribution of this study in improving current understanding of variable past environments and the functioning and importance of contemporary vleï systems.
- (g) To conclude the study in contradiction to or in support of prevailing theory.

In attempting to fulfill these objectives, this study endeavours to provide a reasonably accurate chronology of environmental (including climatic) changes for this area on a local scale, possibly forming associations with data from similar studies on a regional scale in southern Africa.

1.3 SUMMARY

Given that these vleï features contain complicated stratigraphical sequences, interpretation of past geomorphic history is often difficult and confusing. It is thus necessary to interrelate all spheres of the environment when attempting to understand the stratigraphy and to estimate the extent to which sediments reflect changes in one or more of these spheres. A stratigraphic sequence offers a wide source of information on the processes acting upon the system through time, and these processes affect and are affected by one another, thus they result in a complicated range of effects. The ultimate objective of this study is to understand what processes are operating in the contemporary environment, and those which may have operated in the past. In this way, this study can supply data to the quest for a palaeoenvironmental pattern for the late Quaternary in southern Africa.

CHAPTER 2

ENVIRONMENTAL CHANGE: A REVIEW

2.1 INTRODUCTION

Many biogeographical and geomorphological studies have resulted in evidence that the past environment has been extremely variable during the Quaternary period. This evidence, widespread and conflicting in many spheres, has resulted in a complex scenario of data lacking either synthesis or structure. Gradually, compilation of available data has led to the suggestion of links both spatially and temporally, and the appearance of a global pattern of environmental change. The nature of evidence for southern Africa is chaotic and random, hindering any attempt at organization, being variable in concepts of place, time and climate. In addition, the area has thus far been insufficiently researched. An evaluation of existing data is needed to establish some chronological pattern for this large and heterogeneous sub-continent: a daunting task!

The concept of multiple glacials and interglacials was confirmed by extensive studies conducted in the latter half of the 19th Century, once the convergence of opinion from numerous sources was accepted and finally incorporated into the existing ideology. Louis Agassiz in 1837 (Imbrie and Imbrie, 1979) postulated that, in the geological past, the Earth's climate was much cooler than the present day, and spreading of the polar ice sheets took place. In the 1860's, research showed this spread of glaciers not to have been a single event but rather that it took place in various stages, with glacials separated by warmer interglacials, during which the climate was at least as warm as that experienced during the present day (Goudie, 1983; Meadows, 1985). In other words, a complicated picture had been sketched of climatic oscillations involving temperature as well as other factors captured beneath

the umbrella term "climate". These oscillations are of great interest to scientists of many disciplines in that they constitute one of the strongest influences on contemporary plant and animal distribution patterns. Since these distributions are largely controlled by environmental factors (Stott, 1981), any significant change in any one parameter will cause an expansion or contraction in their ranges and may even affect speciation and the course of evolution itself.

As climatic variation is linked most closely to the expansion and contraction of the major ice sheets, it was originally thought that the climate of the tropical areas of the globe remained fairly static and that equatorial latitudes were relatively unscathed (Flenley, 1979). Contemporary scientists have re-examined the concept of climatically "stable tropics", showing Quaternary fluctuations to have been global. It is this 2 million year Quaternary period which is most important and for which records are most complete, as it is the unique period when human populations achieved ecological dominance on Earth (Tivy, 1982; Goudie, 1983). Accumulated evidence for environmental change has been proposed from fields such as geomorphology, limnology and palynology. Although this chapter is a review of selected evidence, it is not the author's intention to examine all published data. This has been adequately covered by several workers (Deacon *et al.*, 1984; Vogel, 1984; Tyson, 1986), and this chapter attempts rather to outline the major patterns postulated for the late Quaternary in Africa and southern Africa. In addition, this review hopefully indicates where sedimentology can and has played a role in elucidating environmental change.

2.2 TECHNIQUES FOR DETERMINING ENVIRONMENTAL CHANGE

Various techniques have been developed to study past environmental change. New isotopic dating techniques have improved the accuracy of micro-fossil dating, the aims being to provide accurate dates for environmental

change with the objective of establishing a spatial pattern for global Quaternary climates. The interpretation of changes in plant and animal distributions during the past 30 to 40 000 years in particular, have also been made possible (Meadows, 1985). Thus, newly developed techniques have tended to supplement rather than replace more traditional methods.

2.2.1 BIOLOGICAL EVIDENCE

Much of the emphasis of research reconstructing past environments, has been placed on biological evidence, and palynological information on vegetation changes through time is widely available, as well as the analysis of micro-fossils including pollen (Faegri and Iversen, 1975; Moore and Webb, 1978), molluscs, diatoms and beetles (Parmenter and Folger, 1974).

Information on Quaternary climates has been derived from deep-sea coring techniques, the ocean floor offering a more lengthy and continuous stratigraphic record than terrestrial deposits. Oxygen isotope ratios of fossil Foramenifera reflect changes in sea temperatures and thus indicate the periodicity of major glacial and interglacial episodes (Shackleton, 1969). Freshwater diatoms and phytoliths from the sea floor can also be used to assess arid, moist, warm and cold phases of the Quaternary (Parmenter and Folger, 1974). Although the contribution to knowledge of past environments from biological sources has been and continues to be enormous, geographical expansion of research has led to a broader spectrum of sources of data. In areas of arid or semi-arid climates, sites for preservation of pollen are scarce, and attention has focused on other sources of information including geomorphology.

2.2.2 GEOMORPHOLOGICAL EVIDENCE

Some chronological estimates have been obtained by the relative dating of geomorphological features such as moraines, degraded slopes, alluvial terraces, weathering products, soil development and pan deposits

(Cooke, 1975; Butzer, 1974, 1984). Geomorphological changes that have occurred through time depend on a variety of factors (geology, climate etc.) and provide an indication of the relative age of features found in an area in the absence of absolute dating (Birkeland, 1974).

One of the most satisfactory methods of using landforms to reconstruct the dry or interpluvial phases of the Quaternary, is the study of the former extent of major dunefields by fossil forms (Flint and Bond, 1968; Lancaster, 1979). For example, active sand dunes are found in areas where precipitation levels are below 100-200mm per annum, while above this level, sand movement is reduced by increasing vegetation cover. Fossil dunes in areas of higher rainfall suggest rainfall levels have increased since the time of dune formation (Goudie, 1983).

Changing lake levels provide data on the balance between rainfall inputs, evaporation and surface area. If temperatures for the Quaternary can be estimated, additional rainfall inputs necessary to account for lake volumes at levels higher than present can be calculated. For example, in East Africa, assuming temperatures in the early Holocene (9 000 to 6 000 BP) were 2 to 3^o C lower than the present, rainfall totals would still have to have been 165% of the present day totals (Butzer et al., 1972; Grove et al., 1975).

Of importance here is the link between climate and geomorphological processes, in that climate influences not only the processes but the landforms resulting from these processes. In this way, landforms and geomorphological features can be used as evidence for climatic conditions which may have occurred in the past.

2.2.3 PEDOLOGICAL EVIDENCE

Soil formation indicates relative geomorphic stability in that deep palaeosols represent minimal or decreased

environmental fluctuation. The development of a soil, reliant on climate, fauna and flora, parent material and the erosive or depositional effects of the environment, requires time, and the characteristics that these soils develop, such as gleying and laterite development, alter to parallel environmental changes (West, 1972; Goudie, 1983).

Soil deterioration can also be used to assess changes in climatic conditions. For example, intense leaching under warm, moist conditions leads to the development of podzols and hardpans. These hardpans restrict drainage and encourage waterlogging, bog formation, and the accumulation of peat deposits (Pearsall, 1964). Palaeosols serve as pedostratigraphic units to identify periods of stability interrupting a sedimentary sequence, and one can characterize some of the environmental conditions of these periods of landscape instability (Jungerius, 1976).

2.2.4 SEDIMENTOLOGICAL EVIDENCE

Isotopic dating has allowed for very precisely dated sediment chronologies, and has been applied to soil carbonates as well as soil organic colloids. The carbon-14 technique has been refined and is used to date materials to 30 000 years BP with considerable accuracy (Butzer *et al.*, 1972; Birkeland, 1974). Dates have been derived up to 50 000 years, although the risk of inaccuracy increases dramatically beyond 40 000 years BP (Grootes, 1978). Reliability of radiocarbon dates is reduced by sample contamination, leaching or precipitation of carbonates, and deeply penetrating rootlets in spring peat materials (Scott, 1982).

Radiometric dating of sediments and studies of the lithological and stratigraphic characteristics of core sediments provides information on sediment sources, the periodicity of major glacial episodes as well as palaeotemperature changes throughout the late Quaternary (Bousman *et al.*, in press; Dardis, in press). Other evidence includes information from

deep-sea cores, fluvial palaeosediments at river mouths and in lakes, aeolian deposits and various clay minerals within the ocean floor strata, all indicative of weathering regimes on land (Street and Grove, 1976; Lancaster, 1979; Goudie, 1983; de Villiers, in press).

Organic sediments and peats, their development promoted by waterlogged situations and a continual addition of organic debris, contain pollen, derived from both regional and local vegetation. Shifts in vegetation boundaries, caused by changes in environmental factors influencing their distribution, is mirrored by pollen grains found in such sediments. Species associations found to exist are a reflection of environmental conditions as plants are characteristic of their habitat. Palynology, a product of sedimentology, has become an extremely valuable tool in the elucidation of past environments.

However, when attempting to interpret such data, it should be remembered that climate, only one of the soil-forming factors (Jenny, 1941), is an extremely variable and complex factor. Although soils and sediments do provide valuable evidence, including the preservation of pollen, caution should be employed during interpretation of results (Birkeland, 1974). This becomes even more problematic in areas such as South Africa where the climate is extremely diverse and any attempt at synthesis remains complicated.

2.3 AFRICAN EVENTS DURING THE LATE QUATERNARY

The late Quaternary was characterised by two extreme periods: a cold phase termed the last glacial maximum at +18 000 years BP, of extremely low temperatures and greatest ice sheet extents, and a warmer phase spanning the Holocene through to the present. Following the last glacial episode, deglaciation is thought to have begun at +14 000 years BP in the northern hemisphere (Goudie, 1983). However, maximum extent of glaciation may not have occurred around 18 000 BP in all

latitudes. It is thought that amelioration in the southern hemisphere may have commenced at $\pm 16\ 000$ BP, and even as late as $13\ 000$ BP in South Africa, several thousand years in advance of the northern hemisphere with its more extensive continental ice sheets (Deacon *et al.*, 1984). This suggestion had been disputed earlier by Williams and Adamson (1980) supporting a synchronicity of glacial episodes between the two hemispheres as evidenced by a colder phase culminating around $17\ 000$ BP in Antarctica in the south and Greenland in the north. This idea was further supported by Livingstone (1975) who suggested glaciation in East African mountains ceased some $15\ 000$ years BP, the same time as the Wurm glaciers in Europe started to wane, although this warming effect only affected vegetation some $2\ 400$ years later.

Goudie (1983) views southern hemispherical glaciation as less important in areal extent than the northern hemisphere, and suggests this as the reason for the chronological disparity. This interpretation may be too simplistic as, latitudinally, much of the southern hemisphere, where one would expect glaciers to have formed, is ocean, giving a bias to the available evidence from the two hemispheres. The idea that the southern hemisphere was relatively unscathed by glacial activity is also contradicted by recent evidence from Lewis and Dardis (1985), a number of studies suggesting that higher mountainous regions of South Africa were affected by extremely cold conditions during the Quaternary period. Evidence from the north-eastern Cape Province establishes that at least two cold phases existed in the southern Drakensberg during which permafrost and ice-wedges developed. Supporting stratigraphic evidence has been proposed by other workers in South Africa (Sparrow, 1967; Harper, 1969; Fitzpatrick, 1978). As suggested by Harper (1969), a first cold phase induced lower mean annual temperatures (MAT of -6° C, 19° C below present day temperatures) than the second, with past MAT of 14° C, although this later cold phase was still cold enough to initiate

solifluction and head deposits in the area (Lewis and Dardis, 1985).

2.3.1 AFRICAN PALYNOLOGICAL EVIDENCE

Much global evidence has suggested that the greatest extent of ice sheets was concurrent with the glacial maximum around 18 000 \pm 3 000 BP. Detailed pollen analyses from numerous tropical mountain swamps and lakes in East Africa show colder and drier conditions than at present (Livingstone and Van Der Hammen, 1978). Evidence from pollen types in equatorial Kenya and Uganda was documented by Van Zinderen Bakker (1962, 1964) and Coetzee (1972), and Hamilton (1974) respectively, confirming that climates at the last glacial maximum were drier and colder (6° C lower) than at present. Montane biome boundaries moved altitudinally from 1 500m to 700m during the glacials when temperatures were depressed. The forest/alpine zone boundary on tropical mountains reached a very low point between 18 000 and 15 000 BP (equivalent to the glacial maximum in higher latitudes), rose steeply as climates ameliorated from 14 000 to 9 000 BP and reached modern altitudes by 7 000 BP (Flenley, 1979). Similar studies in Uganda (Morrison, 1968) also indicate a dry climate and a MAT 6° C lower than the present around 18 000 BP.

Pollen spectra from East Africa indicate a colder, drier climate than the present between the glacial maximum and 12 500 BP, which implies that there were times of relative drought in much of Africa north of the equator (Adamson *et al.*, 1980). Pollen spectra from Malawi indicate that grassland species dominated widely since the end of the last glacial to \pm 12 000 BP, forest vegetation having declined due to increasing aridity in southern Africa (Meadows, 1982). In support of this evidence, deep-sea cores off Senegal show low levels of tropical pollen around 18 000 BP and increasing aridity through to 12 500 BP when tropical pollen was totally absent (Rossignol-Strick and Duzer, 1980). Late Quaternary climatic change has had

extremely marked effects on terrestrial ecosystems, including the equatorial forest biome, previously considered environmentally stable (Hamilton, 1976, 1982). Palynological evidence appears to support that much of the interior, at least, of Africa was colder and more arid than the present, from the last glacial through to 12 000 BP.

2.3.2 LAKE LEVEL EVIDENCE

Aridity during the last glacial phase can also be inferred from lake level evidence, particularly the low stands experienced by almost all tropical African lakes (Butzer *et al.*, 1972). A core from Lake Victoria (Kendall, 1969) shows massive dessication around 14 730 BP with a level 26m below contemporary levels. Later coring by Livingstone (1975) showed a low of 75m below current levels. The total depth being only 79m, the lake may have dried out completely around 14 000 BP. Evidence suggests refilling began at 13 000 BP and overflow at 12 000 BP. A temporary arid interruption occurred around 10 000 BP and a last high 'wet' peak at 6 700 to 3 700 BP. Comparable overflow occurred at Lake Tana into the Blue Nile around 13 000 BP. This implies that the last glacial phase was a time of drought in much of north Africa (Goudie, 1983).

Evidence for a simple model linking glacial periods with pluvial conditions (cold and moist) remains contradictory and shifting windbelts and pressure systems probably led to moisture availability varying widely between regions (Sarnthein and Koopman, 1980). Street and Grove (1976) correlated wet and dry phases in Africa using available radiocarbon dates, by plotting lake levels, indicators of net water balance rather than specific hydrological variables. Sequences from the tropics have three features in common: (a) a long period of high levels prior to the glacial maximum; (b) a pronounced minimum from 21 000 to 12 500 BP, the last glacial phase, (c) recovery of lake levels after 12 500 BP (Street, 1981). Differences between curves reflect only regional palaeoclimatic events

(Street and Grove, 1976; 1979; Figures 1 and 2). The prolonged low stand during the glacial maximum must reflect reduced rainfall (estimated decrease in annual input of 9 to 36%) despite a lowering of temperatures. At the glacial maximum, many tropical areas experienced increased aridity. However, exceptions do exist as in the highland areas of eastern Sahara (Williams and Adamson, 1980), northern (Rognon, 1981) and southern extremities of the African continent (Van Zinderen Bakker, 1982).

Opposition between the pattern of change for the extra- and intratropical areas is apparent. At the glacial maximum ($18\ 000 \pm 3\ 000$ BP), areas north of the Sahara and the southern African interior showed high lake levels (moister conditions) from 18 000 to 12 000 BP, a period of great aridity in East Africa (Street, 1981). Lakes in Ethiopia dessicated completely around 17 000 BP, while high stands in equatorial Africa were attributed to the influence of the Indian Ocean. Such geomorphic evidence demonstrates the vast climatic fluctuations in the tropical regions of Africa for the late Pleistocene and Holocene (Fairbridge, 1976), and suggests that the search for a pattern must continue.

2.3.3 OTHER GEOMORPHOLOGICAL EVIDENCE

Fossil dunefields, as found in the Sudan (Grove and Warren, 1968), supply evidence of shifts in wind and rainfall belts. Fossil dunes described in West Africa (Grove, 1958) and north-west Africa (Sarnthein and Diester-Haass, 1977) showed that marked changes have occurred in vegetation and rainfall conditions in many tropical areas, resulting in dune shifts of up to 1 200km. This movement has been attributed to drier glacial phases and higher trade-wind velocities (Parkin and Shackleton, 1973). Sarnthein (1978) states that deserts and dunes were more widespread at $+18\ 000$ BP, covering almost 50% of the area between 30°N and 30°S . Today they occupy only 10% of the area (Figure 3).

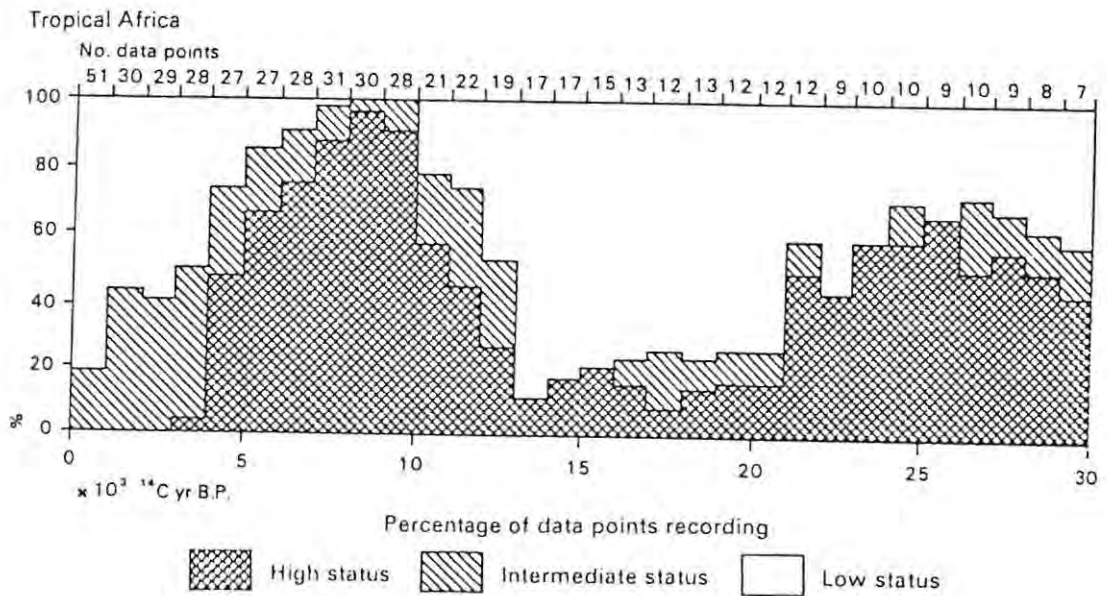


Figure 1: Lake-level histogram for the period 30 000 BP to the present day for intertropical Africa (modified from Street and Grove, 1979).

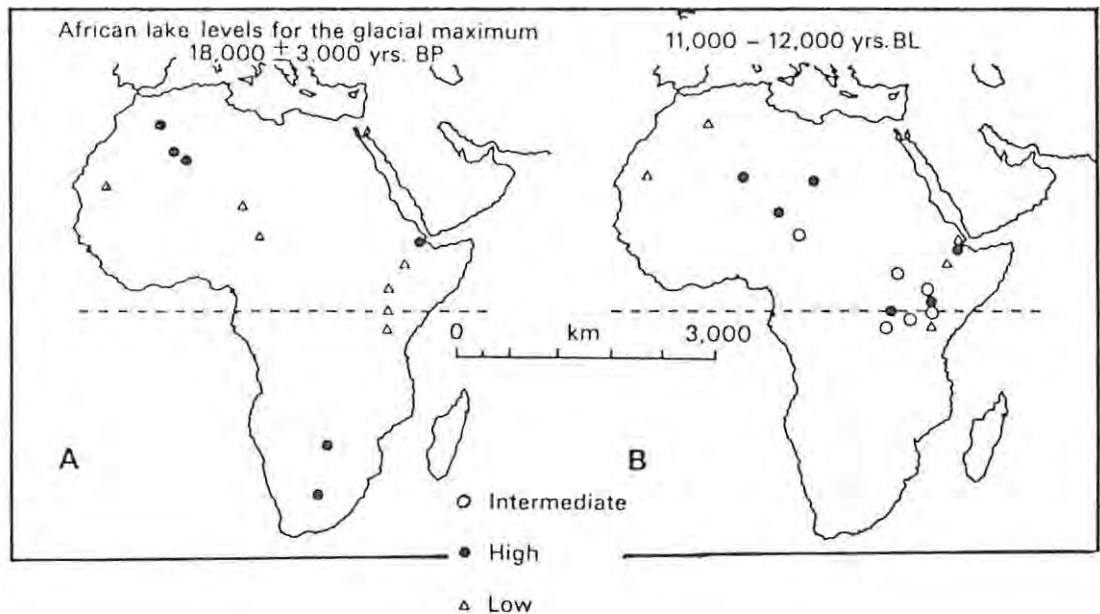


Figure 2: Radiocarbon dated lake level fluctuations in Africa: A. the glacial maximum; B. 11 000 to 12 000 BP (modified from Street and Grove, 1976).

Speculations of aridity during the last glacial maximum have been supported by desert dust in deep-sea cores in the upwelling area off northwest Africa (Street, 1981). Such episodes of maximum dust transport coincide with times of greatest ice build-up in the northern hemisphere. Along the lower White Nile, aeolian processes led to dune formation from 20 000 BP onwards. This cold, dry period ended around 12 500 BP, coinciding with overflow from Lake Victoria and higher rainfall in Ethiopia, leading to floods down the Nile (Adamson et al., 1980; Livingstone, 1980).

Screes of angular debris and similar periglacial deposits have been used to suggest depression of temperatures during glacial periods of up to 11° C in the Cape Province, South Africa (Goudie, 1983). This glacial aridity was confirmed from sedimentological studies off West Africa, where river-borne sediments were found to be absent around 18 000 BP, and aeolian silt fractions were high (Diester-Haass, 1980). A high percentage of quartz from atmospheric dusts was also found off the Sahara and Sahel region (Kolla et al., 1979). Many forms of geomorphic and terrestrial evidence thus retain the imprint of Quaternary fluctuations (Street, 1981) and indicate the dramatic effect these have had on the environment.

2.4 HOLOCENE ENVIRONMENTS IN AFRICA

The most recent geological time period is the Holocene, occupying about the last 10 000 years, and this has received the most attention worldwide. The concept of a stable Holocene environment was supported by Raikes (1967) who believed that since +9 000 BP, all global climatic changes had been 'local, random and of short duration' (Goudie, 1983, p.102). He largely ignored substantial evidence supporting a variable Holocene climate and claimed that climates were essentially similar to those of the present day. Compilation of radiocarbon dated faunal/floral remains, sediments, geomorphological and biogeographical evidence from all

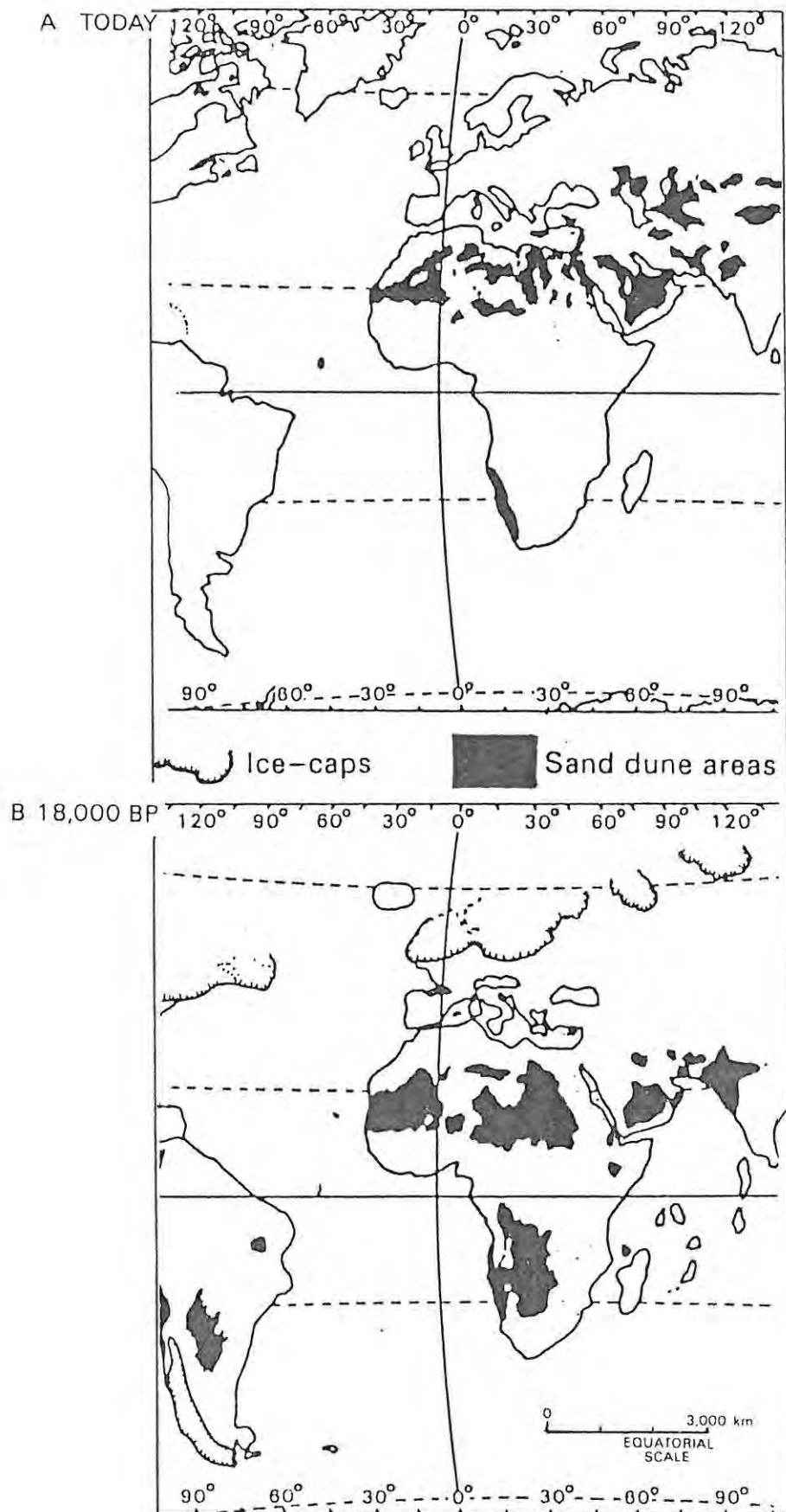


Figure 3: The distribution of active sand dune areas A. today; B. at $\pm 18\,000$ BP (modified from Sarnthein, 1978).

parts of the globe, appears to contradict the 'stable Holocene' concept, although the magnitude of temperature changes in the last 10 000 years (as compared to the Pleistocene) has been relatively smaller (Deacon and Thackeray, 1984).

For human populations, the increase in both temperature and moisture availability with the commencement of the Holocene, and concurrent recession of the ice sheets, introduced rapid and advantageous changes in the environment (Deacon and Thackeray, 1984). Land at higher latitudes became available for human, plant and animal habitation. Massive changes in African biomes were generated by changes in humidity and precipitation as well as temperature, and diversification of species into new habitats was controlled by several climatic oscillations. For most of Africa, the early Holocene was characterised by a moderation of environmental conditions from 11 000 BP, which continued to fluctuate through to the present day.

2.4.1 TROPICAL AFRICAN EVIDENCE

During the early Holocene in tropical Africa, when temperatures were similar to contemporary values, increases in rainfall were estimated at 25 to 85% (Lancaster, 1979; Street, 1981), the largest percentage being the Sahara's southern edge where totals today are low (Street and Grove, 1976). The reversal of temperatures from extreme lows during the last glacial, resulted in rapid oceanic warming, particularly between 9 000 and 8 000 BP. Tropical lakes e.g. in Chad, showed very high stands, and tropical river systems such as the Niger and those of the Sahel, changed from narrow streams to large drainage networks (Street, 1981). A very high strandline has been identified around 10 000 years BP in north-east Nigeria (Grove and Warren, 1968). It appears that many tropical lakes and river systems experienced high levels during the early Holocene. These pluvial conditions then declined somewhat to leave lakes at low and intermediate levels

(Figure 4), similar to those of the present day (Street and Grove, 1976).

2.4.2 NORTHERN AFRICAN EVIDENCE

Goudie (1983) provides evidence of three lacustral phases in the Sahara and adjacent areas from rock-paintings, pollen analyses and sediment dating. These occurred before 8 500 BP, from 7 050 to 4 150 BP, and from 3 550 to 2 450 BP and vegetation was denser than at present. Dry phases reactivated dune formation in semi-arid areas (Grove and Warren, 1968) and where the Sahara encroached, under the influence of more arid climates, the southward movement of vegetation belts must have induced massive effects on fauna and flora both in terms of species composition and their distribution patterns (Charney et al., 1975; Ripley, 1976).

Between 12 000 and 10 000 BP, the Nile filled rapidly due to overflow from lakes above, and became a perennial and permanent feature (Adamson et al., 1980). Rapid oceanic warming around 9 000 BP accompanied the largest expansion of lacustrine conditions in recent geological history, while the early and mid Holocene showed considerable fluctuations in levels, with many parts of Africa north of the equator experiencing low levels between 8 000 and 6 500 BP. From 5 000 BP through to the present day appears to have been one of the more arid phases, although dust transport was lower than during the glacial maximum (Street, 1981).

2.4.3 EAST AFRICAN EVIDENCE

Studies in East African mountains utilizing pollen analysis have shown good correlations between late- and post-glacial events in East Africa and European sequences (Table 1). The strength of this relationship has been questioned by Livingstone (1967): 'there is no discernible basis for a detailed correlation of either vegetation or climate on a zone-to-zone basis..' (Goudie, 1983, p.119). Nevertheless, Hamilton (1981) showed that forest elements spread along the western

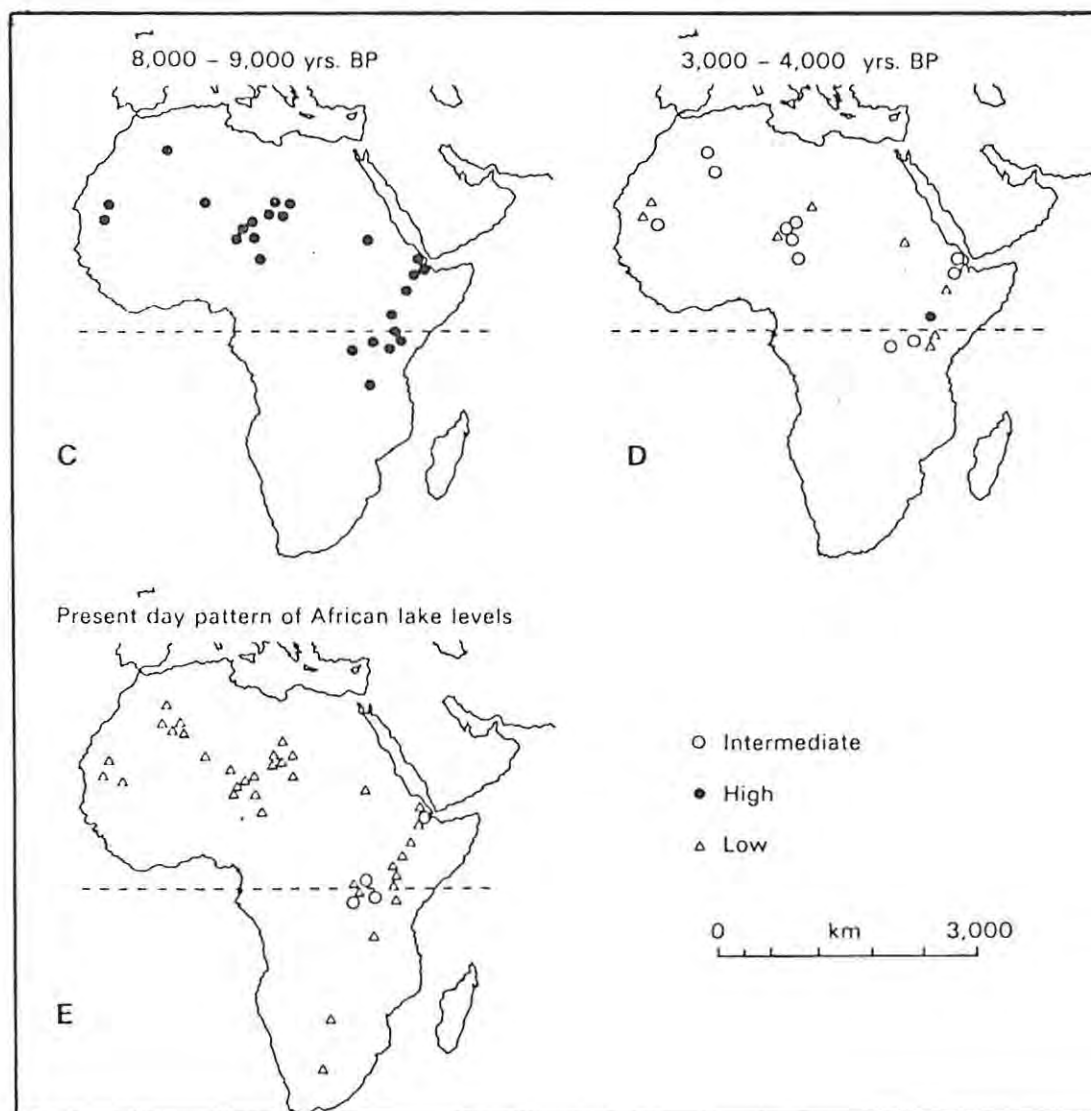


Figure 4: Radiocarbon dated lake level fluctuations in Africa: C. early Holocene, D. late Holocene; E. today (modified from Street and Grove, 1976).

Rift Valley, in support of evidence from a core from Sacred Lake of Mount Kenya (Coetzee, 1967) which indicates montane forest expansion after 10 583 BP, replacing grass/heathland. This pluvial phase correlates with the end of the late Glacial in Europe. A core from Muchoya Swamp, Uganda (Morrison, 1968) shows a change from grass/heathland to Hagenia forest at + 11 000 BP, implying a climatic transition from cold/dry to moist/warm. Van Zinderen Bakker (1983) also suggests that temperatures in the East African mountains were similar to the present day around 10 500 BP, followed by a lowering around 10 200 BP and very high lake levels from 10 000 to 7 500 BP. Forest vegetation in East Africa was more widespread from 5 000 and 2 650 BP than at present with a similar trend indicated in Europe (Goudie, 1983).

The effect of an unstable Holocene on vegetation, climate and fauna has aroused scientists within southern Africa to attempt to produce a description of climatic change for this subcontinent, an area thus far insufficiently researched.

2.5 LATE QUATERNARY ENVIRONMENTS IN SOUTHERN AFRICA

Van Zinderen Bakker pioneered ideas on recent climatic history in southern Africa in the 1950's but, by the 1960's, palaeoclimatic data was still seriously lacking for this region. The major reason was the deficit of suitable and accurately dated deposits, a problem on a subcontinent lacking significant basins of deposition, leaving the record incomplete (Deacon and Thackeray, 1984). Evidence from several sites has since been published, but has added to, rather than clarified, the chaotic and confusing nature of evidence. For this reason, selected evidence is described below in chronological divisions in an attempt to add structure to the jumble of available data.

2.5.1 LAST GLACIAL MAXIMUM: +18 000 BP

Table 1: Holocene environmental change in the East African mountains.

<i>Date BP</i>	<i>Vegetation</i>	<i>Climate at 2926 m</i>	<i>European equivalent</i>
2650	Tree-line descending	Wet, becoming colder	Sub-Atlantic
	Forest maxima	Warmer Colder	Sub-Boreal
4960	Forest closed, tree-line rising	Warmer and wet. Temperature rising	Atlantic
7740	Tree-line round swamp	Warmer and relatively wetter	Boreal
8625	Open vegetation, Tree-line below swamp	Cold and dry	Younger Dryas
9990	Compositae maximum	Warmer and dry	Allerød Older Dryas
10 790	Maxima of grasses and ericaceous belt	Cold and dry	Bölling
	Alpine vegetation	Very cold and dry	

(After Van Zinderen Bakker, 1962)

Possibly the most significant climatic event of the late Quaternary was its coldest interval around 18 000 \pm 3 000 BP for southern Africa. Several investigations of pans, lakes, caves and other landforms in central southern Africa, have permitted an overview of late Quaternary history for this region. The data presented is tenuous and complex, but some trends can be established.

Around 19 000 BP, low lake levels occurred in and around the Makgadikgadi basin (Botswana), due to decreased rainfall under a cold dry glacial climate (Heine, 1982; Vogel, 1982). The alignment of dunes formed during this period allows inferences of strong winds across the interior under anticyclonic weather conditions (Lancaster, 1979, 1981). The Kalahari and central southern Africa received limited rainfall which resulted in diminished woodland and savanna vegetation, and initiated the expansion of drier fynbos elements in the SW Cape (Van Zinderen Bakker, 1982).

Aeolian, fluvial and lacustrine landforms of northern Botswana were investigated by Cooke (1975, 1983) and results paralleled evidence for increased aridity and low lake levels for Lake Chilwa and Malawi (Malawi) and Lake Rukwa (Tanzania) around 16 000 \pm 3 000 BP (Street and Grove, 1976; Meadows, 1982, 1985). Isotopic studies in South Africa (Vogel, 1983) indicate MAT 5°C to 9°C lower than present at \pm 18 000 BP, having severe effects on plant and animal population numbers and their distributions. A continuous pollen diagram from Wonderkrater (Central Transvaal) shows evidence of cold and moist conditions from ericaceous elements dated at the last glacial maximum, followed by more ameliorating climates (Scott, 1983). Isotopic and sedimentological studies indicate that cool, dry conditions paralleled the glacial maximum in the southern Cape, as evidenced at Boomplaas (Deacon et al., 1984) and Cango caves (Vogel, 1983). Montane forest vanished to be replaced by grassland on the Orange Free State plateau, temperatures being 5 to 6°C lower (Van Zinderen

Bakker, 1976). A mean annual air temperature lowering of up to 19° C was suggested for higher altitude areas of the Drakensberg by Lewis and Dardis (1985).

Much evidence suggests colder and more arid climates in southern Africa around 18 000 BP, not moister conditions as suggested by Tyson (1986), although some data does appear to be inconsistent, such as the evidence from the southwestern Cape. Also inconsistent is the fossil pollen record at Wonderkrater which shows cool, moist conditions around 18 000 BP, bushveld having changed to grassland with alpine elements (Scott and Vogel, 1978). After 16 000 BP, a gradual amelioration of temperatures took place with a concurrent increase in humidity, ending at about 11 000 BP.

2.5.2 THE LATE GLACIAL: +16 000 to 10 000 BP

The cold dry glacial period was followed by a moister climate around 16 000 BP, suggesting more humid conditions for the late Glacial. At Urwi pan (Botswana), the existence of lacustrine deposits 2m thick and with an area two to three times the present pan, indicates former increased rainfall and reduced evaporation from 17 000 to 15 000 BP (Lancaster, 1979; Shaw, 1986; 1987). In the southern Kalahari, a three to four-fold increase in rainfall would be necessary to sustain permanent lakes. With lower temperatures, this would be at least twice the present day inputs (600 to 800mm per annum). Increases in rainfall of this order were suggested by Cooke (1975) for north-west Botswana. Sinter growth occurred at Kwihabe between 16 000 and 13 000 BP and estimates of a threefold increase in rainfall were made (Cooke and Verhaggen, 1977). The Kalahari must thus have experienced reduced evaporation and/or increased late Glacial rainfall.

Evidence for a humid phase for the majority of the interior south of the Equator, is further supported by carbon-14 dates from the Namib for a humid phase between 16 000 and 13 000 BP (Rust et al, 1984), the

Etosha Pan at +13 700 BP (Heine, 1982) and the Gaap Escarpment from 21 000 to 14 000 BP (Butzer et al., 1978). The Ngami and Mababe lakes rose to very high levels between 16 000 and 13 000 BP (Shaw, 1986). Lancaster (1979) suggests the rainfall increase was as a result of a more southerly position of the Inter Tropical Convergence Zone (ITCZ). Rust et al. (1984) propose that the increasing humidity was as a result of higher summer and winter rainfall on a seasonal basis. Butzer et al. (1973) suggest that rainfall at Alexandersfontein pan near Kimberley around 16 000 BP was 123% of present day figures, indicating a moister environment enhanced by lower evaporation rates from a 6° C drop in temperature. A variety of sites in the interior of South Africa thus suggest moister conditions between 16 000 and 12 000 years BP (Partridge, 1985).

Around 12 000 BP, when precipitation in much of northern Botswana was decreasing and dessication was setting in (Shaw, 1986), the climate further south was becoming warmer and summer rainfall was increasing, encouraging the southward migration of tropical vegetation into the Kalahari (Van Zinderen Bakker, 1983). Indeed, many areas indicated such dramatic amelioration that organic deposition occurred. An analysis of evidence for southern Africa by Meadows (1988) suggests an initiation of peat accumulation at +12 500 BP, and thus moister conditions, across much of the subcontinent. Peat accumulation dates indicate a moister phase from 12 600 BP to 10 300 BP, at sites as far apart as Cape Hangklip (Schalke, 1973), Winterberg vleis, Eastern Cape (Meadows et al., 1987), Rietvlei, Transvaal (Scott and Vogel, 1983), Inyanga, Zimbabwe (Tomlinson, 1973) and Nyika, Malawi (Meadows, 1983).

Palynological evidence from Lesotho, the Karoo and grassland regions (Van Zinderen Bakker, 1957; Coetzee, 1967; Butzer et al., 1973; Scott, 1983) and woodland areas (Butzer et al., 1978; Scott, 1982) has led to the discovery of several promising sites. Evidence from

Aliwal North indicates cooler, wetter conditions around 12 600 BP, followed by drier local conditions from 12 000 BP to 11 000 BP. Increases in moisture availability reappear at +11 000 BP and 9 650 BP (Coetzee, 1967). Evidence relating to the nature and distributional changes of late Pleistocene vegetation types and palynology has also been forthcoming from sites in Malawi (Meadows, 1984), Zambia (Cole, 1963; Clarke et al., 1964; Livingstone, 1967, 1975), East Africa (Lind and Morrison, 1974; Hamilton, 1982) and Zimbabwe (Tomlinson, 1973, Whitlow, 1984).

Some areas did not correspond with this change to a moister climate at +13 000 BP, such as the coastal regions of the southern Cape, the Namib desert areas and further south along the Orange River, which became more arid (Butzer et al., 1978). At Wonderkrater in the east, bushveld expanded in response to a semi-arid and cool climate (Scott, 1982). This was later replaced by woodland in the late Holocene in response to warmer, wetter conditions. Fossil pollen studied from Rietvlei spring deposits (southern Transvaal) also indicate cool, dry climates up to 11 000 BP which gave rise to open grassland with macchia elements, temperatures being an mean average of 5° to 6° C lower than the present day.

The humid phase below the Equator does not coincide with palaeo-lake studies north of the equator, where a number of sites showed low levels under a cool, dry climate from 17 000 to 12 500 BP (Street-Perrot and Roberts, 1983) and thus the pattern emerging for southern Africa appears to be at variance with the rest of the continent. Similarly, the western Cape coastal belt appears out of phase with cold temperatures and high winds resulting in more winter rain and open vegetation. On the southern coast, the late Glacial was characterized by cold, dry conditions, ameliorating between 13 000 and 12 000 BP, closed vegetation replacing grassland. This occurred while the interior experienced a decrease in precipitation, as indicated

by calcretes in the Kwihabe Valley and falling lake levels in the Mababe Depression (Shaw, 1986).

Attempts to quantify climatic change in southern Africa have not been consistent or reliable, and temporal as well as spatial variations have produced a complicated and somewhat unclear picture. Although no general palaeoenvironmental conclusions can yet be drawn for this large area, data from these sites will expand current knowledge of South Africa's past environment. Information is, in general, too limited to allow detailed reconstruction of past vegetation types by pollen analysis, and only isolated pieces of information are thus far provided. However, the major pattern for the late Glacial in the southern African interior appears to be widespread wetter and cooler conditions than the present.

2.6. HOLOCENE ENVIRONMENTAL CHANGE IN SOUTHERN AFRICA

The abstruseness of Holocene changes in southern Africa can be attributed, in part, to the relative scarcity of evidence and adequately dated palaeoenvironmental sequences. For much of the interior, cooler wetter conditions prevailed until $\pm 10\ 000$ BP, when drier conditions set in. Numerous palynological studies indicate that the early Holocene was drier and warmer than the later Holocene in parts of the interior (Van Zinderen Bakker, 1982). Undisturbed sediment sequences, particularly when waterlogged and plant decomposition occurs to produce peat-like sediments, are recognized as valuable sites for pollen preservation and stratigraphic information. Satisfactory pollen diagrams produced from such sequences throughout southern Africa, have been identified and many of them radiocarbon dated, for example in the eastern Cape Province, Drakensberg, Lesotho, Zimbabwe, Zambia and Malawi (Meadows, 1988).

