by

ERIC ARTHUR WOLFERSTAN TORDIFFE

Submitted in fulfilment of the requirements for the degree of

DOCTOR OF PHILOSOPHY

in the
Faculty of Science,
Department of Geology,
University of the Orange Free State

December, 1978

Promotor:
Professor B.J.V. Botha

ABSTRACT

The aim of this study was to examine some of the major aspects responsible for the chemical quality of the groundwater in the Great Fish River Basin and its influence on the irrigation water. Approximately 18 000 ha of land are at present irrigated from several weirs down the river.

The section of the Great Fish River Basin under discussion comprises an area of approximately 25 000 km^2 located between longitudes 25^oE to 27^oE and latitudes 31^o 15′S to 33^o 15′S. This area is devided into the following main geomorphologic provinces: The Marginal Region (lower than 760 m), the Great Escarpment (750 - 1070 m), the Headbasin (1070 - 1370 m) and the Interior Plateau (higher than 1370 m). Each of these provinces play an important part in controling the movement and the chemical quality of groundwater in the area.

Most of the annual precipitation (350 - 450 mm) occurs between February and March when evapotranspiration is at its highest. Runoff from the entire basin amounts to only 3 percent of the annual precipitation. The rest of the water either evaporates immediately because of the semi-arid climatic conditions, or is temporarily stored in the soil before it is lost to the atmosphere by means of evapotranspiration. It is also pointed out that apart from periods of extreme precipitation, the monthly evapotranspiration always exceeds the monthly precipitation.

Such semi-arid climatic conditions, as well as the nature of

the soils in the area prohibit a fast infiltration of meteoric water and it is therefore doubted whether as much as 5 percent of the annual precipitation ever reaches the groundwater table.

The area under discussion is underlain by sedimentary rocks of the Karoo Sequence beginning with the glacial deposits of the Dwyka Tillite Formation (680 m) at the bottom, followed by the marine deposits of the Ecca Group (2340 m), the transitional deposits of the Koonap Formation (980 m) and the fluvial deposits of the Beaufort Group (4540 m). Because the Koonap Formation represents the transition between the marine (deltaic) deposits of the Ecca Group and the fluvial deposits of the Beaufort Group, it is regarded as a separate formation not belonging to either group. The Beaufort Group on account of the environment in which the sediments were deposited, is subdivided into the Adelaide Subgroup (reducing environment) and the Tarkastad Subgroup (oxidizing environment). Red mudstone is regarded as indicative of an oxidizing environment and is present only in patches in the Middleton Formation, which forms the lower part of the Adelaide Subgroup. No red mudstone is present in the Balfour Formation, which forms the top half of this subgroup, but becomes very prominent in the Katberg and Burgersdorp Formations of the Tarkastad Subgroup. The Balfour Formation, on lithologic grounds, is subdivided into the Oudeberg Sandstone Member (180 m), The Daggaboersnek Member (1200 m), the Barberskrans Sandstone Member (190 m) and the Elandsberg Member (320 m). It is suggested that the arenaceous units of the Beaufort Group, i.e. the Oudeberg Sandstone Member, the Barberskrans

Sandstone Member and the Katberg Formation represent periods of major tectonic activity in the provenance which was located to the south-east. During such activity vast amounts of coarse-grained material were transported and deposited at a relatively fast rate.

Owing to the semi-arid climatic conditions, which prevail in the area, the soils tend to be rather alkaline with a high clay content and the poor development of an A-horizon. Calcrete or caliche occurs at or near the surface of most of the soils.

Dolerite has intruded the sedimentary strata as concordant and conical sills, as well as near-vertical dykes. The dykes in the south of the area have an orientation of approximately 290°, coinciding with the Cape Fold Belt, whilst farther north a prominent northerly trend with a weaker easterly trend is observed. In the extreme north, where the sedimentary strata is at its thickest, an almost random orientation is present. Various types of dolerite are encountered in the area and of particular interest is the occurrence of quartz dolerite which has intruded a sill of normal dolerite near Speelmanskop. This leucocratic body is probably the result of magmatic differentiation lower down in the crust, whilst limited differentiation within the body itself, both from floor to roof and in an "up-dip" direction, must have occured.

The intrusion of the dolerite is of particular importance because of the fracture zones it causes in the adjacent sedimentary rocks. Such zones are normally open to circulating groundwater. Where the dolerite itself is not fractured it may act as an impervious barrier when crossing the regional flow path of the groundwater. In such cases groundwater compartments are developed.

Weathering of the provenance and of the various rock-types in the area, diagenetic processes which proceeded the deposition of the sediments in the Karoo Basin and the adsorption and ion exchange during the interaction of the surface and groundwater with the surrounding rocks, are considered to be the main geochemical factors responsible for changes in the chemical quality of the groundwater in the area. During the chemical weathering of the rock-forming minerals cations such as Na^+, K^+, Mg^{++} and Ca^{++} are released to solution in the groundwater, whilst compounds such as SiO_2 and Al_2O_3 regroup to form residual clay minerals such as montmorillonite. Weathering of the sedimentary rocks is, however, limited because of the fact that the primary minerals which constitute such rocks have already withstood at least one cycle of weathering in the provenance. In areas where leaching is vigorous, K^+ is, however, removed from illite in the mudstone with the result that this clay mineral adopts swelling features similar to montmorillonite, thus causing the rock to crumble. Dolerite in turn, because of its igneous origin, is more prone to chemical weathering.

As a result of compaction the porosity and permeability of the sediments in the Karoo Basin was reduced to extremely low values. The chemistry of the interstitial waters was also altered by this process because of the diagenetic alteration of montmorillonite to illite during which K^+ is removed from the water, whilst SiO_2, H_2O, Na^+, Ca^{++}, Mg^{++} and Fe^{++} are added to the water. During the compaction process, Cl^- was accumulated in the remaining water in the lower strata as result of ultra-filtration as the formation water was squeezed through clay-rich mudstone layers.

Because of its small ionic radius and high electrical charge, Ca^{++} is adsorbed by the clay minerals in the mudrock of the area to a far greater extent than any of the other cations present. The maximum concentration of adsorbed Ca^{++} is observed in the Oudeberg Sandstone Member, which suggests that this unit represents a geochemical marker. A gradual increase in the CEC of the mudrock from the lower strata to this unit is furthermore observed. Sodium concentrations increase toward the south of the study area, therefore suggesting an influence of the palaeomarine environment on the adsorbed cations.

Groundwater in the Great Fish River Basin is restricted mainly to joints in the sedimentary rocks and to fracture zones caused by the intrusion of dolerite. The water levels in most of the bore-holes therefore represent a pressure or piezometric surface rather than an actual water table. Such levels, however, regionally represent a surface which closely resembles the surface topography, whilst the flow of groundwater is down the regional slope and the rivers act as effluent drainage canals for the groundwater.

Although the groundwater is recharged in the higher lying areas by circulating meteoric water, there appears to be no direct relationship between the seasonal precipitation and the groundwater levels.

As far as the origin of the major ions in the groundwater is concerned, the cations are derived mainly from the weathering of primary rock-forming minerals, whilst the anions accumulate from non-lithologic sources. Generally, the groundwater in the areas of recharge, i.e. the higher lying areas, has a pronounced Ca^{++} and HCO_3^- - character, whilst in the stagnant low-lying areas Na^+ and Cl^- are the predominant ions. In between the two extremes, groundwater with a prominent Mg^{++} and $SC_4^=$ - character is encountered. This trend corresponds well with the normal metamorphism of natural waters and appears to be controlled largely by the topography of the area. Groundwater with a distinctly high Na^+ and Cl^- - concentration also has a high salinity concentration. The pH in turn is highest in the areas of high Ca^{++} and HCO_3^- - concentrations and lowest in the areas of high Na^+ and Cl^- - concentrations. All the water of the area is, however, oversaturated in relation to $CaCO_3$ and, where conditions are suitable, calcrete is precipitated.