Southern Africa has experienced irregular and frequent fluctuations in Holocene climates, compounded by the added complexity of seemingly random spatial changes.

Two main moist phases have been recognised as forming the major pattern of Holocene environments, one in the early Holocene around 8 000 BP and the other in the mid to late Holocene around 5 000 BP, with highest temperatures occurring in the mid-Holocene (Vogel, 1983).

2.6.1 EARLY TO MID HOLOCENE FLUCTUATIONS

Colluvia in Swaziland (Price Williams *et al.*, 1982) suggest a period of aggradation around 10 000 BP with more arid climates than at present, reduced vegetation cover and increased removal of weathered bedrock from slopes. Although this proposal does not accord with moister conditions reported for the northern Kalahari (Cooke, 1975; Lancaster, 1979), it follows a trend for an arid phase at +10 000 BP in East Africa (Peterson *et al.*, 1979) and in the Sahara (Grove, 1958). By 9 500 BP, conditions similar to the present occurred in northern Botswana, with semi-aridity and high temperatures prevailing until +6 000 BP. Similarly, pollen studies at Aliwal North indicated more arid conditions +9 600 BP (Coetzee, 1967). It was at this time that areas between the equator and the northern tropic were experiencing high lake levels, possibly due to a northern shift in the ITCZ (Kutzbach, 1983). Evidence for northern Botswana is also at variance with that for wetter climates in the southern and western Kalahari (Heine, 1982). Much evidence from arid zones appears to be frequently contradictory to results found in the more humid eastern parts of South Africa; the idea that moisture availability may have varied widely between regions is a definite probability.

Humid conditions, as shown by tufa deposits, were experienced at the Gaap Escarpment between 9 200 and 7 600 BP. Wonderwerk Cave (northern Cape) indicates cooler moister conditions at 9 200 BP from pollen studies, while at 8 000 and 6 000 BP, slightly drier conditions appeared to exist at this and many other sites where open savanna vegetation prevailed. An expansion of the bushveld elements accompanied later

changes identified as the warmest part of the Holocene at Rietvlei, Transvaal. These results compare favourably with those of Wonderkrater, but the highest temperatures here occurred around 8 000 to 5 000 BP (Scott and Vogel, 1978; 1983). Organic sediment accumulation took place between 11 000 to 6 000 BP at Rietvlei, Transvaal (Scott and Vogel, 1983) and Dunedin, eastern Cape (Meadows *et al.*, 1987), suggesting warmer and moister conditions. After this moister phase, bushveld began declining from 4 000 to 2 000 BP, due to lower temperatures and frost, as evidenced by peat units from Wonderkrater (Scott and Vogel, 1978; Scott, 1982) and Scot (Scott, 1982).

2.6.2 LATE HOLOCENE ENVIRONMENTAL VARIATIONS

The environmental fluctuations for the late Holocene are not well known for southern Africa. A moister phase is thought to have occurred in the middle to late Holocene, commencing at $+5\ 000$ BP, possibly earlier in the Transvaal (Scott, 1983) and as late as $3\ 500$ BP in the southern Cape (Scholze, 1983). The existence of this moister phase is supported by evidence from Lesotho, where grassland in the highland regions, possibly indicative of drier hotter conditions in the early Holocene, gave way to expanding woodland under warmer, wetter conditions in the late Holocene. At the north-western Cape/Orange Free State border and in southern Transvaal, forests became prominent only in the late Holocene at $+2\ 000$ BP. Evidence in both the Cape and Transvaal regions suggests the present vegetation pattern only developed during the last 2 000 years. This moister phase, commencing between $+5\ 000$ and $3\ 000$ BP is also evidenced by palaeosols, alluvial and colluvial deposits from the south-western Cape and Orange Free State (Butzer, 1984). Deposition on level pediments and downcutting of steep valley heads in Swaziland also represent the onset of modern climates (Price Williams *et al.*, 1982).

There is some evidence that small oscillations in temperature and moisture occurred which randomly

changed the composition of the vegetation within the Transvaal during the late Holocene. Evidence from sites in this region indicates warmer conditions at ± 5 000 BP, leading to cooler, moister conditions with more closed vegetation in the late Holocene (Scott, 1984). Table 2 lists various sites and radiocarbon dates of evidence for Southern Africa (Meadows, 1988). Moderate peat accumulation occurred between 5 000 and 3 500 BP in the Central and northern parts of southern Africa and Malawi, while Inyanga, Zimbabwe (Tomlinson, 1973) showed an acceleration in peat accumulation at this time. Organic accumulation also occurred at Wonderkrater (Scott, 1982) after 5 000 BP. Evidence from the southern Cape Province does not closely parallel that from the interior. This disparity could be due to moister climatic conditions reaching the southern tip of the subcontinent somewhat later than in the north, with peat accumulating from 2 000 BP onwards at Cape Hangklip (Schalke, 1973), Nuweveldberg, and Central Cape (Meadows and Sugden, in prep.). It seems this moister phase has continued through to the present day, with peat-like deposits still accumulating in several sites throughout southern Africa (Meadows, 1988). Although available evidence indicates a longer lasting moist period in the mid to late Holocene, this phase may have involved a series of minor fluctuations throughout the subcontinent. However, these phases were small and varied between different climatic zones.

Evidence from tufa and lacustrine deposits in the arid northern Cape indicate drier climates during this period, illustrating the variation in climatic conditions across the breadth of the region (Butzer et al., 1978). It is quite possible that moisture availability as well as temperatures varied widely from region to region within this area. A synthesis of Holocene environmental change is difficult to make in a subcontinent with such diverse climates and topography (Meadows, 1988).

Table 2: Radiocarbon dated palaeo-environmental evidence in Southern Africa.

SITE	NATURE OF DEPOSIT/EVIDENCE	CLIMATIC IMPLICATIONS
Cape Hangklip, SW Cape	Peat accumulates 11000-6000BP and 2000BP onwards	Moister 11000-6000BP; 2000BP onwards
Aliwal North, NE Cape	Organic clays commence accumulation 12600BP, peat commences growth 4320BP	Moister 12600BP, and, 12200BP Moister again 4320BP onwards
Alexandersfontein, OFS	Peat commences growth 4075BP	Moister 4075BP onwards
Wonderkrater, TvI	Peat commences growth 6330BP	Moister 6330BP onwards
Rietvlei, TvI	Organic clay commences accumulation 10300BP, peat commences growth 'after 7500BP'	Moister 10300BP onwards Moister after 7500BP onwards
Scot, TvI	Peat commences growth 5070/3570 BP	Moister after 5000BP onwards
Inyanga, Zimbabwe	Organic sedimentation begins 11750BP, more organic 4670BP onwards, peat accumulates after 750BP	Moister after 11750BP, moister 4670BP onwards
Nyika, Malawi	Organic sedimentation commences 11678BP, peat commences growth 4900BP	Moister after 11678BP, moister 4900BP onwards
Southwestern Cape	Soils and palaeosols	Moister in 'mid-Holocene'
Groenvlei, S. Cape	Lacustrine sediments and pollen	Forest expansion post 6870BP; moister?
Knysna to Tsitsikama, S. Cape	Palaeosols	Moister after 5000BP
Plorisbad, OFS	Spring and alluvial sediments	Accelerated spring discharge 5530 to 4370BP; moister
Pans, OFS	Lacustrine sediments	Moister 7700 to 6300BP and 3600BP onwards
Bushman Shelter, E. TvI	Tufa deposits	Moister 3670BP to 1400BP
Kathu Vlei, N. Cape	Vlei sediments	Moister 7500-5050BP; 5000-3750BP
Kathu Vlei, N. Cape	Pollen in vlei sediments	Moist/Dry/Moist in mid-Holocene
Gaap Escarpment, N. Cape	Tufa deposits	Moister 9700 to 7600BP; 3200 to 2400BP; 1300-250BP
Wonderwerk Cave, N. Cape	Cave sediments	Moister 10000 to 8000BP; 5500 to 400BP; 1900-800BP
Makgadikgadi, Botswana	Calcretes and mollusca on pan terraces	Dry throughout Holocene, moister 1700 to 1500BP
Maun to L. Ngami, Botswana	Laminated calcretes in fossil valley	Moister 9000BP
Auob Valley, Namibia	Geomorphological evidence	Moister 4500BP, then drier, then moister 3500BP
Auob Valley, Namibia	Calcretes	Moister until 7850BP

Although anomalies do occur, evidence from the northern Cape, Transvaal, Zimbabwe and Malawi supports the findings of Street and Grove (1976) for African lake levels across the entire continent. It should be remembered that although a substantial amount of local palaeoenvironmental information has been derived from palynological studies, sample contamination and inaccurate age determinations can lead to confusing results. Also, it is very difficult, if not impossible, to correlate information across great geographical distances.

2.7 CONCLUSION

Meadows (1988) states that the late Pleistocene/early Holocene seems to have been a period of great climatic dynamism incorporating rising temperatures and at least one moister phase. He emphasizes that data may represent a complex series of moister phases rather than a single event, especially since many of the sites at which evidence is found experience markedly different contemporary climatic regimes. Although interpretation of evidence from various sources is fraught with difficulty, it has allowed for the tentative reconstruction of a pattern of climatic change. Rewarding links have been forged with data within and outside southern Africa.

Although trends in data maintain a similar direction of change, temporal and spatial differences do arise. Butzer (1984, p.63) states: "the most striking aspect of the climatic history of the interior [of S.A.] is the rapidity of change...and the complexity of the evidence...that makes synthesis so difficult. In this sense the palaeoclimatic continuum resembles that of tropical Africa, with its wildly fluctuating lake levels and unpredictable spatial patterns, far more than it does that of Europe or North America." Butzer (1984) remains sceptical of attempts to define a glacial pattern of climate for any part of Africa, due

to the complexity and irregular periodicity of changes outlined by available evidence.

What is clearly required in Southern Africa is a comprehensive evaluation of the available radiocarbon dated evidence, as well as continued research for additional accurately dated deposits. This should be multidisciplinary in approach and integration; an attempt to define more logically the trends of climatic change on a regional scale. The most obvious difficulty here lies in the large and diverse area of this region, so far insufficiently researched and considerably lacking in suitable sites of deposition. Another complicating factor is the apparently random fluctuations of past climatic regimes. Understandably variable over such a large area, it is also an added disadvantage that the best sites for sediment deposition, pollen preservation and floral and faunal remains, rarely coincide. Palaeoenvironmental reconstruction for southern Africa remains an elusive goal at this point, available evidence sketching a complicated scenario of environmental instability, and, thus far, resulting only in a rudimentary outline of its environmental history.

It remains, then, for contemporary workers to focus their research on a more proximal scale, such that the comparison and compilation of results will eventually lead to a more meaningful text. The aim of this study is to add one more piece to the jigsaw of data already available for the Cape Province to substantiate evidence previously unearthed, and hopefully to result in association with data from other regions. The results should provide at least an indication of the frequency and magnitude of environmental changes in the region during the Holocene, to allow comparison with pollen profiles and descriptions from other local sites. At present, South Africa, has a substantial, although random, amount of information on a small-scale, with detailed evidence and good profile descriptions. However, at a synthesis level, this

information is still in a state of confusion. Only by the combination of data, can a more distinct proposal be assembled which has more clarity and definition. This proposal will then have wider applicability, and will improve the chronological and palynological framework of southern African environmental change.

CHAPTER 3

DAMBOS AND VLEIS: A Literature Review

3.1 INTRODUCTION

This review chapter provides an opportunity to define and describe vleis and dambos, and outline selected evidence on these features. Geomorphological phenomena termed dambos were first described by Ackerman in 1936 as periodically inundated grass-covered depressions in the headward region of a drainage system (in a region of dry forest or bush vegetation) (Whitlow, 1984). Such features are encountered in seasonal rainfall areas on gently undulating plateaux in the tropics and subtropics. As geomorphic features, they have only recently been recognized as valuable sites for the study of environmental change. Until 1984, literature describing these features was scarce and fragmented, but a compilation of data was made and published in a special journal issue (*Zeitschrift fur Geomorphologie Suppl.-Bd*, Vol. 52, 1985).

Ballantyne (1956) described dambos as containing seasonally or permanently wet soils or as having a high water table which inhibits the growth of woodland species. Possibly the most representative definition of dambos has been advanced by Mackel (1974; p.329): "shallow, linear depressions in the headwater zone of rivers without a marked stream channel". A battery of colloquial terms has emerged including 'mbugas' (Tanzania), 'fadamas' (Nigeria), 'vleis' (Zimbabwe and South Africa) and 'bolis' (Sierra Leone). These features are described as occurring throughout the tropics and subtropics, but the type-site is thought of as being southern Central Africa where the term dambo is most common (Mackel, 1974; Meadows, 1985).

Within this study, the term 'vlei' will be used to refer to dambo-like features as they appear in South

Africa, in the study area. The term 'vlei' has a much less restrictive usage in South Africa and applies to several geomorphologically developed areas which would fall astride the definitions of dambos discussed. It could be said that all dambos are equivalent to vleis, but not all vleis described are analogous to dambos. The term 'vlei' then, will be substituted for 'dambo' when discussing these features as they appear in the study area, although it is recognized that all vleis are not equivalent to dambos. For the most part, this chapter deals with evidence from areas outside South Africa, and thus the term 'dambo' appears more often.

3.2 THE DAMBO ENVIRONMENT

Various authors have described dambos in terms of geomorphological and hydrological characteristics, although aspects of geology, climate and ecology are also important in their development and preservation.

3.2.1 DAMBO DISTRIBUTION

According to Meadows (1982, 1985), dambos in Malawi develop where the wet seasonal rise of the watertable intersects or lies near the surface of the land, as in the following three situations:

- * watershed site: where watersheds between major river systems are smooth with negligible gradients, water collecting in the wet season forming broad, interconnected dambos;
- * valley site: dambos develop on the sides of gently convex interfluves or along axial drainage lines;
- * inselberg site: high runoff from exposed rock domes collects at the foot of these inselbergs and forms a pincer-shaped drainage pattern, the areas of the 'pincers' being dambo.

The two Winterberg vleis described in this study, Dunedin and Salisbury, are found in the first situation, the watershed sites, as they are located on the gently rolling plateau of the Elandsberg watershed in the Winterberg Range.

Mackel (1974) notes that dambos are usually associated with gently undulating areas at elevations of 1 100 to 1 500m above sea level. The climate of these areas is usually of a seasonal 'savanna' type with a marked wet and dry season, and dambos appear to develop where the mean annual precipitation (MAP) is above 900 to 1 300mm (Meadows, 1985). Their occurrence in an easily defined precipitation range (MAP above 900 to 1 000mm) has led Bond (1967) to suggest that dambos are good climatic indicators, although Meadows (1985) views this as being over-simplistic. Mackel (1974) also points out that there are dambos in Zambia on the central plateau region with annual rainfall differences of 500mm between the northern and southern regions, suggesting that climatic variations would have to be considerable in order to affect the dambo system. Well developed dambos also occur in the southern plateau region, although the MAP does not reach Bond's rainfall limit (1967). Mackel (1974) suggests that relief and seasonality of climate are the main environmental factors controlling dambo development, and this appears to hold true for the vleis of South Africa as well.

Meadows (1985, p.151) refers to dambos in Malawi at elevations between 2 000 and 3 000m (the upland form), where rainfall exceeds 1 000mm per annum and relief is of "rolling down-like montane grassland, with small patches of forest in the valley heads and other sheltered locations". For the most part, these environmental factors apply to vlei areas of the Winterberg; relatively moist areas above 1 500m and where rainfall exceeds 1 000mm per annum. Their distribution, related to relief and climate, is comparable to that described for dambos in upland Malawi, where they are found within a higher MAP than the lower-lying dambos described in Zambia. Much of the rainfall in the Winterberg is derived from convection storms, where high intensity precipitation supplies heavy sediment loads from sheetwash and headwater streams. As a result, these higher-lying vleis have deeper sediment accumulation than those

described in Zambia (Mackel, 1974). Sediment availability would also be affected by local geology, determining the rate of weathering.

Meadows (1985) recognizes two types of dambos: high altitude and low altitude dambos. The low altitude dambos are similar to those described by Mackel (1974) in Zambia and have a distinct 'wash belt' of coarser material at the margins, are broader, but are of shallower depth. These dambos could be compared with shallower vleis found in the more arid Karoo regions of South Africa, e.g. Beaufort West (Meadows and Sugden, in prep.). The high altitude dambos on the Nyika Plateau (Malawi) and Inyanga regions (Zimbabwe) are described as having no discernible wash belt, deeper sediments, grassland vegetation surrounding the dambos and occupying steep-sided valleys. Again, the Winterberg vleis are more typical of these high altitude dambos.

The density or spatial variation of Winterberg vleis is far lower i.e. less surface area is given over to these saturated areas, than on the plateau regions of Malawi (compare Figures 5 and 6). This is possibly due to the lower rainfall, generally lower elevation and original drainage density prior to infill, of the South African inland plateaux.

3.2.2 DAMBO MORPHOLOGY

3.2.2.1 Plan Morphology

Dambos vary widely in surface length and width, reflecting regolith depth and geology, but appear to follow a characteristically linear pattern. Generally, dambos are broader in the headwater zone, narrowing downstream and sometimes containing a permanently flowing stream channel. The generally linear shape is often controlled by faulting and jointing, much like any other drainage system, particularly where the geology is of resistant granitic or doleritic origin (Mackel, 1974; Whitlow, 1985). Winterberg vlei

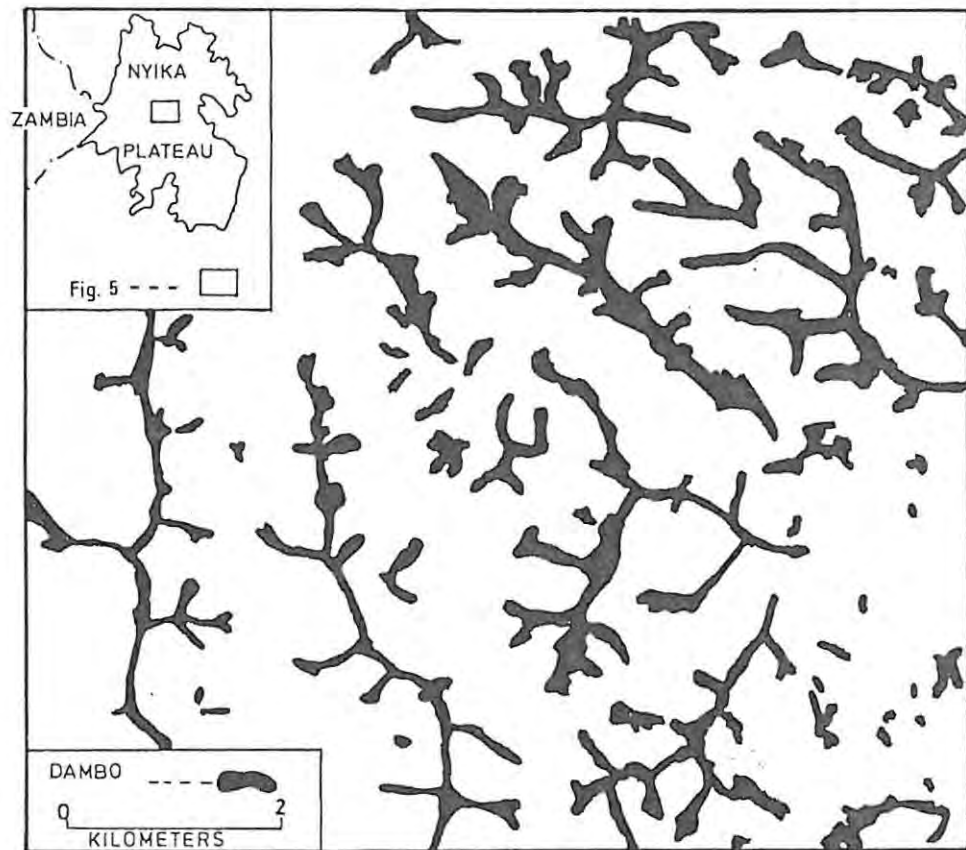


Figure 5: Morphology and density of dambos on the Nyika plateau (modified from Meadows, 1985).

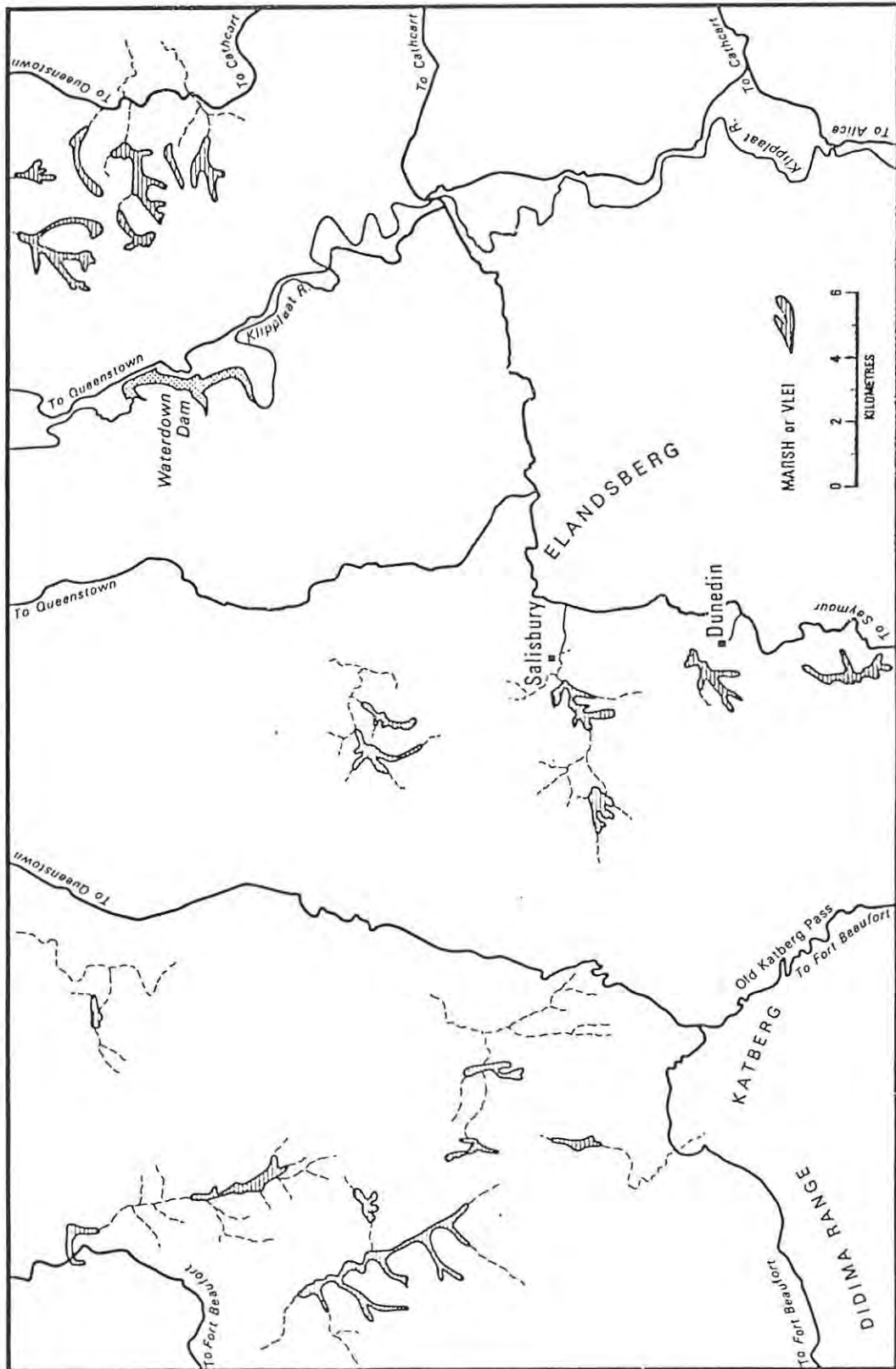


Figure 6: Morphology and density of vleis found on the Winterberg plateau, Eastern Cape Province.

morphology, in terms of linearity and club-headedness, suggests that they are geomorphologically similar features to those described in Malawi.

3.2.2.2 Profile Morphology

Dambo sediments have varying characteristics relating to depth in both longitudinal and transverse planes. Generally, the transverse profile shows increased depth towards the centre. The cross section is usually symmetrical, the axis of greatest depth, 'thalweg', usually well centred. However, this may not be universal, and can be influenced by lithology. An increase in depth away from the headwater zone of the dambo is usually apparent from a longitudinal profile.

3.2.3 DAMBO VEGETATION

Mackel (1974) describes the vegetation surrounding the low altitude dambos in Zambia as either woodland with a dense two-layer formation and closed canopy, or an open grassy woodland affected by anthropogenic factors. Both formations end abruptly at the dambo edges. Although rolling grassland vegetation is dominant on the upland areas of Malawi, montane forest does occur, but is restricted to protected valley heads (Meadows, 1982, 1985). The Winterberg vleis are similarly situated in grass/heathland plains at high altitude, with no woodland anywhere near the vleis themselves.

The dambos in Zambia, Zimbabwe and Malawi are characterised by a total absence of tree growth, and are principally occupied by grasses and sedges (Meadows, 1984). The density of cover and species richness alters with variations in hydrology, with sedges generally increasing in frequency towards the dambo centre. Seasonal changes in vegetal cover usually coincide with changes in hydrological activity; an increase in plant density is noticeable during the rainy season, during which, and for some months following it, dambos are waterlogged or flooded. With the onset of the dry season, the surfaces of the dambos may dry out to some extent (Werger and Coetzee, 1978).

These seasonal changes appear to occur at the Winterberg vleis. Drying out occurs in the Salisbury vlei, while Dunedin vlei remains waterlogged all year round due to a more extensive impeding rock layer at its base. The lack of tree growth and shrubs in dambos and vleis has been attributed to the inhibiting effects of seasonal waterlogging and occasional fires, although these tend to be more frequent on the adjacent interfluves. Restricted infiltration and internal drainage are dependent on sediment texture, soil depth and topography, all of which determine the longevity and extent of waterlogging (Mackel, 1974; Meadows, 1985).

The margin zone (Mackel's upper wash belt) of dambos described in Zambia, Zimbabwe and low-lying areas of Malawi, is usually occupied by Aristida spp., mixed with shrubs and woody species such as Protea spp. and Monocymbium spp. These species grow only where sheetwash is the dominant "fill" process, increasing the amount of sandy well-drained material. The middle zone (Mackel's lower wash belt), adjacent to the margins, where conditions are wetter, are vegetated by Andropogon spp., Sporobolus spp. and Eragrostis spp., with localised patches of Drosera spp. The richness and life span of species is higher in this zone, growth being promoted by greater water availability. This leads to larger annual increments of organic matter in wetter zones. This process forms a vital sediment source within the dambo (Meadows, 1984). The central zone (Mackel's seepage belt), waterlogged throughout the year, is inhabited by sedges of Scleria spp. and Scirpus spp., interspersed with clumps of Phragmites spp. Associated herbs are fairly common, growing amongst dense stands of Carex spp., other Cyperaceae and Juncaceae on the highly organic waterlogged soils (Thompson et al., 1983, Whitlow, 1985).

In the Winterberg vleis, the vegetation transition, in terms of both species and plant density, is not as marked as that of the dambos of Central and Southern

Africa. The species occurring in the Winterberg vleis are similar to those of Malawi and include species of the Cyperaceae, Juncaceae, Crassulaceae and Asteraceae families. In South Africa, Restionaceae are also common. All the species found in these vleis exist in dense, varied populations thriving in the waterlogged environment of these permanently saturated organic soils. This would suggest that organic matter is one of the most important sediment constituents from which to determine climatic change. A dense organic layer, developed from a thriving plant community must reflect moister conditions supporting this vegetation.

3.2.4 DAMBO SEDIMENTS: TEXTURAL ZONATION

Definite soil and sediment patterns exist in dambos as described by various workers (Mackel, 1974; Meadows, 1982, 1985; Whitlow, 1985). Mackel (1974) describes dambos in Zambia as having a distinct zonation of texture, moisture, vegetation and geomorphic processes towards the centre.

The process primarily responsible for sediment transport into dambos, and its consequent differential settling out in zones, is termed Hortonian overland flow. Shallow light grey, sandy textured marginal sediments, generally consist of coarse weathering products. Mackel terms this the "upper wash belt". The transition zone, slightly inward, is distinguished by decreasingly coarse material, increasing fine grained material and higher moisture contents. Lenses of coarser material from dambo sides still occur. This zone usually occupies the largest area of the dambo. The function of this zone is still one of inwash, so it is termed the "lower wash belt". The central part of the dambo is termed the "seepage belt", composed of a series of natural hollows containing the main body of saturation, which are linked to form a type of flow channel (Mackel, 1974; Figure 7).

Vlei sediments of the Winterberg show some zonation from the margin to the centre, but this is not as

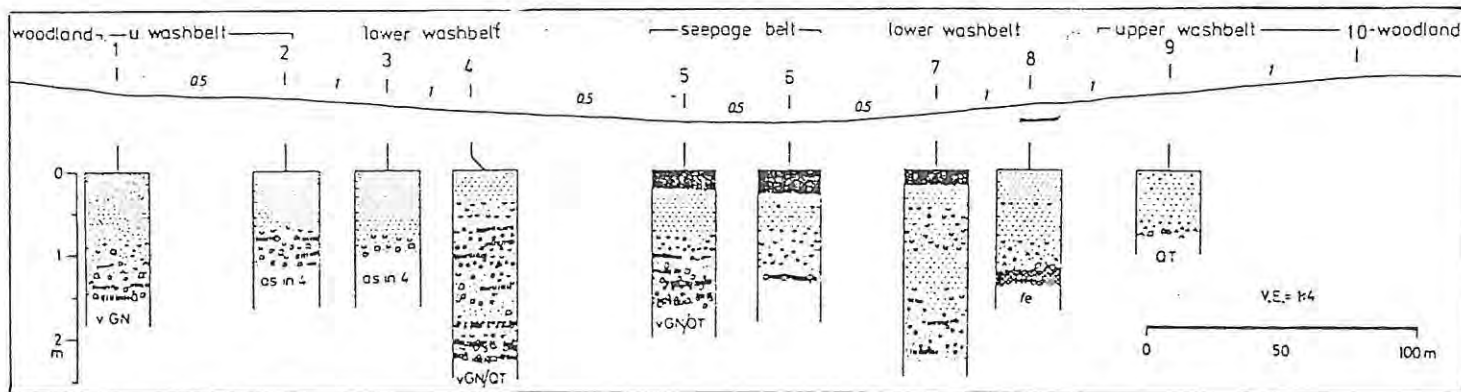


Figure 7: Cross profile of a dambo (Zambia) showing textural zones (after Mäckel, 1974).

marked or clearly defined as those described by Mackel (1974). This is possibly due to overland flow (related to precipitation characteristics of rainfall type and intensity) being different or of lesser importance than at lower altitudes. The zonation of Winterberg sediments is more related to filtering by vegetation of headwater inflow and the mechanics of hydraulic flow within the vlei. The sediments do show a fining in texture towards the centre with a concomittant increase in clay content. The "upper wash belt" along the margins is not marked. Although more sandy material is found, this is probably a result of sporadic inwash off adjacent slopes, but the lack of a distinct belt is related to the fine grained weathering products of dolerite. This marginal zone is possibly more comparable to the transition zone in Mackel's description of sediment zonation.

3.2.5 GEOMORPHIC PROCESSES

3.2.5.1 External sediment sources

A variety of sediment sources exist in and around dambos, while a hydromorphic influence exists within the dambo itself, and is closely related to vegetation. An important relationship in the availability of sediment from external sources is that which exists between climate, runoff processes and the sediment generated.

A simplistic trend that annual runoff increases as annual precipitation increases, except where temperature increases raise evaporation and evapotranspiration losses, is shown graphically by Langbein and Schuum (1958). However, a major influence in this type of relationship, excluding factors of slope, geology and many others of equally great importance, is that of vegetation. The two-phase process of erosion i.e. detachment and transport, is greatly controlled by the effectiveness of plant cover. Hortonian overland flow and rainsplash erosion are both reduced under dense plant cover which serves to

increase interception, infiltration and decrease flow velocity (Morgan, 1980). Vegetative cover is thus an important factor determining the amount and ability of runoff water to detach and transport particles. The effectiveness of vegetation in stabilising soil particles increases with an increasing percentage of ground surface cover (Elwell and Stocking, 1976; Hudson, 1981).

The relationship between erosion, transport and deposition of particles of various sizes, which is reliant on flow velocity, is demonstrated by Hjulström's curves (1935) (Figure 8). Here critical erosion velocity is shown to increase with increasing grain size. For grains smaller than 0.5mm, velocity increases with decreasing grain size due to the cohesiveness of clay minerals, causing them to behave as larger particles.

The origin of coarser grain sizes occurring in the wash belt is either directly from weathering rock below or washed in from nearby slopes. Parent material is available as colluvium, moved into the dambo system by mass movement or sheet wash, or as alluvium from the dambo stream system when in flood. Surface wash entrains particles of all size fractions from adjacent slopes, which are deposited selectively due to the reduction in carrying capacity of water with a change in gradient (Morisawa, 1968). The margins also derive material from in situ weathering rock. With water entering the vlei, coarser particles are deposited near the edge, while finer sands, silts and clays are carried towards the centre of the dambo, requiring a lower transporting velocity once entrained.

The empirical nature of many of the above relationships has been cause for much disagreement. However, it is accepted that vegetation is one of the most influential controlling factors in erosional processes. Changes in vegetation type and ground cover, corresponding to shifts in climate, are thus of fundamental importance

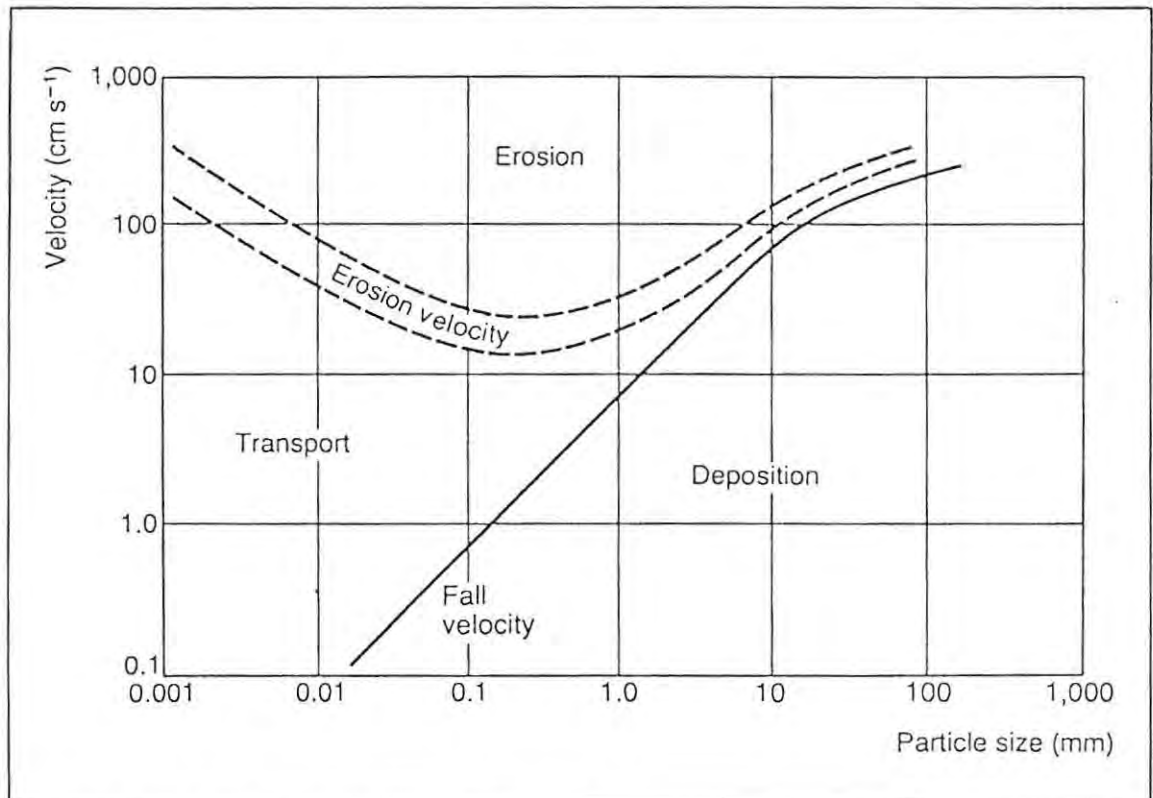


Figure 8: Critical velocities for erosion, transport and deposition of various particle sizes (after Hjulström, 1935).

in the proportion of sediment input to vleis systems on the Winterberg.

During the late Quaternary, climatic changes occurred in regions both within and outside the tropics, and evidence suggests that much of Africa was drier between 21 000 and 12 000 years BP than at present. Dambo sediments are thought to be laid down by cyclic processes dependent on the prevailing environment at the time of deposition. Meadows (1982) describes the dambos of Malawi as being occupied by colluvial and alluvial material from a series of "cut" and "fill" phases. Here it is suggested that during dry periods (around 20 000 to 12 000 BP), the percentage vegetation cover was reduced, increasing sediment removal from the upper reaches of dambo catchments due to overland flow. This resulted in a net removal of sediment from the plateau and a "cut" in the dambo itself. When moister conditions returned between 11 500 and 12 000 BP, sediment accumulation would have occurred within the valleys due to entrapment by the denser vegetation growth resulting in a "fill" process (Meadows, 1985). Further accretions of organic matter are derived from rapidly growing dense vegetation and litter falls, particularly within the last 5 000 years BP. This recent period is thought to have been moist enough to support relatively dense surface cover on the plateau, so that surface wash effects were reduced, the infill process within dambos being due to peat development and organic accretion.

This "cut and fill" process is further supported by soil erosion studies indicating that vegetation is very efficient in reducing runoff, soil loss and sediment transport downslope (Ward, 1975; Morgan, 1980). Although Winterberg vegetation during this cold arid phase may not necessarily have been completely removed, percentage cover was surely reduced, particularly with the retraction of montane forest to protected valley heads, and grassland dominating the upland slopes. Sediment yield during these phases was probably

increased from upper slopes, and excavation or a 'cut' made in the vlei sediments, with a net removal of sediment from the upland plateau to lower-lying slopes on the foothills.

When moister, warmer conditions commenced from around 12 000 BP onwards, the vleis accumulated sediment or underwent a 'fill' phase. Although there may have been reduced loads from adjacent slopes due to denser cover, vlei vegetation acted as a sieve to trap even marginal clastic inputs entering the system. From +7 to 8 000 years BP, organic accumulation took precedence through to the present day. The decrease in mineral inputs may be due to the continually ameliorating climate reducing the rate of weathering as well as better establishment of the macchia-type vegetation providing adequate soil cover.

3.2.5.2 Internal sedimentation

The effect of vegetation on vlei sedimentation does not end on the adjacent slopes, as hydromorphic species exert a direct influence on the mechanics of deposition within the vlei. The mechanics of transport and deposition are described by several laws and formulae (Morisawa, 1968; Morgan, 1980; Hudson, 1981) but essentially, particles entrained and transported by water are deposited when the current is no longer competent to carry them. Deposition occurs as a result of decreasing gradients, decreasing velocity, increasing load or where a stream reaches a lake or level plain. Particles will settle out differentially according to their density and size. Thus coarser particles settle first at the edges of the vlei, while finer grains are carried into the central parts. The effect is almost deltaic, with coarser textures near the margins, grading into fine sands, being overlain by silts and clays. The centre contains very fine colloidal material including organic matter.

The sorting of particles downstream within vleis occurs as a result of three effects: (a) vegetation increases

in density away from margin and headwater zones and reduces flow; (b) a concurrent loss of velocity and volume occurs due to stream outflow and channel widening downstream; and (c) the vlei is slowly aggrading and steepening its gradient with gradual inputs where headwater streams occur. As a result of these effects, sediment sorting takes place within the Winterberg vleis. Vlei vegetation plays a role in the latter two instances and thus appears to be the major controlling factor in sediment gradation within the vlei. This effect occurs in addition to the even greater effect that vegetation plays in the accumulation of organic sediment which forms a large proportion of the total sediment.

Sediment texture appears to vary with underlying or neighbouring parent material e.g. clay soils on doleritic or limestone material and sandy soils on quartzite, sandstone or granite. Changes in textural classes also occur with increasing depth and away from the dambo margins. The clay/silt component generally increases with increasing depth and laterally towards the centre due to the filtering effect of vegetation and water movement through the profile. The geomorphic agent for the transport and accumulation of clays in the lower centre is seepage water or, dambo wash. The degree and duration of waterlogging is a critical factor in fine particle migration within the dambo sediments (Mackel, 1974).

Clays also develop in situ, secondary minerals being formed and translocated by subsurface transport. Their vertical movement is traceable by cutans on peds and accumulations on organic material lower down in the profile. Clay translocation involves the removal of dispersed clays by water moving vertically and laterally through a coarser grained matrix. Active throughflow and dispersed clays are both necessary for this process, which has only recently been described in vleis in Zimbabwe (Whitlow, 1985). Occasionally, a layer of loose fine sands may be washed into the centre

after a torrential storm, forming a pale lens over the clays and organic matter.

The creation of organic sediments and peats arises from the accumulation of organic debris under suitably waterlogged conditions. The highly favourable environment encourages the growth of hydromorphic species which, generation after generation, thrive, die and subside into the water in which they grew. Decay is achieved largely through anaerobic bacteria, and results in a brown or black organic material. The deposition of successive layers of this organic debris as time passes, forms a profile of strata displaying different characteristics, depending on the type of plant, degree of decomposition (Tivy, 1982) and the possible addition of mineral sediments.

Many peat deposits are autochthonous, formed almost entirely from land plants, or, in situ material which grows and accumulates under suitable conditions. Sediments may also be allochthonous in nature, derived from areas of the catchment and carried to the place of deposition in valley bottoms. This sediment differs not only in the organic components, but also in the proportions of the inorganic components, when compared with autochthonous consisting primarily of plant remains. Peat-like materials have high water-holding capacity and are almost invariably acidic. The progressive development of these organic materials is reliant on a stable watertable and constant environmental conditions. If there are radical changes in the watertable e.g. flooding, a shift in climate to reduce rainfall, or conditions unfavourable for plant growth, accumulation of debris and thus the succession, ceases.

The chemical characteristics of these sediments are important in that the surface area and colloidal nature is far higher than that of mineral soils, resulting in a far higher fertility or cation exchange capacity. These sediments are also generally high in nitrogen and

calcium contents, both in forms available for exchange. Phosphorus and potassium contents are generally low and magnesium content similar to that of a mineral soil (Buckman and Brady, 1969). These characteristics make these sediments valuable as agricultural resources, particularly in an economic climate that constantly strives for higher production. For this reason, many of these fragile areas are being drained and used for cultivation, with little appreciation of the loss of a valuable and complex ecosystem.

3.2.5.3 Sediment stratigraphy

Most dambos, whether developed on coarse-grained or fine-grained parent material, have an organic or peat-like top layer, which appears to increase in depth with increasing altitude and/or precipitation. The organic content of this layer may, in some cases, not rise above 10% due to more rapid decomposition under aerated conditions during the dry season (Meadows, 1985). The high watertable found in most dambos, however, results in the accumulation of organic matter and gleyed, even vertisolic sediments towards the dambo centre. Dambos also display a variety of transitional soil types with mottling, plinthic and ferricrete layers formed by internal seepage. These characteristics usually increase with depth to the base (Mackel, 1974; Young, 1976; Millington *et al.*, 1985).

Within the the organic surface horizons which generally become deeper from the shallow margins to the wetter centre, the organic material is well decomposed on drier hummocks, but more fibrous and peaty in saturated hollows. The accumulation of less decomposed material is due to limited rates of microbial activity caused by acidic waterlogged conditions (Meadows, 1985; Whitlow, 1985). True peats are rare in Southern Africa, and even organic soils are uncommon, according to Thompson *et al.* (1983). Deposits described as peats (Mackel 1974; Meadows, 1982, 1985; Whitlow, 1985) in Southern and Central Africa, have organic carbon percentages far lower than peats described elsewhere (Gorham, 1953;

Chapman, 1964; Thompson et al., 1983). True organic sediments are classified as having organic carbon percentages as high as 80%.

Peat forming processes are critical to the understanding of dambo stratigraphy and have important bearing on the interpretation of environmental changes. The actual growth of bog areas in temperate latitudes has been described by Barber (1981) who proposed the Phasic Theory of bog growth. Although developed for temperate peat bogs, the theory may prove to be more widely applicable. This theory describes the upward growth of bogs in a series of "hummock and pool" features, the overall impression being one in which the whole bog responds, in essence, to variations in surface moisture, although topography is also important. The rate of peat growth is thus faster during moister climatic periods and slower during periods of aridity. In this way, the organic content of the sediments may be used as a sensitive index of prevailing and changing climates, the rate of peat growth, and therefore organic content, being dependent on the availability of moisture.