Chloride is the dominant anion in the lower strata of the Karoo Sequence and is attributed mainly to the retention of this ion during the migration of the formation waters through the argillaceous material. High salinities, as a result of high Na^+ and Cl^- - concentrations, prevail in the groundwater up to the Daggaboersnek Member. From the Barberskrans Sandstone Member upward, the concentration of these ions decrease sharply. The cation percentages in the groundwater of the upper strata, however, vary considerably, thus indicating the influence of chemical weathering. There is more $SO_4^=$ in the groundwater of the lower strata, which was deposited under reducing conditions, than in the upper strata, which was deposited under oxidizing conditions. This is attributed to the formation of pyrite under reducing conditions, which can later oxidize to release $SO_4^=$ to the water.

During the periods of extreme precipitation a considerable amount of meteoric water infiltrates down to the groundwater level, dissolving precipitated salts on its way down. This naturally causes an increase in the salinity of the groundwater and is the result of an increase in Na^+ and Cl^-.

Seepage water in the Great Fish River contains Na^+ as the main cation and increases gradually in concentration farther downstream. To the north of Cradock HCO_3^- is the dominant anion but it decreases rapidly farther downstram, with a concurrent sharp increase in the Cl^- - concentration. The increase in the Na^+ and Cl^- - concentration coincides with an increase in the total salt load farther downstream.

A similar trend is observed in the change in groundwater quality down the Great Fish River. This is conclusive proof of the influence of groundwater on the seepage water in the river. The groundwater compartments caused by dolerite intrusions also have a marked influence on the quality of the seepage water.

During a single irrigation lead from Grassridge Dam the initial irrigation water reaching the consecutive weirs along the river possessed an extremely high salinity load as a result of the solution of precipitated salts in the river bed as well as the flushing of saline water from stagnant pools. The duration of the saline head increased at each consecutive weir downstream. Such conditions present a serious threat to the irrigable land along the Great Fish River and therefore measures will have to be taken to either prevent such contamination of the irrigation water or to limit the application of such contaminated water by allowing the saline head to pass the various weirs.

CONTENTS

page

FIGURES

Figure | Page

TABLES

PLATES

PLATE

1. INTRODUCTION

1. 1. GENERAL REVIEW

It is believed that the Great Fish River was originally named Rio do Infante after Juan Infante, a captain of one of the ships of Bartholomew Diaz. Juan Infante was the first to step ashore when these ships anchored at a river mouth on the east coast of South Africa in 1486.
The name Great Fish River is believed to have been derived from the Hottentot word "ou-b" or "au-b" meaning "fish" or from the Xhosa name "i-Nxuba" meaning "exit of the fish" (Du Plessis,1973, p. 229).

During their search for better pastures for their ever expanding herds of stock, the white colonists trekked further east and eventually met with the southward trekking Xhosas in 1778 in the Great Fish River Basin. Clashes between these two groups were inevitable as both were mainly dependant on their stock for survival.
Thus the first "Kaffir War" occured in 1779 and another eight were to follow from then until 1877 (Du Toit, 1954, p. 2). After the first conflict the Great Fish, the Baviaans and the Tarka Rivers were declared as the eastern boundary of the Cape Colony on the 14th November 1780 (Van der Walt _et al_., 1966, p. 98.)

Governors of the Cape Colony like Sir John Cradock then realised that the white population along the Fish River had to be increased in order to strengthen the defence of the eastern border against the Xhosas.

Lord Charles Somerset, who succeeded Cradock in 1814, also realised the importance of a strong white population on the eastern frontier and therefore many of the British Settlers of 1820 were given land in this area.

The Dutch Colonists, however, were discontented with the British rule of the Cape Colony. This dissatisfaction soon led to uprisingslike the Slagtersnek Rebellion in 1815 and eventually to the Great Trek in 1838.

The potential of irrigation farming along the banks of the Great Fish River was soon realised and as early as 1908 a number of small schemes, which made use of flood-water, were in operation. Some of the schemes, which are still in operation to this day are:- Katkop Dam (Knutsford Irrigation Board), Marlow Irrigation Project (Marlow Irrigation Board) and Middleton Irrigation Works (Middleton Irrigation Board) (Director of Irrigation, 1920).
Because of the inconsistent water supply to the schemes the Department of Irrigation (later known as the Department of Water Affairs) undertook the building of Grassridge Dam in the "Grootbrak River" and Lake Arthur in the Tarka River between 1923 and 1925, (Director of Irrigation, 1926, p. 18). Lake Arthur was named after Prince Arthur of Connaught who laid the foundationstone on the 3rd November 1923 (Director of Irrigation, 1925, p. 20). The irrigation above the confluence of the Tarka River with the Great Fish was to be supplied by Grassridge Dam, whilst Lake Arthur

was to supply the area below the confluence (Director of Irrigation, 1922, p. 35). The storage capacity of these two dams has been reduced considerably by silting in spite of the raising of the walls. Lake Arthur for instance had a storage capacity of 78,32 million m^3 in 1924 and in spite of the raising of the wall by 1,8 m in 1937 and a further raising of 0,9 m in 1946, the storage capacity was estimated at 29,68 million m^3 in 1969 (Department of Water Affairs, 1969, p. 41).

This means that within forty-five years the storage capacity of the dam was reduced by 62 percent although the wall was raised by 2,7 m. Instead of raising the wall any further, the Kommandodrift Dam was built further upstream with a present capacity of 67,8 million m^3 (Department of Water Affairs, 1969, p. 41).

The silting of the storage dams resulted in an insufficient supply of water to the irrigation schemes. This problem was already forseen in 1928 by Dr. A.D. Lewis, then Director of the Department of Irrigation (later to be known as the Department of Water Affairs), when he submitted a suggestion that the water be diverted via a cannal and tunnel system from the Orange River to the Great Fish River Basin (Du Plessis 1972, p. 35). The need for such a scheme gradually escalated and eventually resulted in several reconnaissance surveys starting in 1944. On the 23rd March 1962 the Hon. P.M.K. Le Roux, then Minister of Water Affairs, announced in the Senate that the Cabinet had approved the schemes to utilize the water of the Orange River (Du Plessis, 1972, p. 36).

Included in this scheme was the building of the Orange-Fish Tunnel, which would divert water from the Orange River near Venterstad to the Theebusspruit near Steynsburg. The tunnel would be 82,9 km long with a capacity of 33,98 m^3/s and would supply additional water to the 18 000 ha of developed irrigable land along the Great Fish River and 9 000 ha of developed land in the lower Sundays River Valley.
This tunnel was completed in October 1975, and for the first time, water flowed from the Orange River into the Great Fish River Basin.

1.2. NATURE AND SCOPE OF INVESTIGATION

The ever increasing shortage of irrigation water, due to the silting of the storage dams, as well as the erratic rainfall in the Great Fish River Basin, reached a climax in 1961 when a voluntary delisting programme of the irrigable land was implemented by the Government. From then until 1971 about 5 800 ha of irrigable land were delisted in the valley.