Dambo sediments are thus usually composed of a sequence of both clastic and organic deposits. However, the major process for rapid sediment accumulation in these saturated areas remains one of organic or 'peat' accumulation and decomposition (Meadows, 1982). The complexity of the resultant sediment stratigraphy is thus a function of all the geomorphological processes responsible for dambo infill and development. Although changes within the stratigraphy are many and appear to be fairly marked, these may be only minor adjustments to much larger changes in the external environment. It has been emphasised by various workers (Schumm, 1965, 1969; Stocking and Elwell, 1976; Hudson, 1981) that a great reduction in precipitation results in only a slight lowering of annual runoff losses and therefore sediment yields. Similarly, the type of rainfall and intensity will greatly influence the generation of

sediment (Whitlow, 1985), and, as emphasised by Hudson (1981), the nature of the precipitation is one of the most influential factors, rather than rainfall amount per se.

3.2.6 DAMBO HYDROLOGY

Some hydrological research, mainly of agricultural interest, has been devoted to dambo systems (Balek and Perry, 1973; Millington et al., 1985), although this has been minimal due to increasing protection of these areas by legislation (Whitlow, 1985). Restrictions on dambo utilization in terms of agricultural landuse were enforced in Zimbabwe after inappropriate methods caused increasing silting in reservoirs, and erosion of these sensitive geomorphic features. These developments caused concern that dambo "sponges" may dry out and cause perennial streams to become seasonal by nature (Bell and Roberts, 1986). Vlei destruction of a serious nature has already occurred in South Africa (Whitlow, 1985). Walmsley (1987, p.265) points out that "throughout the world, vast areas of wetland have been converted to farmland. The situation appears to be no different in South Africa, an arid country, which can ill-afford to lose its precious water resources in an almost irreversible fashion."

Dambo formation is favoured by the occurrence of sandy soils on adjacent slopes, gently undulating terrain, and shallow bedrock in valley bottoms. High infiltration capacities of sandy soils promote rapid absorption and throughflow on slopes, but impermeable substrates impede drainage and favour waterlogging in valleys (Whitlow, 1985). These factors allow for the formation of dambos under lower rainfall conditions than designated by Bond (1967).

The water balance in dambos is maintained primarily by inputs of subsurface drainage from adjacent slopes (Millington et al., 1985), although suggestions that direct precipitation is the major form of input have been made from studies in Zambia (Balek and Perry,

1973). Mackel (1974) suggests that surface wash (overland flow) and throughflow of groundwater from adjacent interfluves are the greatest suppliers of both moisture and sediment to the dambo.

Meadows (1982) acknowledges that surface wash could be an important source of input to the dambo system, but states that this would only be the case if conditions of aridity and sparse vegetative cover prevailed. The erosive potential of hillslope runoff is greatly reduced by dense vegetation, while infiltration time is increased, thus aiding the reduction of both runoff velocity and amount (Morgan, 1980). However, Meadows' studies (1985) in Central Africa, show that under contemporary conditions (the last 2 000 years at least), it would appear that surface wash is relatively unimportant as a sediment source, although the runoff factor would still be the main source of moisture input. Meadows (1982, p.131) writes: "Surface wash, then, would appear to be currently unimportant in terms of dambo sedimentation although it could have been more effective in times of slightly drier climate and reduced vegetation cover and the current process responsible for dambo infill is peat formation."

3.3 THE GEOMORPHOLOGICAL SIGNIFICANCE OF DAMBOS

The interpretation of chemical and physical changes throughout dambo profiles, remains complex and problematical. If these characteristics reflect changing vegetation patterns, changing hydrological conditions and therefore changes in the prevailing climate, dambo sediments can be considered sensitive environmental indices. Radiocarbon dating techniques on dambo sediments elucidate the chronology of past conditions, and dates have been obtained from Malawi (Meadows, 1985), several sites in South Africa (Meadows, 1988) and Zimbabwe (Tomlinson, 1973). Although there are a number of problems with the interpretation, radiocarbon dates do assist in accurately dating basal and other organic samples.

These may provide a time-span of sediment accumulation, possibly identifying periods of increased moisture or aridity.

In comparison, dambo and vlei features (linear, waterlogged and club-headed with no marked stream channel) appear to be consistent throughout Central, Eastern and Southern Africa, with only local variations in stratigraphy and morphology. Their sediments provide information on recent and historical geomorphic processes from which the relation between similar sites can be established. Radiocarbon dates from dambos throughout Southern Africa do imply some sort of synchronism, suggesting that dambos have responded to broad environmental changes. In terms of chronological identification and sedimentological analyses, these features should prove useful and responsive indicators of environmental change.

Research on vlei systems has, thus far, taken place randomly, without any defined or integrated framework. Descriptions of these geomorphological phenomena suggest that they are complex systems and that processes acting on and within them are, so far, poorly understood. With the compilation of data and standardization of ideas, it becomes clear that more work is needed on the hydrology, sedimentology and geomorphology of these systems, so as to minimize degradation and even the ultimate destruction of these fragile phenomena.

Human populations have become increasingly involved in both the development and destruction of dambo areas. Many of these areas are being subjected to arable landuses in several countries for at least part of the year (Mackel, 1974; Meadows, 1985). As dambo soils are fertile but fragile, they are extremely vulnerable to degradation under cultivation. Loss of and damage to these hydromorphic sediments results in the loss of the 'sponge' effect or water-holding qualities of the sediments. This reduces the slow release of water as

base-flow from the vleis during dry seasons, and increases runoff in the wet season (Whitlow, 1985; Breen and Begg, 1987). Due to serious soil losses, cultivation of these areas has been severely restricted by legislation in vlei areas of Zimbabwe and South Africa (Bell and Roberts, 1986).

CHAPTER 4

VLEI ENVIRONMENTS ON THE WINTERBERG

4.1 INTRODUCTION

Palaeoenvironmental reconstruction of the Holocene period within South Africa has been severely hampered for various reasons. The restricted number of suitable study sites as well as the lack of full palynological and chronological analyses at available sites has led to a scarcity of data with respect to other areas within Southern Africa. Evidence is particularly scarce for the Eastern Cape. Sediments which are peat-like in nature and suitable for pollen preservation are unusual within this area where contemporary climatic conditions are generally too arid for extensive accumulation of organic sediments.

Some of the most productive sites to date in South Africa have been rare swampy areas, like permanent springs with organic deposits (Scott, 1984). Such peaty sediments have also been found in various vleis which exist on the humid tablelands of the Winterberg Range, bordering the more arid Karoo region. The vleis waterlogged condition, and the development of a stratigraphic profile, provides an opportunity to search and analyse for evidence of climatic change. Sediments, by their reliance on contemporary environmental controls of erosion and deposition, are useful indices of a change in any of these controls. Within these sediments, preserved pollen provides an additional source of evidence, and this aspect of study has received much attention.

The general characteristics of Winterberg vleis appear to correspond closely to similar geomorphological features described on Nyika Plateau, Malawi (Meadows, 1982, 1984; 1985) and Inyanga, Zimbabwe (Bond, 1964,

1965, 1967; Whitlow, 1985). These vleis provide a valuable opportunity to study a palaeodepositional sedimentary sequence in South Africa. Radiocarbon dates obtained from the dambos of the Nyika Plateau indicated that the sediments within them represented a depositional sequence of about 10 000 years (Meadows, 1982). A radiocarbon date of $\pm 12\ 600$ BP, fixed on a cored sample from the Dunedin vlei on the Winterberg after a pilot study in March 1985, indicated that these sediments span the late Pleistocene/Holocene periods. This justifies a suitable interest in the sedimentology of these vleis for extensive analysis of morphology, sediment characteristics and present and past processes of deposition under changing environmental conditions.

An added advantage of this study site is that an opportunity for comparison exists as two vleis are situated near one another, and are therefore subjected to broadly similar climatic influences. This enables a sedimentary history to be drawn up from the two sites within the same catchment, providing substantiated evidence for environmental fluctuations (if any) within the area. If the sediments of both vleis are similar, this indicates that the environmental change influencing the sediments was relatively widespread in that it affected the entire catchment area. If not, local variations in topography, geology or vegetation may have caused disparity between the two vleis. The use of both vleis also enables wider comparison with other areas to establish whether the Winterberg vleis respond like the dambos described elsewhere, further evidence for environmental change of a more regional nature.

The two Winterberg sites, Dunedin and Salisbury, are large, branched waterlogged vleis (Plates 1, 2 and 3) similar in morphology to those described on Nyika Plateau and Inyanga, but do not conform with those of Central Malawi, Zambia and low-altitude Zimbabwe. The Winterberg vleis exhibit extensive deposits of peat-like sediments several metres deep. It was thought that analysis of these sediments would enable

the establishment of a chronological sedimentary sequence, providing information on the late Pleistocene and Holocene environments of the area. Information of this type may improve the understanding of contemporary environments of such geomorphological phenomena as well as late Pleistocene climatic and environmental changes for the region. Data can be correlated with research results from other areas, and extrapolated to provide a more extensive view of man-environment relationships through time in the Eastern Cape. The dambos of the Nyika Plateau (Malawi), in particular, have been chosen for comparison, due to the wider availability of information relating to these features, and their similarities in sediment type and morphology.

4.2 LOCATION OF THE STUDY SITE

The Winterberg highlands are composed of a series of resistant peaks, the Elandsberg, Hogsback, Katberg and Gaika's Kop, forming a boundary between the inland plateau and the low-lying coastal areas of the Eastern Cape coast. Several likely sites for peat accumulation exist on the Winterberg Range at high altitude, where suitable climatic conditions maintain the moist nature of large vlei areas. The study area encompasses two sites forming part of a system of vleis on the summit of the Elandsberg, within the borders of the Ciskei homeland (Figure 9). These vleis occur on two farms, Dunedin and Salisbury, and for ease of reference, have been so named. The grid reference for Dunedin farm is $26^{\circ} 46' 05''$ east and $32^{\circ} 27' 04''$ south, while the Salisbury farm location is $26^{\circ} 45' 05''$ east and $32^{\circ} 25'$ south on the 1:50 000 topographic map 3226 BD. The position of the two vleis is shown in Figure 10.

4.3 CLIMATE

The study site lies within the Warm Temperate zone between 32° and 33° south, and is classified by Schulze (1965) as being intermediate to the D (Drakensberg) and Ss (Southern Steppe) regions, the upland influence

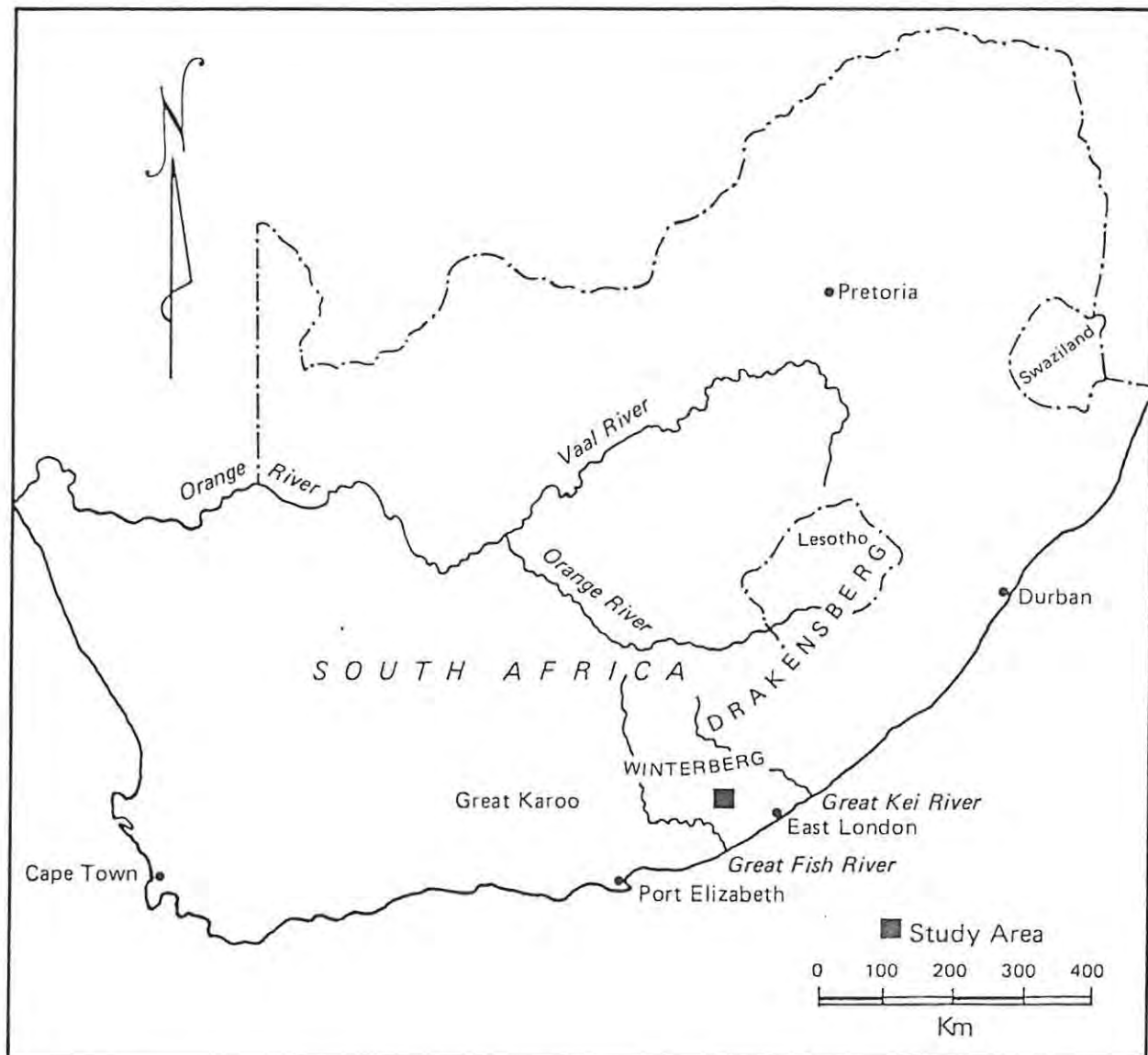


Figure 9: Location of Winterberg study site.

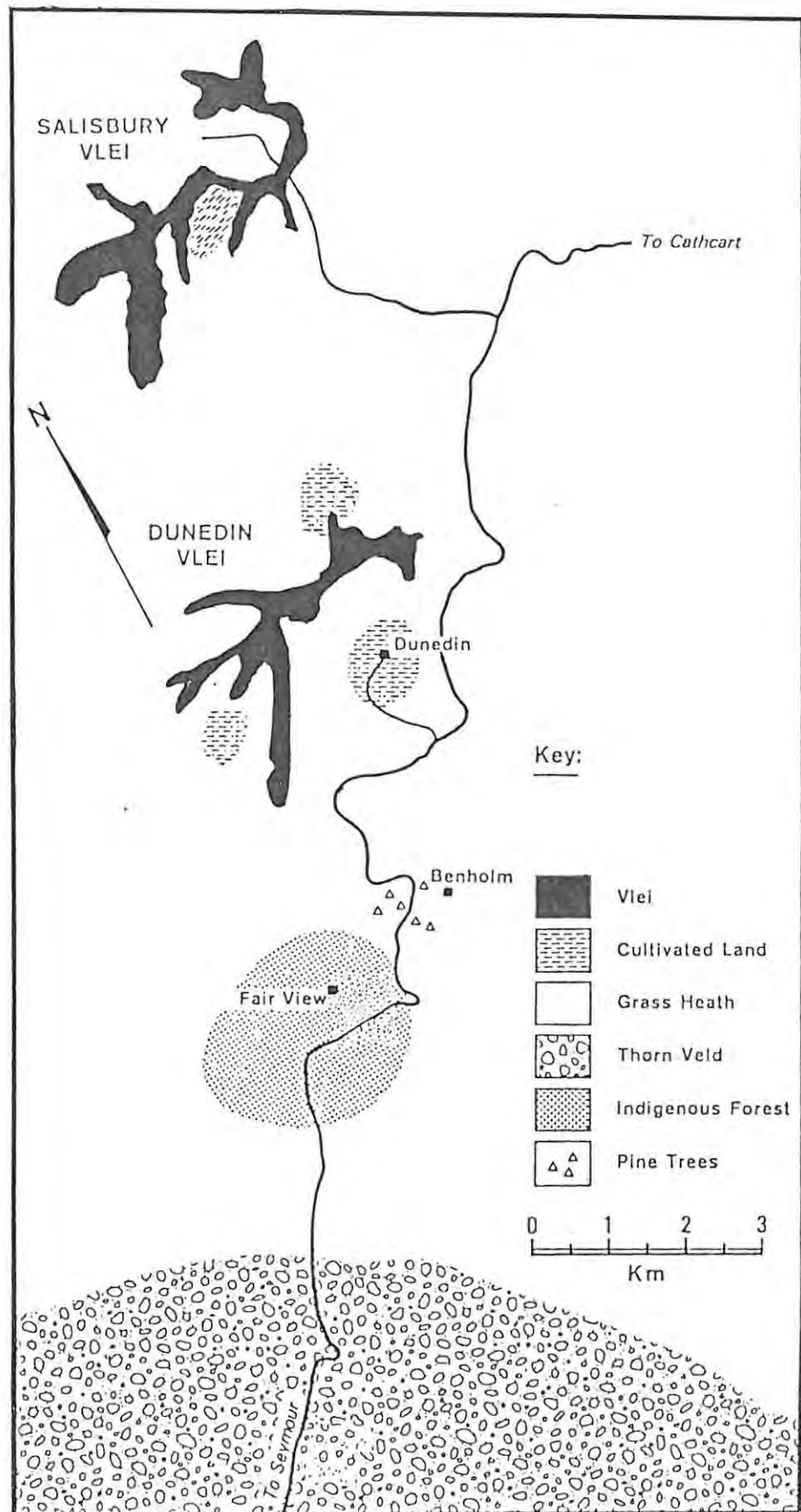


Figure 10: Morphology of Dunedin and Salisbury vleis.



Plate 1. Longitudinal view of Dunedin vleei (dry season).



Plate 2. View of a branch of Dunedin vleei (wet season).



Plate 3. Longitudinal view of Salisbury vlei.

being exerted by the southern tail of the escarpment. There is some difficulty in classifying the climate as it appears to be transitional in this area, the site falling on the boundary of two climatic zones. This makes the area an excellent site for examining environmental change, in that, a change in climate would result in the movement of these boundaries latitudinally and longitudinally, and the site would then be influenced mostly by one set of climatic conditions. For example, if the site fell entirely into the D region, frost, large diurnal and seasonal temperature fluctuations and orographic rains would have a vastly different effect to summer rains, a lengthy dry season and periodic fires of the Ss region. The effects on vegetation, and thus sediment would be marked, both being dependent on climatic conditions existing at that time. Fluctuations of the climatic boundaries, whether a shift in the whole belt or simply an expansion or contraction of either region, would have quantitative effects on the vlei surrounds.

4.3.1 RAINFALL

The study site enjoys a summer rainfall maximum with totals of 1050mm given for Katberg at 1051,5m, rising sharply to over 1600mm with further increases in altitude (Childs, 1971). More humid conditions prevail over the easternmost mountains due to the influence of the warm Indian Ocean currents. Orographic precipitation dominates, although convection storms do build up in hot humid summers. Mist and drizzle are also frequent in summer. Higher rainfall increments are important sources of input for youthful streams flowing southward which etch deep-sided valleys into the mountainsides. Frost is frequent in winter between April and September and snow may fall on the peaks above 900m. Childs (1971) maintains that a decrease in the total precipitation towards the summits results in a "thinning" effect on the vegetation, such that montane forest gives way to tussocky grassland and macchia species. However, Meadows (1985) suggests that precipitation continues to increase with altitude, and

that it is the exposure to extreme conditions which is greatest on the summits: hence the tussock grassland.

The importance of rainfall in terms of sediment input to the vlei is very great, as water is the main transporting agent. The intensity and amount of rainfall affects the amount of sediment carried into the vlei from nearby slopes and headwater areas. High intensity storms will provide for large inputs of mixed clays and medium/coarse sands, whereas a gentle rain will provide fine silts and clays by throughflow, leaching and lower energy runoff. Rainfall will have an additional influence on sediment availability by its interaction with vegetation. The availability of water for growth will determine the density of ground cover and therefore the surface area susceptible to erosion. For example: dry lower-lying slopes are dotted with Acacia karoo and Aloe spp., interspersed with grasses which provide only seasonal cover for the soil surface, while higher-lying valley heads support dense montane forest due to a higher moisture regime and protection from drying berg winds (compare Plates 18 and 20).

4.3.2 TEMPERATURE

Due to the position of the site deep in the interior, and its extreme elevation above sea level (in excess of 1800m), temperatures show great fluctuations both seasonally and diurnally. Temperatures vary from a mean monthly maximum of 30°C in summer and a mean monthly minimum of 0°C or below in winter (Schulze, 1965). As various species are limited in their tolerance of temperature extremes, these oscillations with season have a significant effect on the species richness and plant density, especially the occurrence of frost. The temperature tolerance of various species leads to changes in plant populations within the area and therefore affects sediment availability on adjacent slopes. These vegetation changes alter the sediment supply, which are reflected in the stratigraphy of the vleis, and differences in eg. organic content may reflect changes in temperature.

Great diurnal temperature changes induce mechanical weathering of doleritic rocks on nearby slopes, also increasing sediment availability for transport. Doleritic rocks are particularly susceptible to freeze/thaw action resulting in spheroidal weathering and columnar fracturing of large exposed domes (Plate 4) (Johnson, 1966).

4.3.3 WIND

South-westerly winds accompany half the annual precipitation in the area, resulting in sharp drops in temperature with rain and snow. Berg winds bring high temperatures and low humidities, accelerating evaporation from the thin soils developed on steep slopes, dominating in winter months (Childs, 1971).

4.3.4 SUMMARY

The climatic conditions existing on the Winterberg are conducive to a range of weathering processes which generate sediment for vleis input. Such processes are influenced by the extreme conditions of temperature, inducing mechanical weathering of parent material as well as the effects of rainfall resulting in chemical breakdown.

All climatic factors can be directly related to vegetation cover which undoubtedly reduces erosion (Hudson, 1981) and dissipates the energy of raindrop splash as well as that of running water (Ward, 1975; Morgan, 1980). Conditions important in determining sediment input in vleis might be attributed to rainfall type and frequency. For example: high intensity storms encourage the shifting of erodible sediment from the sideslopes into the vleis, soils being particularly susceptible after natural fires (Plate 5). Greater moisture and humidity would induce more rapid chemical activity in soils and parent material. For peat-like sediments to develop, conditions must be cool and moist, where local topography results in poor drainage. Thus, various climatic factors are interrelated to form a single complex process which affects the rate and



Plate 4. Columnar fracturing of dolerite. Scale 1:60



Plate 5. Signs of erosion from adjacent slopes after a fire. Scale 1:5

type of weathering and its related products and induces more rapid accumulation of organic material within the vleis itself.

4.4 GEOLOGY

The parent material underlying a given soil has one of the most important influences on the type of soil developing from it. Du Toit (1954, p.476) comments: "the geological formation merely provides the raw material from which the soil becomes evolved under the superimposed influences of position, climate and organic life". Parent material affects the texture, colour and structure of the soil as well as the chemical and physical properties of that soil. Parent material not only provides the basis for soil formation but will also influence the inherent properties of that soil.

The entire region surrounding the Great Winterberg Range is composed of sedimentary rocks of the Karoo System, dissected throughout by doleritic intrusions. The area lies on the south-east margin of the great Karoo basin, confining the geology to the Ecca and Beaufort series of late Palaeozoic and early Mesozoic age (Childs, 1971; Johnson, 1976). The Ecca series is of little importance in the study area, the major geological influence being formed by the more resistant Beaufort Series.

4.4.1 BEAUFORT SERIES

This series is divided into three zones which have not been mapped in detail in the Winterberg area, although a description has been given by Mountain (1956) and Johnson (1966). The Lower Beaufort (L.B.) group forms the area below 1615m and is composed of alternating shales, mudstones and sandstones (Johnson, 1966). Haughton (1928), Mountain (1946) and others used the first appearance of mudstone as a basis for defining the Ecca/Beaufort contact. The shales of the L.B., due to their splintery fracture, are usually reduced to

irregular fragments and the sandstone decomposed by spheroidal weathering (Childs, 1971).

The Middle Beaufort is distinguished by horizons of coarse, pale blue sandstone with maroon mudstones and shales above 1620m on the Winterberg (Mountain, 1946; Marais *et al.*, 1962). Du Toit (1954) attributes the scarps and plateaux in the area to thick beds of felspathic sandstone, becoming increasingly prominent up the range. This group is also known as the Katberg sandstone, the name derived by Johnson (1976) from the Katberg Pass in the Winterberg mountains. Beyond 1680m, sandstones are intruded by doleritic dykes and sills of late Jurassic age, giving rise to resistant hills and mountains (Haughton, 1970; Childs, 1971).

The Upper Beaufort appears above approximately 1920m and is of very limited extent (Haughton, 1970). The only area exceeding this altitude (excluding the Elandsberg summit) is in the immediate vicinity of the Great Winterberg (Childs, 1971).

4.4.2 DOLERITE INTRUSIONS

As early as 1905, Rogers made the observation that all ranges in the Eastern Province owed their existence to the presence of thick sheets of dolerite that protected the underlying sedimentary rocks from destruction. The highest points are invariably composed of dolerite. The columnar structure of these basic crystalline rocks can be seen capping the summits of the Elandsberg, Hogsback, Katberg and Winterberg, and are thought to be remnants of once simultaneously intruded, or continuous sheets of dolerite now dissected by erosion (Du Toit, 1905, 1917; Mountain, 1956).

The doleritic cappings are characterised by huge domes and boulders, with a cross-jointing weathering system, which break down to form fertile black or dark reddish brown clay soils (Childs, 1971; Johnson, 1976). The Elandsberg plateau has a capping sill of dolerite several hundreds of feet thick maintaining a maximum

elevation of 2018.9m (Plate 6). This altitude is only exceeded by the Great Winterberg mesa which reaches 2371m (Childs, 1971).

The underlying geology of the two vleis consists of thick doleritic sills which have eroded through a process of cross-jointing to give rounded boulders. These boulders lie scattered about the slopes and ridges above the vleis and by weathering and erosion have given rise to dark-brown and red loams overlying heavier blocky clays (Plate 7). At Salisbury vlei, one slope adjacent to the vlei has been cleared of boulders and used for agricultural cultivation. However, remnants of dolerite boulders are found on the other surrounding slopes and are exposed on the streambed at the bottom end of the vlei. The Dunedin vlei is in a year-round waterlogged condition, maintained primarily by the remnants of a dolerite dyke which extends across the stream channel leaving the vlei (Plate 8). This dyke retards flow, causing moisture to back up throughout the area, raising the water table.

The surrounding geology is of great importance as this forms the major source of material for sedimentation processes in the vlei. The rate of weathering is controlled by climate, topography and biological influences, and indirectly determines the availability of sediment for transport and deposition by the agents of erosion. The weathering of dolerite generally produces fine-grained particles, mostly of the fine sand and clay size fraction (Buckman and Brady, 1968; Childs, 1971).

4.4.3 ALLUVIUM AND OTHER SUPERFICIAL DEPOSITS

Due to the high altitude and steep slopes of the Winterberg peaks, streams in this area create high-energy environments in which erosion, rather than deposition, is the dominant process. However, where relief is undulating, as on the upper shelves of these tablelands, drainage may be impeded, promoting the development of waterlogged vleis.



Plate 6. Dolerite outcrops protecting higher peaks.



Plate 7. Dark red loams. Note wash (below profile) after rain. Scale 1:25



Plate 8. Remnant dolerite dyke and scattered boulders on adjacent slopes.



Plate 9. Agricultural land irrigated from vlei seepage.

Dunedin consists of three branches which join downstream where drainage is impeded by the dolerite dyke. Drainage within this vlei has been altered somewhat by man in that several shallow drains have been made to encourage flow downstream for use on agricultural lands below (Plate 9). Continuous base flow resulting from the sponge-like effect of these organic sediments, provides a perennial source of water to areas downstream (Plates 10, 11). If drainage is indiscriminate and uncontrolled, vlei surfaces may dry out and the water-holding capacity be reduced. With continuous demands such as these, the entire system could be destroyed, and the danger of turning the perennial supply into an intermittent one becomes very real. This form of disturbance has raised a storm of controversy elsewhere, and legislation has been passed in Zimbabwe to prevent further damage to these systems (Whitlow, 1985). The morphology of Salisbury vlei is less branched and less waterlogged due to a more freely draining situation.

Within both vleis, organic and clastic sediments have accumulated through various drainage processes such as inwash from headwater streams and slope sheetwash, as well as accumulating organic material from the dense hydromorphic species growing in the vleis themselves. These layers of sediment form a chronological record and so provide evidence of past environmental conditions.

4.5 GEOMORPHOLOGY

The Great Winterberg mesa extends some 48km in an east-west direction. This residual surface can be traced by the crests of the Katberg (1831m), the Elandsberg tableland (2019m), Gaikaskop monadnock (1963m) and the three Hogsback ridges (1938m) as shown by Plate 6 and Figure 11 (Haughton, 1970). These remnant summits were first believed to form part of L.C. King's "Gondwana cyclic surface" (Agnew, 1958), but King (1962) regards the Gondwana surface as being



Plate 10. Surface flow — lines converge at dyke.



Plate 11. Shallow man-made drains ensure perennial flow.

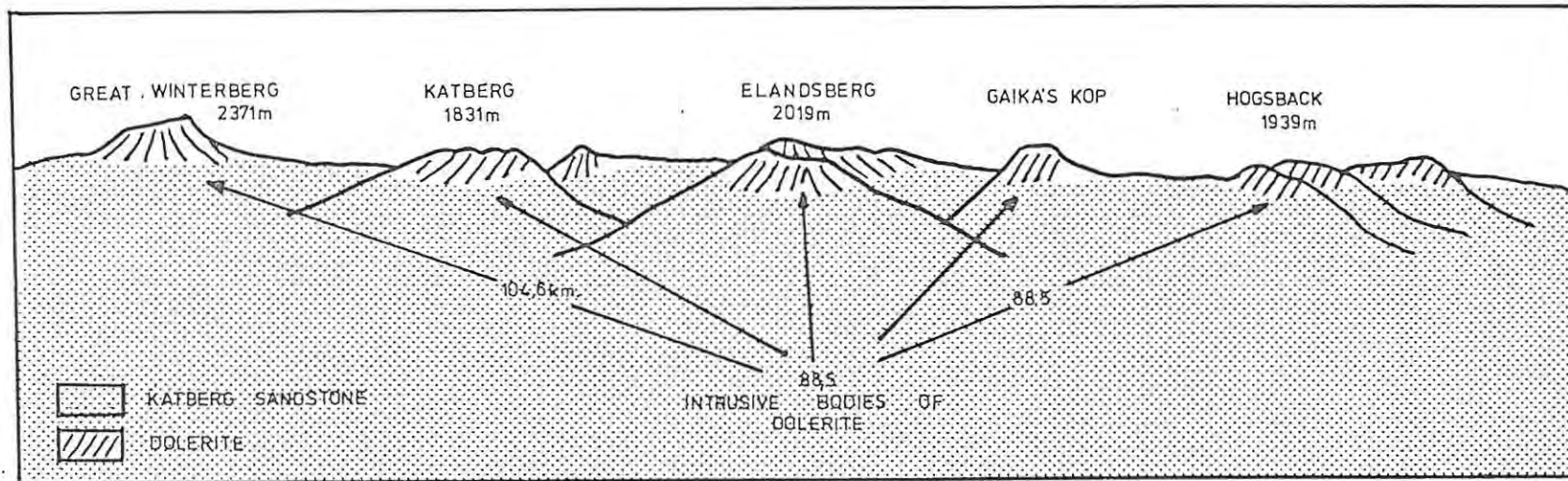


Figure 11: Residual surface of the Great Winterberg mesa.

far more limited in extent. Evidence from the Transvaal and Natal highland surfaces of 1830m have been dated as early Tertiary or "African" in age, and tend rather to classify the Amatola-Winterberg summits as the remains of this African surface (King and King, 1959). The lower surface (1220 to 1525m) is apparently post-African or Quaternary in age (King, 1963). The Dunedin and Salisbury vleis are thus resting upon remnants of the African-dated highlands.

The rolling hills on the summit of the Winterberg range which extends inland to the west, is a region of very diversified relief (Plate 12). These highlands are considerably dissected by north/south valleys, separated by grass-covered spurs or interfluves. This high-lying surface owes its preservation to a dolerite sill and the absence of a summit remnant on the eastern slopes of the Katberg is due to the absence of such a doleritic intrusion (Childs, 1971).

4.6 SOILS

Soils are a product of their environment, arising from the interaction of several processes acting upon them over time. It is the element of time with which this study is most concerned. As soils are formed by various cycles of erosion and deposition, their profiles take on various characteristics which can be analysed to provide information on the environment at the time of soil formation. In this way, a soil profile can, by its stratigraphy, supply valuable data on past environmental conditions.

4.6.1 SOILS OF THE SURROUNDING AREA

The soils found on the steep slopes of the Winterberg are skeletal due to vigorous sheet erosion and are closely related to the sedimentary parent rock of the Beaufort series or dolerite. On sedimentary parent rock, the relatively high rainfall results in heavy leaching of minerals, and thus these soils are generally low in nutrient status and highly erodible.



Plate 12. Rolling topography characteristic of Beaufort sediments.

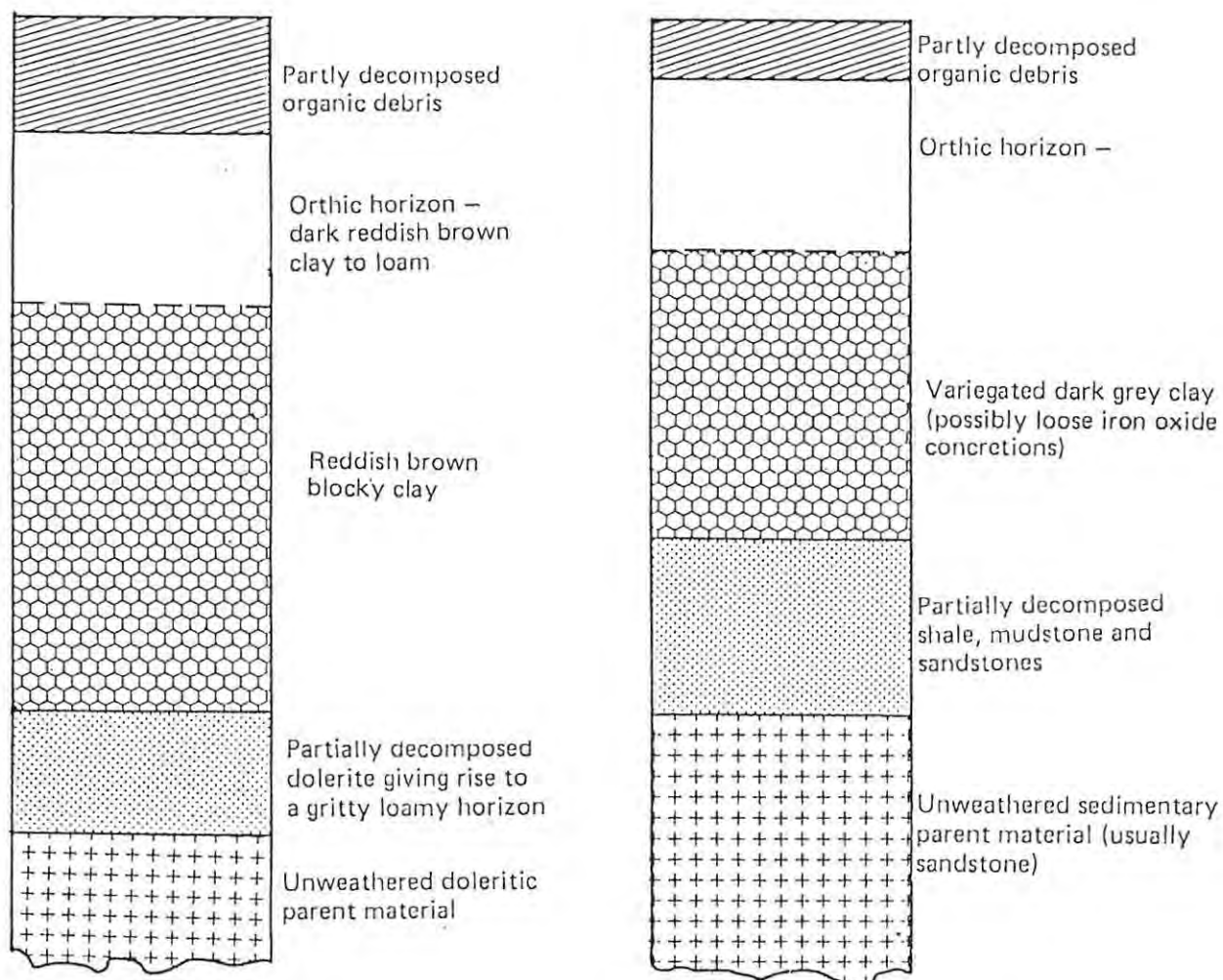
A diagrammatic soil profile is given by Figure 12 and typical soil forms are those of the Mispah, Swartland (Plate 13) and Valsrivier forms (MacVicar *et al.*, 1977). The profile exhibits a shallow erodible A horizon which is often removed by sheet erosion occurring particularly after burning or overgrazing has reduced the vegetation cover, thus exposing the B horizon in which dongas are likely to develop. These soils are acidic with pH values ranging from 3.8 to 5.3 at 1220m on the Katberg (Childs, 1971).

Where dolerite is the underlying rock type, deep uniform red or black clays develop (Plate 14). These soils exhibit a typical profile as shown in Figure 12 and Plate 15. This loamy soil type tends to retain more moisture than the shallower profiles of soils developing on sedimentary rocks, and small pockets of forest are frequently associated with soils developed on doleritic parent material. Due to the high mineral content and basic composition of dolerite, these soils are of superior fertility to those formed on the Beaufort series and representative soil forms are those of the Bonheim and Shortlands forms (Plate 16). The pH values of Katberg soil samples were found to range from 4.7 to 6.2 (Childs, 1971).

4.6.2 SEDIMENTS WITHIN VLEI AREAS

Both the Dunedin and Salisbury vlei sediments appear to be similar in textural and colour characteristics. Much of the doleritic parent material available for deposition is colluvium derived from weathered bedrock on adjacent slopes, carried in by sheetwash. Some of the sediment in the vlei is composed of alluvium which enters the system at the headwaters by streams in flood. The black or dark-brown colour of the sediment is mainly due to the decomposition of organic matter under waterlogged reducing conditions within the profile.

The Dunedin vlei sediments, waterlogged year-round which influences pedogenesis, are hydromorphic clays,



Soil profile of a deep uniform fertile red clay developed on doleritic sills and dykes.

eg. Shortlands

Soil profile of skeletal podzolic soils on steep slopes or sedimentary strata.

eg. Swartland

Figure 12: Typical profile diagrams of Winterberg soils.



Plate 13. Swartland form soil on lower slopes. Note grey erodible C horizon. Scale 1:50



Plate 14. Dolerite decomposing into dark red loams. Note soil tongues in saprolite. Scale 1:40



Plate 15. Blocky structure of Shortlands form soil.



Plate 16. Shortlands form soil profile. Note deep rooting depth.

relatively high in organic matter. Organic matter accumulates in the topsoil, and limited gleying occurs in the lower sediment layers due to the reducing effects of permanent saturation. The Salisbury vlei is not waterlogged all year round due to its topographic position, fewer mists, greater berg wind influence, being better drained as well as being influenced by farming practices on adjacent slopes. Thus the surface layers dry out somewhat during the winter months, resulting in a superficial horizon of lower organic content (usually less than 15%) due to oxidation and decomposition by aerobic bacteria.

The sediments in both vleis are clearly subjected to irregular inputs of colluvium and alluvium which follow high intensity, short duration rainfall events. This input is sandy loam material and corresponds closely to a "washbelt" of coarser material found near the edges of dambos (Mackel, 1974). Finer silts and clays appear to filter through this material to the centre where they accumulate and mix with organic matter derived from the dense hydrophytic vegetation to form a black matrix fairly high in decomposing plant remains. These organic strata become more fibrous towards the surface where deposition has been more recent, and are comprised of various distinguishable plant remains intermixed with fine clays (Plate 17).

Coarser sandy material also increases, relative to fine material, towards the headwater zones of the vlei, where deposition of larger particles occurs first due to the reduction in velocity of water entering the valley bottom. Where this deposition takes place by flood waters, some disruption of sediment layers occurs, making the boundaries between layers less clear. The sediment stratigraphy is recognisable by characteristics of colour, texture and organic matter, and can be used to identify periods of accelerated or diminished sedimentation of colluvial and alluvial inputs.



Plate 17. Black organic surface sediments with plant remains visible.

4.7 VEGETATION

A vegetation map produced by Acocks (1953) recognizes six major ecological "veld types" occurring in the Eastern Province region (Figure 13). However, much of Acocks' classification has been revised on a smaller scale by White (1983) due to the inaccuracy of uniting structurally unrelated types, and to devise a biogeographical map rather than an ecologically based one. Childs (1971) describes eleven vegetation "types" occurring in the Eastern Cape, based on the plant species present. On the cooler highlands of the Winterberg Range and its adjacent foothills where the MAP exceeds 600mm, three major vegetation types can be identified:

4.7.1 FALSE THORNVELD OF THE EASTERN PROVINCE

(Acocks, 1953, 1975).

On the Elandsberg foothills, the lower drier slopes are covered in low tussocky grasses interspersed with stunted individuals of Acacia karoo. This vegetation type is thought to have developed from pressures on the open grassland formations by overgrazing, trampling, sheet erosion and frequent natural fires. The true grassland formation has succumbed to encroachment of Acacia karoo and has been thinned out to reveal patches of bare ground, exacerbated by the on-going drought of the past three years (Plate 18).

This thornveld formation is relatively widespread when compared to the limited distribution of montane forest vegetation. The lengthy dry season and restricted summer rainfall when evaporation and evapotranspiration are high, are important ecological factors in the maintenance of this "open" almost treeless vegetation type. Due to these factors, the density and biomass of this vegetation type is lower than that of the higher slopes and valley heads (Meadows, 1985). The Eastern Province Thornveld extends its scattered appearance along the steep exposed interfluves of the Elandsberg. The dominant grass species are Themeda triandra

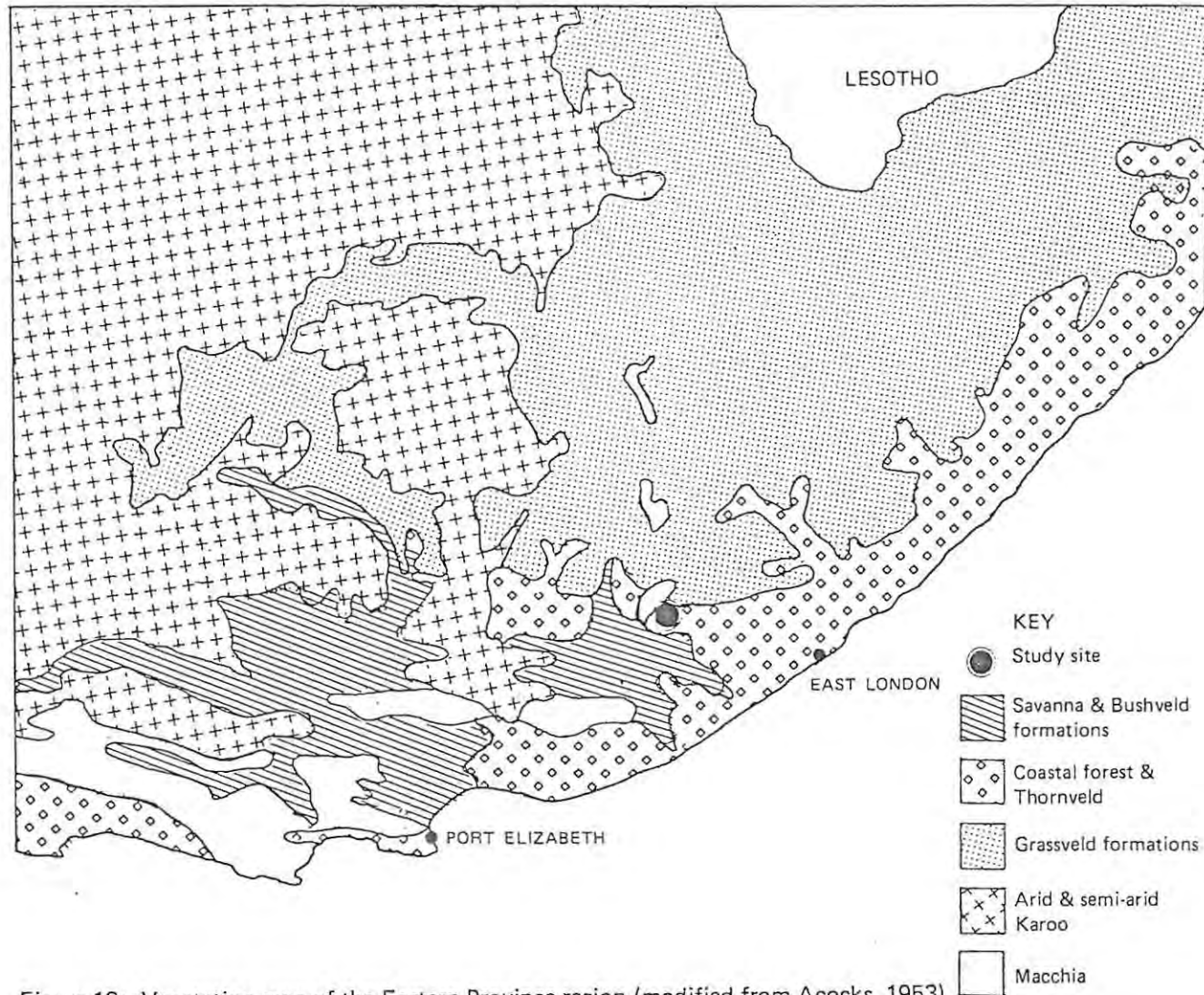


Figure 13: Vegetation map of the Eastern Province region (modified from Acocks, 1953)



Plate 18. False Thornveld (on Ecca shales).