The soils and the groundwater tend to be rather saline in semi-arid climatic environments such as the Great Fish River Basin. These conditions are normally due to evapo-transpiration exceeding the average annual rainfall, thus resulting in an accumulation of salts in the soil. During the application of water, either by rainfall or irrigation, a certain percentage of the water seeps through the soils in order to reach the groundwater table or seeps back into the river before reaching the groundwater table. On its way through the soil this water naturally dissolves some of the accumulated salts, thus increasing the mineral content of the environment into which it seeps.

The irrigation along the Great Fish River presents a rather complicated problem in the sense that the river is used as a canal for the irrigation water, whilst groundwater as well as the seepage-water from the irrigation has free access to the river. A progressive deterioration in the quality of the water farther downstream is therefore to be expected. Not only does this water of poor quality present a hazard to the soil and the crops lower downstream, but it can also become unsuitable for industrial and domestic use.

The question which arised was: what extent will an increased frequency of irrigation, as well as an increase in the irrigation area by a re-enlisting of the irrigable land, have on the quality of the water in the river?

The Hydrochemical Working Group for the Orange River Project of the C.S.I.R. initiated a thorough investigation of the mineralization of the water in the Great Fish River.
A grant was presented by the C.S.I.R. to the Geology Department of the University of the Orange Free State in order to investigate the possible influence of the groundwater on the quality of the irrigation water in this river.
The author was therefore employed by the above Geology Department to study this problem.
The overall purpose of this investigation, was:

a. To investigate the main factors which control the quality of the groundwater in the area.

b. To compile a geological map of the Basin, including a study of various geological units in terms of their lithology, mineralogy, geochemistry and physical properties.

c. To determine the changes in the quality of the groundwater and to relate these changes to the above-mentioned factors.

d. To relate the changes in the groundwater quality to the changes in the quality of the runoff in the river.

The investigation is based entirely on qualitative data because of the lack of adequate measuring facilities along the river and because of the extent of the area.

1. 3. LOCATION AND EXTENT OF AREA

The part of the Great Fish River Basin which was investigated, is located between longitudes 25^{o}E to 27^{o}E and latitudes 31^{o} 15′S to 33^{o} 15′S. (Figure 1 - 1). This part of the basin represents a map-area of about 25 000 km^2, with a length of 250 km and a width of 100 km. Table 1 - 1 lists the main tributaries of the river, indicating the basin area of each tributary.

The sub-basins therefore comprise about 70 percent of the total area whilst the remaining 30 percent consists of the main river and minor tributary basins. An interesting

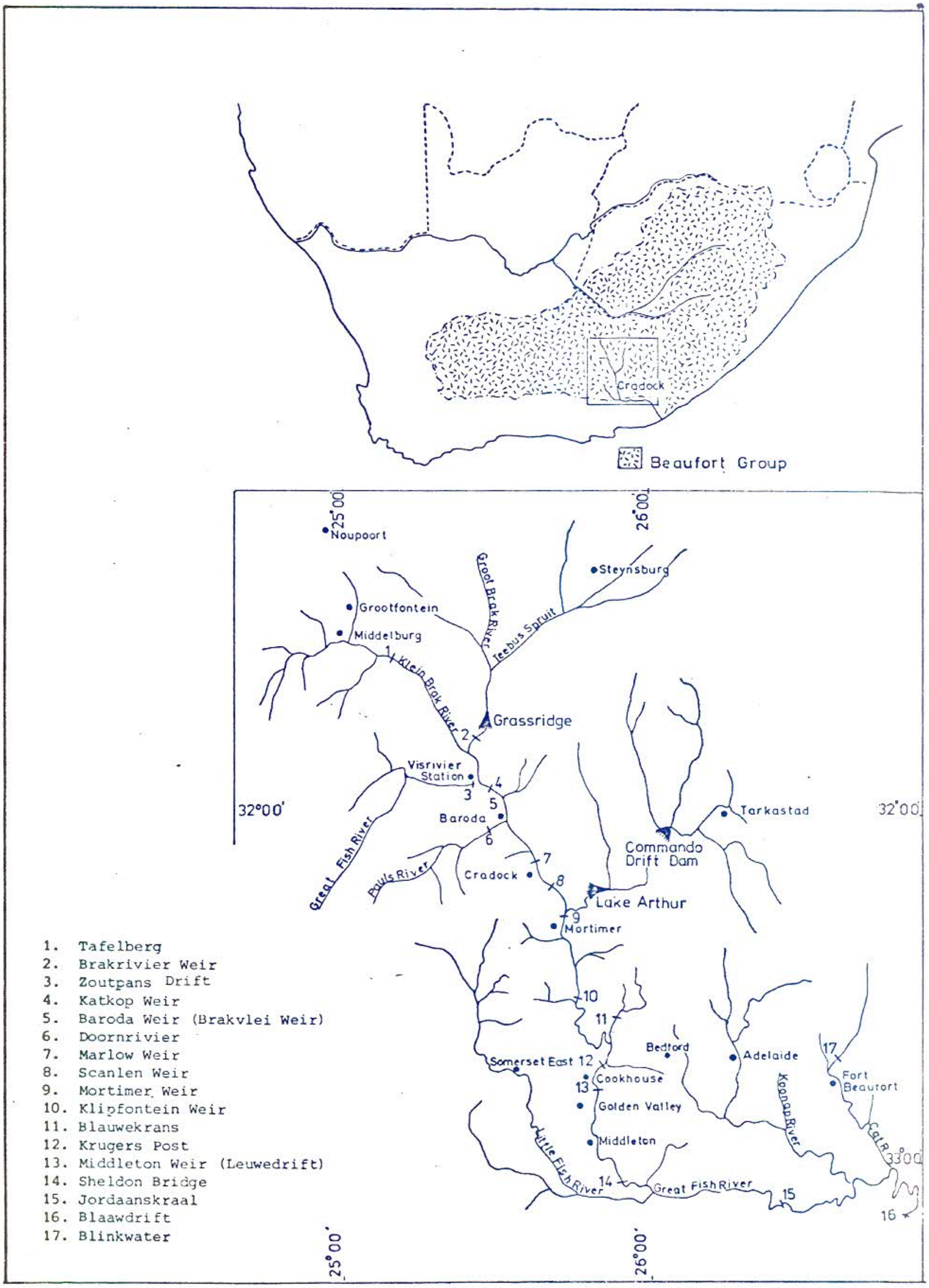


Fig. 1 - 1 Locality map of the Great Fish River Basin.

feature is the Teviot Basin which will be discussed more thoroughly in the proceeding chapters.

TABLE 1 - 1. Basin Areas Of The Main Tributaries Of The Great Fish River

Tributary Basin	Map Area (km^2)
Tarka River	3750
Groot Brak River	2850
Koonap River	2850
Little Fish River	2775
Klein Brak River	2175
Cat River	1575
Paul's River	675
Teviot Basin	600
	17250

From its source in the Nardousberg, east of Graaff-Reinet, the Great Fish River follows a course of about 630 km down to Rocky Point 25 km north of Port Alfred, where it flows into the Indian Ocean. The area investigated, however, extends only as far as the confluence of the Cat River, which is 500 km downstream from the source.

Irrigation in the Basin can be divided into two main areas, i.e. the Grassridge area which extends as far as the confluence of the Tarka River, and the Kommandodrift / Lake Arthur area which extend down to the Sheldon Bridge south

TABLE 1 - 2. Irrigation Boards Controlled By The Great Fish River Irrigation Board

Grassridge Dam Area	Area (ha)
Private Irrigator	128
Brak River Irrigation Board	728
Private Irrigator	171
Private Irrigator	128
Knutsford Irrigation Board	1616
Baroda Irrigation Board	1583
Private Irrigator	87
Marlow Irrigation Board	272
Cradock Municipality	194
Sub Total	4907

Kommandodrift Dam/Lake Arthur Area	
Kommandodrift Irrigation Board	385
Marlow Irrigation Board	1100
Scanlen Irrigation Board	1390
Private Irrigator	422
Gannavlakte Irrigation Board	497
Tarka Bridge Irrigation Board	1393
Mortimer Irrigation Board	1271
Klipfontein Irrigation Board	1245
Renfield Irrigation Board	758
Private Irrigator	17
Hougham Abrahamson Irrigation Board	3076
Middleton Irrigation Board	1946
Private Irrigator	86
Sub Total	13586

of Middleton (Figure 1 - 1). All the irrigation along the river is controlled by the Great Fish River Irrigation Board. Table 1 - 2 is a list of irrigation boards which are controlled by the above Board. In future, however, only the Kommandodrift, Gannavlakte and Tarka Bridge Irrigation Boards will be supplied by the Kommandodrift Dam and Lake Arthur; the rest will receive water from the Orange River.

1. 4. PREVIOUS WORK

1. 4. 1 Geological Investigations

As early as 1801 travellers and geologists explored the geology of the hinterland of the Cape Colony. One of the pioneers of South African geology, Andrew Geddes Bain in 1856 published the first attempt at unraveling the stratigraphy of the area. In this publication (Bain, 1856) the term "Fort Beaufort Grit" is used for the first time and can be regarded as the precursor of the term "Beaufort Group" which is presently used. It is needless to say that this stratigraphical unit is represented nearly throughout the entire Great Fish River Basin. R.N. Rubidge (1857) and R. Pinchin (1875) also contributed to the early unravelling of stratigraphy of the Eastern Cape Province.

Johnson (1966 and 1976) describes the stratigraphy and sedimentology of the Cape and Karoo sequences in the Eastern Cape Province thoroughly. The stratigraphy and sedimentology of the Ecca Group in the Eastern Cape

Province was investigated by Kingsley (1977).

Although the above investigations cover the entire area, no detailed geological maps were produced. An unpublished photogeological map of the area from longitudes 24°E to 27°E and latitudes 31° 30′S to 33°S was compiled in 1966 by Geomap S.A. on behalf of SOEKOR. In 1974 the Geological Survey published the geological map 3226 King William's Town (scale 1 : 250 000), which covers a large part of the area east of longitude 26°E. To the west Van Niekerk (1977) has mapped the area around Graaff-Reinet.

The first South African fossil vertebrate was discovered near Fort Beaufort in 1838 by Bain (1845). Since then quite a wealth of fossils have been found in the rocks of the Beaufort Group and Kitching (1977) describes the distribution of the Karoo vertebrate fauna. Originally the stratigraphycal subdivision of the Karoo sequence was made only on the occurrence of fossils and Kitching (1977) presents a biostratigraphical map indicating the fossil localities.

1. 4. 2 Groundwater Investigations

Bond (1946), in his geochemical survey of the groundwater supplies of the Union of South Africa, gives a thorough account of the groundwater quality of the Karoo sequence. Many of the samples discussed in the above investigation were collected from the present area. At that stage, however, very little was known about the lithostratigraphy of the area and therefore some of the lithological units from which the samples were allegedly taken, are doubted.

Frommurze (1937), in his investigation on the water-bearing properties of some of the geological formations in the Union of South Africa, also considers the properties of the "Karoo System". Some of the factors which control the quality and occurrence of groundwater in arid and semi-arid regions in the Union of South Africa and Angola were also investigated (Frommurze, 1953).

Authors such as Du Toit (1915) and Kent (1949) describe the porosity of the rocks of the Karoo System in South Africa and the thermal waters of the Union of South Africa and South West Africa respectively. In both publications, samples from the area in review were investigated. Young (1913) describes a tidal phenomenon in bore-holes near Cradock.

The sulphuretted spring just north of Cradock, although examined by Kent (1949) was already mentioned by Krauss (1843).

2. PHYSIOGRAPHY

The amount and type of dissolved substances occuring in natural waters depends on the environment and the extent of equilibrium reached between the water and the components of that environment. Environmental factors considered in this study are: geology, relief, drainage, climate and vegetation. The geology will be discussed in the following chapter.

2. 1. RELIEF

All the groundwater occuring in the Great Fish River Basin is considered to be moving to some greater or lesser extent. Hem (1970, p. 54) states that a solution moving through a porous solid, where equilibrium is attained, may display behaviour like that of solutions moving through ion-exchange or chromatographic columns. Where equilibrium is not attained the composition of the groundwater will be influenced both by the movement rates and by the reaction rates.

During the weathering of rocks the soluble products are removed in solution by the perculating groundwater. Therefore a close relation exists between rock-weathering and groundwater quality. Lukashev (1970, p. 34) emphasises the fact that water is the most important factor in the cycle of matter on the earth's surface, resulting in the tremendous geochemical activity in the weathering crust.

Hawkes and Webb (1962, p. 84 - 85) give a brief summary of the relationship between relief, groundwater perculation and rock weathering. In very mountainous terrain, physical erosion removes the rock debris faster than it can be decomposed chemically, whereas in areas of moderate to strong relief an extreme variability in the depth to the water table is encountered. Under these conditions chemical decomposition is most active beneath the crests of ridges, where the water table tends to lie at maximum depth below the surface. A shallower water table is encountered in the lower-lying valleys between the hills and often near-surface zones of permanent saturation are limited to the immediate vicinity of springs and drainage channels. Between the ridge crests and the drainage channels the circulation of groundwater is most vigorous and therefore processes of weathering and solution of the more mobile constituents are favoured.

In the flat-lying terrain the groundwater table is relatively shallow and the movement of the groundwater is sluggish. Here equilibrium is soon reached and weathering processes come to a virtual standstill.

An outstanding feature of the South African physiography is the Great Escarpment which resulted from the African Cycle of erosion of the Post-Gondwana landscape (King, 1963, p. 222). The Post-Gondwana landscape started developing from the Middle Jurassic to Early Cretaceous Era when Grondwanaland drifted apart (King, 1963, p. 206). During this era the Great Fish River must have started its development. This river has played an important role in the geomorphological evolution of the Eastern Cape Province.

The Great Fish River Basin may be sub-divided into the following main geomorphologic provinces: The low-lying Marginal Region (< 760 m), the Great Escarpment (760 - 1070 m) the Headbasin (1070 - 1370 m) and the Interior Plateau (> 1370) (PLATE I).

2. 1. 1 The Marginal Region

Between the Great Escarpment and the Cape Fold Belt in this area (approximately 50 km wide) lies a rather undulating landscape of medium to low relief. The foot of the escarpment is taken as the 760 m topographic contour and is defined approximately from west to east by the towns Somerset East, Cookhouse, Bedford, Adelaide and Fort Beaufort (PLATE I).

The tops of the undulating hills in the Marginal Region represent the older African and post-African surfaces, which has been disected by younger erosion cycles.

An interesting feature of this region is the meandering nature of the rivers, e.g. the Great Fish, Little Fish, Koonap and Cat Rivers.

According to King (1963, p. 252) this area belongs to the Karoo province of the Marginal Region. A southerly extention of the Bruintjieshoogte range, however, clearly separates this region from the typical Karoo landscape to the west.