Plate 19. Transition into montane forest in valley heads (on dolerite).

(rooigras), Eragrostis capensis and Aristida congesta. Shrubs are frequent and invaders such as Opuntia spp. are common on the lower dry slopes which are subjected to heavy grazing. Aloe ferox and Aloe arborescens are also numerous on the rocky doleritic hillslopes.

4.7.2 TEMPERATE EVERGREEN FOREST (Acocks, 1953; Childs, 1971).

The dense woodland found on the higher slopes and enclosed protected valley heads up to 1500m has been described, accompanied by extensive species lists, by several local workers (Dyer, 1937; Adamson, 1938; Story, 1952; Martin and Noel, 1960; Moll and White, 1978). This forest formation is found primarily on south-east aspects, sheltered from harsh, drying effects of hot berg wind conditions. These slopes receive cool moist effects of mist, drizzle and precipitation (Plate 19), enabling the support of higher biomass and species richness than the more arid lower areas (Plate 20). Dominant tree species are Podocarpus falcatus, Ptaeroxylon obliquum and Olea capensis. Alongside the forest/grassland transition, pioneer species such as Rapanea melanophloeos thrive. A profusion of spiny species like Asparagus spp. and Fagara capensis are found entangled within the dense groundflora (Plate 21), the only limitation on plant density being light, reduced by the evergreen nature of the tree canopies. The relatively infrequent occurrence of xeric and spinescent characteristics and the abundance of ferns, mosses and lichens indicates that there is abundant moisture in these cool dark forest areas (Plates 22 and 23) (Board, 1962; Dewey, 1984). However, in dry seasons, and more particularly due to the recent on-going drought in the area, direct moisture input to the forest soils are reduced by canopy interception. This, in addition to the higher evaporative losses by increased biomass, lowers the moisture availability to groundflora considerably, particularly nearer to forest edges. There is some evidence to suggest that forest areas in the Eastern Cape are shrinking rather than expanding due to man's



Plate 20. High biomass and plant density of montane forest.



Plate 21. Tangled undergrowth is light limited. Scale 1:30



Plate 22. Lichens, ferns and mosses. Scale 1:10



Plate 23. Mosses and fungi indicate moisture availability.

intervention (White, 1983), frequent burning (Meadows and Dewey, 1986) and a seemingly increased aridity in the area.

Clumps of plantation species of Pinus radiata and P. pinaster are found as windbreaks alongside fields and in scattered clumps on farmland areas. These exotic species easily withstand the harsh conditions on exposed upland slopes. Due to the acidic nature of the soils beneath these trees, groundflora is limited to a few tolerant species (Dewey, 1983; Plate 24).

4.7.3 GRASS/HEATHLAND OR MACCHIA VEGETATION

The highest peaks and slopes of the Winterberg are exposed to harsh winters with snow, frost and strong winds. Temperature extremes as well as other adverse climatic factors severely reduces the height and density of vegetation. Growth at these high altitudes is also retarded by acidic shallow soils which offer limited anchorage and nutrient supplies for larger plants, resulting in a varied vegetation mosaic of grassland and macchia species (Plate 25). Macchia is generally a more xerophytic type of vegetation capable of withstanding dessication by wind. These plants found on the Winterberg are generally green blue-grey shrubs, profusely branched and less than 3m in height (Plate 26). Common families are Ericaceae, Asteraceae and Restionaceae (Plate 27). Surrounding the vleis on adjacent slopes is dense tussock grassland interspersed with macchia species (Plate 28).

There has been much speculation as to the exact classification of this South African macchia or "fynbos" vegetation. The problem of definition lies in the fact that "most formations are very diverse and single-species dominance is not common" (Meadows, 1985, p.128). The presence of this vegetation is often due to the existence of several environmental controls and, as such, the definition should possibly be one of a biogeographical nature rather than purely ecological. The presence of this vegetation type on the Elandsberg



Plate 24. Limited groundflora under pine plantation.



Plate 25. Shrublike grass/heathland on exposed plateau.



Plate 26. Mosaic of grey-green shrubs. Scale 1:10



Plate 27. Profusion of heathland species. Scale 1:10

is due mainly to the response of the plant community to the gradients of environmental factors acting upon it, these factors including soil depth, geology, aspect and altitude.

4.7.4 VLEI VEGETATION

The dense vegetation associated with poorly drained valley bottoms is a form of mire grassland. Shrubs and grasses of Crassulaceae, Asteraceae, Monocymbium spp., Agrostis spp., and Aristida spp. form a sparser population along the margins, preferring the seasonally wet, freely-draining sandy wash belt (Plate 29). The more hydromorphic species of Knifophia and Mentha spp., Juncus spp., and sedges such as Scirpus spp., Carex spp., and other members of the Cyperaceae family thrive in a varied and dense community towards the waterlogged centre. These species are interspersed with odd patches of Phragmites spp., which form taller stands in the centre of the vlei, where their basal parts are completely submerged in water (Plate 30).

These vlei species are mostly small and tufted and are marsh or hydrophytic species. The importance of these dense stands of water-loving species cannot be overstated as they provide a direct and very important source of organic input to the vleis. The changes in sediment characteristics over time reflect not only changes in hydrological conditions, but also changing vegetation in the vleis. Different peat-like sediments can be explained in terms of changing moisture conditions, where drier or wetter conditions alter the vegetation type and rate of decomposition. Saturated conditions, as in the vleis, high precipitation and acidic parent material all facilitate the continued accumulation of plant material, a process which is largely uninterrupted all year round. Clays and sands are eroded and deposited primarily under drier conditions, either as seasonal or large-scale changes in climate, producing narrow or wider mineral bands respectively. Mineral sediment is derived mainly when



Plate 28. Tussock grassland on adjacent slopes.



Plate 29. Intermixed vleis species of sandy margin decrease in density to vleis edge.



Plate 30. Dense taller stands of hydromorphic species towards vlei centre, including Phragmites and Scirpus spp.

vegetation cover is reduced and overland flow results in increased sediment entrainment.

4.8 SUMMARY

Vegetation boundaries are dynamic and respond to gradients of environmental factors acting upon them. As such, they can act as indicators of environmental change. The significance of potentially changing vegetation boundaries and types is important in terms of sediment supply to the vleis and valley areas. The effectiveness of plant cover in reducing erosion depends upon the height and continuity of the canopy, the density of cover and root development and is therefore of great importance in determining the availability of sediment for transport processes. If different vegetation community boundaries and soil boundaries are found to be parallel, it can be assumed that soil characteristics are largely determined by the vegetation types existing on them (Meadows, 1982, 1983; Dewey, 1984).

Any change in distribution of vegetation communities on the Winterberg would significantly affect sediment input (organic and inorganic) to the vleis. For example: if the boundary of the forest vegetation was near the edge of the vlei, sediment input would be limited due to the dense undergrowth and full canopy, as the soil has greater protection against the forces of erosion. In this case, the main source of sediment to the vlei would be from the vlei vegetation itself depositing organic debris. If arid conditions were to prevail, forest species would retreat to protected valley heads and tussock grassland would dominate, leaving bare patches of soil. This would result in an increase in the rate and amount of inorganic sediment from adjacent slopes.

Vegetation boundaries are in flux with climatic conditions, and sediment input and stratigraphic development, can be directly related to vegetation

cover, already known to be fairly sensitive to climatic conditions (Meadows and Dewey, 1986). It is likely that there have been considerable changes in past climates on the Winterberg, causing fluctuations in vegetation distributions. The sediment strata found within the vleis of this area should thus contain varying amounts of organic matter, different cation exchange capacities and particle size distributions. From these characteristics the environment at the time of sediment deposition can be inferred.

It is possible at this point to create a number of hypotheses:

- (a) vlei sediment characteristics have a pattern similar to that described in other areas;
- (b) increasing organic matter levels correspond to moister conditions;
- (c) vlei sediments are fairly sensitive indicators of environmental change.

These hypotheses can be tested by analysing the sediments both in the field and laboratory to establish whether these features are indeed useful for interpreting past conditions, as have been the damboes of Malawi, Zimbabwe and Zambia.

CHAPTER 5

TECHNIQUES OF DATA COLLECTION

5.1 INTRODUCTION

To describe the sediments of the two Winterberg vleis, a three-phase sequence of analysis was employed. Initially, the samples were described on site after extraction, then subjected to laboratory analysis. This included physical and chemical measurements, selected to identify differences in character between samples. Once the analytical data was tabulated, statistical procedures were employed to establish the significance of relationships between variables. As the study was primarily a descriptive and comparative account of the vleis, it was felt that the nature of the data did not warrant sophisticated statistical analysis. For this reason, simple statistics were applied only to test the strength of relationships already noted between variables. The final analysis of data enabled inferences to be made on sediment chronology and environmental change in the area. A description of techniques used in this study follows, and all analytical results appear in Appendix 1.

5.2 ANALYTICAL TECHNIQUES

The sediments which have accumulated in the Winterberg vleis are peat-like in appearance and are permanently waterlogged, making description and sampling of exposed sections impossible. It was decided that undisturbed cores of sediment would be sampled using a suitable borer, so that a stratigraphic sample of the entire sediment sequence could be extracted. During a pilot study in March, 1985, a Hiller borer (Smith and Atkinson, 1975) was initially selected (Figure 14; Plate 31), but it was found that this type of sampler was impractical and gained only limited penetration on reaching compacted coarse sediments found below the

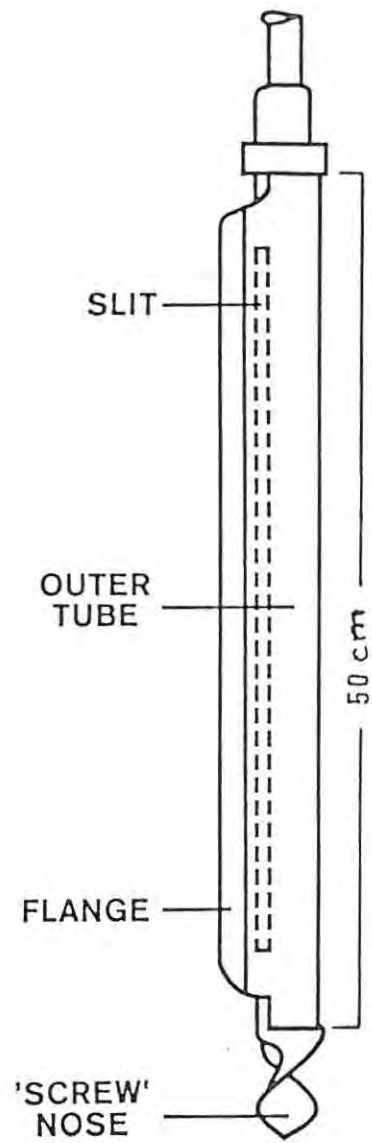


Figure 14: A Hiller borer.

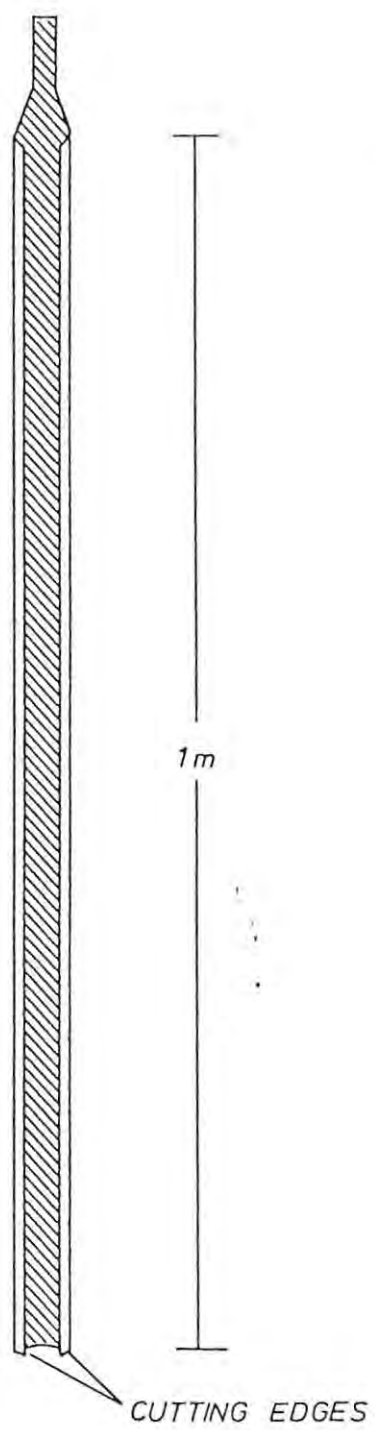


Figure 15: Diagram of a gouge auger.



Plate 31. Hiller borer — restricted sampling depth.



Plate 32. Gouge auger and extension rods. Scale 1:12

superficial organic sediments. Similar problems were encountered by Meadows (1982) in his study of dambos in Malawi. A gouge auger allowed for deeper penetration beneath the black peat-like sediments reaching depths of up to 3.60m (Figure 15; Plate 32). A slight disadvantage with this method is the higher possibility of sequence distortion. That is, the sediments can be subjected to either compaction or expansion within the column of the auger during penetration or extraction of the core. Careful sampling can minimize this aspect of the sampling procedure.

The sampling procedure was to use three transects run across the vleis. Although a grid sampling pattern was initially selected, time and the number of samples required made this procedure impractical. Transects, although not ideal, were selected as they enabled representative samples to be taken showing changes in stratigraphy with depth and in longitudinal and transverse directions. Their position was determined by accessibility, the large extent of the vleis and sampling time. Three transects were placed across the Dunedin and Salisbury vleis, with cores extracted at regular intervals. These cored samples were then used for chemical and physical analysis at a soils laboratory. Soil samples from adjacent slopes bordering the vleis were sampled using a Edelman soil auger, subjected to similar analyses for information on sediment supply to the vleis. Both the peat borer and soil auger, manufactured by a Dutch company, Eijelkamp, were light-weight, easy to assemble and had compatible extension rods.

5.2.1 FIELD SAMPLING TECHNIQUES

Dunedin vlei was selected as the principal study site as it had a number of advantages, outlined in Table 3. Two transects were run across the Dunedin vlei along which samples were extracted. A longitudinal transect was run the length of the valley and one transverse transect spanned the width of the vlei, with cores taken at regular intervals (Plate 33). The positions

Table 3: Criteria for Selection of Study Site

Accessibility	- the vlei was accessible by road which facilitated the removal of sediment cores.
Extent	- the vlei was one of the largest in the area, over a kilometer in length and branched.
Deep sedimentary sequence	- coring was possible to a depth of 3,6 m.
Least disturbed	- there had been little human interference within the vlei.
Radiocarbon dating	- dating was possible on basal and midsection strata, and indicated an undisturbed chronological sequence.



Plate 33. Longitudinal and transverse section of Dunedin.

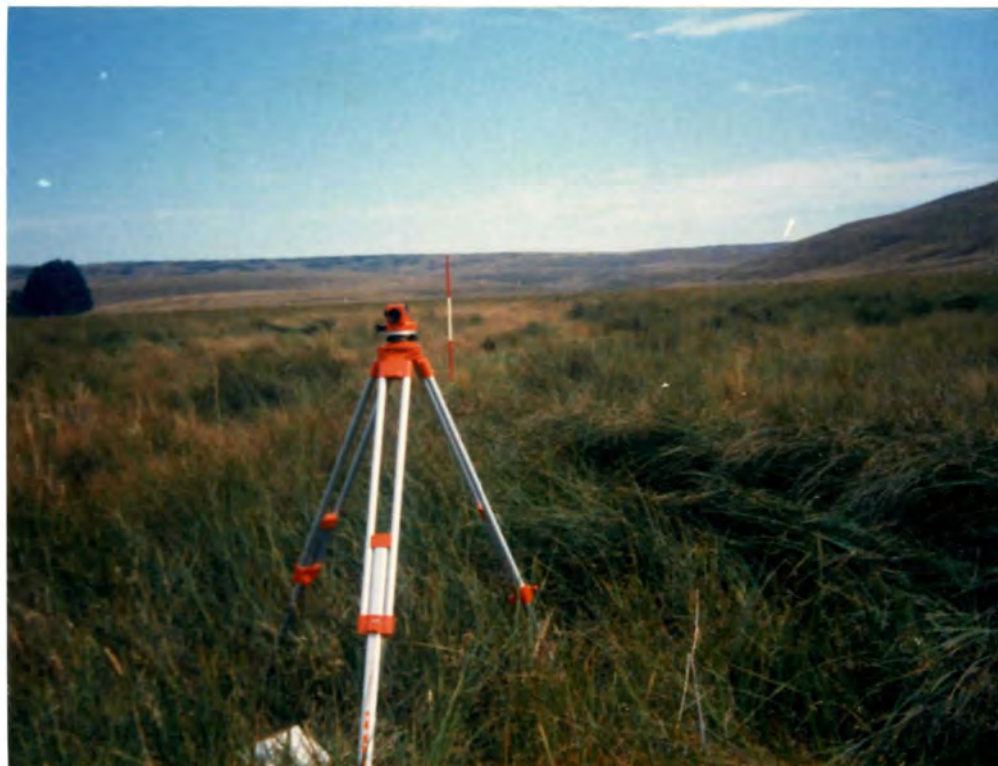


Plate 34. Dumpy level for profile examination.

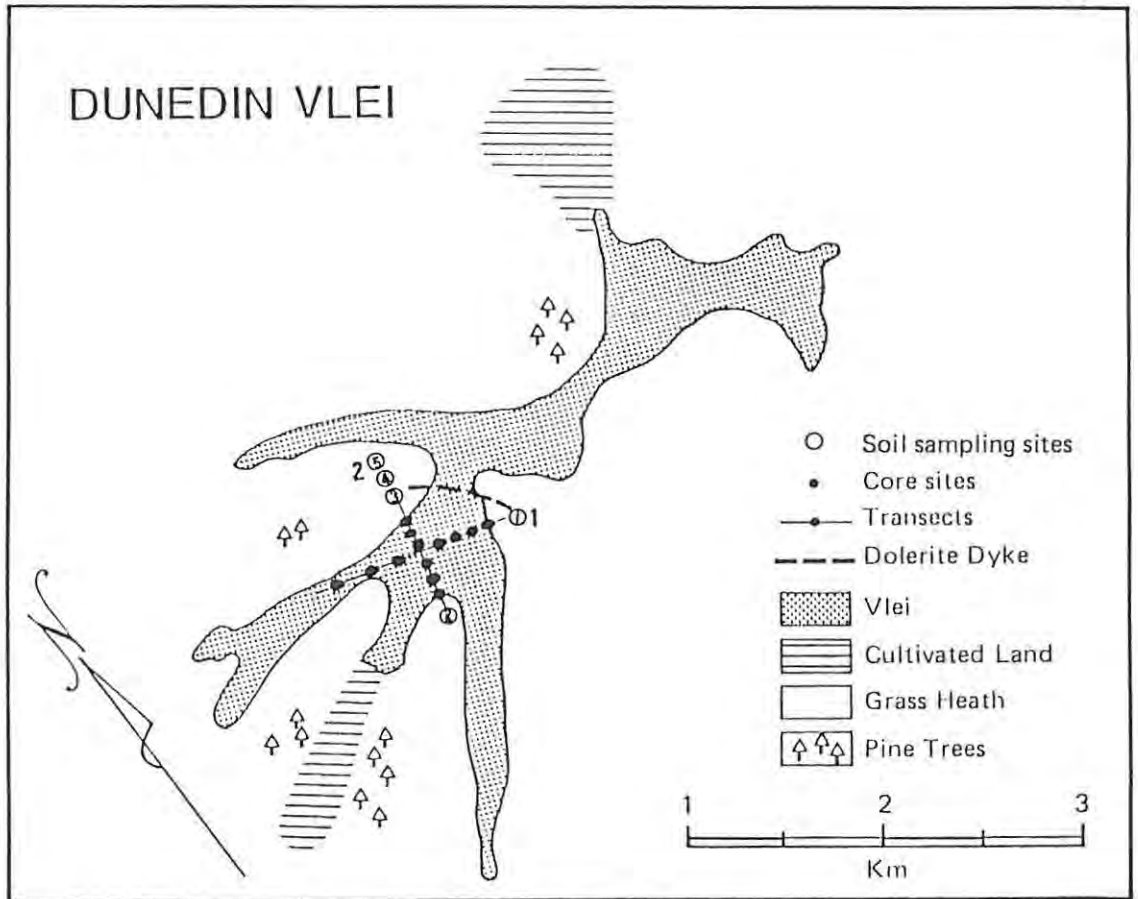


Figure 16a: Transect and coring positions at Dunedin vlei.

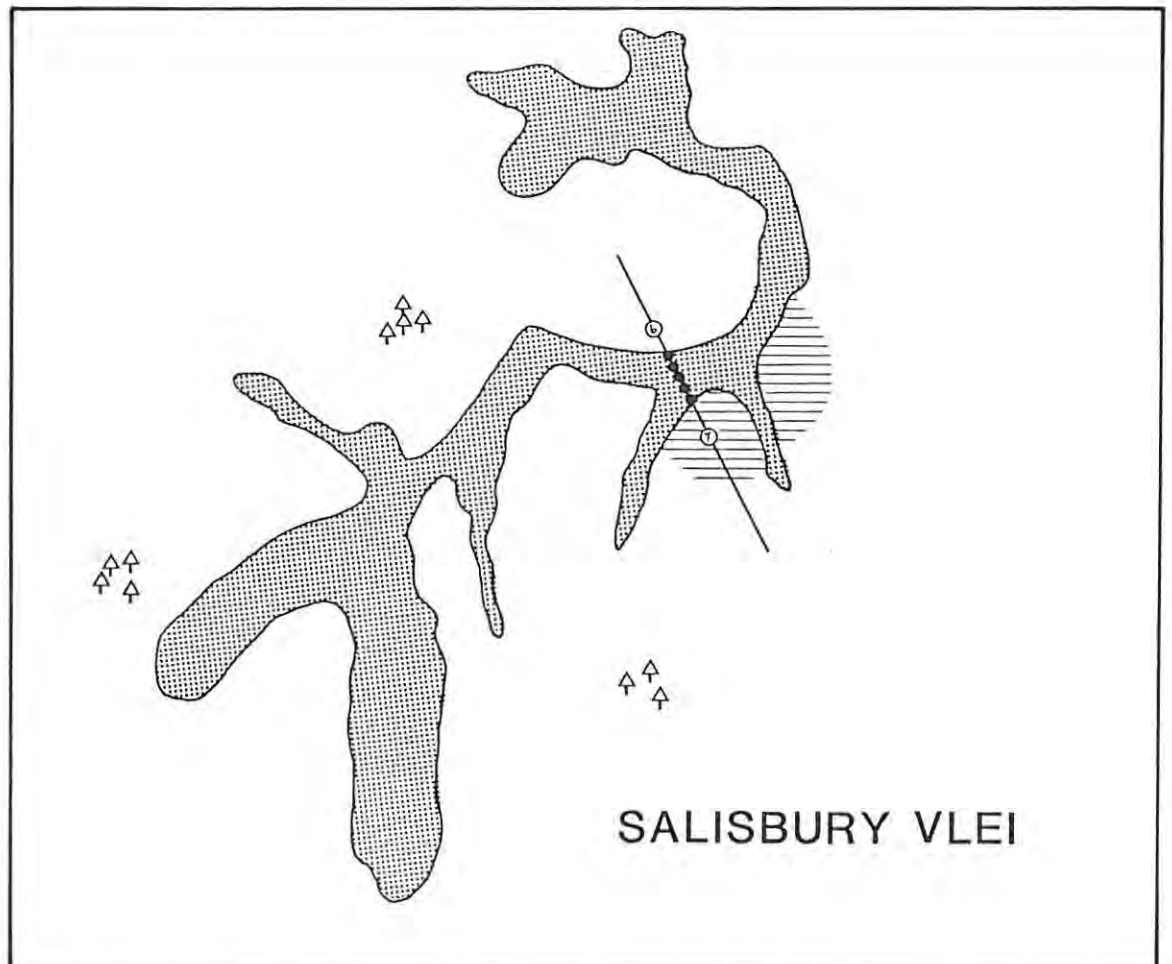


Figure 16b: Transect and coring positions at Salisbury vlei.

of the transects and coring sites are shown in Figure 16a. The longitudinal transect (transect 1) was 750m long, cored at 14 sites (every 50m) for description to bottomset beds, and sampled at 7 sites (every 100m) for laboratory analysis, including one soil sample from the slope at the bottom end of the vlei. The 300m long transverse transect (transect 2) was cored and sampled at 30m intervals. This resulted in 7 cores within the vlei itself and 4 soil samples from adjacent slopes. In all, 25 cores were extracted, 19 of which were subjected to laboratory analysis. A radiocarbon date, obtained from 330cm depth, indicated the exact period, 12 500 \pm 160 years BP (Lab no. Pta - 4207), of deposition in the vlei. The transects were levelled using a Dumpy level (Plate 34) and stratigraphic sequences drawn. Soil profiles, exposed on the adjacent slopes to identify parent material and enable accurate identification to soil form.

The extracted samples were placed in stiff plastic which was tightly sealed and maintained in a rigid position to prevent distortion of stratigraphic units, loss of moisture or contamination (Plate 35). The surface 15cm was sampled by hand as the saturation of this layer prevented coring. Samples were carefully transported to a nearby vehicle where the rigidity of the plastic was maintained to forestall confusion between stratigraphic units. The limitation of corer penetration by compact beds was partially overcome by the introduction of the narrower, less time-consuming and more practical gouge auger, enabling samples of 1m lengths to be extracted.

The sampling procedure for Salisbury vlei was less intensive than that for Dunedin, the latter being the principal site under study. Salisbury vlei was thought to be less indicative of 'natural' vlei conditions as one of the adjacent slopes had previously been under extensive cultivation. These practices may have altered natural sediment input to the system and disturbed the sedimentary sequence to some degree.



Plate 35. Sample extraction and sealing.



Plate 36. Transverse view of Salisbury. Scale 1:20

However, Salisbury vlei showed similar morphological characteristics to Dunedin, and, being in the same catchment area, was thought to be a useful index for comparison. A radiocarbon date for these somewhat shallower sediments, 11 800 \pm 120 BP (Lab no. Pta - 4318) correlated well with the Dunedin date, signifying that changes occurring in the environment on a regional basis must have affected both sites. The gouge auger proved compatible under all sampling conditions at this vlei, even though some areas were considerably drier than Dunedin. One transect was run diagonally across the lower end of the vlei for 240m (Figure 16b; Plate 36). Cores were extracted every 40m within the vlei, of which 5 were for laboratory analysis. Two soil samples were taken from the adjacent slopes.

A measure of vegetation density was made using a 1m^2 grid to estimate the percentage cover within the vlei. Twenty replicates were made along the margins and repeated in the centre, and a mean measure of cover resulted.

Using the Munsell Colour Chart (1975), fresh sediment colour was analysed and recorded upon extraction of the core before sealing. Due to the saturated nature of the samples, and the very dark colour of the organic sediments, colour distinctions were difficult. For this reason, colour analyses were repeated in the laboratory after air-drying. The colour of the 250g soil samples, after extraction from the auger head, was noted before being sealed and labelled in polythene bags. Upon opening under laboratory conditions, colour was again recorded and checked against previous notes. The depth of each core was noted on the sample unit and recorded on a chart (Plates 37 and 38).

5.2.2 LABORATORY TECHNIQUES

To interpret the significance of these vleis in qualifying environmental change, all samples taken during the field investigation were subjected to physical and chemical analyses. These analyses,

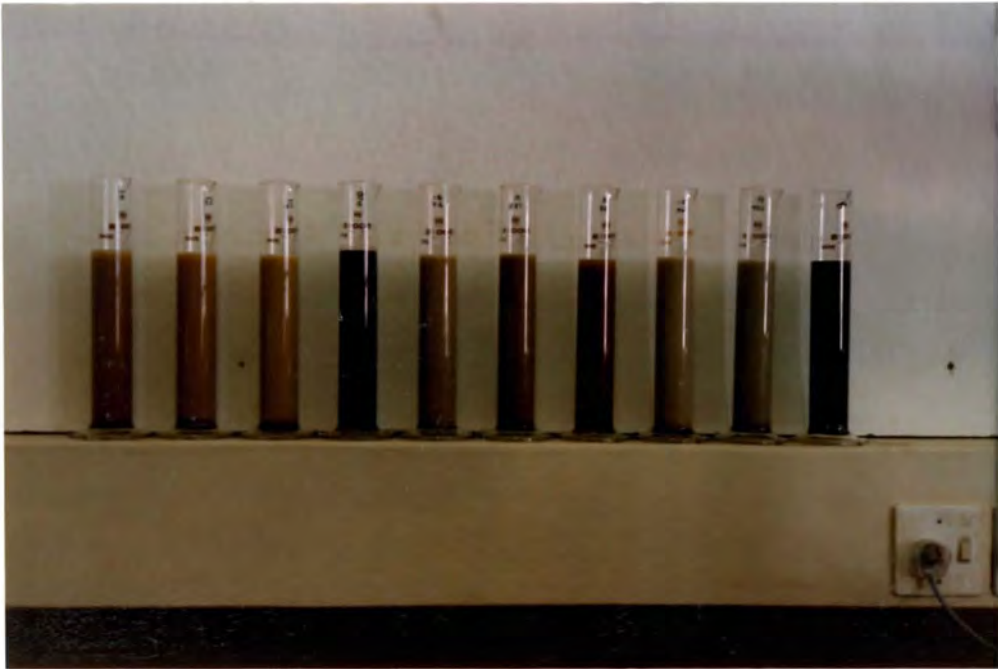


Plate 37. Row of sedimentation cylinders. Note colour variation. Scale 1 : 14



Plate 38. Black surface sediments contrast with pale grey basal sands. Scale 1 : 5

barring the radiocarbon dating, were carried out in the Soils Laboratory at Rhodes University Geography Department. The techniques used were identical to those stipulated by the British Standards Institute manual (BSI,1975).

The physical and chemical characteristics of the sediments were thought to offer clues as to their origin e.g. related to geology or vegetation type, and thus on the environmental conditions at the time of their deposition. The choice of parameters for laboratory measurement was made on the basis of those characteristics which could be related to external factors of the environment, as listed in Table 4.

The number of measurements and parameters selected were limited by time and small sample size. Each core was subsampled at 15cm intervals, and these subsamples were subjected to analysis. Excluding moisture content analyses, all samples were air dried over several days, and then ground to pass through a 2mm sieve. The crushing of samples after oven-drying proved difficult with samples of high clay content, as the rock-like samples totally resisted breakdown with a plastic covered pestle. A mullite (ceramic) pestle was thus employed, possibly affecting particle size distribution by slightly exaggerating the finer size fractions.

5.2.2.1 Physical analyses

The physical analysis of samples facilitated the investigation of changing sediment size characteristics across the vleis. Previous studies have indicated a fining of sediments from the sides to the vlei centre, as well as with an increase in depth (Mackel, 1974; Meadows, 1985).

Moisture content determination was made prior to any other disturbance of the samples to avoid loss of moisture to the atmosphere once samples were unsealed. About 10g of sample was accurately weighed into a weighing dish of known mass, and oven-dried for at

Table 4 : Sediment parameters selected for analysis

Physical	Chemical
Moisture content	pH
Particle size : clay %	Organic matter
silt %	Macro-cations
fine sand %	Exchangeable hydrogen
medium sand %	Cation exchange capacity
coarse sand %	Total exchangeable bases
	Base saturation

least 6 hours at 105°C. The sample was then cooled in a dessicator and reweighed (Smith and Atkinson, 1975). The loss in mass was taken as a measure of soil moisture.

Particle size analysis using the Pipette technique was made, after air-drying, on all subsamples obtained using a sample splitter to achieve as near-to-representative subsamples as possible. These samples were then oxidised using concentrated hydrogen peroxide to remove organic matter (Plate 39). Particle size was then accurately determined using the standard pipette analysis described in the British Standards Institute manual (Plate 40) (BSI, 1975). Problems were, however, encountered with the highly organic samples and those with a high clay fraction. After oven-drying, these samples were very resistant to crushing and breakdown. Excess dispersion agent, sodium hexa-meta-phosphate, had to be added to the solutions even after oxidation with hydrogen peroxide.

5.2.2.2 Chemical analyses

(a) pH

pH was measured in a 1 : 2.5 ratio of a 10g soil sample to 25mm of de-ionised water, which was stirred thoroughly every 15 minutes for one hour. The pH was then read on a Labion Model 17 Specific Ion Counter, three readings taken for each sample, and a mean pH calculated for each sample. This procedure was repeated with a 1 : 2.5 ratio of soil to KCl and the readings between the two methods compared. Variations of between 0.5 and 1.75 pH units occurred. The KCl readings were used as the final pH readings as these varied less between successive readings and were more consistent throughout. It has been established that pH readings in KCl solution are less influenced by changes in biological and meteorological conditions than water solutions, and tend to give a more representative characteristic of the soil than a pH measured in a water solution (Peech, 1967).



Plate 39. Sedimentation cylinders of dispersed sediment. Scale 1:13



Plate 40. Sampling pipette for particle size analysis.

(b) Organic matter

The total carbonaceous material in a given soil sample was measured by a method following that originally proposed by Walkley and Black (1934). Due to inconsistent results and extremely low values, weight loss by ignition was attempted for comparison. Suspiciously high results showed values of up to 21% higher by the ignition method, although it is known that this method is non-specific to organic matter and thus often produces exaggerated results (Dean, 1974). It was then discovered that the ammonium ferrous sulphate used in the Walkley-Black method had become hydrated and had started to degenerate. Fresh chemicals were obtained, after which results were higher and appeared more consistent with anticipated results.

In the Walkley-Black method, a known mass of soil is digested by concentrated sulphuric acid after the addition of 1N potassium dichromate solution. Orthophosphoric acid is added as a catalyst and titrated against 0.5N ammonium ferrous sulphate, using sodium diphenylaminesulphonate as indicator. The amount of organic carbon in the sample is proportional to the quantity of reacted dichromate, and total organic carbon is expressed as the product of the oxidisable organic carbon and a conversion factor of 1.33 (Smith and Atkinson, 1975). A constant of 1.724 converts organic carbon to organic matter (Meadows, 1982).

(c) Exchangeable cations

Cation availability and exchange is a function vital for the supply of nutrients for various plant metabolic and growth processes. To measure these nutrients, the most representative method would be the one which most closely resembles the activities of plant roots under given environmental conditions. There is a need to distinguish between those nutrients which are freely exchangeable and those which are in a fixed form and thus unavailable for plant uptake. This, under

artificial conditions, is dependent on the type of extractant used. The amount of any one element leached by an extractant is a function of the nature, concentration, soil to extractant ratio and the length of time available to the extractant (Smith and Atkinson, 1975). In the laboratory, artificial conditions result in several problems:

- * various extractants give rise to a change in dominance of one form of chemical bonding to another;
- * extractants fully saturate the soil particles, reaching most particle surfaces, while root systems are in contact with only a fractional particle surface area;
- * environmental factors constantly alter nutrient status with changes in e.g. temperature, where low temperatures reduce the availability of all cations (Peech, 1967; Viets, 1967).

The method of extraction in this study for the four macro-cations (Ca^{2+} , Mg^{2+} , Na^+ and K^+) was with neutral ammonium acetate solution ($\text{CH}_3\text{COONH}_4$). Details on the preparation of the reagents and method of extraction are given in Appendix 2. After extraction, the leachate was analysed by atomic absorption spectroscopy using Lanthanum chloride (LaCl_3) as an internal standard, and are expressed as milliequivalents per 100g of soil (m.e/100g).

(d) Exchangeable hydrogen

The ammonium acetate leachate used for the determination of exchangeable cations, is used to assess the concentration of exchangeable hydrogen. From the original leachate sample, a 25ml aliquot is withdrawn by pipette into a conical flask and titrated back to pH 7 using 0.2N NH_4OH with bromothymol blue as indicator.

$$\text{Exchangeable H}^+ = (\text{ml NH}_4\text{OH} \times 0,2 \times \frac{250}{25} \times \frac{100}{\text{Soil mass}}) \text{ meq/100 g}$$

(Smith and Atkinson, 1975).

A curve derived by Brown in 1943 (Jackson, 1958) from an equation, plots the pH of NH_4OAc against total exchangeable H^+ in m.e./100g of soil (Figure 17). This curve was used for comparison of the samples determined by titration, and results appeared to correspond fairly closely.

(e) Cation exchange capacity (CEC) and total exchangeable bases (TEB)

The CEC of a sample can be taken as a measure of the nutrient status of a soil at a given time for a given condition (Chapman, 1964). For this study, the CEC for the mineral fraction was taken simply as the sum of the exchangeable cations added to the exchangeable hydrogen concentration. The CEC of the whole soil, including the very colloidal organic portion, is far higher due to the surface area being far greater than that of a clay particle (even of montmorillonitic type). It is suggested that the CEC of organic material is approximately 2.0 m.e. for every 1% organic matter (Buckman and Brady, 1969). The total CEC is thus the sum of the mineral CEC and twice the organic percentage. The total concentration of bases present in the soil, attached to clay and organic particles as well as in the soil solution, are termed exchangeable as they are available for exchange into the soil solution and therefore for plant uptake by osmosis via the roots. CEC, when calculated for the whole soil is far more representative of the nutritional status of that soil under natural conditions. TEB is calculated as the sum of the major cations ($\text{Na}^{2+} + \text{K}^+ + \text{Ca}^{2+} + \text{Mg}^{2+}$) and is also expressed in m.e./100g of soil.

(f) Base saturation (BS)

BS is the percentage of the CEC occupied by available exchangeable bases. That is, the number of negative charges sites filled by cations, not including hydrogen ions, from various sources as a percentage of the total number of negative charge sites available.

$$\% \text{ base saturation} = \frac{\text{TEB}}{\text{CEC}} \times \frac{100}{1}$$

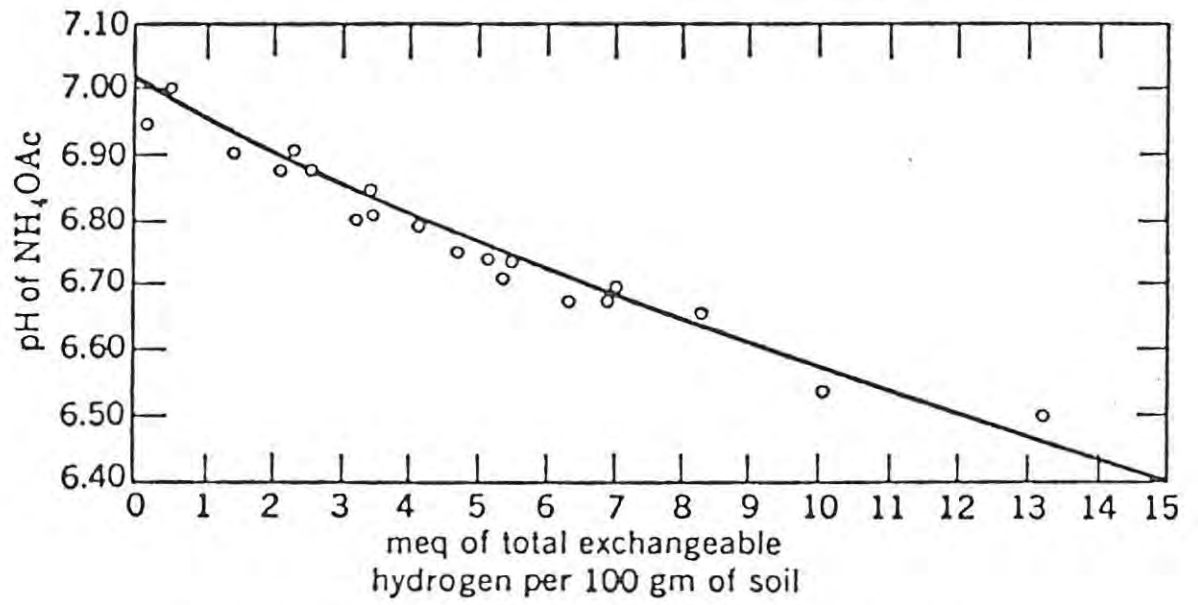


Figure 17: Relation of pH of NH₄OAc solution to exchangeable hydrogen (after Brown, 1943).

5.3 STATISTICAL TECHNIQUES

The statistical analyses performed on the raw data set were simple and used to substantiate trends described thus far. The main objective of the study was to describe and use laboratory analytical procedures to characterise the sediments and facilitate comparison to other similar features. For this reason, the statistical analyses employed were purely to determine the degree of confidence to be placed on relationships described from this analytical data. Although these relationships have been shown from the raw data, the significance and strength of such relationships had to be quantitatively determined. For example: a relationship between organic matter and pH may seem unquestionable, but how dependent pH is on organic matter is unknown. In addition, statistical testing of the hypotheses outlined within the study was necessary.

To test for various relationships between variables and describe the analytical results statistically, correlation analysis was employed. This analysis assessed the various associations, if any, between variables. It is important to remember that an association does not infer dependence or causation, but merely that the two variables appear to be associated. Regression analysis was employed to graphically describe the various relationships which were statistically significant, and to show whether the variables were positively or negatively correlated. The scatter plots indicate the degree of variation between points and the position of the regression line.

Association analysis was also employed in an attempt to group variables on the basis of a similar characteristic, thus forming a pattern within the data where variables are selected according to set criteria. For example, variables were sorted according to a threshold organic value and the degree of variation about a point. These variables were plotted around the threshold value, giving an indication of the

associations between samples taken from widely differing points.

5.4 SUMMARY

All data collected from the laboratory analyses were tabulated and graphs drawn to show significant relationships between variables. Photographs are included as a visual assessment aid to the degree of variation between soil and vegetation types.

From the various analyses conducted, some indication of contemporary hydrology and sediment types accumulating in the vleis can be described. Changes in sediment characteristics, for example, an increase in organic matter, may indicate a period of accelerated organic accumulation with concurrent increases in moisture availability and temperature. To accurately determine the chronology of the various sediments, radiocarbon dating was conducted on selected samples at CSIR, Pretoria. The number of samples which could be sent for radiocarbon dating were limited by the low organic content in some of the stratigraphic bands. The only sections from the cores which could be accurately dated were those fairly high in organic matter (above approximately 15%). There was also danger of contamination and distortion of the cores, although this was hopefully minimized by sealing and careful transportation of samples. The radiocarbon dates of basal and other samples provided a time-span for sediment accumulation and assisted in the identification of drier and moister periods through time (Chapter 6).

As sediment characteristics reflect an interaction of several environmental controls (Jenny, 1941), the comparison between Dunedin and Salisbury vleis is made possible due to the constancy of some of these factors of formation. The geology (doleritic parent material), geomorphology (rolling hills) and climate may be taken as constant for both vleis, although Salisbury vlei

conditions are drier for some months of the year. The sediments are assumed to have developed over a similar time span, confirmed by radiocarbon dates and due to similarities in extent, morphology and depth. Vegetation in both vleis is of similar structure and function (hydromorphic species). The two vleis are thus directly comparable. The chemical and physical characteristics of the sediments should offer information enabling the elucidation of environmental change and some inferences as to the frequency and magnitude of these changes. Correlation between various characteristics and between samples from the two vleis, may lead to substantiation of a sediment history of the area.

CHAPTER 6

ANALYTICAL AND STATISTICAL RESULTS

All analytical results are listed in sequence in Appendix 2. Tables and figures of results are presented in chronological order at the end of this chapter for ease of reference, while explanations are given under the appropriate headings. Characteristics of sediments and soils are described from the youngest (surface) to the oldest (bottom beds) layers, except when describing the stratigraphy. Here, sediments are dealt with in order of deposition through time, the bottomset beds being described first (as the first layer deposited) moving upwards to the surface.

6.1 SOILS SURROUNDING THE WINTERBERG VLEIS

6.1.1 INTRODUCTION

The soils at the foothills of the Elandsberg were not analysed as they bear no significance in vlei sedimentation. The soils at the summit of the Winterberg, developing from decomposing dolerite, do form an integral part of the sedimentation processes in the vleis. The texture, colour and amount of sediment, determined by the parent material from which it is weathered, is supplied to the system by erosional and drainage processes.

6.1.2 SOIL FORM AND DEPTH CHARACTERISTICS

Soil forming processes on the Winterberg at high altitude, are limited by low soil temperatures and a summer rainfall regime. However, chemical and physical weathering are promoted by extreme climatic conditions and leaching processes which occur mainly during the wet summer months. Soils are fairly shallow, 7 profiles averaging only 95cm. Exposed soil profiles enabled the identification of these soils as the Shortlands form, according to the Binomial system of classification (MacVicar et al., 1977). The effective

rooting depth is greater than a metre where bedrock is not limiting.