2. 1. 2 The Great Escarpment

King (1963, p. 222) states that although this feature is not an escarpment in the strict sense of the term, it, however, always constitutes a relatively sharp rise from the Marginal Region to the high Interior Plateau. This description is also applicable to the present area. Here the Great Fish River has gnawed at the Great Escarpment at a far greater pace than its own tributaries (Little Fish, Koonap and Cat Rivers) and some of the smaller neighbouring rivers. The result of this being the capture of a large part of the Interior Plateau in the form of a headbasin behind the escarpment.

For the purpose of this study, the Great Escarpment is regarded as passing through the Tandjiesberge, Coetzeesberge, Grootbruintjieshoogte, Bosberg, Baviaansrivierberge and Winterberge (PLATE I). King (1963, p. 224), however, regards the above mountain ranges as the former position of the escarpment and that it now passes through Tembuland (Northern part of the Republic of Transkei) and the Suurberg, which forms the water devide between the Orange River and the Great Fish River Basin.

The Great Escarpment is an area of high relief lying between the 760 m and 1370 m topographic contours (PLATE I).

Parts of this province, which are not well clad by bush and other vegetation, are easily eroded, the result being the development of colluvial pediments at the base.

2. 1. 3 The Headbasin

The Great Fish River as previously mentioned has eroded part of the Interior Plateau behind the Great Escarpment. As a result an almost circular headbasin of about 100 km in diameter has developed.

Mountains such as Lootsberg, Wapadsberg, Kommetjiesberg, Graatjiesberg, Tandjiesberg, Bankberg, Gannahoekberg, Winterberg, Toorberg, Bamboesberg, Suurberg, Kikvorsberg, Carltonhill and Agter-Renosterberg surround the basin in an anticlock-wise direction from the west (PLATE I).
An escarpment toward the basin margin is defined by these mountain ranges. When crossing over the watershed between the Orange and the Great Fish River at the Carltonhills one can clearly see this escarpment.

At the base of the above escarpment lies a vast expanse of colluvium-covered plains which present an area of relatively low relief. On these plains bahadas have developed between scattered inselberge, mesas and buttes, which are the remains of the Interior Plateau. Tafelberg near, Middleburg and Koffiebus and Theebus, near Steynsburg, are the better known of these features.

Resistant dykes of dolerite stand out as prominent elongated ridges throughout the area. One such dyke runs in a north-south direction from Middelburg, past Cradock and can clearly be seen from Witkransnek on the main road between the two towns.

King (1963, p. 252) regards this basin as part of the Eastern Uplands which belong to the Marginal Region.

The largest part of this headbasin lies between the 1060 m and 1370 m topographic contours.

2. 1. 4 The Interior Plateau

This geomorphologic province constitutes the water devide between the Great Fish River Basin and the Orange, Kei and Sundays Rivers.

Bevelling by the Africa Cycle of erosion has resulted in scattered mountain ranges of which the dolerite-capped tops represent remnants of the Post-Gondwana Plateau. The only part of the area which can still clearly be recognised as a plateau, is that to the north of Steynsburg, represented by the Suurberg and Kikvorsberg.

Relative to the Marginal Region and the Headbasin, this province has a high relief.

2. 2. HYDROLOGY

It is stated by Frommuize (1953, p. 61) that the groundwater of South Africa is chiefly of meteoric origin. One can therefore accept that the total amount of groundwater in the area is a function of the annual precipitation, amount of runoff (either surface or subsurface), amount of evapotranspiration and the amount of water retained by the soil before infiltration to the groundwater table can occur.

Lawrence (1975, p. 9) gives the following equation for the hydrological cycle:

$$E = P - R - U$$

where E is evapotranspiration

P is precipitation

R is runoff

U is infiltration

The above equation is prefered because in the case of the Great Fish River Basin many of the parameters of the hydrological cycle had to be estimated due to a lack of sufficient data covering a long enough period. In some cases no data at all exist. Factors such as the interception of rainfall by vegetation, which according to Whitmore (1961, p. 5) can account for 5 - 15 percent of the annual rainfall in subhumid areas, are discarded here, not only because of the absence of data, but also because of the sparseness of the vegetation in the area.

2. 2. 1 Precipitation

The mean monthly precipitation at six different stations throughout the area is presented in Table 2 - 1. It is evident from the above table that the area can be devided into three main rainfall zones. Steynsburg and Tarkastad are located in the area between the Headbasin and the Interior Plateau, whilst Grootfontein and Cradock lie in the Headbasin. There appears to be some discrepancy between the rainfall at Somerset East and Adelaide although both stations are located at the foot of the Great Escarpment. The average annual precipitation for the area is 439,9 mm, with averages of 430,9 mm, 350,7 mm and 535,2 mm respectively for the zones mentioned above.

Figure 2 - 1 illustrates the mean monthly rainfall and evaporation at Grootfontein over the period 1935 to 1967 (Kriel, 1970, p. 181). The rainfall and evaporation over the period October 1973 to September 1974 is also presented in the same figure. During March 1974 disasterous floods were experienced in the Great Fish River Basin. Following the floods the groundwater table rose to such an extent that water flowed out of bore-holes and springs which had been dry for a considerable number of years. In one case the groundwater flowed out from a bore-hole in which the water table is normally 6 m below the surface.

Both Table 2 - 1 and Figure 2 - 1 reveal a gradual build-up of rainfall from August to March and a sharp decline to May. The entire basin therefore falls within a summer rainfall-cycle.

TABLE 2 - 1 The Mean Monthly Precipitation At Stations In The Great Fish River Basin

STATION	PRECIPITATION (mm)												TOTAL (mm)
	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	
Steynsburg	52,3	59,1	71,0	35,1	24,7	10,4	9,3	10,9	21,6	27,2	40,1	47,8	409,5
Tarkastad	58,9	64,5	71,8	36,8	28,0	13,8	13,4	15,1	23,6	32,6	43,4	50,3	452,2
Grootfontein	44,4	58,0	64,8	29,6	18,2	7,6	10,6	8,3	18,7	23,8	38,2	38,3	360,5
Cradock	44,1	54,8	55,7	30,1	20,9	9,1	8,5	9,4	17,2	25,3	31,2	34,6	340,9
Somerset East	65,7	71,9	82,4	48,5	35,4	18,0	22,6	23,4	45,2	59,1	63,0	68,0	603,2
Adelaide	46,5	58,3	69,7	40,8	26,6	16,3	13,9	17,1	36,1	45,7	47,4	48,8	467,2

(obtained from the Monthly Rainfall Data published by the Weather Bureau.)