The Shortlands form soil profile consists of a weakly structured orthic topsoil overlying a moderately strong blocky subsoil. On the Winterberg, the topsoils are humic-phased in that they contain a fairly high organic matter content ($> 5\%$) (Figures 18 and 19). For this reason, the new form name would be the Lusiki form, as specified by the revised Binomial system (MacVicar et al., in prep.). Weathered bedrock, or saprolite, is encountered at approximately 1m depth. These loamy soils are relatively mature. Although distinction between horizons is unclear due to the characteristic uniformity of colour, changes in texture and structure with depth are marked by a rapid increase in clay content. Such changes with depth occur due to leaching or eluviation, of nutrients and clays from the topsoil which accumulate in the subsoil. The presence of distinct horizons reflects pedogenic processes which mark the development of a maturing stable profile.

The water balance within these soils is affected by depth and slope. The available moisture capacity, or water-holding property, of these blocky loams is moderate to high due to high clay and organic matter contents. Even during winter months, the moisture content varied between 17 and 35%. These soils have a medium to high infiltration rate into the permeable topsoil horizon, good drainage and a low runoff and erosion hazard. These are important factors in the supply of sediment by overland flow and subsurface drainage into the vleis at the base of these slopes.

6.1.3 SOIL COLOUR

Colour development in these soils (dark reddish-brown loams), is due to good drainage and aeration related to their upland position. The topsoil horizon is slightly darker in colour due to a higher organic content, but colour is generally uniform throughout the profile. This is characteristic of the Shortlands form soil.

The red colours are due mainly to the presence of iron and aluminium oxides coating individual soil particles. Colour ranged from 2.5 YR 3/4 to 10 R 4/3 in the top- and subsoil layers to 5 YR 4/8 at depth where tongues of soil are found penetrating into weathered saprolite, giving a variegated yellow colour to the profile.

Mottling was absent, indicative of free drainage and good aeration common to upland soils, although variegated colours were found where saprolite occurred. Shiny clayskins or 'cutans' were found coating ped surfaces, giving them a slightly darker appearance. These cutans are common in blocky loams, particularly where clay contents exceed 35% and eluviation from surface layers is commonplace. A similar process of clay translocation is thought to occur in the vleis itself, clay particles being washed through the profile by seepage waters to accumulate in fibrous organic bands below.

6.1.4 ORGANIC CONTENT

The physical and chemical properties of soils are greatly influenced by the proportion of organic matter occurring in them. Organic material is a major source of plant nutrients, released during decomposition, and particularly of nitrogen, this being the only form available for plant uptake. Organic matter is also an integral part of the clay-humus colloidal complex, able to absorb and retain large quantities of moisture as well as supplying other sites for nutrient adsorption (Buckman and Brady, 1969; Tivy, 1982). Organic matter also serves to improve the stability of soils by acting as a binding agent, strengthening soil peds and structure and reducing the erosive effects of water.

Under grass/heathland vegetation, as found on the Winterberg, organic matter is supplied not so much by litterfall, as beneath forest communities, but by decomposition of the dense adventitious rooting networks found below the soil surface.

Organic matter varied between 5 and 14% under the grass/heathland conditions, but values decreased sharply with depth. The high surface values were related primarily to the highest density of roots occurring in the uppermost layers. Nearing bedrock, the values were reduced to between 0.5 and 3% (Figures 18 and 19). The lower surface organic percentages under grassland conditions as compared to those under dense vlei vegetation, are due to lower productivity, improved aeration, more rapid decomposition, extremes in temperature and frequent burning. Natural fires are common on the Winterberg during winter months, often spurred by hot dry berg winds (Plate 41). Surrounding slopes and the uppermost aerial leaves of vlei vegetation were burned by an extensive natural veld fire in the winter of 1985 (Plate 42), fortunately following completion of the sampling programme. Major changes in vegetation cover, such as that following a widespread fire, will greatly affect the amount of sediment available for erosion. The first rain after a fire or excessively dry winter will result in the highest sediment yield, and could supply a "flush" of mineral sediment of all particle sizes into the vlei system.

The trend of variable organic content in the Winterberg soils is comparable with available data from Malawi (Meadows, 1982) where similar organic percentages were found on the slopes adjacent to the dambos of the Nyika Plateau. Here, Ah horizon values ranged from 6.2 - 14.6% under grassland, falling to 0.3 - 6.8% at 1m depth. Far higher values were recorded under nearby forest communities on the Nyika plateau.

6.1.5 CHEMICAL CHARACTERISTICS

The physical weathering of bedrock is promoted in this area of temperature extremes and where large diurnal fluctuations occur. Chemical decomposition of rock is assisted where drainage is improved by steeper slopes and higher insolation levels, but is generally retarded during low winter temperatures. Dominant clay minerals



Plate 41. Emerging plants after natural fire (winter 1985).



Plate 42. Tops of vlei vegetation burned to water-level. Note micro-topography of hummocks around plants, and hollows between.

within these soils as a result of these weathering processes, are illite and montmorillonite, both of which have higher cation exchange capacities (CEC) than the kaolinitic clay types. Van der Merwe and Weber, (1965) note that at higher altitudes, due to relatively low soil temperatures and low aeration in expanding clay soils during the rainy summer season (as is often the case on the Winterberg), the initial weathering of dolerite produces predominantly montmorillonite which remains in a relatively unweathered form.

The CEC (m.e/100g of colloid) was generally moderate throughout the profile, although values were elevated in the surface horizon by higher organic matter contents, and varied from 40 to 9 m.e/100g at the base. A strongly positive relationship existed between organic matter and CEC ($r = 0.744$) in all profiles (Table 5), expected with the larger adsorptive surface offered by organic material. The regression line is shown in Figure 20.

Calcium and exchangeable hydrogen appeared to be the major cations contributing to the CEC, although a moderate supply of potassium also existed. Magnesium was well supplied and these fertile red loams are also known to mineralize considerable proportions of nitrogen. The relatively low levels of sodium are as a result of dolerite being well buffered against a build-up of sodicity because of the moderate degree of leaching which takes place and good drainage characteristics. The Winterberg CEC values were higher, but compared well with those of the Nyika plateau. This could be explained by differences in parent material, the soils on Nyika plateau overlying granites and gneisses, somewhat lower than dolerite in their constituent bases.

PH was variable throughout the profile, and showed an inverse relationship with organic matter ($r = -0.625$, Table 5). The regression line is shown in Figure 21. The profiles were all moderately acidic, indicative of

well-drained and well-aerated porous soils. The pH ranged from 4.59 to 6.89, and tended to be most acidic in the uppermost 5cm, possibly due to the decomposition of litter incorporated in the topsoil layer. The variability of pH throughout the profile was not as marked with depth as in the vleï stratigraphy, the coefficient of variation being 3%. The pH increased with depth through the subsoil (B) horizon, the highest values (although still acidic) were found overlying weathered parent material. Dolerite is known to be less acidic than other materials such as granite and basalt (Van der Merwe and Weber, 1965).

Dolerite, of basic origin and generally fairly high in major bases, gave rise to soils of moderate base saturation. Base saturation was variable in all soil profiles, being highest at the base and decreasing towards the surface. This was expected as the number of available exchange sites will increase with increasing organic matter. The surface layer is also the zone of most intensive leaching or eluviation, bases being carried into the illuviation zone in the subsoil. As a result of this process, a higher number of exchange sites are left unoccupied by available bases (calcium, magnesium, potassium and sodium), giving a lower percentage base saturation.

6.1.6 PARTICLE SIZE VARIATION

The Winterberg soils were sampled to enable particle size determinations to be made throughout each profile. The topsoil or A horizon of these soils was a reddish-brown loam overlying a red blocky structured clay in the B horizon. Such soils, developing from parent material of inherently fine-grained nature, consist of weathering products composed primarily of clay-sized minerals, montmorillinite and illite being the most common clay types. Clay percentages in these soils varied from 41% to 68% for the topsoil and subsoil horizons respectively. It was expected that clay would increase with depth into decomposing parent material (Figures 18 and 19).

Clay was also found to have a fairly significant inverse relationship with organic matter ($r = -0.632$). This was expected with root densities decreasing towards bedrock, and an increase in organic litter occurring in surface layers (Table 5). The sand fraction consisted mainly of fine and medium grained particle sizes, with only negligible amounts of coarse sand in the samples (Appendix 1).

The relationship between clay and organic matter for profile 7 (Figure 19), situated in a cultivated land alongside Salisbury vleis, had a mean correlation coefficient of $r = -0.590$. This relationship was of low statistical significance. This could possibly be explained by past cultivation of this field resulting in the profile being overturned and disrupted by ploughing practices. If this is true, the clay percentages throughout the profile would have been altered due to the intermixing of subsoil (high clay %) and topsoil (lower clay %) horizons.

The physical properties of these soils include high infiltration rates, high available moisture capacities and good drainage, further improved if found in an upslope position. Even with the continued drought in the area, soil moisture held during winter months was above 17%. These characteristics are important in that these soils provide almost all the mineral matter available for vleis sedimentation. It is thought therefore, that if the various mineral characteristics of some of the sediment strata are similar to those of the adjacent soils in textural definition, it could be assumed that the sediments are, at least in part, derived from these adjacent soils and that infill has occurred from external erosional processes at some stage in the past.

6.2 VLEI SEDIMENT CHARACTERISTICS AND STRATIGRAPHIC CHANGES

In an attempt to outline past environmental conditions for the Winterberg area, the sediments of Dunedin,

primarily, and Salisbury vleis were intensively sampled using a gouge auger. Several chemical and physical analyses made on these samples, using the techniques described in Chapter 5, led to the identification and description of the characteristics of the various sediment types found.

6.2.1 SEDIMENT COLOUR

As in the adjacent soil samples, sediment colour varied with organic content and textural changes throughout the vlei profile, although far more markedly within the sediments. A dark black organic surface layer (2.5 Y 2/1 and 10 YR 2/1) overlay an amorphous peat-like layer, similar in colour to the surface organic material. Below this amorphous layer was a transition band of moderately high organic material, paler in colour due to the more sandy texture (2.5 Y 3/1; 5 Y 3/2 to 10 YR 3/1). The next layer was a very black oily organic band, similar to the true peats described from other sites. Dark clay minerals were also thought to add to the dark colour of this band which contained few recognisable plant remains. A dark brown loamy sand followed, of low clay and organic content (2.5 Y 3/2 to 10 YR 5/2). Nearer to bedrock, this sandy layer graded into inert mottled pale sands (10 YR 7/2 to 5 Y 7/1).

There was limited evidence of gleying in the vlei profiles. This suggested that stagnant conditions were uncommon and that water movement was continuous through the saturated mass, probably in concentrated flow patterns. Aerated veins sometimes occurred along decayed root passages, resulting in a mottling of bright orange (oxidised) colour development.

6.2.2 SEDIMENT DEPTH AND SURFACE GRADIENTS

The Winterberg vleis are elongated branched marshes in broad shallow valleys, longer than they are wide. The slopes in the valley bottoms are gentle, consistent with the undulating nature of the entire plateau surface. The longitudinal sediment surface gradient was uniform and shallow with a maximum fall of 2% from

the headwaters to the foot for Dunedin. The transverse profiles varied more, with falls of between 1% and 3% from the margins to the centre for both vleis. These surface gradients are shown in the stratigraphic diagrams (Figures 31 to 33). Small mounds were found to have formed from continually expanding populations of sedges and grasses, particularly as in the Salisbury vlei, where drying out had caused sediments between these mounds to shrink. The depth of Dunedin vlei sediments reached a maximum of 3,6m in the centre, while Salisbury reached only 2,6m, this being a broad shallow basin, limited by lithology.

The south-facing slope adjacent to Dunedin vlei rises sharply with a grade of slope of 16%, but other interfluves between the remaining branches and at Salisbury vlei are of more gentle gradient. One of the slopes adjacent to Salisbury vlei had been cultivated and ploughed, no doubt influencing the sediment availability to some degree. Other slopes were undisturbed by man, bearing dense grass/heathland vegetation cover, sporadically reduced by natural burning.

6.2.3 MOISTURE CONTENT

Sampling was primarily conducted during the winter months to facilitate movement through the dense vegetation, and, as a result, moisture contents could be rather conservative, particularly for the drier surface layers of Salisbury vlei. The sediments of Dunedin vlei are characteristically waterlogged all year round, and the moisture content of these samples varied from as high as 90% in the surface samples to approximately 50% in the sandier basal layers. Salisbury sediments were somewhat less moist due to the more freely draining nature of the vlei, which resulted in drying out surface layers during the drier winter months. The moisture content of these samples varied with depth from 68% in surface layers to as low as 25% at depth. During moister months, these values may be considerably elevated.

Moisture content with depth in the vlei profiles was higher where organic matter was higher due to the porous sponge-like nature of this material. High moisture levels also corresponded with higher clay layers, moisture being retained within these strata by low hydraulic permeability and the strong hygroscopic bonds between water molecules and clay surfaces. In general, moisture content was very high throughout the sediments when compared to the adjacent soil profiles.

6.2.4 ORGANIC MATTER

6.2.4.1 Lateral variations with distance.

Organic matter, being that fraction of the sediments vital for radiocarbon dating and a direct sediment input from vlei vegetation, was thought to be one of the most important characteristics from which to determine environmental change. Variations in organic matter were marked throughout all the vlei profiles. Organic matter increased laterally from the margins to the centre and longitudinally from the headwaters to the foot.

The histogram (Figure 22) shows decreasing mean organic matter percentages with distance away from the foot of the vlei. This change, particularly in a lateral direction in both Dunedin and Salisbury, appeared to parallel changes in vegetation density which responded to the increasing moisture gradient towards the centre (Figures 23, 24). The percentage bare surface (in Dunedin, open water surface was used) near the margins varied between 15 and 30% per 1m^2 grid, and 0 to 8% in the centres, being the mean of twenty measurements made in each zone. These measurements indicated that plant density increased markedly towards the centre of the vleis. Grasses and sedges were thicker, taller and coarser, making movement through the vleis extremely difficult. Species richness was considerably lower in the Salisbury vlei. Although the number of plants per square metre compared well with Dunedin, vegetation tended to be almost monospecific, particularly in the

winter months. The more seasonal nature of waterlogging at this site appeared to reduce the number of hydromorphic species found within this vlei. Phragmites spp. were not found, the dominant species appearing to be tougher plants such as Carex spp. and Juncus spp., those better able to withstand drying out during winter months. The surface organic percentages of this vlei were slightly lower than Dunedin, supporting the hypothesis that organic matter is indicative of moister conditions.

6.2.4.2 Variations with depth (stratigraphic)

Organic matter varied markedly with depth down all the extracted cores. In both vleis, organic matter values showed a mean decrease with depth, varying from around 32% in surface layers to negligible percentages nearing bedrock. Both vleis showed highly organic material throughout the profiles with darkly coloured fibrous organic horizons interbedded with lighter sandy loam layers. In such valley bottoms, and especially where drainage was impeded, as in Dunedin, the sediments tended to be far deeper (>3.5m, only 2.6m at Salisbury) than at the margins (about 60cm) and consisted of a complex stratigraphy of colluvial, alluvial and accumulated organic matter.

Three forms of organic, peat-like sediments were found, shown most clearly in the stratigraphic diagrams (Figures 31 to 33). A surface fibrous layer (0 - 75cm) of low bulk density was composed of irregular but recognisable plant remains in various stages of decay. This overlay a second layer of amorphous sandy textured organic material of slightly lower organic content (75 to 200cm). This layer contained well decomposed organic matter. The texture was coarser than the clayey surface matrix. Lower in the profile (220 to 330cm in Dunedin; 180 to 220cm in Salisbury) was a black well-decomposed oily organic material of similar organic percentage to the surface horizon. This organic band most closely resembled a true sedimentary peat, situated low in the profile, in comparatively

deep water and being derived from plant material which was completely humified. These materials were olive-green to black in colour and high in organic matter.

In the Dunedin longitudinal (Figure 25) and transverse (Figure 26) transects, the mean organic percentages all showed a decrease with depth. Changes with depth for all Dunedin cores (longitudinal and transverse samples together) are shown in Figure 27. In the Salisbury vlei (Figure 28), organic percentages showed a similar trend, although the lower organic band occurred at shallower depth (200 to 220cm). The above-mentioned organic sediment types are thought to be representative of an environment when organic accumulation was favoured by moister conditions and concurrently higher vegetation densities.

Using the organic matter values, association analysis was used to statistically group samples according to their organic content, and correlation coefficients at ten degrees freedom. The resultant curves from this analysis were binomial parabolic distributions when plotted against distance across the two vleis (Figures 29 and 30). The mean organic matter value was superimposed on each of these curves of associated samples (where 50% of the samples lie above this value and the other 50% below). In both cases, samples with higher organic matter values (those which exceeded the mean or threshold limit) were all samples from the central region of the vlei, while those lying below this value towards the tail ends of the parabolic binomial curves, were from the margin zones. These curves indicate that where the degree of variability between samples is lowest, and the correlation coefficient highest, organic values were higher. Similarly, those samples, grouped using the highest correlation coefficients, also coincided with those samples from the central region of the vlei. If the respective vlei margins are visualised as being the y and z axes of these graphs, a trend of increasing

organic matter towards the central zone is clearly shown. The association of variables is more clear in Figure 30, where the highest organic matter values, and least variability between samples, results in a group of samples in the central zone of the vlei.

6.2.5 pH

The Winterberg sediments were characteristically strongly acidic, values varying from highly acidic (3.3) in the humifying organic layers and tending towards neutral (6.5) in the more sandy layers (Figures 25, 26 and 28). Acidity is typical of organic or peat-like sediments, invariably due to a higher percentage of hydrogen ions in solution, displaced from colloidal surfaces by hydrolysis, or microbial and organic decomposition (McLean, 1973). The more organic, and thus darker, sediment layers had a higher hydrogen ion concentration, associated with lower base concentrations. It was thus expected that the pH values would be highly acidic where organic matter was high. The significant inverse relationship has a mean correlation coefficient of $r = -0.72$ (Table 6). This relationship for the Dunedin longitudinal transect is shown Figure 34. The surface fibrous organic layer was most acidic in all cores, (mean pH = 4.31) and was of low bulk density due to the high proportion of undecomposed plant remains.

Organic matter showed large fluctuations with depth in all profiles and, as expected, the opposite trend was reflected in the pH values, changes in organic matter showing similar but opposite trends in pH for all profiles (Figures 25, 26 and 28). Sandier lenses, being more freely draining, were thought to be more strongly leached by percolating drainage water. This process would serve to remove organic particles and thus yield a slightly less acidic pH than the organic layers. The mean pH of the basal sands was 6.17, approximately 3 units higher than the mean of the organic bands.

6.2.6 BASE STATUS AND CATION EXCHANGE CAPACITY

Prior to describing the chemical characteristics of these sediments, a note on the distinction between total CEC and mineral CEC is felt necessary to forestall confusion between these terms. The total CEC refers to the CEC of the whole sample, including the very colloidal organic matter, and for this reason, these CEC values are far higher than the mineral CEC, which excludes organic matter. The total CEC is thought to be more representative of the true nutrient status of the sediments, as the organic content is a very important constituent in the determination of sediment fertility status and base saturation.

In both the Dunedin and Salisbury vleis, the chemical analyses yielded data on the major cations (Na^+ , Mg^{2+} , Ca^{2+} , and K^+), cation exchange capacity and base saturation. Where sediments had a high total CEC (those high in both clay and organic matter), base saturation was generally much lower. Base saturation was very weakly correlated to CEC, indicating that although the capacity for exchange was large, the number of exchange sites occupied was very low. That is, the large number of negatively charged sites available on organic and clay particle surfaces were poorly saturated. This was particularly true of the uppermost organic layer where loss of cations by leaching and surface water movement occurred the most, and where organic matter was highest. The inverse relationship between base saturation and organic matter was not statistically significant at the 5% confidence interval, in any of the core profiles, the mean correlation coefficient being $r = -0.415$ (Table 6). This is possibly due to the base saturation percentage being reliant on the organic matter and clay fractions occurring together, rather than on organic matter alone.

The base status was higher in lower layers, the number of sites being limited by reduced clay and organic content, sediments being composed primarily of

chemically inert sand particles (Figures 25, 26 and 28). In these layers, the majority of available charge sites are filled by the calcium, magnesium, potassium and phosphorous ions leached from higher layers and existing in solution in seepage water. Generally, base saturation was fairly low, available charge sites being occupied by hydrogen ions or other toxic elements such as aluminium, particularly where the total exchangeable macronutrients was low.

Cation exchange capacity decreased with depth, approaching zero at the basal sandy beds (Figures 25, 26 and 28). This was expected as these sands were virtually inert for cation exchange, having a limited number of exchange sites. The total CEC (of the whole sample) was found to have a highly positive relationship with organic matter ($r = 0.915$) and fairly strong positive relationship with clay ($r = 0.67$) (Table 6; Figures 35 and 36). Both these relationships were significant. This follows for the extremely high colloidal nature of both clay and organic materials. The clay and organic matter formed a minimum of 65% of material within the organic bands. The main contributors to the CEC of these sediment bands were exchangeable hydrogen, very high due to release during decomposition, and calcium. The chemistry of the mineral constituents of the stratigraphy was greatly influenced by the nature of the bedrock from which they were formed. Bases of calcium, magnesium and potassium, being high in the adjacent soils weathered from dolerite, were leached and washed into the vlei system.

6.2.7 TEXTURAL VARIATIONS

Changes in texture were marked throughout the vlei profile with depth and from margins and headwaters to the centre. These variations were thought to reflect to some degree the nature of depositional processes which operated in the vlei throughout the late Pleistocene and Holocene. There appeared to be two zones of textural variation within the vleis, differences between the two being quite marked. In

both vleis, one zone, the margin and headwater areas, varied markedly from the second central zone. The margin cores consisted of black fibrous organic matter overlying amorphous organic material, which in turn graded into pale sands (Figures 32 and 33). The bottomset sandy layer appeared to be deeper near the margins, and was certainly paler in colour than the sandy lenses found nearer the centre. Towards the headwater zone (Figure 31), a similar sandy deposit existed, possibly laid down by heavy sediment loads during storm flows. In the central zone, the stratigraphy was more complex and showed a great variation between successive strata. The organic and clay contents were far higher and sandy lenses were narrower and less prominent.

The distinction between the stratigraphy of the margins and headwater zones, and that of the centre can be made for almost all analyses and in all core samples when compared with distance (Figures 22, 23 and 24). In all cores, organic matter and clay showed an increase in % towards the centre, while the total sand and pH showed a decrease. A positive relationship existed between organic matter and clay content, both being higher in the centre than at the margins. This trend could be explained by the fact that during moist phases, dense vegetation growing in the vlei would have trapped sediment more efficiently, and resulted in a loss of transporting ability. In addition, the sediments entering the vlei would have consisted of finer particle sizes due to the good cover on adjacent slopes, erosion being reduced by vegetation.

Variations in clay content with depth was confusing, as clay decreased in occurrence to the base of the profile in all cores, following a similar trend to organic matter (Figures 25, 26 and 28). Being a product of weathering dolerite, it was expected that clay would increase nearing the base. The opposite occurred, fine and medium grained sands forming the bottomset bed. An explanation for this could be that the majority of

material in the vlei is from external sources and not formed in situ from weathering parent material. Hand augering with a gouge auger also did not allow sufficient penetration to sample through this coarse sandy layer. Solid bedrock may thus not have been reached due to auger limitation, resulting in samples consisting of the coarse sediment brought into the vlei by erosion.

6.2.8 SEDIMENT STRATIGRAPHY

From the textural analyses of sediments, there appear to have been two phases of clastic, or mineralogic, deposition, separated by two main phases of organic accumulation. Dunedin stratigraphy (Salisbury, although very similar, was somewhat shallower) consisted of various sediment layers and is described in sequence below (Figure 31). Inert pale grey sands (360 to 345cm) formed the basal beds, possibly overlying weathering parent material, or even another set of sediments. Above this layer was a dense brown sandy loam layer with negligible clay and organic contents (345 to 330cm), followed by an oily black peat-like material of very high organic matter (330 to 285cm). This black layer contained no recognisable plant remains, and was very high in clay content. The radiocarbon date from this layer confirmed the age of $12\ 500 \pm 160$ years BP (Lab. no. Pta-4207), the first major sign of rapid organic accumulation for the late Pleistocene at this site. Above this peat-like layer was a transition layer of moderate clay and organic content, and an increased fine sand content (285 to 200cm). In this layer the mineral components formed the major proportion of sediment, possibly indicative of an environment less favourable for organic accumulation when compared with the highly organic layer found beneath it. A radiocarbon date for a sample of this sediment type at 220cm depth, where the organic matter was once again higher, gave a date of $7\ 990 \pm 200$ BP (Lab. no Pta-4318). This layer graded into an amorphous organic band at a depth of 200 to 75cm. This amorphous band consisted of very high

organic and clay contents and was topped directly by a shallow fibrous organic layer (0 to 75 cm), forming the modern day organic accumulation. The radiocarbon dates are of utmost importance in elucidating an accurate age and chronological history of the deposits in vlei features. The number of absolute dates available was limited by the relatively low organic contents (when compared with true peats) and danger of contamination during sampling, but the two dates above confirmed the age of the sediments and that they were undisturbed layers following in chronological order.

6.3 COMMENTS

The vleis of the Winterberg show two main stratigraphic zones, a margin zone and a central seepage zone. The margins are comprised mainly of sandy material and less organic matter, while the clay fractions and organic matter percentages increase greatly towards the centre. This distinction is thought to be due to various processes operating in the vlei systems in response to the external environmental conditions. Analysis of the processes which are acting and have acted on the Winterberg vleis remains problematic due to the complex nature of their stratigraphy, reflecting not only changing environments in the catchment, but changes in vegetation and local hydrology as well. A simple explanation for the development of such sediment characteristics is unrealistic, and it is apparent that any interpretation should be of a multi-disciplinary nature. However, some conclusions from the analytical data have been drawn, supported by radiocarbon dates which facilitated the elucidation of the chronology of sediment accumulation, which represents the late Pleistocene and Holocene.

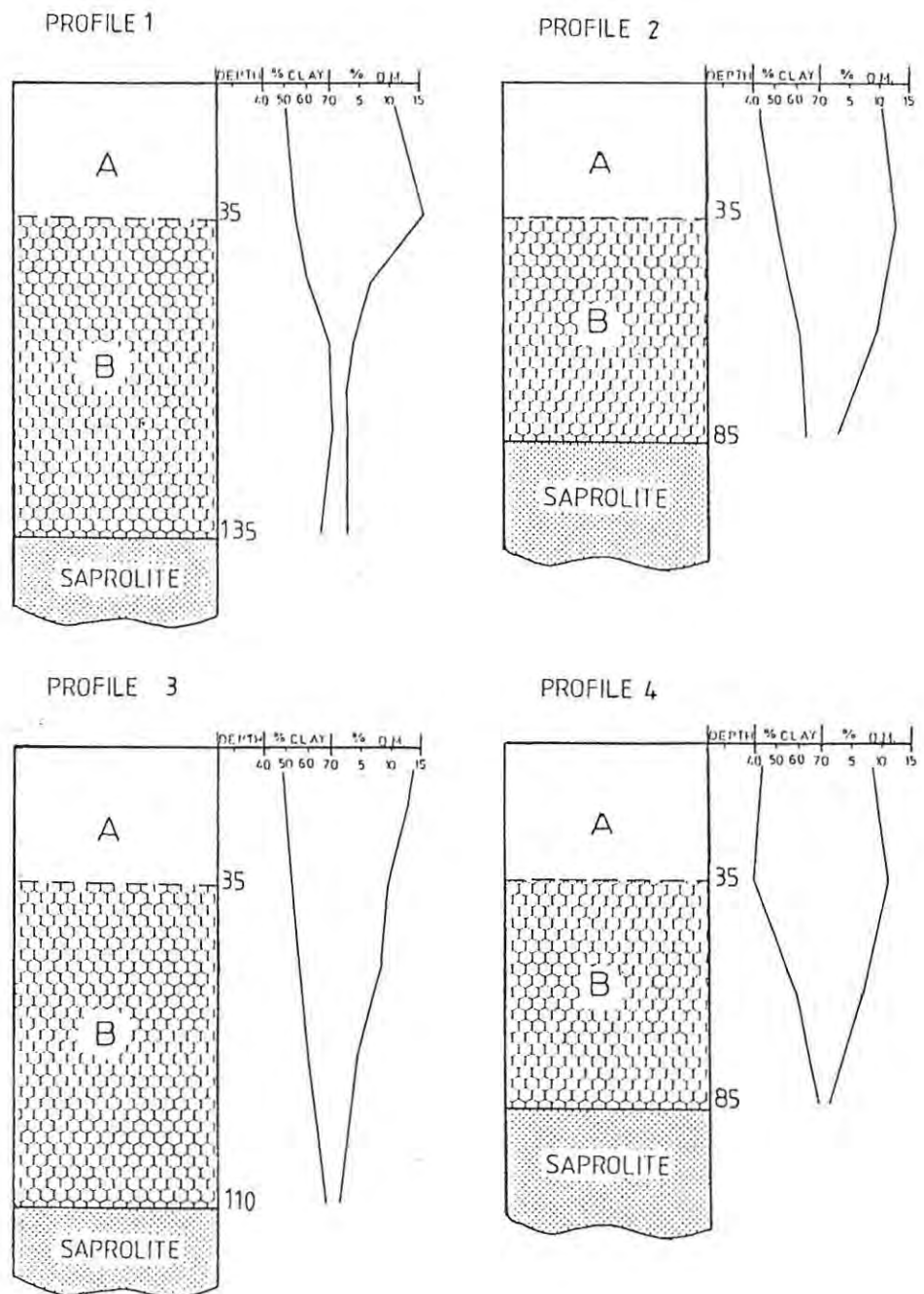


Figure 18: Soil profile characteristics of profiles 1 to 4 at Dunedin vleij.

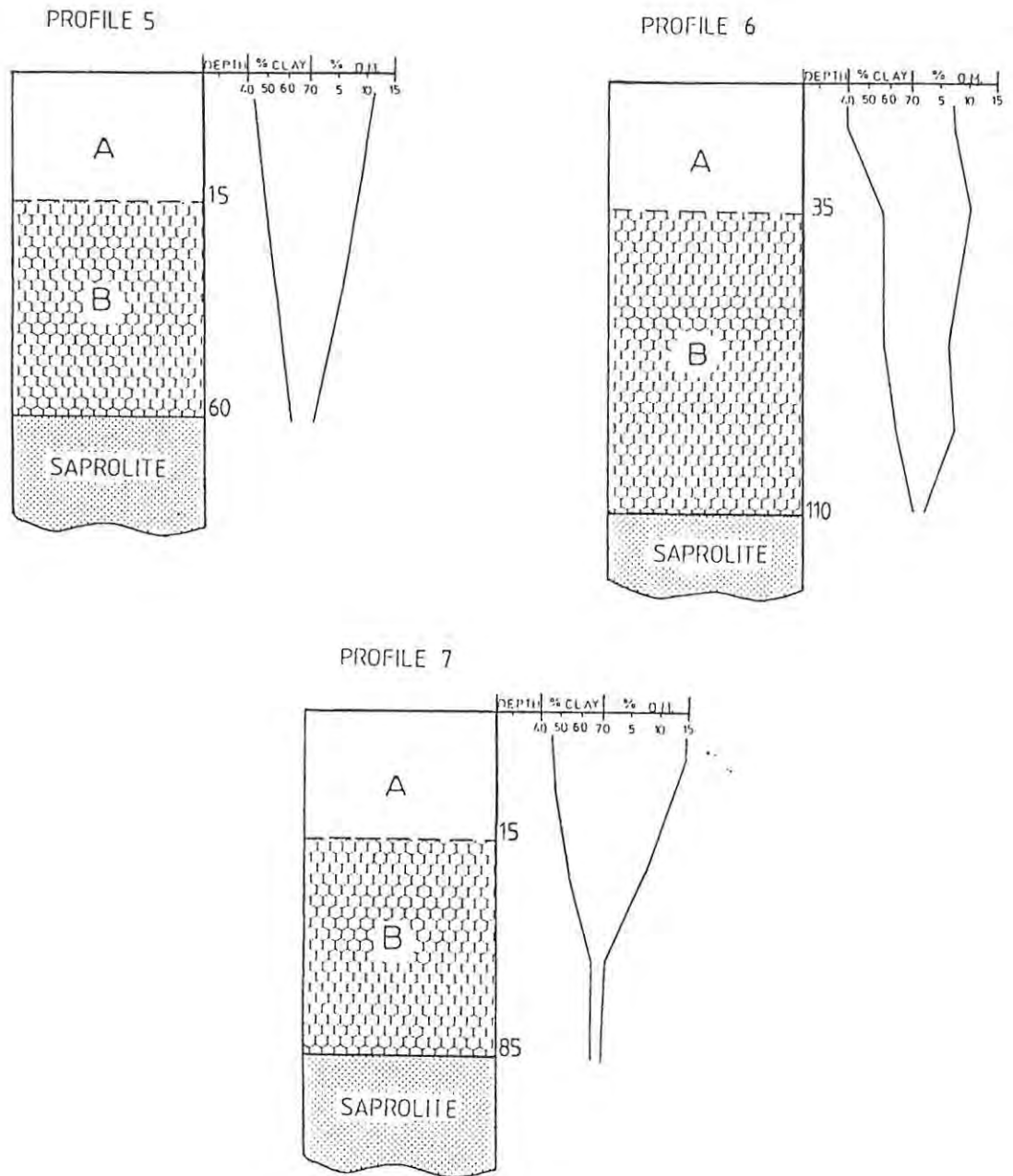


Figure 19: Soil profile characteristics of profiles 5 to 7.

Table 5. Correlation coefficients of soil variables.

Variable	Soil profile	O.M.	Significance @ 95% CI/10 d.f.)	Direction
1. pH	1	0,644	Fair	-(ve)
	2	0,608	"	-(ve)
	3	0,621	"	-(ve)
	4	0,639	"	-(ve)
	5	0,640	"	-(ve)
	6	0,629	"	-(ve)
	7	0,593	Low	-(ve)
2. CEC	1	0,743	High	+(ve)
	2	0,768	"	+(ve)
	3	0,741	"	+(ve)
	4	0,728	Fair	+(ve)
	5	0,702	"	+(ve)
	6	0,773	High	+(ve)
	7	0,755	"	+(ve)
3. Clay	1	0,683	"	-(ve)
	2	0,609	Fair	-(ve)
	3	0,676	"	-(ve)
	4	0,602	Low	-(ve)
	5	0,633	Fair	-(ve)
	6	0,628	"	-(ve)
	7	0,590	Low	-(ve)

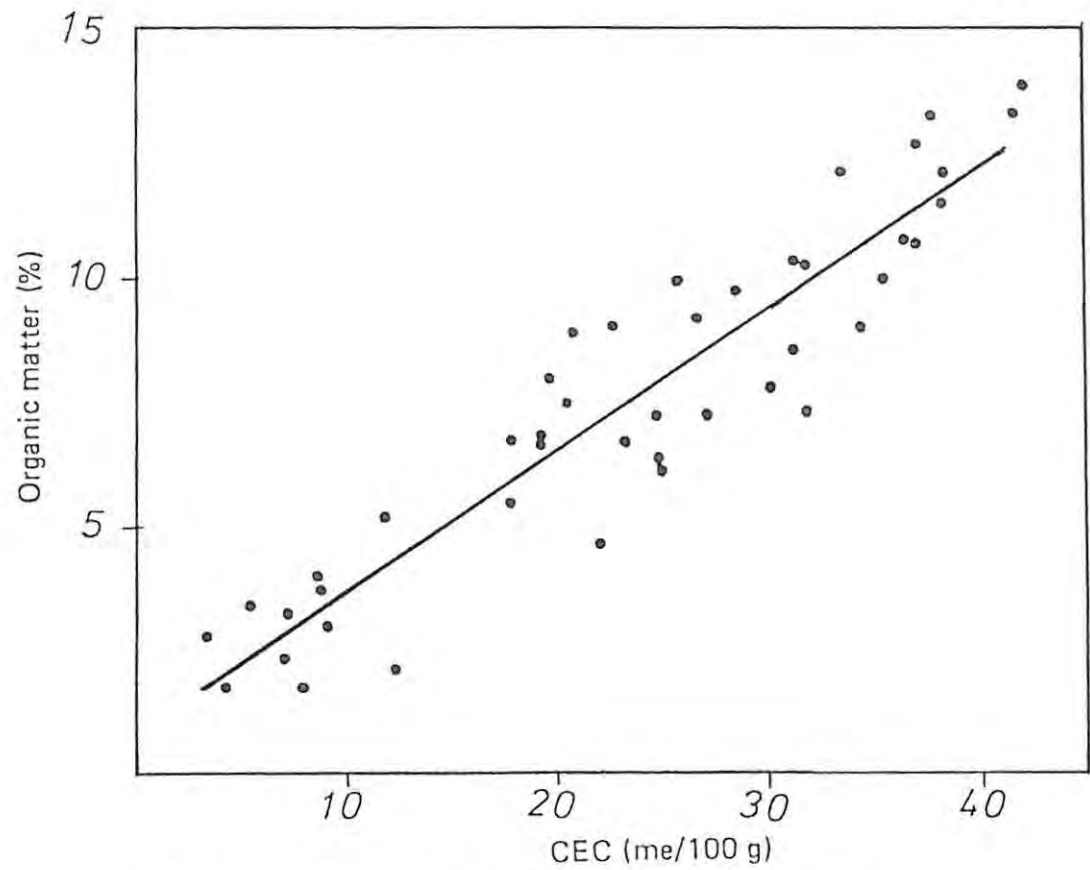


Figure 20: Regression line of soil CEC and organic matter (mean of 7 profiles).

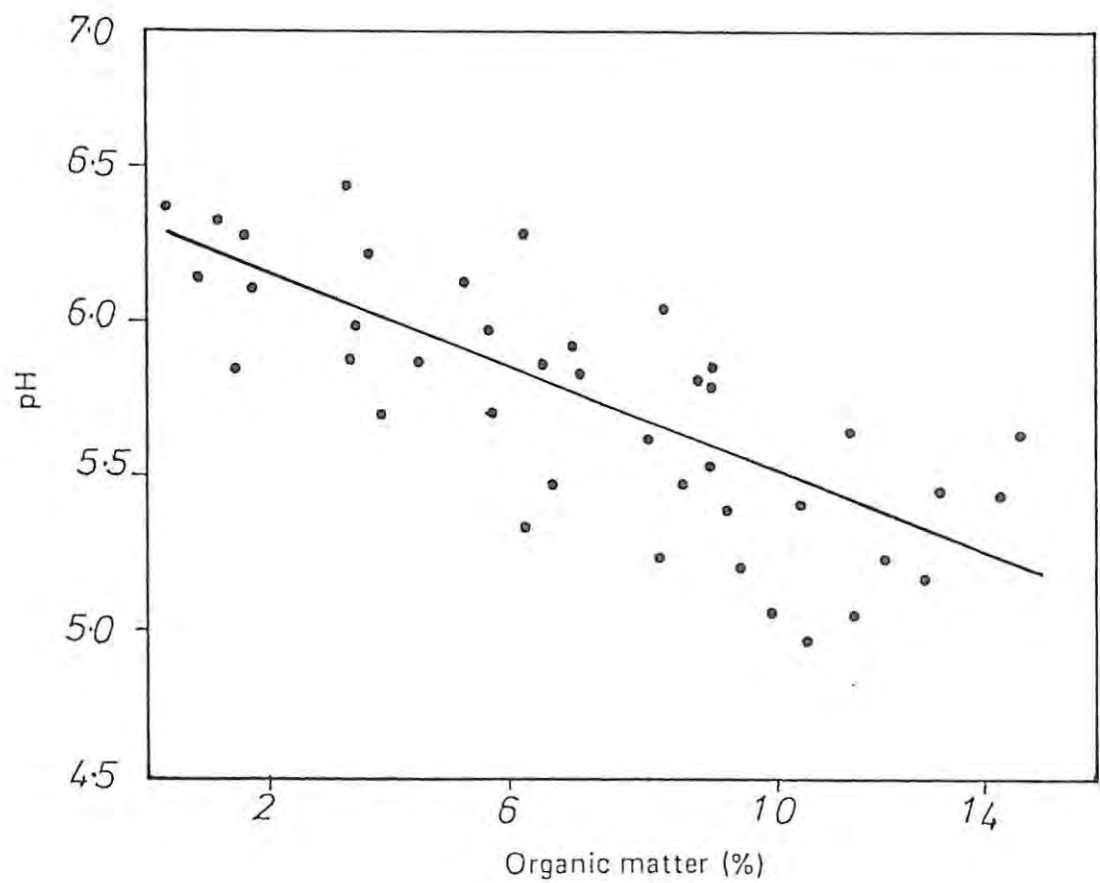


Figure 21: Regression line of soil pH and organic matter (mean of 7 profiles).

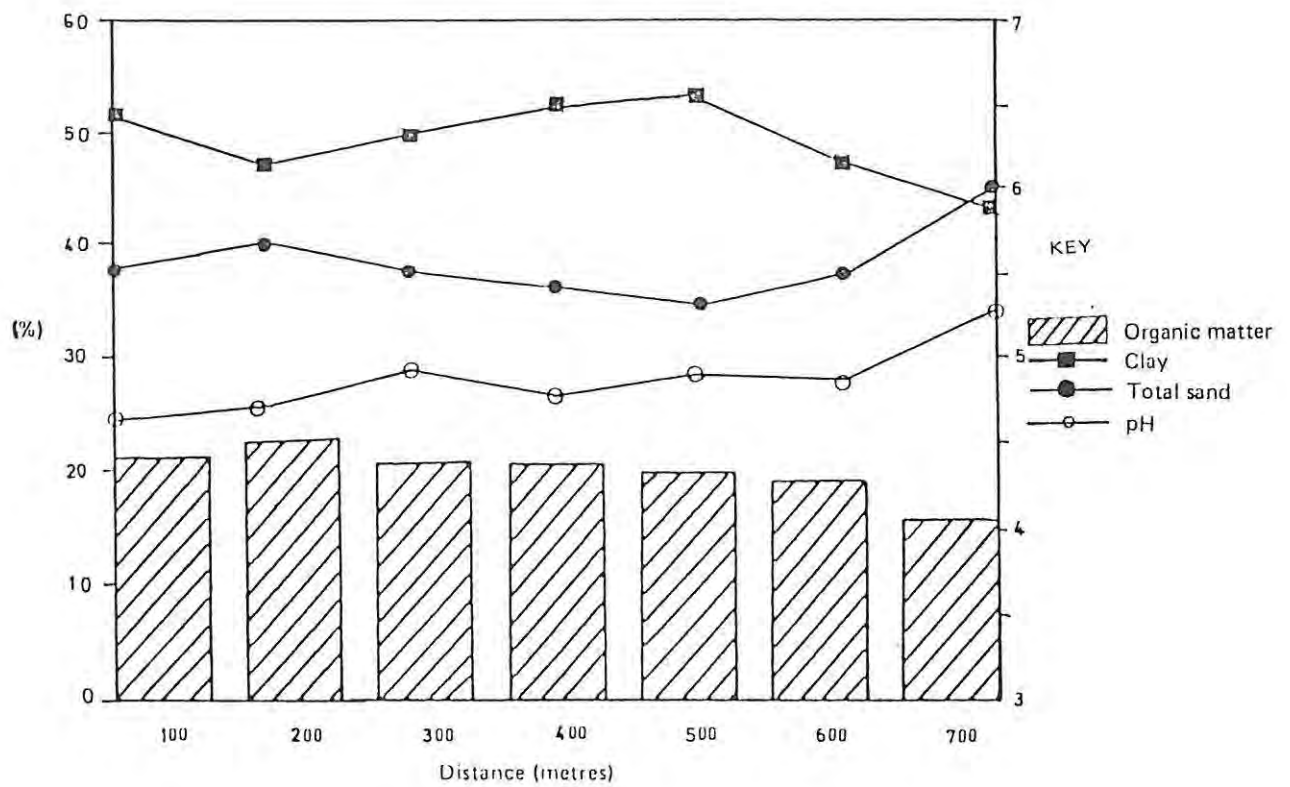


Figure 22: Changes in sediment characteristics with depth from the foot to the headwater zone of Dunedin.

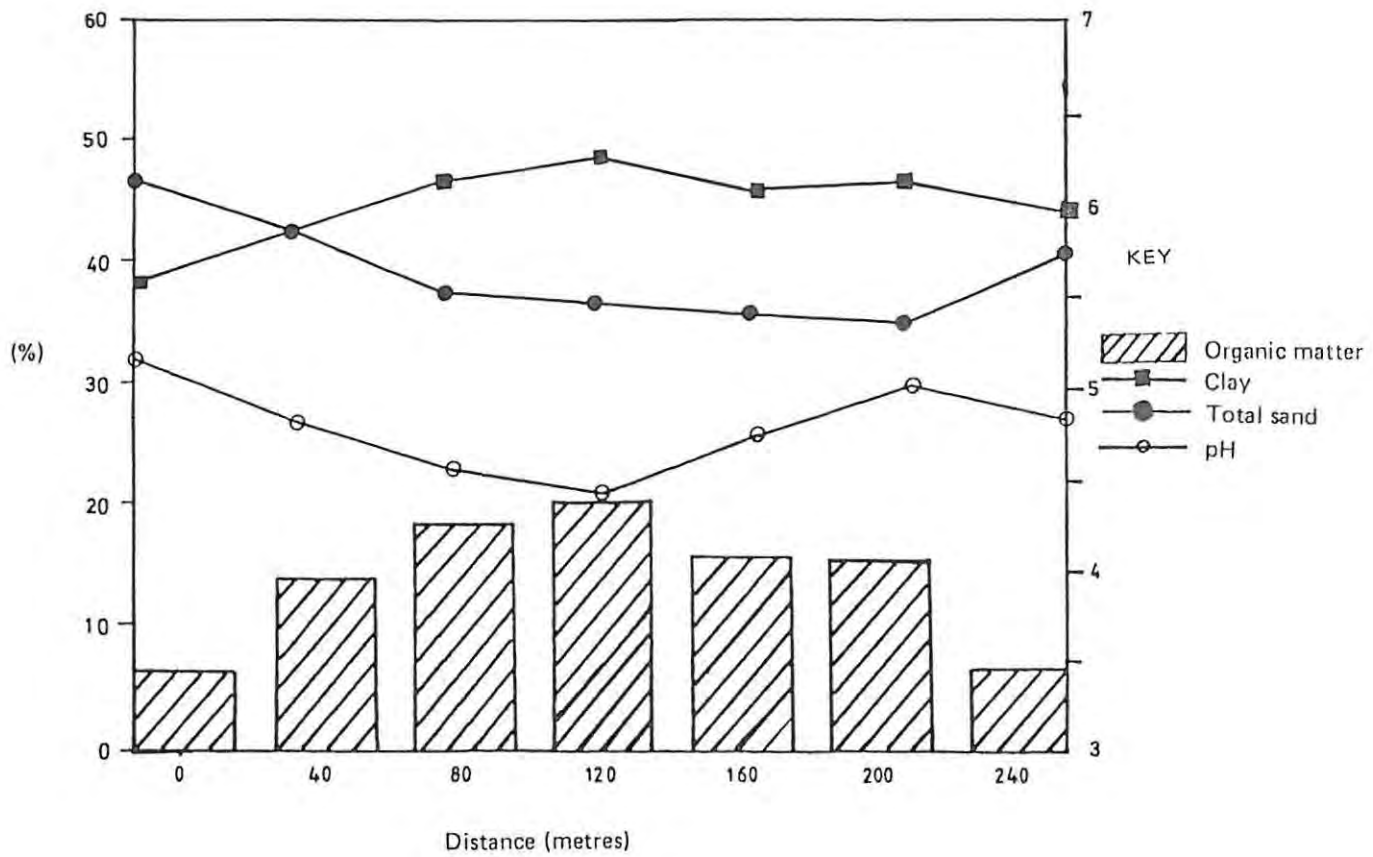


Figure 23: Changes in sediment characteristics with distance across Dunedin (mean transverse values).

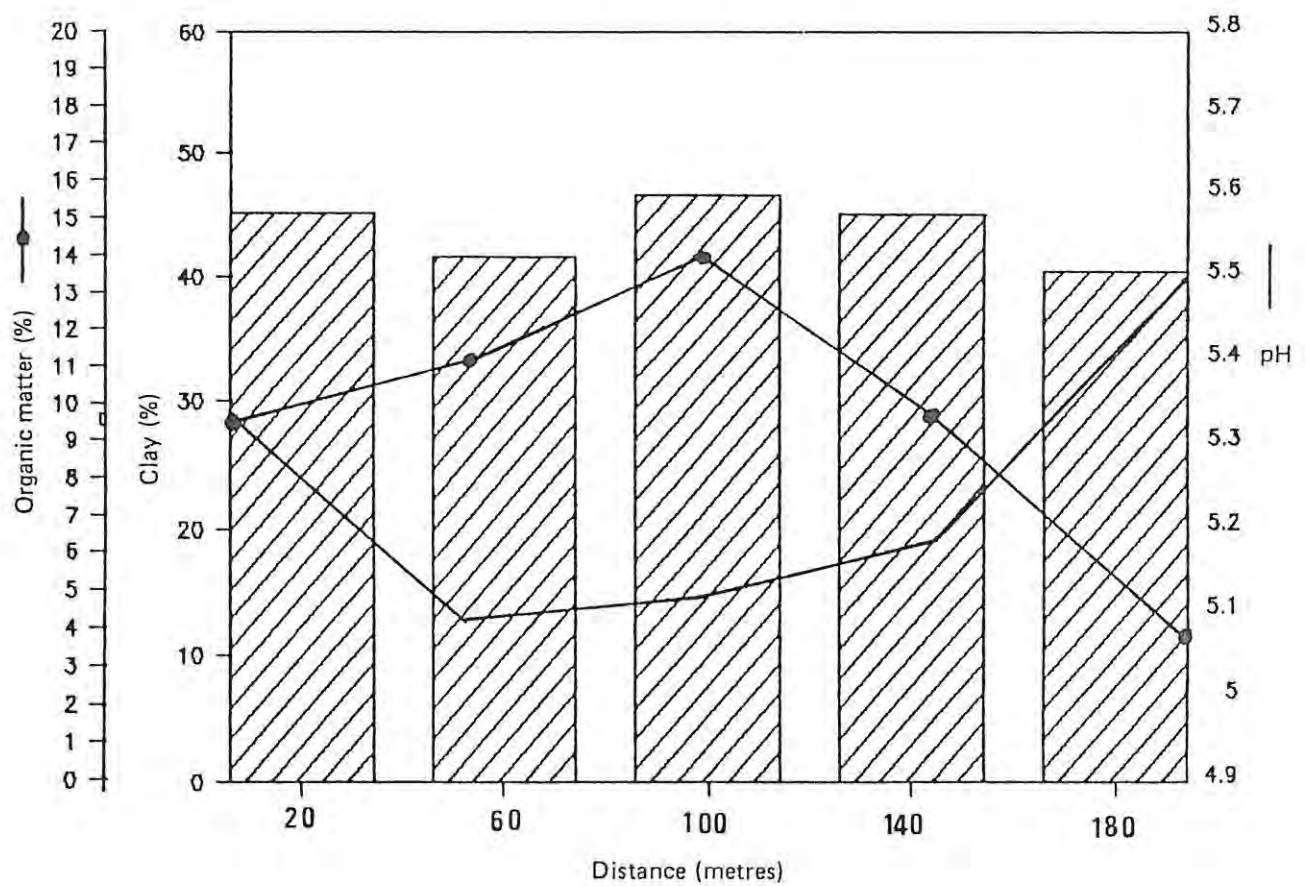


Figure 24: Sediment changes with distance along the Salisbury transverse transect.

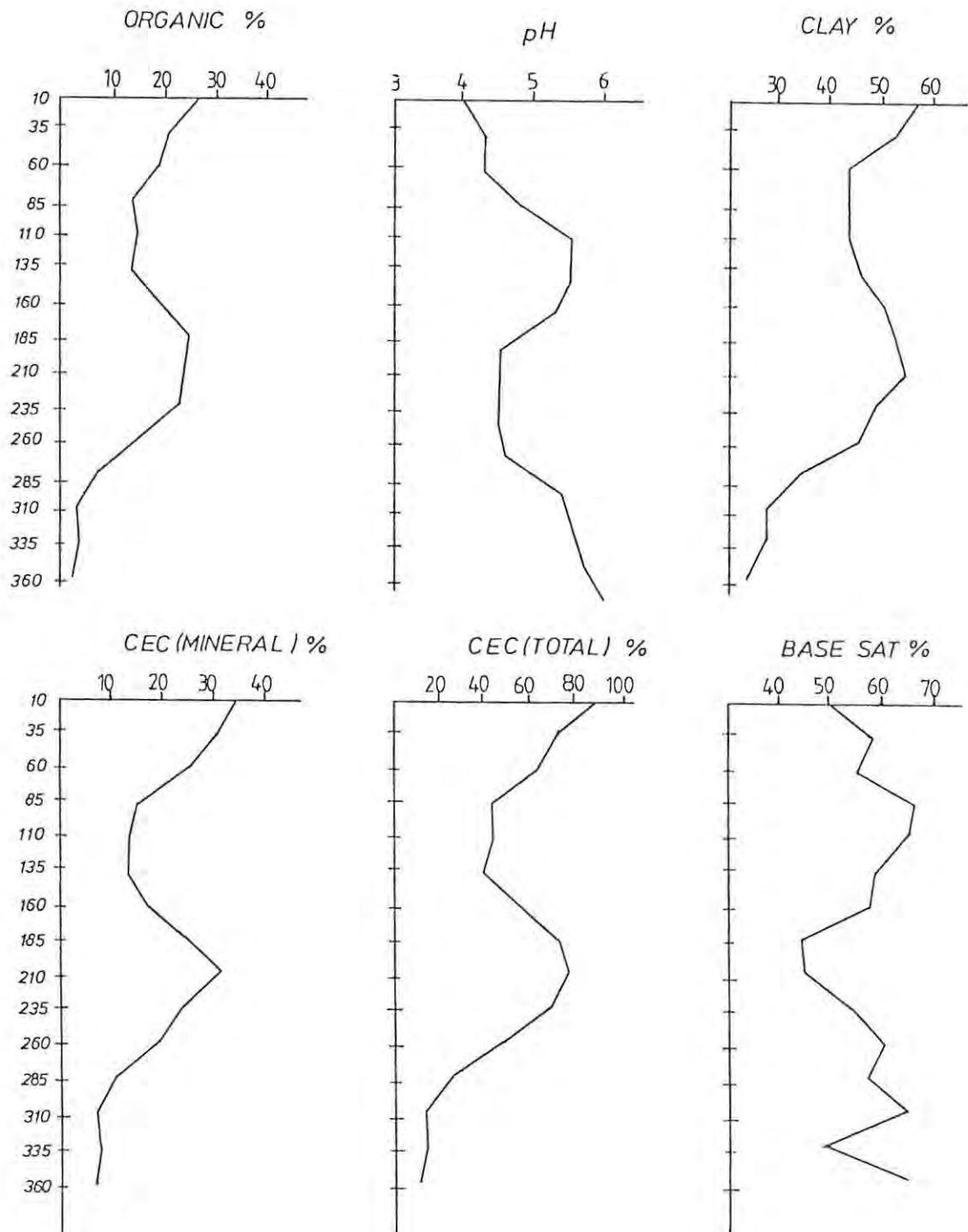


Figure 25: Changes in mean sediment characteristics with depth for Dunedin longitudinal transect.

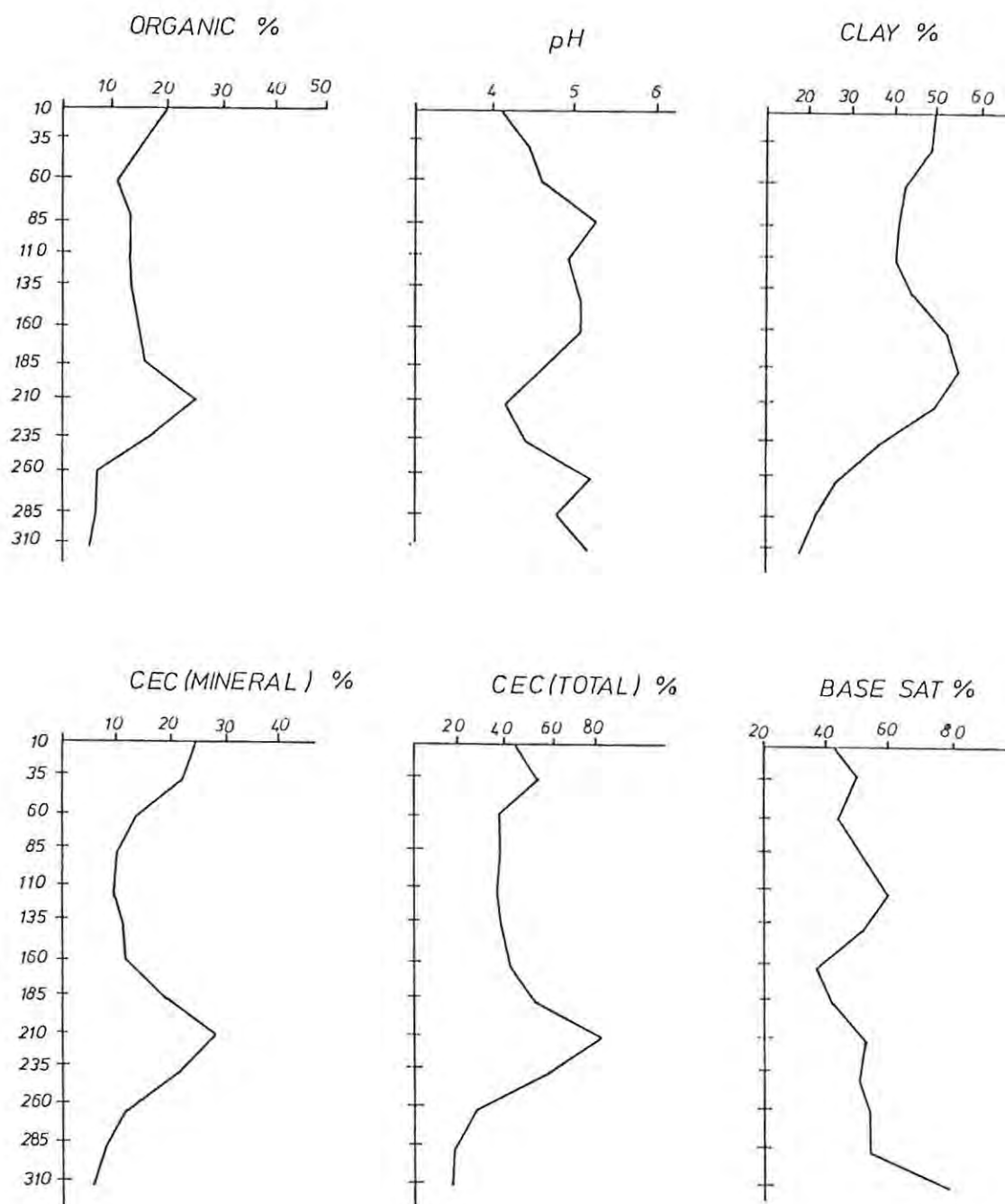


Figure 26: Changes in mean sediment characteristics with depth for Dunedin transverse transect.

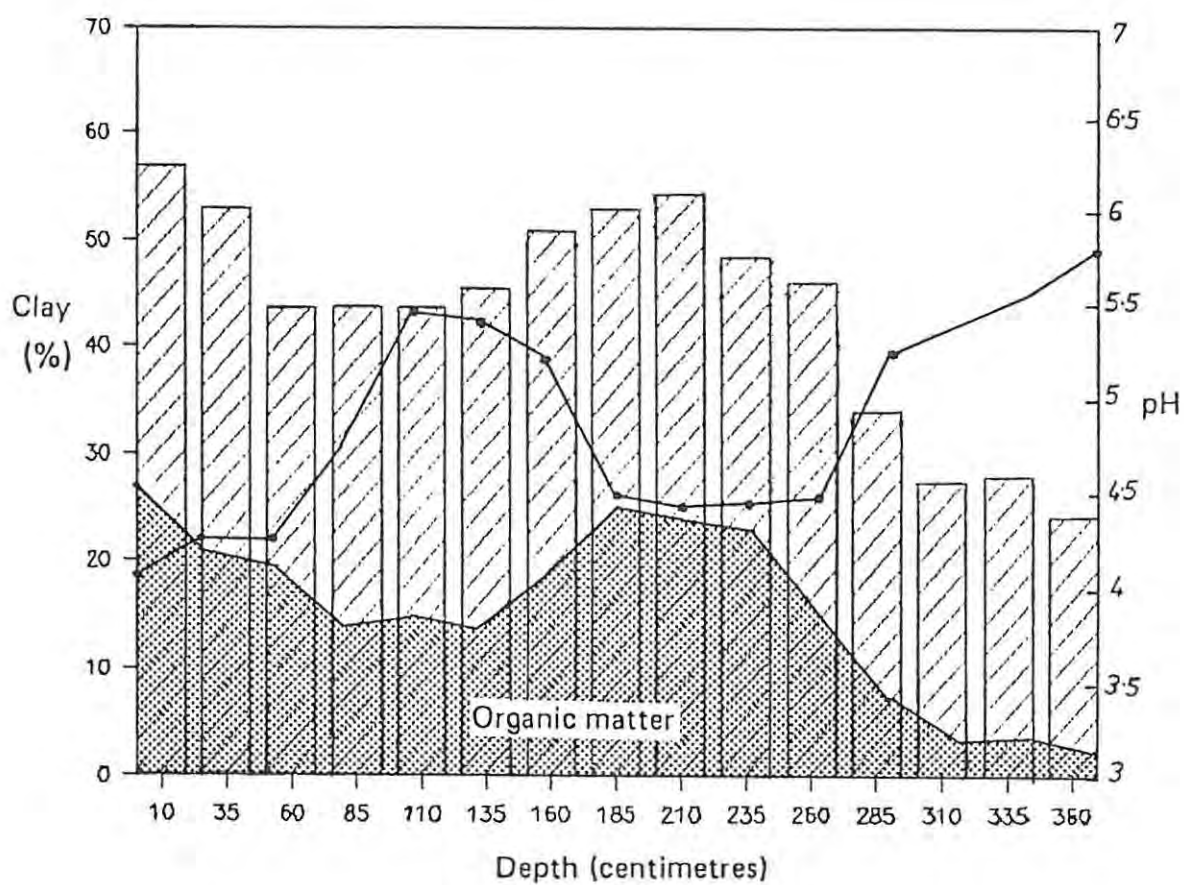


Figure 27: Sediment variations with depth for Dunedin vlei (mean of all core samples).

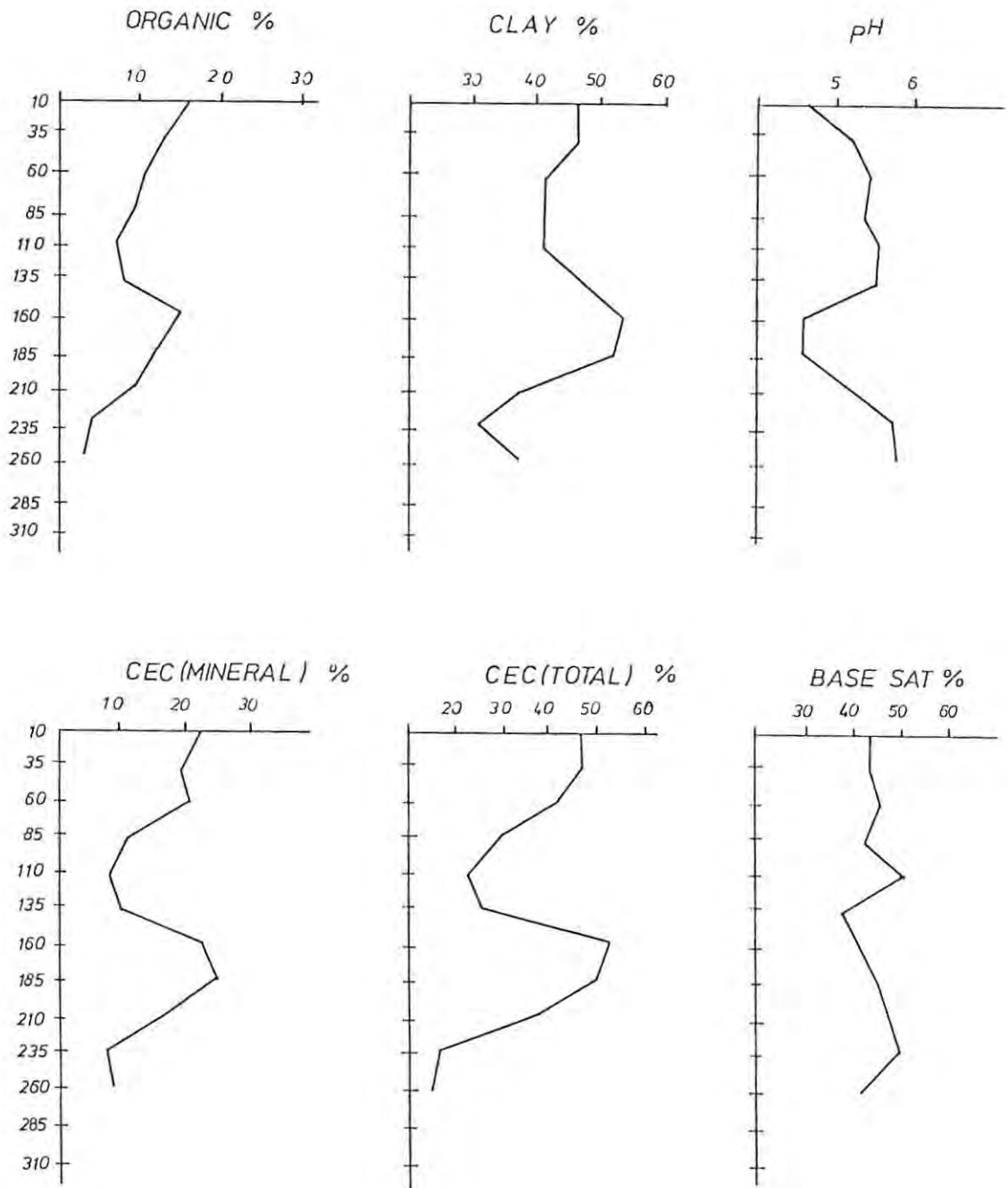


Figure 28: Mean sediment characteristic changes with depth for Salisbury vlei.

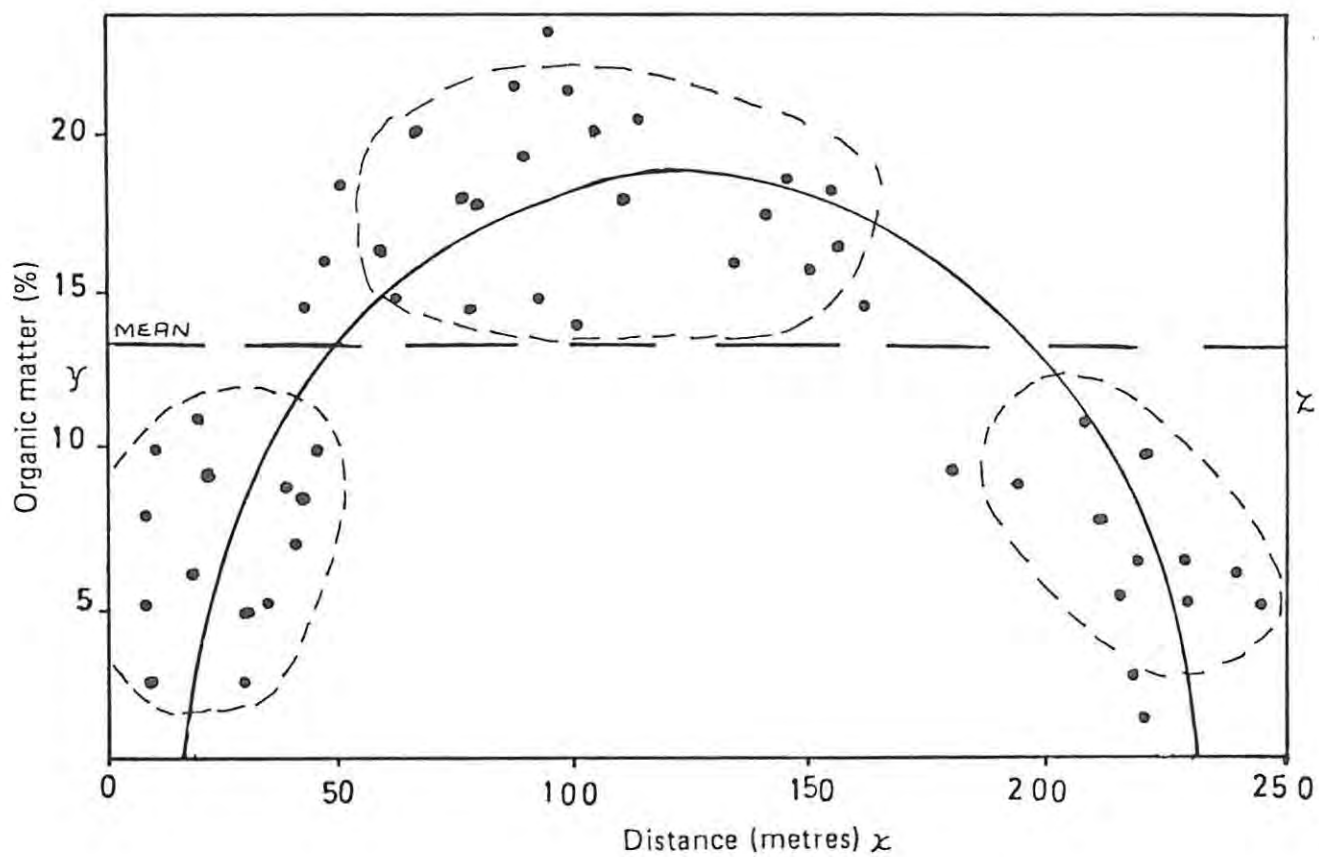


Figure 29: Association analysis showing parabolic distribution of organic matter values plotted against distance across Dunedin vlei.

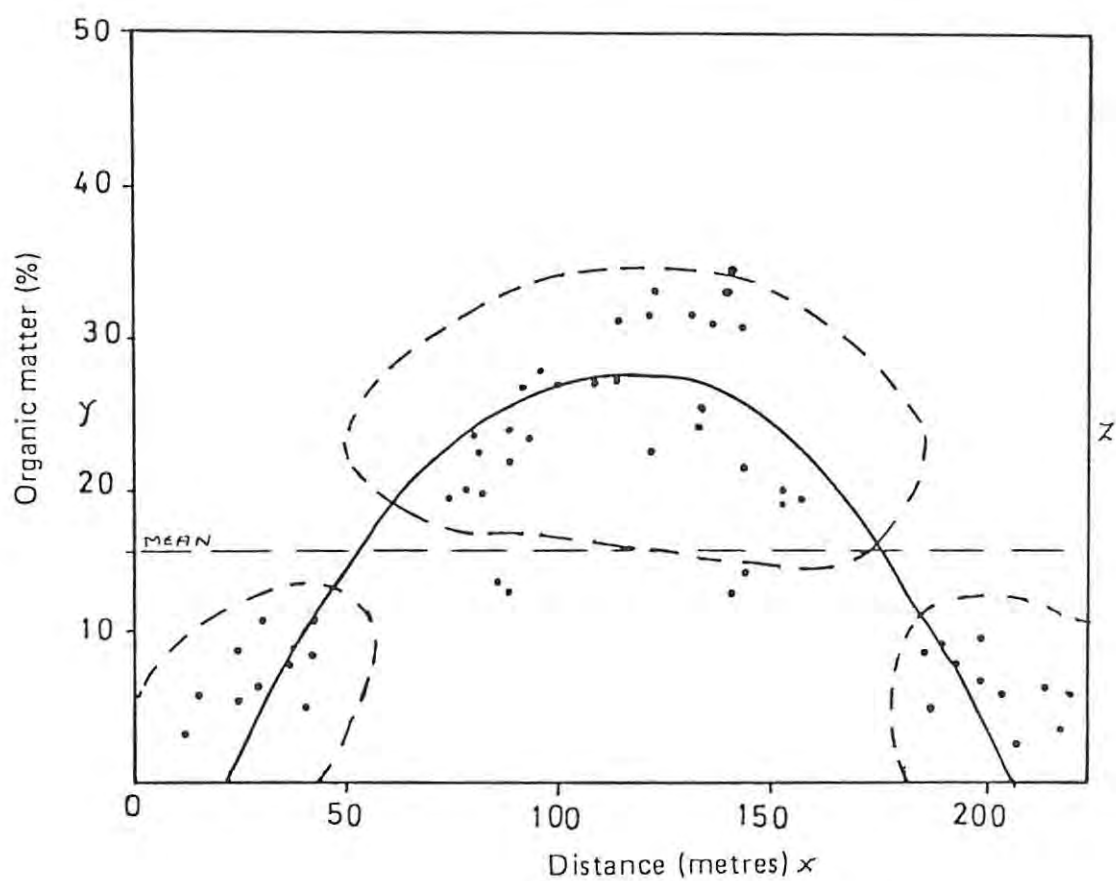


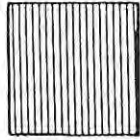
Figure 30: Association analysis and parabolic distribution of organic matter values across Salisbury vlei.

Table6: Correlation coefficients of selected variables.

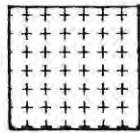
Variable	Transect	Organic matter	r - value	Significance (at 95% C.I/ 10 d.f.)	Direction
1. pH	DT	0,658		Fair	-(ve)
	DL	0,658		Fair	-(ve)
	DTV	0,694		Fair	-(ve)
	S	0,773		High	-(ve)
2. CEC(Min)	DT	0,698		Fair	+(ve)
	DL	0,697		Fair	+(ve)
	DTV	0,676		Fair	+(ve)
	S	0,596		Low	+(ve)
3. CEC(total)	DT	0,953		High	+(ve)
	DL	0,950		High	+(ve)
	DTV	0,955		High	+(ve)
	S	0,957		High	+(ve)
4. Clay	DT	0,693		Fair	+(ve)
	DL	0,741		High	+(ve)
	DTV	0,607		Low	+(ve)
	S	0,592		Low	+(ve)
5. Base saturation	DT	0,426		None	-(ve)
	DL	0,470		None	-(ve)
	DTV	0,447		None	-(ve)
	S	0,317		None	-(ve)
		CEC			
6. Clay	DT	0,683		Fair	+(ve)
	DL	0,728		High	+(ve)
	DTV	0,602		Low	+(ve)
	S	0,560		None	+(ve)

Where: DT = Dunedin Vlei - total no. of samples
DL = " " - longitudinal samples only
DTV = " " - transverse samples only
S = Salisbury Vlei - all samples

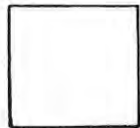
KEY TO STRATIGRAPHIC DIAGRAMS



Black fibrous highly organic material



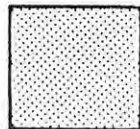
Dark amorphous organic material



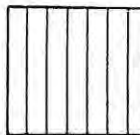
Sandy layer of low organic content



Black oily peat-like band



Pale basal sands



Transition band of sandy material
of moderate organic content.

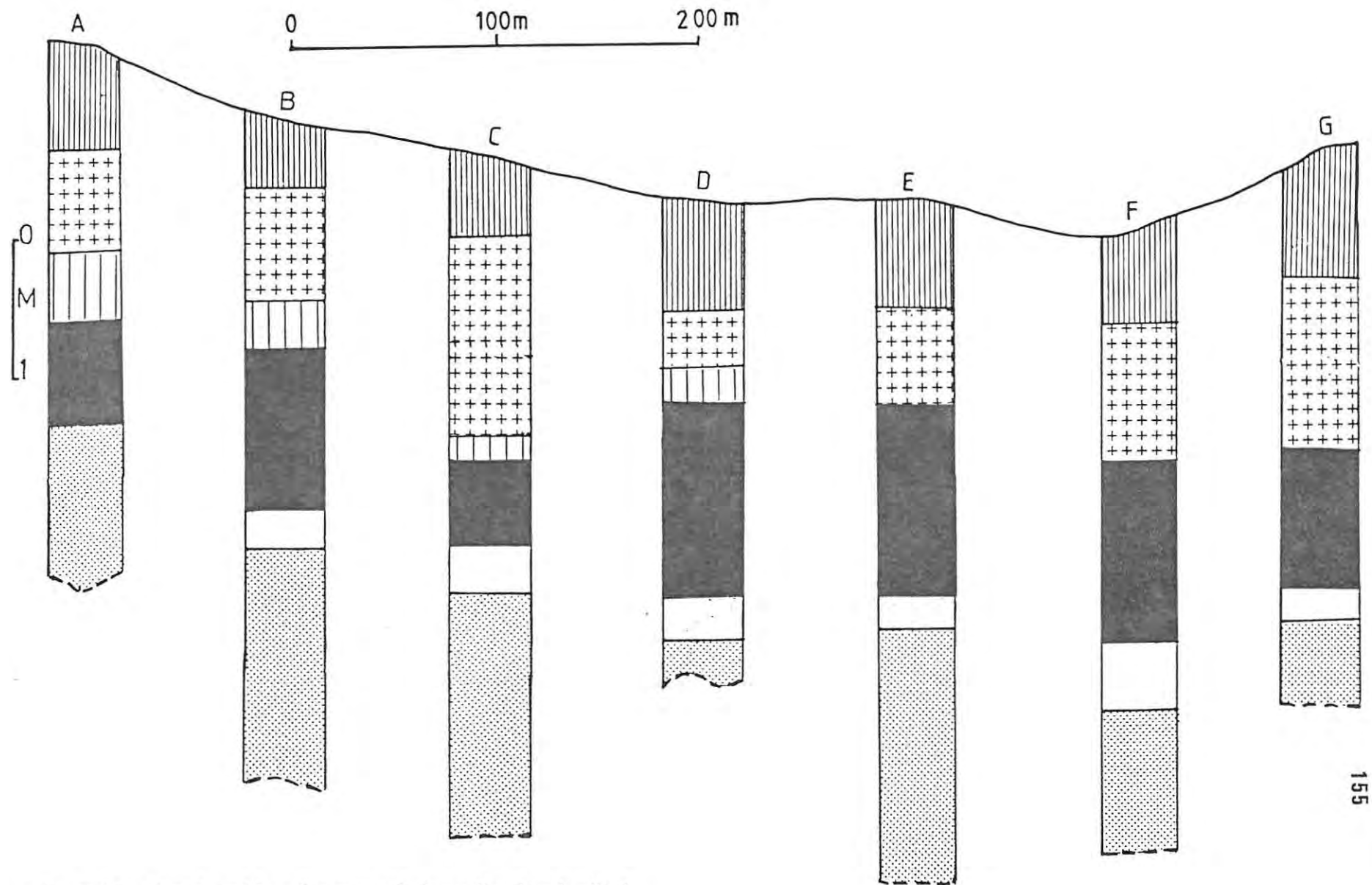


Figure 31: Stratigraphic diagram of Dunedin longitudinal section.

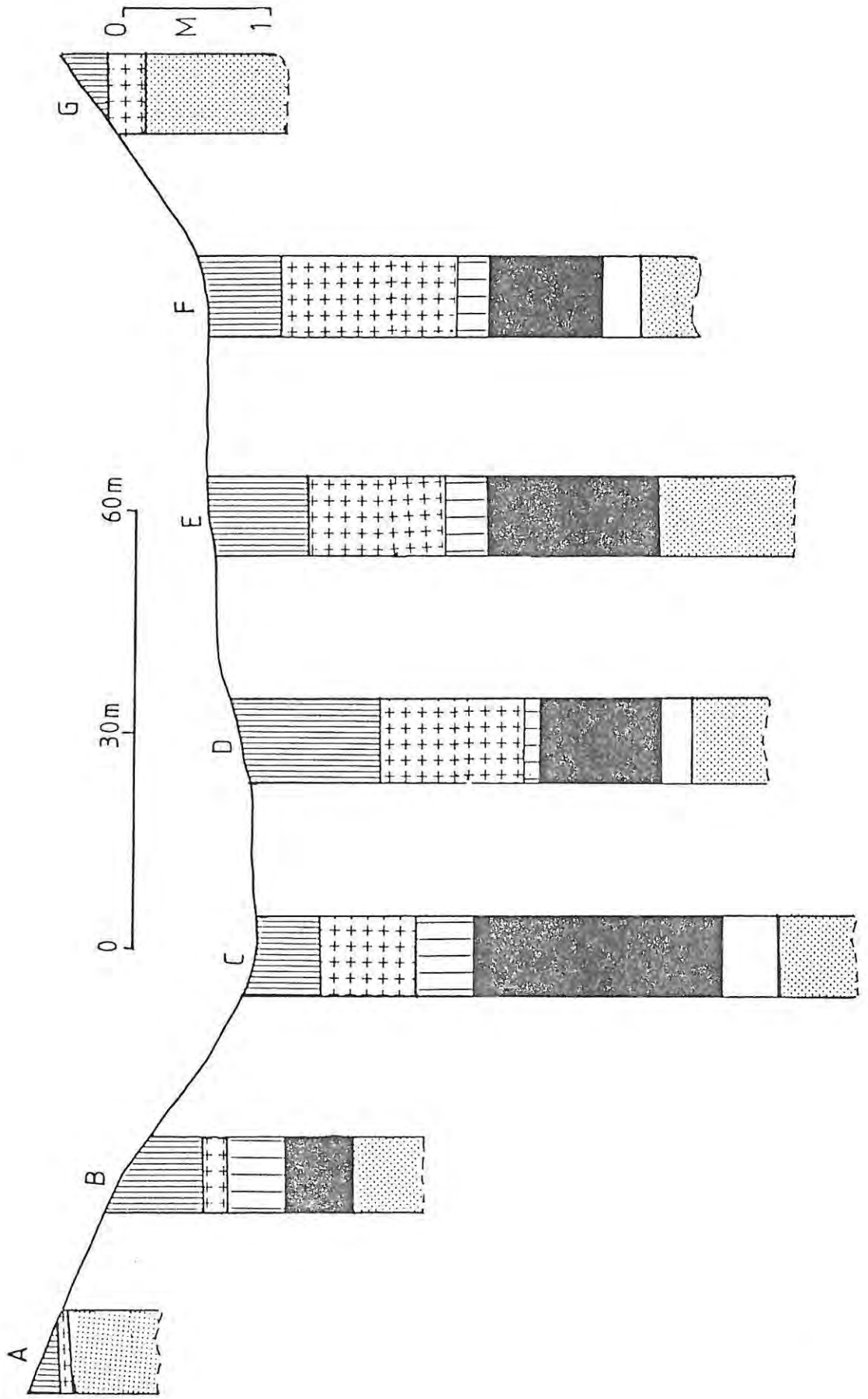


Figure 32: Stratigraphic diagram of Dunedin transverse section.

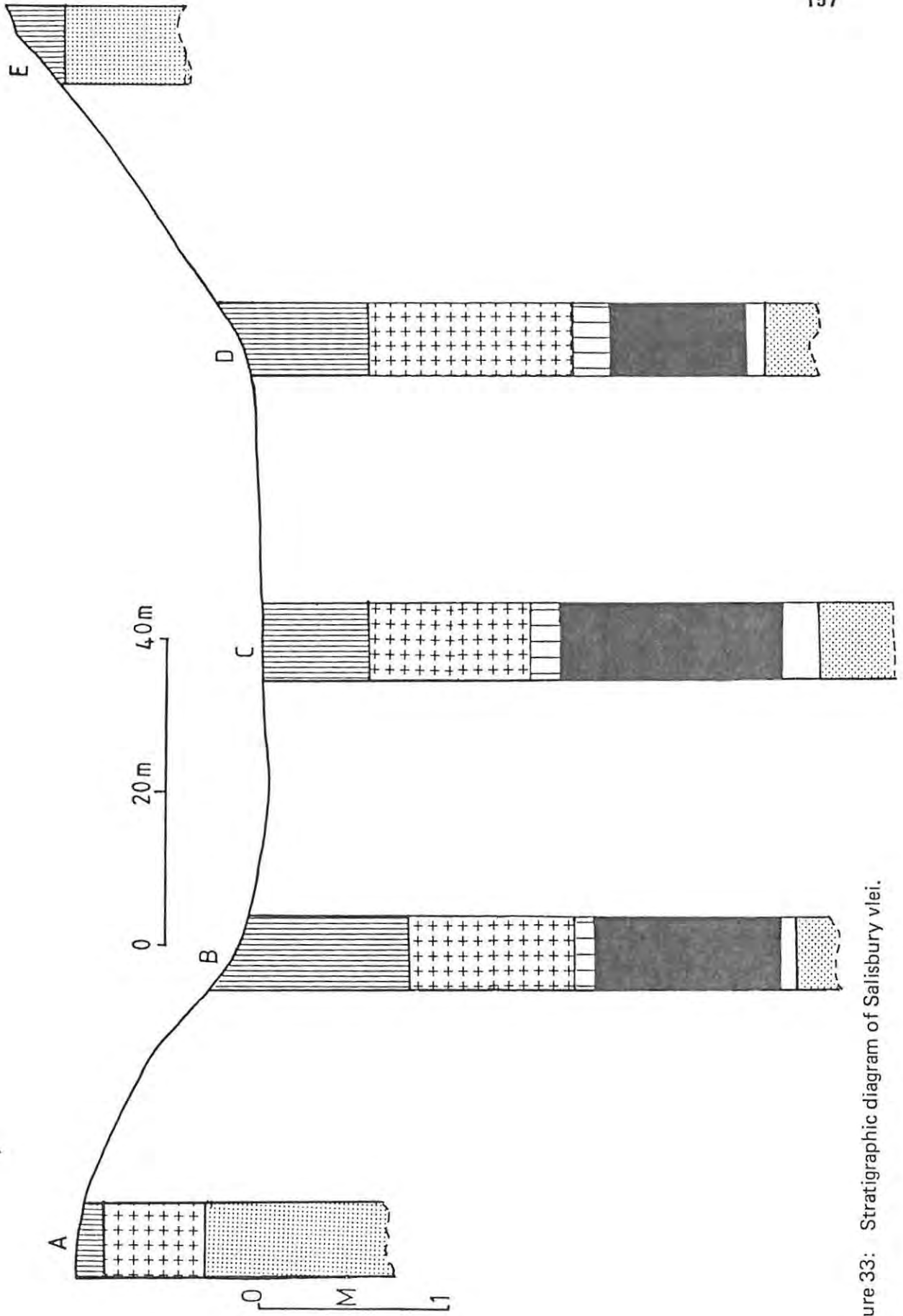


Figure 33: Stratigraphic diagram of Salisbury vlei.

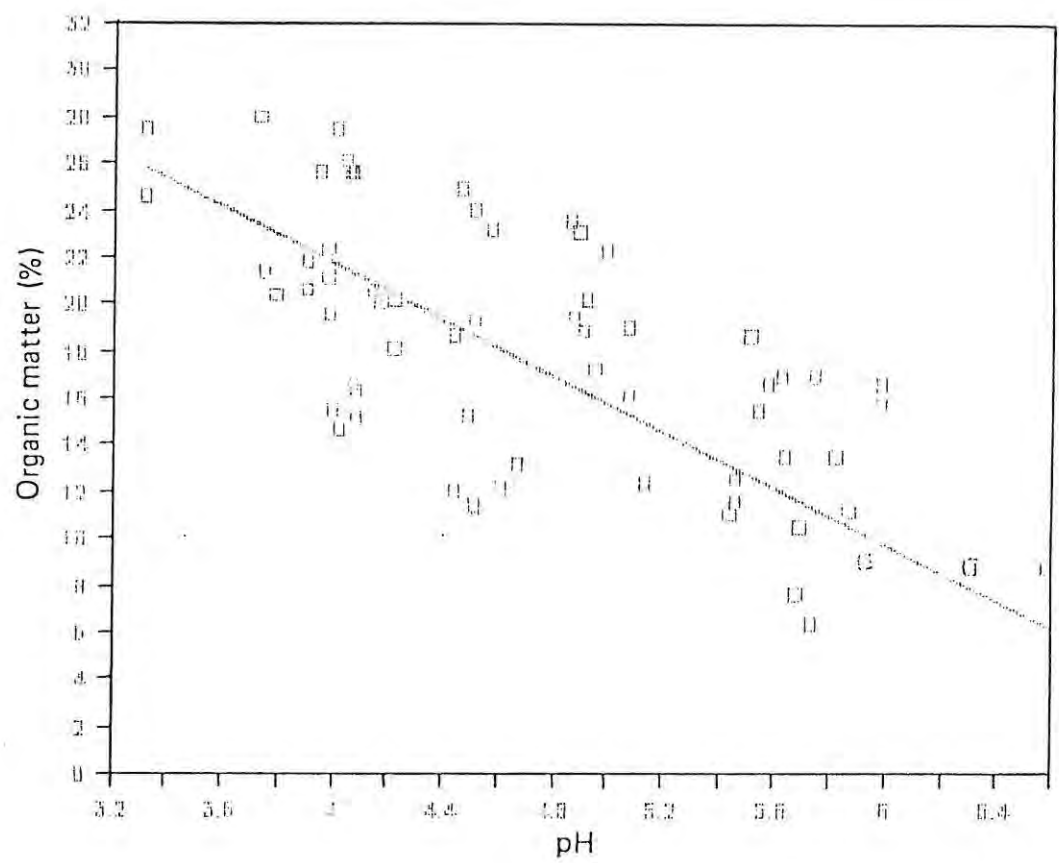


Figure 34: Inverse regression line of Dunedin longitudinal organic matter and pH values.