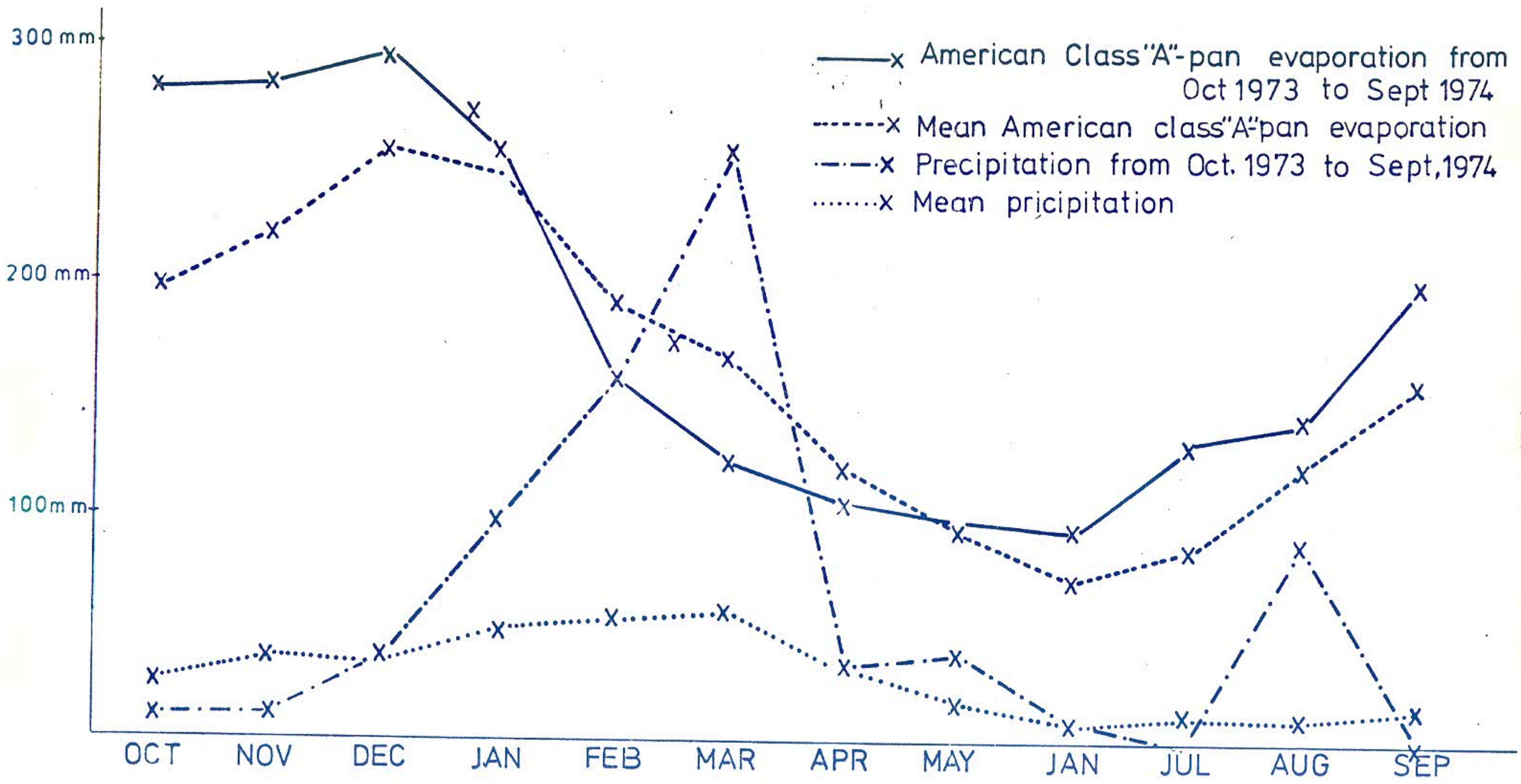


Fig. 2 - 1 The monthly mean rainfall and evaporation from an American Class"A"- pan at Grootfontein (C.P.).

Tyson *et al.* (1975) suggest that since 1880 there appears to have been a decline in the annual rainfall at some stations in South Africa. In the Vryburg District, which also has its highest rainfall during the summer, Hodgson, (1975) noticed a similar declining trend since 1935. The same declining trend is displayed by Figure 2 - 2 for the Grootfontein area. Tyson *et al.* (1975), however, states that there is little conclusive evidence to support the view that South Africa has undergone progressive dessication over the period 1880 to 1972.

The oscillating curve in Figure 2 - 2 was constructed by means of the five-yearly moving averages, because the individual values are too scattered. According to the above figure there appears to be a ten to twelve year cycle in the rainfall. The years of high precipitation having been 1941, 1950, 1962 and 1974, whilst the years of minimum rainfall were 1946, 1955 and 1969. If the above cycles are correct, then a period of minimum rainfall can be expected in 1979, from whence a gradual increase should occur.

Hodgson (1975, p. 5) emphasises the fact that the period of 40 years over which the rainfall data are available, is far too short to attempt any accurate prediction. The possibility does exist that the negative trend in the annual rainfall has reached its lowest ebb and that a positive trend might be encountered in future.

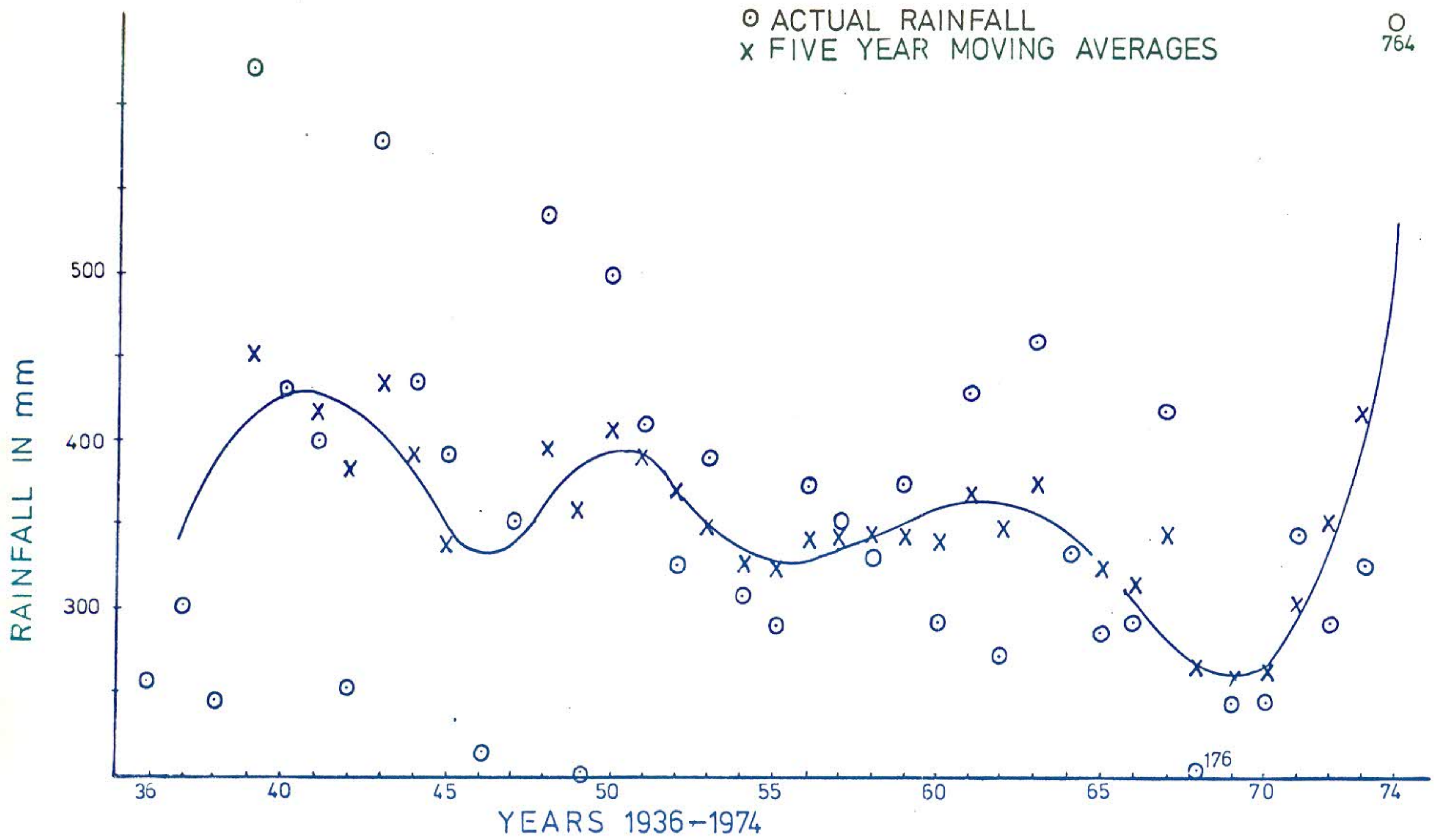


Fig. 2 - 2 Fluctuation in the annual precipitation at Grootfontein over the period 1936 to 1974.