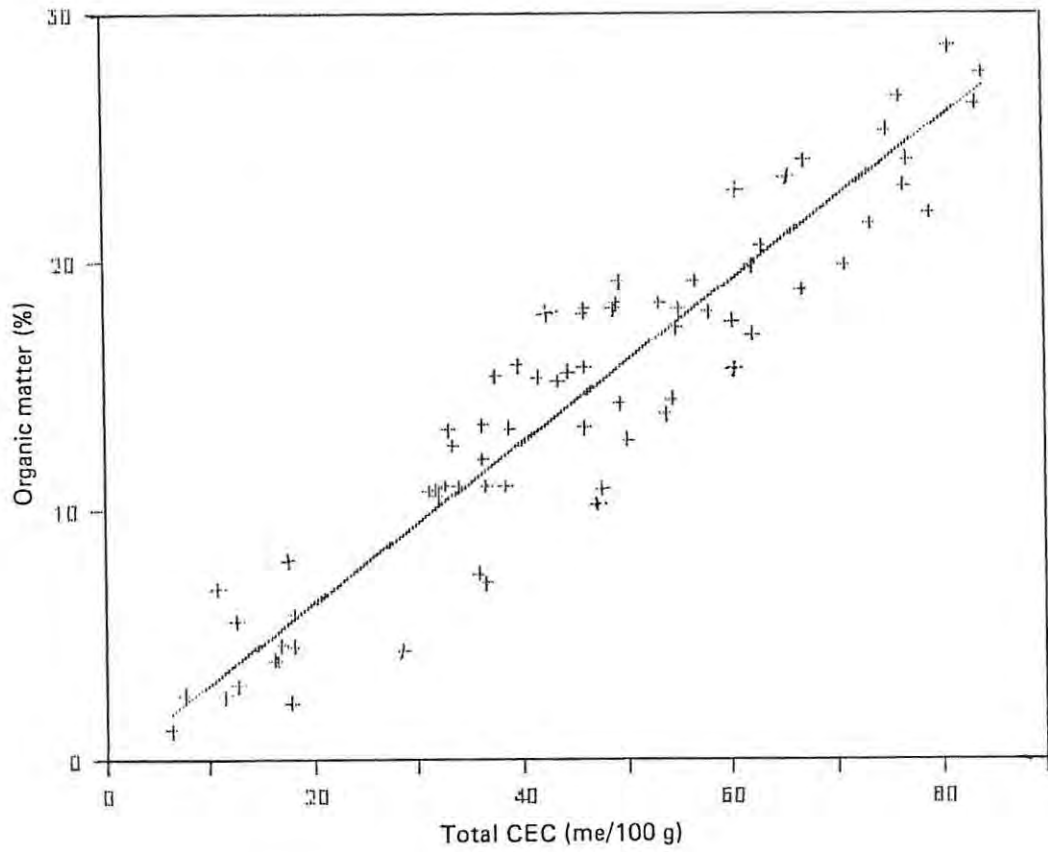


Figure 35: Regression line of Dunedin transverse total CEC and organic matter values.

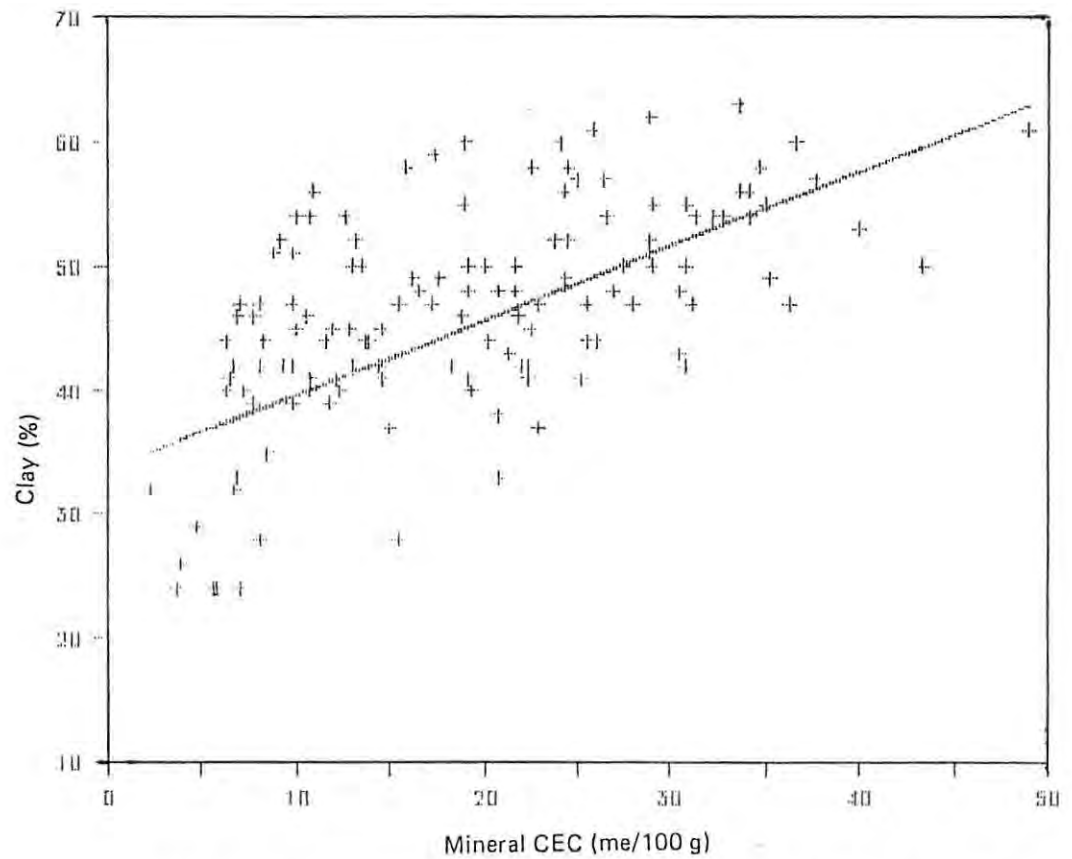


Figure 36: Regression line of mean CEC (mineral) and mean clay for all Dunedin core samples.

CHAPTER 7

VLEI SEDIMENTS AND ENVIRONMENTAL CHANGE: A DISCUSSION

7.1 INTRODUCTION

The broad objective of this study was to uncover and attempt to contribute evidence from the Eastern Cape Province to the paucity of data on environmental change for this area. The study progressed to encompass a number of aspects of the environment as it varied through the late Pleistocene and Holocene periods. The Winterberg was chosen because of its suitability for a study of this nature, possessing very definite vegetation boundaries, and relatively deep and undisturbed vleis sediments.

7.2 WINTERBERG SOIL AND VEGETATION PATTERNS

The soils occurring at the foothills of the Elandsberg were not sampled or analysed during this study, as these, being so much lower-lying, were thought to have little bearing on sedimentation processes in the vleis. However, a brief description of the soil profiles is given for comparison with the soils occurring on the plateau slopes adjacent to the vleis, as an example of the importance of parent material in generating sediment of specific characteristics.

The soils formed on the low-lying foothills are poorly structured sandy soils such as the Swartland and Valsrivier forms. These soils have a weak erodible A horizon, overlying a structured B horizon of slightly higher clay content. As seen in Plate 13, these soils are weathered from Beaufort series parent material, and are easily eroded under a savanna-like climate which supports only a discontinuous vegetation cover. Due to the shallow sandy nature of these soils, and fairly low nutrient status and water-holding capacity, they support grassland with intermittent hardy shrub and

tree species such as Acacia karoo on the more arid low-lying slopes.

The change in soil type with altitude is dramatic, as Beaufort sediments give way to dolerite which weathers into deep fertile loams. The vegetation transition is similarly clear, semi-arid savanna abruptly changing to dense montane forest of exceptionally high biomass. The multi-layered canopy and dense groundflora of this vegetation type lends itself as excellent protection for the soil against rainsplash and erosion. In addition, this dense cover aids infiltration of water into the soil and thus reduces the amount of runoff water. The forest becomes denser in protected valleys where soils are deepest and moisture availability highest.

Nearing the summit, the montane forest changes to shrub vegetation less than 3m in height. Montane forest is not restricted at this altitude by a change in parent material, but rather by harsher environmental conditions. These conditions, in turn, restrict the rate of weathering and thus soils derived here are shallower and less capable of supporting deep-rooted plants. Only short shallow-rooted xerophytic heathland and tussock grassland species survive on top of the plateau. The soils, still developing from decomposing dolerite, form an integral part of the sedimentation process affecting the vleis. The texture, colour and amount of mineral sediment is determined by the nature of the parent material from which it is weathered, as well as the rate of weathering and erosion taking place. The soil/sediment relationship is thought to be closely related, as inferred from the results. The sediments reflect several characteristics similar to those of the adjacent soils. This, in turn, can provide a great deal of information on the erosion processes operating in the area.

Possibly the most important factor affecting the rate of erosion, and thus sediment entrainment, is the

vegetation type existing on these soils. The definite vegetation boundaries, as they exist on the Winterberg today, may well have fluctuated markedly in the past. It is thought that the montane forest type would have expanded towards the summits in wetter and possibly warmer conditions (warmer to enable various forest species to survive, particularly those of a tropical nature, such as *Olea woodiana*) and retreated to its contemporary position under slightly less favourable conditions.

The suggestion that the rolling heathland vegetation such as that found on the summit of the Winterberg, is a derivation of such montane communities has been made in the past (Acocks, 1975). A number of studies on the relationship between vegetation type and other environmental variables have attempted to establish the history of such communities (Lundgren and Lundgren, 1972; Meadows, 1983; Meadows and Dewey, 1986). Indeed, soil and vegetation patterns are closely linked and a change in vegetation will result in dramatic changes in the soils beneath it. It is possible that the heathland was derived from montane forest in the distant past, although the soils formed beneath these communities appear to have established various characteristics (organic matter, depth, pH etc) which suggest an equilibrium has been achieved. This does not exclude the possibility that montane forest once existed on the plateau (in this case this would probably have been around 8 000 years BP), simply that since that time, the soils beneath the respective vegetation types have gained some sort of stability. In addition, it is difficult to estimate just how much time is required for such an equilibrium status to be achieved after a change in vegetation or climate has occurred. The spread of montane forest is possibly also restricted by periodic fires, which, along with the harsher climatic and moister conditions of the exposed peaks, assist in maintaining these definite boundaries.

If the vegetation boundaries have indeed fluctuated in the past, the extremely delicate balance between soil, vegetation and sediment yield must have been affected. If so, any change in vegetation should be mirrored by sediment characteristics deposited over this time period. From this study, the relationship between soils, vegetation and sediments has been tested on the basis that, if the sediment characteristics did correspond to the adjacent soils and their protection by different vegetation types, a change in the erosional regime will be noted from corresponding changes in the sediments.

The vleis scattered across the Winterberg appear to exist in a catenary position in the topography, where soils are of similar age and exist under similar climatic conditions, but differ in characteristics due to differences in relief and/or drainage. The top members of the interfluves are red clay soils, well drained and aerated, weathered exclusively from bedrock. The soil profiles lower downslope are younger and less well developed than those of the crest, being less weathered and less leached. At the bottom of these slopes is a sandy colluvium, changing to waterlogged black clays and peat-like sediments in the valley bottoms. This is an important feature of these vleis as the soils existing adjacent to the vleis are fundamental to the amount and type of sediment available for deposition. The sediments do in fact, reflect to a large degree that they are derived, at least in part, from the soil profiles, having many similarities in clay content and texture.

7.3 WINTERBERG VLEI MORPHOLOGY

The accepted definition of a dambo (Mackel, 1974) states that no permanent channel exists in these features. This appears to hold true for the Winterberg vleis, particularly Dunedin where surface channels are not clearly visible, although constant subsurface seepage does occur. Towards the foot, where the vlei

narrows, surface flow appears where the water level rises to flow over and through the remnant dolerite dyke. The size and shape of the Winterberg vleis is dependent on geology, particularly resistant doleritic bedrock. The Winterberg vlei profiles on a transverse and longitudinal gradient are gentle and the basins are symmetrical.

Although varying in frequency, size and shape to some degree, the vleis scattered across the Winterberg escarpment do have several features in common with those described as existing elsewhere in southern Africa, notably:

- * they are treeless waterlogged depressions;
- * their club-headed shape;
- * their situation on undulating plains at high altitude;
- * their layered stratigraphy containing peat-like material.

In all the above aspects and in the distinct pattern of sediment, moisture, vegetation and geomorphic zones, these vleis are similar to the dambos of Inyanga, Zimbabwe and Nyika, Malawi. The only area in which these features differ, is that the other high-altitude forms (Zimbabwe and Malawi) are of higher spatial density than those of the Winterberg. This factor is possibly related to latitudinal differences in climate, particularly rainfall, and topography. However, a number of similarities suggest that their sediments are comparable and that the Winterberg vlei sediments offer similar information on the environmental history of the region.

7.4 GEOMORPHIC PROCESSES IN VLEI SEDIMENTATION

The process of vlei development has been described as a complicated series of 'cut and fill' events, a 'cut' occurring when sediments and upland catchment slopes are scoured away by erosion in times of low vegetation cover, and 'fill' processes operating when vegetation

cover is plentiful, acting as a trap to sediment laden waters. In this way, layers of sediment are deposited, only to be scoured out later with a change in the controlling environment. Sediments thus display a record of repeated cut and fill cycles.

It appears that the Winterberg sediments have developed from a series of 'fill' processes whereby clastic, but mainly organic material has been deposited, either in situ from vlei vegetation or eroded from headwater areas and adjacent slopes. How many cut and fill cycles have occurred on the Winterberg prior to the deposition of this particular set of sediments will possibly never be known. However, a relatively continuous and chronologically undisturbed set of sediments appears to exist there now, various older layers underlying younger deposits, as identified from radiocarbon dating.

The complex process of colluviation, eluviation, in situ weathering and organic accumulation has probably been interrupted by periods of active removal in the past. However, as infill continued in valleys where free drainage was restricted, waterlogged conditions prevailed, exacerbated by thriving vegetation communities. This dense vegetation led, in turn, to an increase in the efficiency of sediment entrapment within the vlei area and supplied organic increments, relative to the density and diversity of the plant community, to the sediment profile. Sediments were thus continually deposited from both minerogenic and organic origins. The bulk of the Winterberg sediment stratigraphy is composed of peat-like plant remains and clay (50 to 64%), separated by lenses or narrow bands of sandy mineral sediment.

Three infill processes appear to have operated in the Winterberg vleis, each of which has resulted in sediments with unique characteristics; (a) infill of coarser mineral material, primarily by erosion from overland flow and colluvial action, providing the bulk

of the sand sized particles; (b) subsurface seepage resulting in the translocation and leaching of finer particles such as silts and clays by lateral flow through the profile, particularly downslope; (c) organic accumulation from the dense vegetation growing in the vleis resulted in an organic matrix of peat-like quality. In order to understand the mechanisms by which the sediments have been carried to and deposited in the vleis, inferences have been made from the sediment analyses, in addition to observations of the contemporary environment. It is assumed, however, that similar infill processes occurred in the past, varying only in the degree of importance under a particular environmental set.

7.4.1 EXTERNAL PROCESSES

The origin of the sediments found in the Winterberg vleis is, at present, unclear and not well understood. Studies elsewhere have attempted to clarify and describe processes operating in and around dambos (Mackel, 1974; Millington *et al.*, 1985; Whitlow, 1985), although no clear or undisputed summary has yet been offered. At the Dunedin and Salisbury vleis, there appear to be two distinct external processes affecting infill, which result in very different sediment types. These processes have operated simultaneously in the past, although one process may have achieved dominance over the other given a suitable set of environmental controls. The changeover or alternation between these processes could have been short-term seasonal variations in vegetation cover or rainfall, or long-term climatic changes. These processes have resulted in a complex stratigraphy of mineral material alternating and interspersed with organic accumulations.

The two external processes of infill, one by overland flow and the other by leaching and subsurface seepage, alternate seasonally, as a result of increased vegetation improving infiltration rates and reducing the effects and velocity of overland flow. Where

vegetation cover is dense and offers good protection to the soil, >70% surface cover (Elwell and Stocking, 1976), as on the Winterberg during the wet season, overland flow will be reduced and infiltration increased. Seepage or throughflow will take precedence under these conditions, carrying only fine silt and clay particles and nutrients in suspension into the vlei. A notable difference exists between these two processes; overland flow and subsurface seepage. The material derived and transported by overland flow is of mixed particle size, consisting of heavier loads due to the greater energy of high velocity runoff water, and a far more rapid process. Throughflow is a slower process of far lesser energy, thus carrying only the finest size fractions and nutrients in solution. Accumulation in the vleis of any noticeable depth would thus take hundreds or even thousands of years. Although the Shortlands form soil is well buffered against a build-up of sodicity, which enhances the dispersion of clays, low sodium levels being maintained by periodic leaching through these well drained profiles, some translocation of dispersed clays is evident from cutans and higher clay percentages of the B horizon. Eluviation in these soils is thus a fundamental process in the development of horizons within the soil profile and in the addition of fine sediment to the vlei.

Both of these processes must have occurred in the past, not necessarily independently, but, more realistically, simultaneously, one simply being more pronounced under suitable environmental conditions. The mineral components, derived from either physical or chemical weathering, which results in vastly different weathering products, would be transported either as colluvium or alluvium into the vlei.

The external hydrology occurring around the vleis, which enhances the development and maintenance of vleis even during dry years, is dependent on the soil characteristics of the area (Whitlow, 1985). The

occurrence of well drained profiles with high infiltration rates and undulating terrain promotes rapid absorption of water. However, these profiles are rather shallow, and being situated on slopes, encourage lateral seepage downslope. The presence of impermeable substrates near the surface in valleys restricts free drainage and thus favours the development of waterlogged areas. The issue of sources of water inputs to vlei systems has been raised by several authors, but subsurface seepage, overland flow and direct precipitation all supply moisture. It is felt that at the Winterberg sites, these three sources all occur, simply varying in degree of importance depending on season, rainfall intensity, vegetation and other environmental factors. More evidence is needed at this site to quantitatively establish whether any one process has been more important than another in supplying moisture and sediment to the vlei.

The suggestion made from the results that an increase in mineral sediment occurs as a result of overland flow during drier phases, is supported by observations made during seasonal changes which take place on the Winterberg. Although the Shortlands form soil surrounding the vleis is a well-drained soil with good infiltration rates, seasonal changes in vegetation are obviously of great importance in determining sediment availability and transport, particularly on the steeper slopes. The first rains after a long dry spell, or after an extensive natural fire, resulted in high rates of runoff at the beginning of the rainy season, erosivity and erodibility being highest when vegetation cover was most limited. The topsoil horizon of the Shortlands form soil is more weakly structured than the subsoil, and is more susceptible to erosion. This inherent weakness will be exacerbated by steep slopes and if vegetation cover is limited. Under high intensity storms, such as those which occur on the Winterberg in summer, the major product of erosion would be this sandier topsoil layer, removed from slopes by overland flow and surface wash. High

intensity events are known to produce sediment with lower percentages of clay-size aggregates (Mitchell et al., 1983). In addition, ponding or dense stands of vegetation on the adjacent slopes would result in deposition of larger sized particles, unless storm intensity was high enough to overcome this, or the trapping effectiveness of vegetation was decreased by severe events (Hamlett et al., 1987). This confirms that vegetation cover would have to be sparser to allow larger sized particles to be carried to the vlei.

After the dry season, soluble salts including sodium, having accumulated near the surface during evaporation, would also be present in the runoff water, adding nutrients to the vlei system. Sodium, which could increase the dispersion of already deflocculated clays deposited previously, is a cation readily available from weathered dolerite, and subject to loss from nearby soils by leaching.

The selective deposition of various grain sizes is similar to processes described in the true dambos at lower altitude (Mackel, 1974). Here, surface wash is noted as the most important process of dambo infill, while the significance of this surface wash appears to be greatly reduced on the higher altitude, densely vegetated slopes of the Nyika plateau (Meadows, 1982). The clear distinction between layers of sandy material and those of organic and clay material, is different to that described by Meadows (1982) for Malawi, where organic bands were distinct from bands of mineral matter high in clay. The origin of sediment must thus be different for these two areas, the processes of infill being controlled by rainfall type and intensity, as well as weathering products available for deposition. This process occurs on a more seasonal basis at the Winterberg, particularly after fires and droughts. Here, the deposition of loose sandy material suggests that surface erosion is the dominant process, while clay and organic deposition occur during moister phases when vegetation cover is greater. The

seasonality of climate and therefore vegetation cover is obviously of great importance on the Winterberg. Even so, this process has led to no significant deposits for the last 7 to 8 000 years BP. Organic deposition appears to be uninterrupted for this period.

7.4.2 INTERNAL PROCESSES

The processes operating within vleis have not been thoroughly researched, but the development of a layered stratigraphy due to external inputs alone does not account for the changes which take place within the sediments once deposited. Undoubtedly, such processes as lateral water movement, fluctuating watertables and clay translocation must complicate the stratigraphy to some extent, although quantification of such processes would be difficult. This is where the value of radiocarbon dating, although based primarily on organic matter, is so important in confirming that contemporary sediment strata are indeed in the sequential position in which they were deposited, and that chronological inversion has not occurred. However, changes in the mineral fraction may still go undetected by such dating techniques, and as such, clay translocation and secondary formation of minerals would have to be monitored by another method.

Two main internal processes appear to take place within the sediments of the Winterberg vleis; mineral movement such as clay translocation and organic accumulation. The dark grey to black topsoil layer of these sediments contrasts with the red soils of the surrounding slopes, the colour being due to dark organic particles. The predominance of organic matter in the surface layers suggests that minerogeric inputs play only a minor role in sediment accumulation at present. Organic accumulation appears to be a continuous undisturbed process under contemporary environmental conditions. The relationship between organic matter accumulation and moister environmental conditions is supported by the increased density of hydromorphic species with a higher watertables towards the centre of the vlei.

Meadows (1982) states that a larger annual increment of plant material would result from the longer life-span of those species being supported by higher available moisture. This would explain the higher organic matter content in the central parts of the vleis. In addition, the lack of significant mineral deposits suggests that erosion processes are limited by dense vegetation on surrounding slopes. Thus, a dense layer of organic matter within the stratigraphy must surely indicate a moister environment.

Clay translocation is a less obvious process than surface wash and involves the removal of dispersed clays by water moving through a coarse grained matrix in vlei subsoils (Mackel, 1974; Whitlow, 1985). Clay translocation within sediments is dominant under conditions of a high watertable, clays being removed from porous sandy lenses, to accumulate amongst the dense, colloidal organic masses of the peat-like layers, or be removed from the system entirely. Constant baseflow and seepage from the vleis would ensure that a hydraulic gradient exists and that water movement is constantly taking place. All these conditions exist within the Winterberg vleis, Dunedin in particular, and sodium ions are leached from adjacent soils to promote dispersion of clays. This process is thus distinctly possible within these systems.

There is no mixing of mineral and organic sediments by earthworms or other organisms, which favour better drained and less acidic soils. Thus, it is possible that the majority of the clay fraction found within the organic bands, was deposited simultaneously with organic accumulations, with additional leaching from sandy layers taking place at a later stage. This simultaneous deposition of fine mineral material and organic matter would occur if the moister conditions favouring dense vlei vegetation, also applied to adjacent slopes. Vegetation would afford good soil protection, and seepage would thus be the major

contributor of fine mineral particles. Erosion by surface runoff would be effectively minimized.

Vegetation is known to have a filtering effect on sediment laden water, reducing the velocity of flow and thus resulting in the differential settling out of particle sizes, according to Stokes' Law (Bryan, 1968; Langbein and Schumm, 1958). In this way, clays and fine silts are carried further into the vleis than heavier sand particles. In addition, the chemical decomposition of minerals, once deposited in vleis, into secondary clay types would be enhanced by warm, moist conditions.

The various size fractions of the mineral particles appear to be deposited in distinct zones, between which marked differences in textural definition can be noted. These zones are the sandier headwater and margin zones, similar in many respects, and the centre of the vleis which is highly clayey in nature. This distinction of texture appears to have come about entirely due to the mechanics of hydraulic flow, erosional processes and depositional environments, which all resulted in the differential settling out of various particle sizes. The sorting of sediments was indicative of an environment suitable for the formation of a 'washbelt' near the margins, deltaic formations in the headwater zone and the fining of particulate matter nearing the centre and foot of the vleis. These distinctions were not as clear for Salisbury vleis, possibly due to the influence of agricultural disturbance on one slope, which could well have led to vastly accelerated erosion and sediment input during high intensity storms. From the textural analyses of sediments, distinct patterns or zones of deposition within the vleis are noted. This zonation, although less distinct, appears to correlate well with similar patterns described for dambos elsewhere in Southern Africa and appears to be a characteristic feature of these geomorphological phenomena.

Major sediment characteristics may be noted here.

- (a) Texture is variable due to external erosion and internal sedimentation processes in the vleis.
- (b) Clay and organic matter increase towards the centre, and decrease with depth.
- (c) Sandy material is found around the margins and increases with depth.
- (d) Organic matter accumulates at the surface if the watertable is permanently high.

7.5 WINTERBERG SEDIMENT STRATIGRAPHY

The characteristics of the sediments have now been outlined, as well as the processes leading to their deposition and development. The descriptive part of this study having been accomplished, to what extent can the sediment characteristics yield clues as to the environmental history of the area? In addition, what inferences can be made about the environment itself and its related effects on soils and vegetation in the past? The trends noted from the sediment analyses indicate that the vlei sediments have a number of characteristics which might assist in elucidating past environmental conditions. Also, available radiocarbon dates supply valuable information as to the age and chronology of these deposits. The number of dates obtained were limited by both time and cost, and because the vlei sediments were relatively low in organic matter, and could not be classified as true peats. Peat requires an organic matter content of about 80% (Chapman, 1964; Moore and Bellamy, 1974). However, as they were waterlogged, thus reducing the rate of decomposition, and did have +30% organic matter, they could be classified as organic sediments.

From the stratigraphic diagrams, there appear to have been two phases of clastic deposition, separated by two main phases of organic accumulation. In the lower organic band (330cm), organic percentages were similar to those of surface layers, and appear to identify periods when clastic deposition was reduced by

protective ground cover on slopes and organic deposition encouraged by denser vlei vegetation; a moist phase. If this moist phase resulted in expansion of the montane forest areas to areas surrounding the vlei, and thus drastically reducing the level of erosion, this may indicate a simultaneous warming of conditions if such species as Olea spp. and Rapanea spp. were to survive. Where organic matter is reduced, and strata are composed of coarser sandy material, conditions of sparse vegetation cover such as xerophytic heathland (dry and cool) could have resulted in larger erosional inputs from sideslopes or flood events. A point of clarification is necessary here, for terms of reference such as 'moister' and 'drier' etc. These terms are taken as indicating a shift in conditions related to the period prior to accumulation, and not relative to contemporary climates (Meadows, 1988). This is a constant source of inaccurate interpretation if the frame of reference is not given by the author.

The high clay contents of the sediments are dependent on the fine-grained nature of doleritic parent material. However, the basal sediments do not show the expected increase in clay content with depth, but are underlain by up to 0.5m of dense sands. These sands could have been deposited under more arid conditions during the last interpluvial, prior to the initiation of organic accumulation around 12 000 years BP. The organic clayey horizons possibly prevent, by their fibrous sticky nature, vertical eluviation of fines into the basal sands, clay percentages in these layers being very low.

Sandy layers, found at the base of the vleis with negligible amounts of organic matter, could well be representative of the arid, cooler phase which followed the glacial maximum around 18 000 BP, prior to organic accumulation which began around 12 500 BP, shown by the organic stratum at 330cm in the Dunedin stratigraphy. The lack of organic matter in these sands may indicate

more xeric conditions when organic accumulation was retarded. In addition, when first deposited, these sandy deposits would have been better drained and aerated, being the surface layers at the time and exposed to the atmosphere. This would have resulted in more rapid decomposition of organic matter, particularly if the watertable was lower and drying out of the surface layers occurred, and leaching in these coarser layers would have been a dominant process, seepage water removing both clay and organic particles from the exposed surface layers.

The hypothesis that an increase in organic matter corresponds to increasingly moister environmental conditions, and a subsequent increase in plant density, appears to follow and be true for the sediments in these vleis.

From the stratigraphic diagrams of the transverse sections (Figures 32 and 33), the margin cores all lack a highly peat-like organic band at depth. From Figure 37, the differences in estimated levels of infill between the present day and those of 12 000 BP are shown diagrammatically. The extent of the vlei at the time of this organic phase of accumulation, +12 000 BP, would be narrower and only approximately 30 to 50cm deep. With continual infill, the vlei has aggraded and expanded up the valley sides, such that contemporary margins are far above and outside the level of infill deposited at around 12 000 BP. This suggests that infill has been relatively constant through to the present day, unbroken by severe climatic changes which may have induced a "cut" phase. This supports the suggestion that Holocene environmental changes, although many and varied, did not have as great an impact on ecosystems as those of the Late Quaternary. In addition, these conclusions demonstrate that a state of dynamic equilibrium appears to exist on the Winterberg plateau, which has prevailed for at least the last 6 to 8 000 years. Changes in hydrological

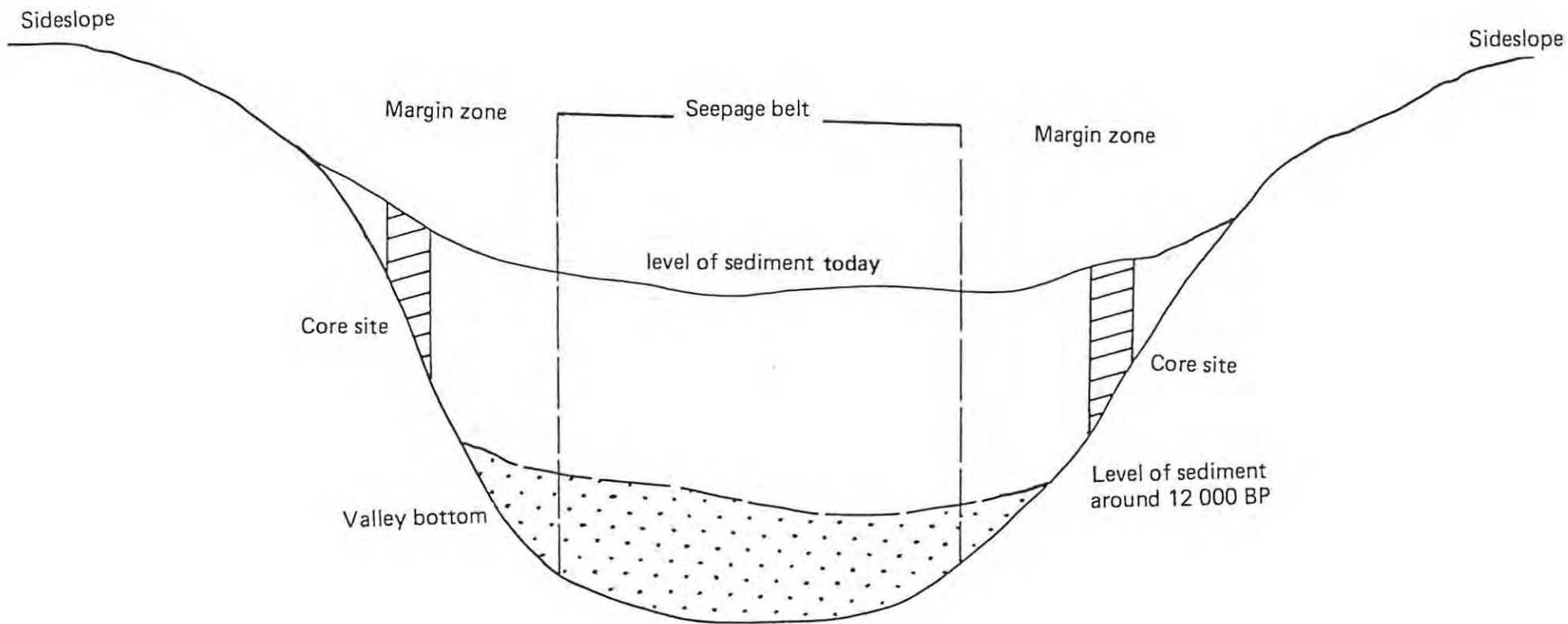


Figure 37: Diagrammatic representation of sediment zones and level of infill at 12 000 BP and today.

processes have been relatively small and not of great influence on vegetation changes.

7.6 CHRONOLOGY AND SIGNIFICANCE OF WINTERBERG VLEI SEDIMENTOLOGY

Meadows (1988) summarises evidence for two moist phases in southern Africa, one at the terminal stage of the Pleistocene, and the other during the mid-Holocene, citing studies from parts of the Cape Province, Transvaal, Malawi and Zimbabwe. The first wetter phase is supported by sedimentological evidence from Dunedin and Salisbury vleis, and pollen analyses from these sites (Meadows et al., 1987). Widespread evidence suggests cooler, moister conditions during the earlier wetter phase than the present, with the initiation of several peat-like or organic deposits throughout Southern Africa. This moist cool phase appears to have been followed by a slightly drier and warmer climate which was maintained through to the early-Holocene.

The second moist phase, around 5 000 to 3 000 years BP, is supported by evidence from the southern Cape (Scholtz, 1987) and several other sites summarised by Meadows (1988). Accumulation of organic sediments during this second phase began somewhat earlier at occasional sites, including Dunedin, dated around 8 000 years BP. As suggested by Meadows, it would be surprising if sites within widely different climatic and geomorphic zones all responded similarly to changes in climate. It is possible too, that at sites such as Dunedin, climatic fluctuations would have had to be of larger scale to have a significant influence on the environmental regime affecting sediment accumulation. Although Holocene fluctuations in climate have occurred, and been fairly frequent, it is possible that these phases have not been dramatic enough to induce a complete change (or reversal) of conditions, such as those of the last glacial maximum.

It is possible that South Africa was less affected by environmental changes than other parts of Africa, although the distinct shortage of evidence for this area renders this conclusion tentative. It is possible that vegetation in the region is not particularly sensitive to climatic changes of relatively low magnitude, such as may have occurred in the late Holocene, requiring far larger shifts in moisture availability to produce measureable changes in the vegetation assemblage. Greater aridity may have produced somewhat lower densities of cover and increased the frequency of fires, but the reduction in precipitation required for a total vegetation change, must be large. The basal date for the commencement of organic infill for Dunedin was 12 500 BP, similar to that of Zimbabwe and Malawi, and coincides with increased precipitation in the tropics (Meadows, 1982). The further increase in organic accumulation at 8 000 BP may have been of localised extent, but since that time, there is little evidence to suggest that there has been a vegetation change, and therefore environmental change, of any great significance. It remains a tentative conclusion, however, as the vegetation in the area may simply not be sensitive to slight changes in precipitation or temperature. The seasonality of rainfall may be more important, rather than total amount. This is opposite to the vlei system, which responds rather to variations between season than annual totals. Further evidence is required from similar sites with a more complete record of the late Pleistocene to support this.

The exact age of the basal sediments of the Winterberg cannot be set from this study, the dense nature of the sandy bottomset beds limiting auger penetration. However, the more organic characteristics of the layer found resting on these sands has been dated at 12 600 years BP, signifying the commencement of organic accumulation. Such sediments can and do reflect changes in the environment, which result in a layered

stratigraphy in the vleis, each band setting a different environmental scene.

Prior to the initiation of organic accumulation on the Winterberg, conditions during the last glacial maximum $\pm 18\ 000$ years BP (possibly represented by sediment layers below those obtained by current coring methods), were drier, having profound effects on the vegetation of the plateau, and thus on sediment and water supply to valleys. Reduced vegetation cover would have given rise to higher erosive energies of runoff water and heavy loads of coarser sediments, containing little organic matter. Evidence for these xeric conditions is found in other areas of southern Africa (Vogel, 1985).

The fairly low organic percentages in the majority of samples limited the number of radiocarbon dates obtained. However, these dates are vital in accurately determining the age of sediments and to construct stratigraphic diagrams representing sediment deposition. Basal sediments from Dunedin at 330cm depth were dated at $12\ 500 \pm 160$ years BP (Lab. no. Pta-4207), while basal sediments of Salisbury at 216 to 225cm depth were dated at $11\ 800 \pm 120$ years BP (Lab. no. Pta-4318). This confirms the supposition that organic accumulation began at roughly a similar time for both vleis, and that the sediments span more than the Holocene. A sample of the black amorphous peat at Dunedin at 220cm depth was dated at $7\ 990 \pm 200$ years BP (Lab. no. Pta-4207), confirming that there was indeed a second phase of moist conditions during the early to mid Holocene. This date is somewhat earlier than the moist phase at other sites, where organic accumulation appears to begin between $3\ 000$ and $5\ 000$ years BP (Meadows, 1988). A radiocarbon date from a vlei further west of Dunedin, Ellerslie, at 150cm depth was dated at $4\ 200 \pm 60$ years BP (Lab. no. Pta-4335), more in keeping with data from the interior (Meadows, 1988). This does tend to complicate the pattern somewhat, although regional differences are to be expected between areas of different climatic characteristics.

The dates received for the two Winterberg vleis compared very well, and suggested that fluctuations in the environment had affected the entire plateau area. In addition, the stratigraphy of sediments in the two vleis correspond closely. This suggests that any differences in sedimentary character must be due to localised topography, but that the period of deposition for both vleis is similar, organic accumulation commencing at roughly the same time for the entire catchment area.

The radiocarbon dates, apart from giving an accurate age for the sediments, also supply information on the rate of accumulation of such sediments. Establishing the accumulation rate of sediments by dating and particle size analyses is desirable to increase the reliability of interpretation. The rate of organic accumulation on the Winterberg appears to have been somewhat slower than that of peat deposits in Nyika plateau dambos. At Nyika, a rate of 1m of sediment per 1000 to 1500 years is suggested, a rate which is similar to temperate latitudes as described by Moore and Bellamy (1974). At Winterberg, Dunedin vlei sediments have accumulated at a rate of only 29cm for every 1000 years, and 22cm per 1000 years at Salisbury. This could be due to the lower rainfall regime and the lower altitude of these vleis when compared with those of Nyika, which will influence the density and floral characteristics of vlei vegetation. Organic matter is also fibrous and bulky until thoroughly decomposed, after which it is highly compressible. Thus, these rates are not entirely accurate. Also, moist phase accumulations are more rapid than those of drier phases, and extreme events also complicate the sequence, a major storm adding or removing a broad band of sediment in the space of hours. Compaction of lower layers will also lead to inaccuracy in estimating stratigraphic depths. However, the rate of accumulation in the Winterberg is lower than that described elsewhere in Southern Africa, possibly due to local differences in topography and

climate. Radiocarbon dates have indicated that the stratigraphy is in the correct chronological sequence, and that no overturning of sediments has occurred.

7.7 SUMMARY OF EVENTS ON THE WINTERBERG ESCARPMENT

What can be inferred from Winterberg sediments, is that environmental changes did accompany the present sedimentary set deposited during the late Pleistocene, between 11 500 and 12 500 years BP. From the analyses conducted, it appears that, as rain fell on the undulating plains of the uplifted Elandsberg escarpment, protected at such elevations by a thick dolerite capping, sediments were washed down the slopes to settle in the lower-lying upper courses of valleys. The environment gradually ameliorated after the glacial maximum to warmer, moister climates. This led to increased vegetation cover on slopes (possibly even dense forest formations) and higher watertables. Where flow along drainage lines was restricted, often geologically, waterlogged conditions arose. This offered a new and different niche to plants, and hydromorphic species began to form part of the vegetation mosaic of rolling heathland surrounding patches of montane forest. Under these moister, warmer conditions, rainfall was sufficient to promote rapid colonization of these areas. These species, in turn, produced organic matter, further stabilizing the sediments in which they grew and resulting in the initiation of peat-like sediment formation around 12 500 BP.

Conditions grew slightly less favourable between 11 000 and 10 000 years BP, slopes were less well protected and erosion was increased. This stage possibly marks the retraction of montane forest to valley heads and the spread of hardy grass/heath shrubs across the cooler and now drier plateau. This is shown by sandy deposits of low organic matter, suggesting that heavy sediment loads and mineral matter comprised the majority of the vlei input. Pollen of such species as

Erica, Cliffortia and Helichrysum became pronounced during this time, with montane species disappearing almost totally (Meadows et al., 1987).

Between 10 000 and 8 000 BP, slopes were once again increasingly protected by vegetation, with a concurrent increase in moisture. Storms still resulted in clastic sediment loads being carried into the vleis. These lenses of mineral origin, although indicating a continuation of simultaneous organic accumulation, are found between the bottomset beds and the second organic band at a depth of 2.0m. Around 8 000 years BP, conditions became even more favourable for further development of these hydromorphic stands, and organic deposition was greatly accelerated, forming the most important source of infill. Around this time, montane forest expanded in accordance with the higher temperatures and moisture availability, resulting in Podocarpus spp. and Olea spp. pollen being preserved in the vlei sediments, and a loss of the grass/heathland pollen (Meadows et al., 1987). In addition to this organic deposition, some clastic sediments were still washed into the vleis following rainfall events, to be trapped and held by the dense network of roots and stems of vlei vegetation. Clays and nutrients slowly filtered through the profile under the influence of subsurface drainage, accumulating and being held primarily by the organic rich colloidal layers. This accumulation of limited clastic and mainly organic sediments appears to have continued through to the present day, suggesting that environmental fluctuations during this period did occur, but were not marked.

The predominance of peat-like material in more recent layers suggests that infill in the last 6 000 years BP was primarily in situ organic accumulation, and that erosion products from nearby slopes had been severely limited. Organic sediment accumulation appears to be relatively undisturbed since initiation around 8 000 BP, suggesting that more stable environmental conditions and an equilibrium between soil and

vegetation has been achieved in the area through to the present day.

7.8 CONCLUSION

The high-altitude vleis existing on the Winterberg have been examined and compared with dambos found elsewhere throughout central and southern Africa. The Winterberg vleis are similar to those described on the Nyika Plateau in Malawi and the upland dambo forms described in Zimbabwe. It would appear that the period of organic accumulation corresponds with that described in Malawi, a major accumulation commencing at about 12 000 years BP, corresponding with a "fill" phase described throughout tropical Africa. A clastic phase of sedimentation followed, possibly representing slightly less favourable climatic conditions until around 8 000 BP when organic accumulation again became the dominant process through to the present day. From the analysis of the stratigraphic changes of these vlei sediments, it appears that these features are indeed useful, easily comparable and sensitive indicators of environmental change, possibly even more so than the more commonly studied fossil vegetation communities.

CHAPTER 8

REVIEW AND CONCLUSION

8.1 REVIEW OF STUDY AIMS AND OBJECTIVES

The broad aim of this study was to attempt to elucidate past environmental change and thereby supply additional data to the existing collection for the late Quaternary in southern Africa. From this broad aim, the focus sharpened into a study of one particular geomorphological phenomenon which indicated environmental change; vlei sediment stratigraphy. An explanation of the complex yet very marked stratigraphy which has developed in the Winterberg vleis was the primary objective of this study.

By detailed investigation of the chemical, physical and morphological characteristics of these sediments, an attempt was made to decipher the time-span and mechanics responsible for this sediment accumulation. The situation of these vleis, being so similar in many features to those described elsewhere, made for a suitable site in which to search for clues of fluctuations in past environments. The only restraints placed on the study were the physical limitations on the depth of extracted samples and the limited number of samples suitable for radiocarbon dating. The accuracy of conclusions was also hampered by the necessity of using inferences, made mainly due to the few absolute radiocarbon dates available.

Although the environmental implications made from the Winterberg vlei sediments are restricted to a local scale, and extension and correlation on a regional scale is difficult, this type of extrapolation is still enormously valuable in the attempt to produce a network of data across the subcontinent. Data from similar sites appear to be very comparable and suggest that inferences made may be more valid than was first

thought. Dunedin and Salisbury vleis allow the reconstruction of the past environmental history of a local catchment sequence. It has been shown that these vleis are indeed sensitive environmental indicators as they operate as profit-and-loss systems, reflecting inputs and outputs and having an internal storage facility, a change in any one influencing variable altering the systems balance and resulting in a measurable change.

Differences between sites within southern Africa are many, but primarily they are differences in the duration of environmental shifts, rather than a total contradiction of one, and in this way, the rather scattered data from southern Africa does appear to be slowly forming a climatic history for this area. The synthesis of evidence is made difficult due to the complexity of evidence, where swift changes in one province are represented by a longer phase in another. An additional problem is that various sites of similar age which could provide detailed evidence of a change, rarely coincide and thus an integrated approach remains, for the moment, unobtainable.

8.2 INTEGRATION OF OBJECTIVES

In fulfilling the aims of this project, it is easier to consider each of the original objectives of this study as separate lines of interest, and present them as an overview of the achievements made within this study.

The objective of describing and characterising the properties of the sediments, has been achieved and valuable relationships established between these variables. Changes in sediment characteristics correspond closely to vegetation changes, although the chronology of accumulation of these sediments through to the present day cannot be absolutely clarified. The Winterberg vleis and their sediments do correspond well to dambos described elsewhere, even when the comparison is made on purely external morphological components.

There are similarities with the true dambos of Malawi, Zimbabwe, as well as other sites described locally, in terms of organic and clastic sediment accumulations, mechanisms of deposition and general stratigraphic characteristics.

Radiocarbon dating has been used to successfully and accurately date the sediments, thereby establishing a chronological sequence of deposition as represented by stratigraphic diagrams. Comparative stratigraphic diagrams were constructed and interpreted for changes in the sedimentary environment. A basal date suggested the initiation of infill at around 12 000 years ago, and this was found to correspond with dates from other sites. Further accelerated organic accumulation followed at about 8 000 years BP, somewhat earlier than is suggested by dates for other sites. This organic accumulation appears to have continued relatively undisturbed through to the present day, and the inferences made from this evidence is supported by data from numerous sources across the region. While the sedimentology of the Winterberg vleis does suggest that removal of sediment is continuing from the plateau slopes, this sediment is being captured and held in an infilling process in valley vleis. These sediments are not being lost from the catchments, simply translocated to lower-lying areas.