2. 2. 2 Runoff

The runoff from a drainage basin depends on the annual precipitation and the frequency thereof, the topography, the nature of the soil and the geology. Therefore in areas with a low rainfall, but medium to high relief, a greater runoff than in the flat-lying areas, with high rainfall, is expected. According to the geomorphology of the Great Fish River Basin, the Headbasin and the Marginal Region represent flat areas in which very little runoff should take place. The Great Escarpment and the area between the Headbasin and Interior Plateau represent a medium to high relief and runoff ought to be high here.

There are two drainage regions in the area, i.e. the Headbasin which is drained mainly by the Grootbrak, Kleinbrak, Paul's and Tarka Rivers and the Great Escarpment and Marginal Region which are drained by the Little Fish, Baviaans, Koonap and Cat Rivers.

The annual rainfall and corresponding runoff at several guaging stations in the area (Kriel, 1970 and Jordaan, 1968) are presented in Table 2 - 2.

To determine the runoff from the Headbasin, the data from the Kruger's Post guaging station are used and from this the total rainfall and runoff in the Baviaans River at Blauwe Krans are substracted. The annual rainfall in the Headbasin, therefore, amounts to $4944,9844 \times 10^6$ m^3 of which $92,728704 \times 10^6$ m^3 is lost through runoff; the latter represents 1,88 percent of the rainfall.

The runoff at Blaawdrift is considered to be representative for the whole basin. In this case the runoff amounts to 3 percent of the total rainfall. This increase in runoff is attributed to the higher runoff from the Great Escarpment. Rivers such as the Koonap and the Cat respectively have a runoff of about 8 percent and 16 percent of the total rainfall.

The rainfall and runoff from the Great Escarpment and the Marginal Region are calculated by subtracting the values for the Headbasin from those determined at Blaawdrift. This leaves a figure of $5568,686 \times 10^6$ m^3 for total precipitation and $217,4975 \times 10^6$ m^3 for runoff, which is 3,19 percent of the total rainfall.

According to Whitmore (1963, p. 5) runoff will not commence until the requirements of interception, surface retention and initial infiltration have been met. The amount of rain involved in this initial retention over the course of the year is a function primarily of the frequency of showers, catchment gradiënt, the nature and density of the vegetation (which controls interception and retardation of runoff), climatic factors affecting evapotranspiration which in turn determines the amount of soil moisture to be replenished during each shower, as well as the infiltration capacity, moisture holding capacity and depth of the soil and purvious substrata.

TABLE 2 - 2 Rainfall And Runoff Data For Several Drainage Basins In The Great Fish River Basin

RIVER	STATION	PERIOD	AREA (km^2)	AVERAGE ANNUAL RAINFALL	TOTAL RAINFALL (m^3/ YEAR)	TOTAL RUNOFF (m^3/ YEAR)	RAINFALL MINUS RUNOFF (m^3/ YEAR
Klein Brak River	Tafelberg Q1M10	21/2/59 to 30/9/60	1950	Grootfontein 360,5 mm	$702,975x10^6$	83 540	$702,89146x10^6$ (99,99%)
Groot Brak River	Klipheuwel Q1M02	1/10/20 to 30/11/23	4543	Grassridge Q1E01 291,6 mm	$1324,7388x10^6$	15923666	$1308,815134x10^6$ (98%)
Great Fish River	Zoutpansdrift Q2M01	1/12/26 to 30/9/41	1700	Grassridge Q1E01 291,6 mm	$495,72x10^6$	29553237	$466,166763x10^6$ (94%)
Great Fish River	Katkop Q1M01	1/3/18 to 30/9/60	9150	Grassridge Q1E01 291,6 mm	$2668,14x10^6$	44649668	$2623,490332x10^6$ (98%)
Paul's River	Doorn River Q3M01	1/12/36 to 31/7/48	862	Grassridge Q1E01 291,6 mm	$251,3592x10^6$	15614003	$235,745197x10^6$ (94%)
Tarka River	Teeken Fontein	12/1/14 to 30/9/24	4470	Lake Arthur Q4E01 241,0 mm	$1077,27x10^6$	64705142	$1012,564858x10^6$ (94%)
Baviaans River	Blauwekrans	1/10/18 to 30/9/37	686	Lake Arthur Q4E01 241,0 mm	$1653,26x10^6$	22452106	$1630,807894x10^6$ (99%)
Great Fish River	Kruger's Post Q7M02	1/8/22 to 30/9/48	18436	Average Fish River 357,9	$6598,2444x10^6$	115180810	$6483,06359x10^6$ (98%)
Great Fish River	Cookhouse Q7M04	1/11/48 to 30/9/60	18474	Average Fish River 357,9 mm	$6611,844600x10^6$	55510103	$6556,334497x10^6$ (99%)
Great Fish River	Leuwedrift Q7M03	1/11/28 to 31/3/48	18503	Average Fish River 357,9 mm	$6622,2237x10^6$	130539155	$6491,684545x10^6$ (98%)
Great Fish River	Middleton Q7M01	1/1/06 to 30/11/28	18954	Average Fish River 357,9 mm	$6783,6366x10^6$	291269083	$6492,367517x10^6$ (96%)
Great Fish River	Jordaans-kraal Q9M12	1/10/35 to 30/9/60	23051	Average Fish River 357,9 mm	$8249,9529x10^6$	136979444	$8112,973456x10^6$ (98%)
Great Fish River	Blaawdrift Q9M10	13/7/30 to 31/3/56	29376	Average Fish River 357,9 mm	$10513,6704x10^6$	310226179	$10203,44422x10^6$ (97%)
Cat River	Blinkwater Q9M08	1/12/21 to 30/9/60	751	Adelaide 467,2 mm	$350,8672x10^6$	54619851	$296,247349x10^6$ (84%)
Little Fish River	Somerset East Q8M02	1/4/57 to 30/9/60	1386	Somerset East Q8E01 548,6 mm	$760,3596x10^6$	2450586	$757,909014x10^6$ (99,7%)
Koonap River	Adelaide Q9M02	1/9/26 to 30/9/60	1267	Adelaide 467,2 mm	$586,2409x10^6$	45562495	$540,678405x10^6$ (92%)

There appears to be a linear relationship between runoff and rainfall in semi-arid areas (Whitmore, 1963, p. 6). This relationship is illustrated by Fig. 2 - 3 which represents rainfall data at Grootfontein and Grassridge and runoff data at Katkop Weir over the period 1935 to 1955 (Table 2 - 3). The curve intersects the Y-axis at a point where no runoff occurs. This point is known as the runoff threshold and indicates the minimum rainfall required for a drainage basin before runoff can take place. For the Headbasin the runoff threshold appears to be 215 mm (Fig. 2 - 3). It is important to know the runoff threshold of a drainage basin, because this figure represents that part of the annual rainfall which is retained in the basin. Such retention may be partly the result of infiltration into the soil and partly the storage of surface water in dams.