The significance of these vleis for environmental change has been shown by the very nature of these pollen-preserving sediments, which have, by all their inherent properties, reflected a pattern of environmental change. It is felt that this study has made a significant contribution to existing data and has improved current understanding of the sensitivity of vlei features to any form of disturbance, particularly agricultural. Effective management of contemporary ecosystems relies heavily on this understanding of past environmental fluctuations. Hopefully one result of this study will be to increase awareness as to the value of these delicately balanced

systems and the importance of their conservation. Although the results from this study on the Winterberg do not accent or dispute trends from other areas, they do retain a similar direction of information, and suggest that fluctuations in the external environment have taken place, and affected the vlei system, throughout the late Pleistocene and Holocene periods.

8.3 FURTHER RESEARCH

There is a need to study other vlei sites in greater detail as they appear to be geomorphological features which have been overlooked to a some degree by researchers in southern Africa. More detailed evidence is needed on the Winterberg and similar sites to confirm the inferences that have been made. A more extensive and complete set of radiocarbon dates is needed to fix the chronology of these sediments more precisely and confirm the longevity of environmental changes, in relation to dates from similar sites elsewhere.

Detailed studies are required on the hydrology and erosional processes of these systems, particularly from an agricultural viewpoint, where these systems are being used for water supplies and agricultural development. In depth study of such contemporary processes can shed light on how and whether these processes have operated similarly in the past. Vleis are areas which have become focal to agricultural landuse, and it is hoped that as information emerges from similar studies, a greater awareness of these ecosystems will arise. Perception of the sensitivity of these systems is fundamental to any form of preservation and the judicious use of the products they have to offer. Thus far, relatively little detailed research has been conducted to estimate the consequences of various landuse practices on vlei soils and hydrology, and it is these aspects that warrant further study. A good starting point in the search for better understanding of vlei development and

preservation, would be to recognize that they are complex functioning systems, hold valuable information of the past and present geomorphic processes and a finite resource.

8.4 SUMMARY

The aims of this study have been met with variable success, some being limited in accuracy, others by physical constraints. Realization of the worth of vleis as sites which offer some explanation of the environmental conditions which prevailed throughout the late Quaternary period, is imperative, and a treasure that dare not be carelessly overlooked. To quote Breen and Begg (1987; p.221): "The tragedy of these lost wetlands is that we had the foresight to promulgate preventative legislation, but not the courage or responsibility to apply it...". It is important that this study be viewed in a global context with a need to preserve these features. The study has allowed the reconstruction of at least a local sequence of events which appears to show fruitful links with data elsewhere. Gradually, such geomorphological evidence can be integrated with data from other fields. This may enable the production of a multidisciplinary collation of palaeoenvironmental evidence for the late Quaternary for southern Africa.

Key: A–G: Dunedin longitudinal samples
 DTA–DTG: Dunedin transverse samples
 STA–STE: Salisbury transverse samples
 1–7: Soil samples

SAMPLE	DEPTH	MOISTURE	ORGANZ	PH	NA	K	HG	CA	TEB	EXCH.H	CEC(MIN)	CEC	B.SZ
A	10	68.670	29.040	4.02	0.250	0.450	3.960	6.130	10.790	13.43	24.228	82.31	44.57
A	35	87.250	21.300	3.75	0.124	0.800	2.805	14.079	17.800	13.42	31.300	74.07	57.14
A	60	85.270	20.390	3.79	0.096	0.915	1.402	5.410	7.823	6.70	14.523	55.30	53.87
A	85	73.810	19.580	3.99	0.099	0.972	2.506	5.312	8.809	1.87	10.759	49.92	82.62
A	110	70.980	17.270	4.95	0.013	0.227	1.390	3.530	5.160	1.14	6.300	40.85	81.93
A	135	82.330	11.620	5.46	0.066	0.523	2.473	1.631	4.693	3.36	8.049	31.29	50.31
A	160	85.490	15.510	5.54	0.061	0.552	1.401	2.372	4.386	4.87	9.256	40.28	47.39
A	185	81.680	31.010	4.80	0.176	0.545	0.436	0.875	2.032	11.49	13.522	75.54	15.03
A	210	67.360	23.100	4.89	0.116	0.284	1.297	4.469	6.166	14.58	20.746	66.95	29.72
A	235	65.320	22.310	4.99	0.178	0.244	3.300	8.554	12.276	13.54	25.816	70.44	47.55
A	260	67.610	11.200	4.10	0.200	0.235	5.121	3.033	8.589	7.33	15.915	38.32	53.97
A	285	67.100	2.130	5.00	0.210	0.201	2.121	1.890	4.422	2.51	6.9344	11.19	63.77
A	310	66.300	1.040	5.12	0.230	0.111	0.125	2.144	2.610	1.07	3.68	5.76	70.92
B	10	74.600	27.450	3.32	0.102	0.333	3.709	16.232	20.456	28.44	48.096	103.00	41.04
B	35	83.900	25.630	4.07	0.276	0.600	2.960	17.109	20.945	22.40	43.345	94.61	48.32
B	60	83.540	20.120	4.17	0.309	0.219	2.732	14.800	18.140	15.90	34.038	74.28	53.29
B	85	88.970	11.360	4.51	0.450	0.767	2.567	7.200	10.992	9.72	20.712	43.43	53.07
B	110	78.700	19.040	5.07	0.376	0.742	1.240	2.701	5.059	16.76	21.819	59.90	23.19
B	135	73.700	13.560	5.82	0.291	0.673	2.800	8.132	11.976	8.32	20.296	47.42	59.01
B	160	82.360	20.130	5.74	0.270	0.430	3.630	7.562	11.900	10.59	22.49	62.75	52.91
B	185	88.430	28.540	4.86	0.287	0.770	4.461	13.763	19.281	11.21	30.491	87.57	63.24
B	210	78.360	31.020	4.85	0.236	0.820	6.021	14.209	21.206	21.10	42.306	104.43	50.22
B	235	88.790	29.620	4.51	0.172	0.604	1.263	9.876	11.915	18.96	30.875	90.12	38.59
B	260	72.360	18.740	4.44	0.245	0.763	0.863	7.301	9.172	13.42	22.592	60.07	40.60
B	285	78.590	16.620	5.58	0.306	0.892	1.403	2.200	4.801	5.44	10.241	43.48	46.88
B	310	69.390	16.420	5.73	0.205	1.132	1.361	3.181	5.879	6.38	12.259	25.10	47.96
B	335	68.760	5.120	5.76	0.386	1.006	1.306	1.543	4.241	4.23	8.471	18.71	50.06
B	360	70.540	2.220	5.81	0.204	0.976	1.032	1.601	3.813	2.00	5.813	10.25	65.59
C	10	85.800	28.560	4.35	0.489	0.361	2.726	11.810	15.306	21.24	36.626	93.75	42.01
C	35	84.870	23.170	4.57	0.120	0.750	4.521	12.555	17.946	11.00	28.946	75.29	62.00
C	60	88.470	18.690	5.51	0.199	0.670	2.254	16.140	19.263	18.73	37.993	75.37	50.70
C	85	75.360	8.760	6.58	0.107	0.543	1.778	6.750	9.178	6.34	15.518	33.04	59.14
C	110	85.460	11.050	5.44	0.195	0.322	2.366	4.329	7.212	2.64	9.852	31.95	73.20
C	135	72.310	12.550	5.46	0.212	0.110	1.575	0.330	2.227	4.73	6.957	32.06	32.01
C	160	83.240	21.030	5.43	0.283	0.209	2.093	2.905	5.570	1.49	7.06	49.12	78.90
C	185	86.540	30.600	3.58	0.348	0.523	1.693	3.600	6.172	10.11	16.282	77.64	37.91
C	210	76.800	27.980	3.73	0.118	0.696	3.717	4.936	9.467	17.01	26.477	82.44	35.76
C	235	88.740	24.930	4.46	0.109	0.065	6.777	9.870	16.021	15.43	32.251	82.11	52.16
C	260	88.000	18.880	4.91	0.297	0.110	5.196	12.232	17.835	9.63	27.465	65.23	64.94
C	285	80.010	7.660	5.68	0.346	1.270	2.301	8.720	12.637	2.40	15.037	30.36	84.04
C	310	72.100	2.150	5.40	0.801	1.071	1.007	2.610	5.409	3.64	9.129	13.43	60.13
C	335	70.740	2.240	5.16	0.794	0.679	0.985	0.780	3.230	2.36	5.590	10.00	57.84
D	10	88.650	26.130	4.05	0.362	0.236	2.431	16.306	19.335	14.34	33.675	85.94	57.42
D	35	85.460	22.350	3.98	0.149	0.007	4.760	17.432	22.356	12.64	34.996	79.70	63.88

SAMPLE	DEPTH	MOISTURE	ORGANZ	PH	NA	K	MG	CA	TEB	EXCH.H	CEC(MIN)	CEC	B.SZ
D	60	91.010	21.780	3.90	0.109	0.312	2.409	4.031	6.861	12.50	19.361	62.92	35.44
D	85	72.430	14.660	4.02	0.260	0.076	0.863	5.307	6.514	7.50	14.014	43.33	46.48
D	110	77.360	12.360	5.13	0.362	0.632	1.036	10.346	12.376	4.89	17.266	41.99	71.68
D	135	84.460	16.570	4.07	0.459	0.660	1.784	7.567	10.470	8.72	19.19	52.33	54.56
D	160	82.540	20.620	3.90	0.276	0.780	2.809	13.609	26.554	13.40	39.954	81.19	66.46
D	185	74.700	21.120	3.98	0.274	0.555	3.562	11.920	16.311	21.32	37.631	79.07	43.34
D	210	79.320	25.630	4.06	0.175	0.201	2.710	11.460	14.546	19.02	33.566	84.83	43.34
D	235	67.390	24.020	4.51	0.250	0.707	1.600	6.843	9.400	8.00	17.4	65.44	54.02
D	260	67.580	15.240	4.49	0.259	0.763	1.436	4.362	6.820	3.25	10.07	40.55	67.73
D	285	68.700	17.130	4.68	0.189	0.730	0.397	0.970	2.286	4.45	6.736	21.00	33.94
D	310	66.320	2.330	5.11	0.246	0.600	0.580	0.990	2.416	2.37	4.786	9.45	50.48
D	335	67.000	3.240	5.78	0.401	0.460	0.840	0.890	2.591	4.44	7.031	13.51	36.85
E	10	94.600	28.250	4.32	0.202	0.333	4.789	12.232	17.556	16.53	34.086	90.59	51.51
E	35	84.500	16.120	5.07	0.276	0.600	2.960	10.100	13.936	10.13	24.066	56.31	57.91
E	60	83.650	20.430	4.17	0.389	0.219	3.323	11.800	15.731	8.79	24.521	65.38	64.15
E	85	88.960	11.450	4.51	0.450	0.769	2.567	7.209	10.995	6.70	17.695	40.60	62.14
E	110	78.700	11.240	5.87	0.479	0.742	1.240	10.701	13.162	5.13	18.292	40.77	71.95
E	135	83.700	17.890	5.82	0.591	0.673	0.880	8.132	10.276	3.40	13.676	49.46	75.14
E	160	82.300	16.960	5.74	0.278	0.430	1.630	3.562	5.900	4.10	10	43.92	59.00
E	185	78.500	23.550	4.86	0.287	0.772	2.461	3.763	7.283	11.32	18.603	65.70	39.15
E	210	84.500	25.610	3.95	0.236	0.820	5.020	14.209	20.285	14.46	34.745	85.97	58.30
E	235	88.790	27.650	4.51	0.172	0.604	3.263	10.876	14.915	12.00	26.915	82.22	55.42
E	260	62.370	12.010	4.44	0.450	0.763	1.877	12.301	15.391	7.50	22.8865	46.91	67.25
E	285	71.336	4.150	5.58	0.306	0.892	0.403	7.200	8.801	6.80	15.601	23.90	56.41
E	310	64.390	5.040	5.73	0.805	0.132	0.361	2.101	3.479	0.30	3.779	13.86	92.06
F	10	88.340	27.480	4.01	0.079	0.236	2.010	15.076	18.201	16.01	34.211	89.17	53.20
F	35	90.760	20.110	4.17	0.103	0.789	3.763	14.234	18.889	13.91	32.799	73.02	57.59
F	60	88.740	15.140	4.00	0.321	0.643	1.401	8.964	11.329	7.86	19.189	49.47	59.04
F	85	74.160	16.980	5.62	0.454	0.990	1.096	6.709	9.249	2.70	11.949	45.91	77.40
F	110	72.650	18.540	5.78	0.363	0.896	0.632	6.276	8.167	4.84	13.007	50.09	62.79
F	135	74.160	10.540	5.69	0.551	0.436	0.841	5.100	6.936	4.64	11.576	32.66	59.92
F	160	83.930	16.660	5.98	0.632	0.206	1.460	6.676	8.974	4.01	12.984	46.30	69.12
F	185	87.000	19.450	4.87	0.341	0.434	2.781	9.730	13.286	13.32	26.606	65.51	49.94
F	210	70.320	20.230	4.92	0.198	0.732	2.790	13.235	16.955	19.32	36.275	76.74	46.74
F	235	69.050	24.550	3.32	0.123	0.401	1.543	11.789	13.856	11.33	25.186	74.29	55.01
F	260	62.320	12.110	4.61	0.204	0.976	1.206	7.432	9.818	5.63	15.440	39.67	63.56
F	285	61.160	7.030	5.01	0.306	0.760	0.310	2.360	3.736	3.41	7.146	21.21	52.28
G	10	84.000	18.120	4.22	0.297	0.362	1.809	13.736	16.204	12.90	29.104	65.34	55.68
G	35	86.790	20.360	4.14	0.202	0.071	2.330	7.204	10.607	8.30	18.907	59.63	56.10
G	60	92.360	16.330	4.00	0.476	0.760	2.760	12.213	16.225	9.33	25.555	58.22	63.49
G	85	80.360	15.460	4.00	0.659	0.609	1.732	8.690	11.690	3.02	14.71	45.63	79.47
G	110	81.490	15.780	5.98	0.674	0.236	1.570	2.400	4.800	2.40	7.28	38.04	67.03
G	135	73.630	13.560	5.64	0.833	0.806	2.800	1.276	5.715	2.63	8.345	35.47	68.48
G	160	72.920	20.250	4.22	0.202	0.239	2.109	0.430	2.900	7.80	10.78	51.28	27.64

SAMPLE	DEPTH	MOISTURE	ORGANZ	PH	NA	K	MG	CA	TEB	EXCH.H	CEC(MIN)	CEC	B.SZ
G	185	69.230	19.330	4.51	0.398	0.768	4.436	7.100	12.702	9.00	21.702	60.36	58.53
G	210	70.130	13.220	4.67	0.350	0.818	2.505	6.860	10.533	10.87	21.403	47.84	49.21
G	235	68.030	4.040	4.89	0.840	0.767	1.286	3.140	6.033	2.06	8.093	16.17	74.55
DTA	10	73.600	15.240	5.02	0.368	0.101	2.345	2.852	5.666	7.30	12.966	43.45	43.70
DTA	35	76.790	2.580	5.14	0.340	0.583	1.686	1.991	4.600	1.73	6.33	11.49	72.67
DTA	60	68.260	1.150	5.16	0.385	0.891	0.534	0.768	2.578	1.40	3.978	6.28	64.81
DTB	10	83.050	20.620	4.09	0.406	0.325	4.679	3.843	9.253	12.47	21.723	62.96	42.60
DTB	35	87.360	14.500	5.32	0.432	0.712	1.987	11.916	15.047	10.42	25.467	54.47	59.08
DTB	60	75.460	15.760	5.29	0.576	0.583	2.179	5.359	8.697	5.76	14.457	45.98	60.16
DTB	85	78.950	10.320	4.90	0.502	0.789	4.412	2.650	8.353	4.03	12.383	49.02	67.46
DTB	110	78.270	10.360	3.62	0.541	0.568	5.372	3.782	10.263	6.40	16.663	53.38	61.59
DTB	135	63.270	7.090	4.51	0.662	0.143	3.325	9.421	13.551	8.89	22.441	36.62	60.39
DTB	160	62.300	4.030	5.73	0.489	0.789	2.232	1.003	4.513	3.65	8.163	16.22	55.29
DTC	10	78.650	21.910	3.80	0.308	0.301	3.851	11.859	16.319	18.92	35.239	79.06	46.31
DTC	35	84.220	19.870	3.41	0.125	0.834	1.658	11.002	13.619	17.50	31.119	70.86	43.76
DTC	60	86.550	11.030	4.21	0.331	0.000	2.542	3.462	6.343	8.23	14.573	36.63	43.53
DTC	85	81.090	17.970	5.32	0.272	0.224	1.046	1.445	2.987	6.90	9.887	45.83	30.21
DTC	110	75.380	13.450	5.63	0.292	0.360	1.240	2.936	4.828	4.50	9.328	36.23	51.76
DTC	135	79.280	13.240	5.89	0.642	0.136	1.470	1.906	4.154	2.40	6.554	33.03	63.38
DTC	160	73.290	18.090	5.54	0.419	0.890	0.762	2.972	5.043	7.60	12.643	48.82	39.89
DTC	185	80.660	13.360	4.60	0.438	0.312	2.632	6.843	10.225	9.00	19.225	45.95	53.19
DTC	210	80.990	28.690	3.10	0.298	0.763	3.606	8.094	12.761	11.01	23.771	81.15	53.68
DTC	235	78.840	26.640	3.70	0.452	0.052	2.904	6.795	10.203	12.75	22.953	76.23	44.45
DTC	260	69.440	19.820	4.19	0.703	0.403	4.863	8.793	14.762	7.60	22.362	62.00	66.01
DTC	285	64.950	10.790	4.12	0.780	0.860	2.765	2.094	6.499	3.03	9.529	31.11	68.20
DTC	310	64.430	5.830	5.25	0.082	0.880	2.156	1.894	5.012	1.30	6.312	17.97	79.40
DTD	10	88.320	26.360	3.01	0.376	0.828	4.552	7.632	13.388	17.40	30.788	83.51	43.48
DTD	35	74.460	25.300	3.98	0.203	0.487	3.443	7.789	11.922	12.42	24.342	74.94	48.98
DTD	60	71.050	18.000	3.90	0.463	0.993	1.518	4.230	7.204	14.73	21.934	57.93	32.84
DTD	85	76.450	13.310	4.82	0.530	0.308	0.969	6.376	8.183	4.01	12.193	38.81	67.11
DTD	110	76.330	15.370	4.51	0.409	0.993	0.815	4.776	6.993	3.75	10.743	41.48	65.09
DTD	135	87.440	18.100	5.01	0.278	1.639	0.793	1.170	3.880	6.03	9.91	46.11	39.15
DTD	160	89.680	19.160	5.17	0.432	0.360	1.439	1.643	3.874	7.11	10.984	49.30	35.27
DTD	185	72.010	24.010	4.97	0.389	0.507	0.433	6.470	7.799	11.32	19.119	67.14	40.79
DTD	210	83.230	27.620	4.02	0.211	0.463	2.888	8.370	11.932	17.00	28.932	84.17	41.24
DTD	235	87.360	17.070	4.22	0.432	0.143	2.760	9.847	13.182	14.80	27.982	62.12	47.11
DTD	260	77.560	2.310	5.35	0.541	0.507	1.800	2.232	5.080	8.14	13.22	17.84	38.43
DTD	285	68.800	3.000	5.39	0.646	0.520	0.959	0.628	2.753	4.00	6.753	12.75	40.77
DTE	10	86.800	24.080	4.32	0.271	0.460	1.222	10.590	12.543	16.40	28.943	77.10	43.34
DTE	35	83.190	18.880	4.63	0.443	0.556	2.207	10.491	13.697	15.30	28.997	66.76	47.24

SAMPLE	DEPTH	MOISTURE	ORGANZ	PH	NA	K	MG	CA	TEB	EXCH.H	CEC(MIN)	CEC	B.S%
DTE	60	71.800	13.880	3.82	0.400	0.918	1.983	4.675	7.976	18.00	25.976	53.74	30.71
DTE	85	70.100	12.100	5.90	0.430	0.455	1.808	3.263	5.956	6.13	12.086	36.29	49.28
DTE	110	88.070	10.980	5.18	0.448	0.979	0.620	2.739	4.786	7.03	11.814	33.77	40.51
DTE	135	80.500	15.830	5.08	0.498	1.019	1.724	0.284	3.525	4.32	7.845	39.51	44.93
DTE	160	82.560	10.630	4.43	0.323	0.885	1.337	0.471	3.016	7.67	10.686	31.95	28.22
DTE	185	72.660	15.570	4.28	0.243	0.630	1.469	3.876	6.218	7.00	13.218	44.36	47.04
DTE	210	76.540	21.480	4.79	0.728	0.079	2.979	15.812	19.598	10.84	30.438	73.40	64.39
DTE	235	65.560	14.310	4.80	0.604	0.940	1.730	6.340	9.614	11.10	20.714	49.33	46.41
DTE	260	67.070	2.590	5.81	0.823	0.440	1.993	0.867	4.123	2.00	6.123	11.30	67.34
DTF	10	72.360	17.320	4.51	0.704	0.139	2.232	4.470	7.545	12.62	20.165	54.81	37.42
DTF	35	82.490	18.090	4.56	0.541	0.446	1.763	5.190	7.940	11.05	18.99	55.17	41.81
DTF	60	62.380	11.010	5.72	0.467	0.279	1.608	0.460	2.814	7.83	10.644	32.66	26.44
DTF	85	70.210	12.620	5.63	0.472	0.232	0.987	0.160	1.851	6.32	8.171	33.41	22.65
DTF	110	77.320	15.390	5.54	0.489	0.099	1.388	1.120	3.096	3.69	6.786	37.57	45.62
DTF	135	88.300	10.910	5.29	0.401	0.672	1.760	2.320	5.153	4.80	9.953	31.77	51.77
DTF	160	81.440	19.200	4.81	0.871	0.090	1.089	3.390	5.440	12.75	18.19	56.59	29.91
DTF	185	71.060	17.630	4.47	0.432	0.080	3.406	3.710	7.628	17.32	24.948	60.21	30.58
DTF	210	76.540	23.030	4.63	0.330	0.383	5.343	9.890	15.946	14.79	30.736	76.00	51.88
DTF	235	81.020	11.020	4.73	0.487	0.671	3.937	6.260	11.355	5.01	16.365	38.41	69.39
DTF	260	75.330	4.000	5.74	0.542	0.763	2.176	0.380	3.861	4.54	8.401	16.40	45.96
DTG	10	70.370	12.870	4.17	0.397	0.907	3.389	5.506	10.199	14.30	24.499	50.24	41.63
DTG	35	72.410	7.480	4.25	0.231	0.800	4.576	4.650	10.257	10.56	20.817	35.78	49.27
DTG	60	76.550	4.570	4.61	0.596	0.096	2.563	1.209	4.464	4.40	8.864	18.00	50.36
DTG	85	61.050	4.620	5.51	0.754	0.982	2.364	1.340	5.440	2.32	7.76	17.00	70.10
DTG	110	60.540	2.630	5.55	0.608	0.114	1.348	0.100	2.170	0.12	2.29	7.55	94.76
STA	10	62.540	12.320	5.12	0.391	0.731	2.631	4.972	8.725	12.10	20.825	45.47	41.90
STA	35	64.630	10.820	5.19	0.387	0.331	2.903	3.631	7.252	10.13	17.382	39.02	41.72
STA	60	57.150	10.910	5.20	0.403	0.576	1.021	2.742	4.742	8.04	12.782	34.60	37.10
STA	85	58.950	7.630	5.64	0.526	0.480	1.332	0.036	2.374	4.20	6.574	21.83	36.11
STA	110	55.320	6.050	5.50	0.529	0.236	0.431	0.987	2.183	1.30	3.483	15.58	62.68
STB	10	52.300	18.550	4.21	0.576	0.236	3.030	7.763	11.605	10.14	21.745	58.85	53.37
STB	35	67.780	16.300	5.08	0.631	0.918	2.906	5.541	9.996	7.60	17.596	50.20	56.81
STB	60	52.300	11.090	5.27	0.602	0.830	1.232	12.236	14.900	16.93	31.83	54.01	46.81
STB	85	48.230	12.320	5.22	0.790	0.432	1.408	4.709	7.339	9.90	17.239	41.88	42.57
STB	110	48.900	8.670	5.57	0.433	0.460	0.892	2.031	3.816	6.83	10.646	27.99	35.84
STB	135	40.000	9.760	5.43	0.416	0.483	0.460	1.642	3.001	6.00	9.001	28.52	33.34
STB	160	54.600	14.560	4.68	0.423	0.132	0.532	11.541	12.628	8.50	21.128	50.25	59.77
STB	185	41.830	12.090	4.59	0.389	0.313	1.409	11.732	13.843	14.00	27.843	52.02	49.72
STB	210	48.630	7.060	5.76	0.241	0.897	1.232	7.916	10.286	11.43	21.716	35.84	47.37
STB	235	48.100	1.950	5.10	0.636	0.362	0.471	2.555	4.024	4.80	8.824	12.72	45.60

SAMPLE	DEPTH	MOISTURE	ORGANZ	PH	NA	K	MG	CA	TEB	EXCH.H	CEC(MIN)	CEC	B.SZ
STC	10	55.750	23.950	4.80	0.689	0.170	3.124	12.232	16.215	21.00	37.215	85.12	43.57
STC	35	58.420	19.980	4.85	0.626	0.584	4.368	14.495	20.073	18.50	38.573	78.53	52.04
STC	60	60.790	18.630	5.39	0.590	0.611	2.967	12.857	17.005	11.06	28.065	65.33	60.59
STC	85	47.090	9.350	5.57	0.858	0.383	0.643	2.690	4.574	7.60	12.174	30.87	37.57
STC	110	48.330	6.570	5.43	0.848	0.879	1.630	0.465	3.822	9.10	12.922	26.06	29.58
STC	135	48.970	5.240	6.20	0.741	0.707	1.457	0.732	3.637	6.40	10.037	20.52	36.24
STC	160	34.650	14.100	4.52	0.370	0.106	1.163	1.845	3.484	16.86	20.344	48.54	17.13
STC	185	39.250	14.680	4.56	0.378	0.783	2.465	3.693	7.319	16.01	23.329	52.69	31.37
STC	210	39.250	19.550	4.23	0.496	0.350	3.787	4.732	9.365	11.20	20.565	59.67	45.54
STC	235	40.230	8.660	5.64	0.593	0.203	0.753	2.295	3.844	4.80	8.644	25.96	44.47
STC	260	38.750	3.180	5.85	0.996	0.073	1.152	1.470	3.691	5.00	8.691	15.05	42.47
STD	10	48.970	18.730	4.89	0.186	0.632	2.889	4.438	8.145	11.23	19.375	56.84	42.04
STD	35	50.180	14.300	4.90	0.264	0.540	1.870	4.236	6.910	11.80	18.71	47.31	36.93
STD	60	48.570	8.760	5.23	0.786	0.763	1.600	3.219	6.368	17.40	23.768	41.29	26.79
STD	85	52.140	7.630	5.26	0.826	0.233	1.340	2.800	5.199	4.20	9.399	24.66	55.31
STD	110	44.560	7.320	5.97	0.636	0.892	1.200	1.630	4.358	1.30	5.658	20.30	77.02
STD	135	44.340	8.730	5.09	0.736	0.791	2.641	0.860	5.028	6.38	11.408	28.87	44.07
STD	160	45.690	16.400	4.32	0.386	0.286	3.201	8.568	12.441	13.43	25.871	58.67	48.09
STD	185	37.480	10.230	4.35	0.230	0.556	3.100	9.793	13.679	10.90	24.579	45.04	55.65
STD	210	36.330	3.190	5.44	0.541	0.724	1.110	1.750	4.125	3.90	8.025	14.41	51.40
STD	235	41.330	2.320	6.42	0.792	0.386	0.700	1.624	3.502	2.21	5.712	10.35	61.31
STE	10	47.980	7.630	4.88	0.304	0.832	3.763	0.763	5.662	8.03	13.692	28.95	41.35
STE	35	32.390	2.840	6.17	0.487	0.763	0.790	0.208	2.248	4.40	6.648	12.33	33.81
STE	60	35.600	2.190	6.21	0.887	0.080	0.436	3.208	4.611	3.11	7.721	12.10	59.72

SAMPLE	DEPTH	MEDIUM SAND	FINE SAND	TOTAL SAND	SILT	CLAY	COLOUR
A	10	9	24	33	11	56	2.5 Y 4/2
A	35	9	25	34	12	54	2.5 YR 4/2
A	60	10	35	45	13	42	2.5 YR 2/1
A	85	14	34	48	11	41	10 YR 2/1
A	110	15	34	49	11	40	10 YR 3/1
A	135	10	28	38	15	47	10 YR 3/2
A	160	8	31	39	9	52	10 YR 2/1
A	185	8	34	42	8	50	5 YR 2.5/1
A	210	7	18	25	7	68	5 YR 2.5/1
A	235	8	19	27	12	61	10 YR 5/2
A	260	7	22	29	13	58	2.5 Y 6/2
A	285	18	33	51	16	33	2.5 Y 6/2
A	310	21	40	61	15	24	2.5 Y 7/1
B	10	2	20	22	17	61	2.5 Y 2/1
B	35	7	27	34	16	50	2.5 Y 2/1
B	60	12	33	45	16	39	10 YR 2/1
B	85	11	36	47	15	38	10 YR 3/1
B	110	12	32	44	10	46	2.5 Y 3/2
B	135	14	34	48	8	44	2.5 Y 3/1
B	160	11	37	48	7	45	2.5 Y 3/1
B	185	13	33	46	6	48	10 YR 2/1
B	210	8	32	40	16	44	10 YR 2/1
B	235	6	21	27	18	55	10 YR 3/1
B	260	7	15	22	20	58	10 YR 3/1
B	285	5	15	20	23	57	2.5 Y 6/2
B	310	11	30	41	18	41	2.5 Y 6/2
B	335	14	38	52	13	35	2.5 Y 6/2
B	360	18	42	60	16	24	2.5 Y 7/1
C	10	3	22	25	15	60	10 YR 2/1
C	35	4	26	30	19	51	10 YR 2/1
C	60	7	36	43	17	40	10 YR 2/1
C	85	10	34	44	9	47	10 YR 3/1
C	110	13	32	45	8	47	7.5 YR 3/0
C	135	12	37	49	5	46	7.5 YR 3/0
C	160	11	35	46	7	47	7.5 YR 2/0
C	185	8	28	36	15	49	7.5 YR 2/0
C	210	5	25	30	13	57	7.5 YR 2/0
C	235	8	20	28	18	54	10 YR 2/1
C	260	8	31	39	11	50	10 YR 2/1
C	285	12	37	49	14	37	10 YR 5/2
C	310	16	42	58	18	24	10 YR 6/2
C	335	19	43	62	14	24	10 YR 7/3
D	10	9	21	30	14	56	10 YR 2/1
D	35	6	25	31	14	55	10 YR 2/1

SAMPLE	DEPTH	MEDIUM SAND	FINE SAND	TOTAL SAND	SILT	CLAY	COLOUR
D	60	8	36	44	16	40	10 YR 2/2
D	85	11	37	48	8	44	10 YR 2/2
D	110	7	36	43	10	47	10 YR 4/2
D	135	10	34	44	8	48	10 YR 4/2
D	160	9	34	43	4	53	10 YR 2/2
D	185	8	21	29	14	57	10 YR 2/1
D	210	7	18	25	12	63	7.5 YR 2/0
D	235	6	20	26	15	59	7.5 YR 2/0
D	260	5	25	30	25	45	7.5 YR 3/2
D	285	9	37	46	22	32	7.5 YR 7/2
D	310	14	42	56	15	29	7.5 YR 7/0
D	335	19	41	60	16	24	7.5 YR 7/0
E	10	8	21	29	15	56	10 YR 2/1
E	35	4	24	28	12	60	10 YR 2/1
E	60	4	24	28	14	58	10 YR 2/2
E	85	13	29	42	9	49	10 YR 2/2
E	110	12	36	48	10	42	10 YR 4/2
E	135	11	36	47	9	44	7.5 YR 3/2
E	160	10	30	40	6	54	7.5 YR 3/2
E	185	8	20	28	8	64	7.5 YR 2/0
E	210	6	21	27	15	58	7.5 YR 2/0
E	235	3	25	28	24	48	10 YR 2/1
E	260	12	27	39	24	37	10 YR 5/2
E	285	15	39	54	20	26	10 YR 6/2
E	310	22	42	64	18	18	10 YR 7/3
F	10	9	21	30	16	54	10 YR 2/1
F	35	4	25	29	17	54	10 YR 2/1
F	60	6	32	38	21	41	10 YR 2/2
F	85	11	30	41	14	45	10 YR 2/2
F	110	13	35	48	10	42	10 YR 2/2
F	135	14	40	54	2	44	10 YR 4/2
F	160	13	29	42	8	50	10 YR 3/2
F	185	9	21	30	16	54	10 YR 2/1
F	210	7	24	31	22	47	10 YR 2/1
F	235	8	23	31	28	41	7.5 YR 2/0
F	260	18	39	57	15	28	7.5 YR 5/2
F	285	24	40	64	18	18	7.5 YR 7/0
G	10	8	20	28	17	55	2.5 Y 2/1
G	35	9	24	33	21	46	2.5 Y 2/1
G	60	10	41	51	5	44	2.5 Y 3/2
G	85	11	39	50	9	41	7.5 YR 4/0
G	110	13	38	51	9	40	7.5 YR 4/0
G	135	12	37	49	7	44	7.5 YR 7/2
G	160	9	27	36	10	54	2.5 Y 2/1

SAMPLE	DEPTH	MEDIUM SAND	FINE SAND	TOTAL SAND	SILT	CLAY	COLOUR
G	185	10	24	34	18	48	2.5 Y 2/1
G	210	12	39	51	6	43	2.5 Y 3/2
G	235	23	42	65	15	20	7.5 YR 7/0
DTA	10	12	25	37	18	45	2.5 Y 3/0
DTA	35	9	36	45	11	44	2.5 Y 7/0
DTA	60	13	44	57	17	26	2.5 Y 7/0
DTB	10	8	23	31	19	50	2.5 Y 2/0
DTB	35	9	25	34	19	47	2.5 Y 3/0
DTB	60	11	38	49	9	42	2.5 Y 3/2
DTB	85	11	37	48	12	40	2.5 Y 5/2
DTB	110	6	30	36	16	48	2.5 Y 5/3
DTB	135	5	38	43	15	42	2.5 Y 6/2
DTB	160	15	41	56	16	28	2.5 Y 6/2
DTC	10	7	26	33	18	49	10 YR 2/1
DTC	35	9	24	33	20	47	10 YR 2/1
DTC	60	8	31	39	16	45	10 YR 3/1
DTC	85	10	42	52	9	39	10 YR 2/1
DTC	110	9	38	47	11	42	10 YR 5/2
DTC	135	10	37	47	12	41	10 YR 5/3
DTC	160	11	28	39	7	54	10 YR 3/1
DTC	185	8	22	30	20	50	10 YR 2/1
DTC	210	4	22	26	22	52	10 YR 2/1
DTC	235	6	23	29	24	47	10 YR 2/1
DTC	260	9	29	38	21	41	10 YR 2/1
DTC	285	15	43	58	24	18	10 YR 5/2
DTC	310	24	41	65	19	16	10 YR 7/2
DTD	10	7	21	28	22	50	10 YR 2/1
DTD	35	5	25	30	21	49	10 YR 2/1
DTD	60	7	36	43	15	42	10 YR 3/1
DTD	85	12	38	50	9	41	2.5 YR 3/2
DTD	110	14	38	52	8	40	2.5 YR 3/2
DTD	135	10	39	49	9	42	2.5 YR 3/2
DTD	160	9	27	36	8	56	10 YR 3/1
DTD	185	8	21	29	11	60	10 YR 2/1
DTD	210	5	20	25	13	62	10 YR 2/1
DTD	235	6	20	26	27	47	10 YR 3/1
DTD	260	17	42	59	16	25	10 YR 6/3
DTD	285	22	42	64	20	16	10 YR 6/3
DTE	10	6	25	31	17	52	10 YR 2/1
DTE	35	8	20	28	22	50	10 YR 2/1

SAMPLE	DEPTH	MEDIUM SAND	FINE SAND	TOTAL SAND	SILT	CLAY	COLOUR
DTE	60	9	27	36	20	44	10 YR 3/1
DTE	85	8	38	46	9	45	10 YR 3/1
DTE	110	13	35	48	13	39	10 YR 4/2
DTE	135	10	34	44	10	46	10 YR 4/2
DTE	160	7	17	24	15	61	10 YR 3/1
DTE	185	8	21	29	19	52	10 YR 2/1
DTE	210	4	20	24	33	43	10 YR 2/1
DTE	235	12	37	49	18	33	10 YR 4/2
DTE	260	23	44	67	16	17	10 YR 7/2
DTF	10	5	21	26	24	50	10 YR 3/1
DTF	35	8	18	26	19	55	10 YR 2/1
DTF	60	7	26	33	21	46	10 YR 3/1
DTF	85	10	39	49	9	42	2.5 YR 3/2
DTF	110	8	34	42	16	42	2.5 YR 3/2
DTF	135	12	26	38	11	51	2.5 Y 4/2
DTF	160	8	22	30	5	65	2.5 Y 3/0
DTF	185	6	24	30	13	57	10 YR 2/1
DTF	210	4	24	28	30	42	10 YR 2/1
DTF	235	13	36	49	31	20	10 YR 3/2
DTF	260	21	44	65	11	24	10 YR 6/2
DTG	10	5	26	31	17	52	10 YR 3/1
DTG	35	6	28	34	18	48	2.5 YR 3/2
DTG	60	9	29	38	11	51	2.5 Y 4/2
DTG	85	15	32	47	14	39	2.5 Y 6/2
DTG	110	19	36	55	13	32	2.5 Y 7.2
STA	10	6	27	33	19	48	10 YR 5/2
STA	35	4	24	28	21	51	10 YR 4/1
STA	60	8	37	45	11	44	10 YR 4/3
STA	85	11	38	49	10	41	10 YR 7/2
STA	110	14	35	49	9	42	10 YR 7/2
STB	10	9	27	36	20	44	10 YR 2/2
STB	35	5	29	34	19	47	10 YR 3/3
STB	60	6	34	40	20	40	10 YR 3/3
STB	85	9	37	46	15	39	10 YR 4/3
STB	110	11	37	48	10	42	2.5 Y 7/2
STB	135	9	26	35	14	51	2.5 Y 6/4
STB	160	8	24	32	12	56	10 YR 3/4
STB	185	11	26	37	18	45	10 YR 3/1
STB	210	16	34	50	19	31	10 YR 7/4
STB	235	25	39	64	14	22	10 YR 8/2

SAMPLE	DEPTH	MEDIUM SAND	FINE SAND	TOTAL SAND	SILT	CLAY	COLOUR
STC	10	5	27	32	17	51	10 YR 2/2
STC	35	7	26	33	19	48	5 Y 3/2
STC	60	8	32	40	16	44	10 YR 2/2
STC	85	12	38	50	5	45	10 YR 4/2
STC	110	10	37	47	11	42	10 YR 7/4
STC	135	11	40	51	3	46	10 YR 7/3
STC	160	6	27	33	17	50	10 YR 4/4
STC	185	7	18	25	18	57	10 YR 3/2
STC	210	8	23	31	25	44	10 YR 4/4
STC	235	17	36	53	7	40	10 YR 8/4
STC	260	23	39	62	0	38	10 YR 8/3
STD	10	7	21	28	25	47	5 Y 2.5/1
STD	35	8	29	37	16	47	5 Y 2.5/2
STD	60	11	32	43	13	44	5 Y 3/2
STD	85	10	39	49	11	40	10 YR 4/2
STD	110	12	38	50	9	41	10 YR 4/1
STD	135	10	27	37	16	47	10 YR 4/1
STD	160	7	24	31	13	56	5 Y 3/1
STD	185	7	21	28	16	56	5 Y 3/2
STD	210	16	35	51	7	42	5 Y 7/1
STD	235	24	39	63	5	32	5 Y 8/2
STE	10	11	27	38	19	43	10 YR 5/2
STE	35	15	34	49	10	41	10 YR 6/2
STE	60	19	37	56	6	38	10 YR 7/2

SAMPLE	DEPTH	MOISTURE	ORGAN%	PH	NA	K	MG	CA	TEB	EXCH.H	CEC(MIN)	CEC	B.5%
1	10	31.063	10.630	5.30	0.325	0.577	1.710	2.708	5.320	3.76	9.08	30.34	58.59
	35	20.262	14.190	5.80	0.330	0.362	1.019	2.487	4.198	2.48	6.678	35.06	62.86
	60	20.459	6.230	5.89	0.362	0.358	1.558	2.694	4.972	3.37	8.342	20.80	59.60
	85	18.897	3.170	6.20	0.472	0.306	1.821	3.590	6.189	1.03	7.219	13.56	85.73
	110	18.860	1.890	6.31	0.368	0.543	2.045	3.052	6.008	2.01	8.018	11.80	74.93
	135	20.630	2.090	6.15	0.427	0.676	2.037	3.030	6.170	1.87	8.04	12.22	76.74
2	10	31.030	10.320	5.21	0.418	0.436	0.891	1.632	3.377	2.00	5.377	26.02	62.80
	35	29.870	12.080	5.44	0.420	0.899	1.536	2.000	4.855	4.30	9.155	33.32	53.03
	60	23.631	9.030	5.57	0.462	1.063	1.808	2.809	6.142	3.54	9.682	27.74	63.44
	85	22.014	2.620	5.68	0.445	0.689	2.132	4.436	7.702	2.80	10.502	15.74	73.34
3	10	28.110	12.120	4.96	0.470	1.203	1.073	1.109	3.855	2.32	6.175	30.42	62.43
	35	15.240	9.110	5.12	0.503	1.468	1.289	1.889	5.149	4.63	9.779	28.00	52.65
	60	19.870	8.860	5.08	0.489	1.876	1.303	2.043	5.711	3.28	8.991	26.71	63.52
	85	21.020	3.320	5.05	0.512	1.299	1.873	2.013	5.697	8.10	13.797	20.44	41.29
	110	29.376	0.630	5.18	0.467	0.109	2.868	3.368	6.812	1.27	8.082	9.34	84.29
4	10	28.340	9.360	5.54	0.480	0.363	0.863	1.711	3.417	6.32	9.737	28.46	35.09
	35	16.360	10.470	5.61	0.412	0.643	1.230	2.400	4.685	7.70	12.385	33.33	37.83
	60	18.790	7.930	6.89	0.417	0.927	1.863	1.876	5.083	2.90	7.983	23.84	63.67
	85	26.920	1.620	6.13	0.362	0.708	2.102	2.901	6.073	2.30	8.373	11.61	72.53
5	10	19.630	9.530	5.01	0.354	0.320	0.888	1.232	2.794	6.20	8.994	28.05	31.07
	35	20.710	5.910	5.03	0.401	0.800	1.768	1.103	4.072	5.60	9.672	21.49	42.10
	60	28.790	0.710	5.26	0.376	0.640	1.970	2.845	5.831	4.00	9.832	11.25	59.31
6	10	22.650	13.430	5.21	0.423	0.894	1.634	1.214	4.165	2.10	6.265	33.13	66.48
	35	18.010	9.760	5.27	0.476	0.963	2.345	2.012	5.796	3.30	9.096	28.62	63.72
	60	22.230	6.330	5.26	0.489	0.876	1.912	2.562	5.839	4.34	10.179	22.84	57.36
	85	28.730	6.810	5.32	0.272	0.576	2.601	3.765	7.214	1.73	8.944	22.56	80.66
	110	27.630	2.200	4.59	0.187	0.580	2.634	3.214	6.615	1.50	8.115	12.52	81.52
7	10	32.360	14.090	5.84	0.894	1.879	2.236	2.108	7.117	4.41	11.527	39.71	61.74
	35	38.790	7.630	5.89	0.808	2.132	2.109	4.576	9.625	2.20	11.825	27.09	81.40
	60	27.320	1.450	6.18	0.663	1.870	4.006	3.231	9.770	1.20	10.97	13.87	89.06
	85	27.600	0.320	6.20	0.076	1.643	3.839	4.896	10.454	1.04	11.494	12.13	90.95

SAMPLE	DEPTH	MEDIUM SAND	FINE SAND	TOTAL SAND	SILT	CLAY
1	10	5	21	26	26	48
	35	4	14	18	30	52
	60	6	16	22	20	58
	85	2	12	14	19	67
	110	1	10	11	21	68
	135	0	11	11	26	63
2	10	4	22	26	31	43
	35	3	18	21	29	50
	60	2	9	11	28	61
	85	2	12	14	22	64
3	10	2	24	26	25	49
	35	3	19	22	32	46
	60	1	21	22	23	55
	85	2	12	14	28	58
	110	1	7	8	29	63
4	10	5	26	31	28	41
	35	4	21	25	35	40
	60	1	17	18	24	58
	85	1	11	12	23	65
5	10	3	23	26	28	46
	35	2	19	21	24	55
	60	2	14	16	24	60
6	10	3	25	28	32	40
	35	4	27	31	13	56
	60	2	13	15	31	54
	85	1	14	15	24	61
	110	2	9	11	21	68
7	10	5	23	28	28	44
	35	4	17	21	25	54
	60	2	15	17	18	65
	85	3	9	12	24	64

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