The runoff of groundwater by means of seepage into the Great Fish River is estimated by Viljoen (1972, p. 1) to be approximately 0,113 to 0,283 cumecs. An average is 0,2 cumecs, which represent an annual runoff of 6307200 m^3. This is 2 percent of the annual runoff at Blaawdrift and 0,06 percent of the annual rainfall in the Basin.

2. 2. 3 Evapotranspiration

Most of the rain water, which is retained in the basin, will be stored temporarily in the soil before returning to the atmosphere through evaporation or transpiration.

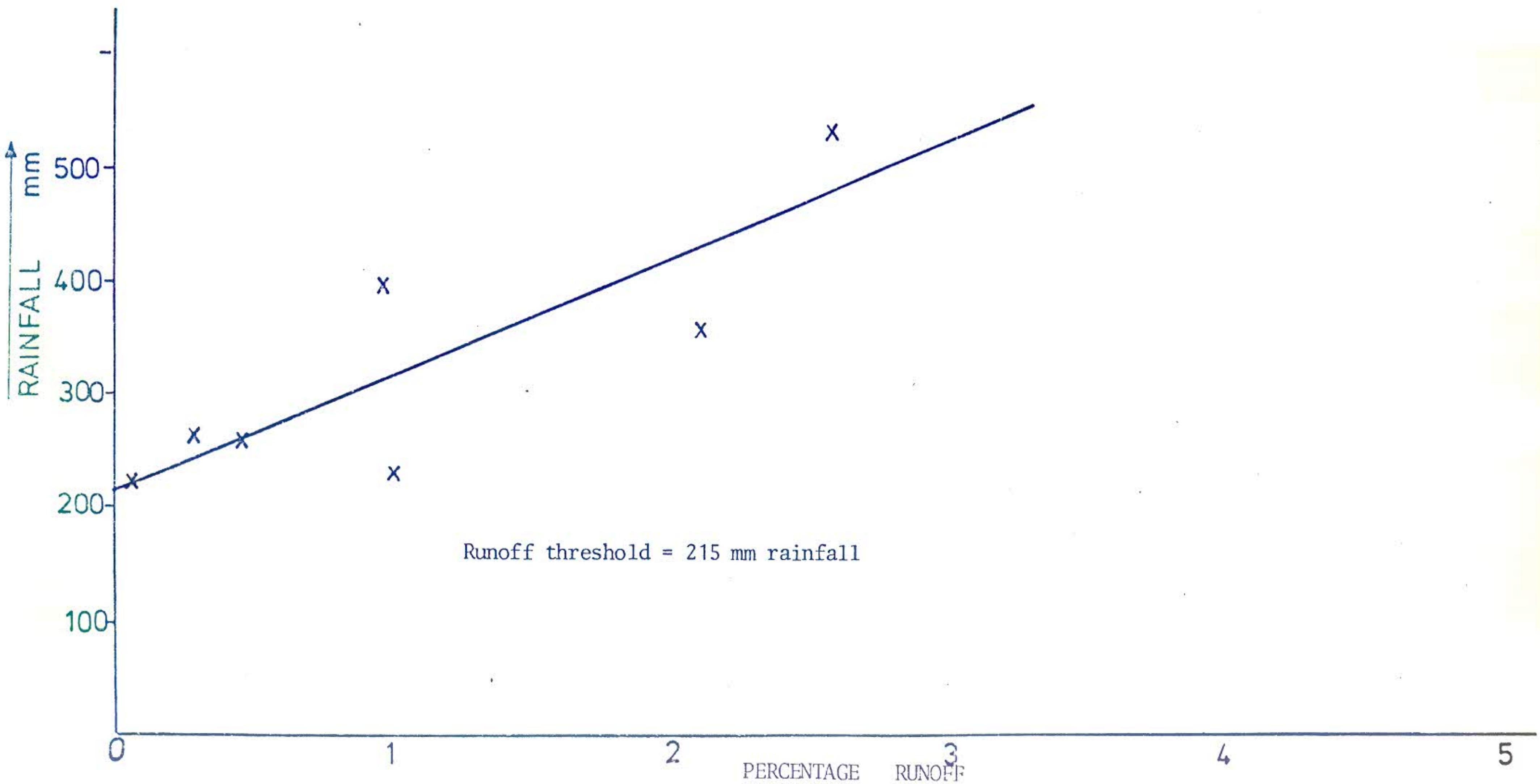


Fig. 2 - 3 The relationship between annual rainfall and the percentage runoff in the Great Fish River Basin.

TABLE 2 - 3 The Relationship Between Rainfall And Runoff In The Headbasin

DATE	MEAN PRECIPITATION AT GROOTFONTEIN (mm)	MEAN PRECIPITATION AT GRASS RIDGE (mm)	AVERAGE (AVE.)	TOTAL m^3 $9150 \times 10^6 \times$	RUNOFF AT KATKOP(m^3)	RETAINED (m^3)	RETAINED (mm)
1935-36	252,73	195,07	223,9	2048685000	1187873	2047497127	223,8
1936-37	299,72	164,59	232,2	2124630000	20799526	2103830473	229,9
1938-39	623,32	444,25	533,8	4884270000	126107735	4758162264	520,0
1940-41	399,54	397,76	398,7	3648105000	35056613	3613048386	394,9
1941-42	257,30	259,59	258,4	2364360000	10528993	2353831006	257,2
1949-50	502,92	388,62	445,8	4079070000	185159385	3893910614	425,6
1956-57	353,06		353,1	3230865000	67609556	3163255443	345,7
1954-55	289,81	238,25	264,0	2415600000	6915770	2408684229	263,2

Although it is relatively simple to determine the evaporation from free water surfaces, it is far more difficult to determine the actual evaporation from the soil. The evaporation loss from soil is a function of the frequency rather than the amount of rain, because the more often the soil surface is wetted by showers, the greater the cumulative evaporation loss.

According to Whitmore (1961, p. 9) evaporation is a superficial phenomenon and about 75 percent of the loss of soil moisture by evaporation over a season takes place in the top 8 cm of the soil.

Fig. 2 - 1 shows that the surface evaporation from American class "A"-pans at Grootfontein is at all times much higher than the monthly precipitation. During the summer months from October to March this difference is greater than in the winter. The importance of the ratio: precipitation to evaporation as far as the direction of perculation of groundwater is concerned, is emphasised by Hawkes and Webb (1962, p. 98). In the case of a low ratio, a rise of soil moisture from the underlying groundwater table is favoured. During the evaporation of this soil moisture an accumulation of salts occurs at certain depths in the soil. The relationship between the depth of lime accumulation and rainfall is illustrated by Hawkes and Webb (1962, p. 111). Table 2 - 4 presents the rainfall and evaporation data from American class "A" - pans at various stations in the area.

TABLE 2 - 4 Average Monthly Rainfall And Evaporation Measured With American "A" - Class Pans At Various Stations In the GREAT FISH RIVER BASIN. (Data From Kriel, 1970).

STATION	PERIOD		OCT	NOV	DEC	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	TOTAL (mm)
Grassridge Q1E01	1/1/26 to 30/9/67	P	28,1	42,6	36,9	51,5	52,5	64,7	35,6	21,2	7,9	11,8	9,9	16,1	378,8
		E	256,6	288,2	346,1	331,2	268,8	237,8	166,4	130,8	101,4	114,5	163,7	208,4	2613,9
Grootfontein Q1E02	1/6/35 to 30/9/67	P	24,4	34,0	34,8	59,7	30,2	56,6	41,1	18,5	13,7	13,5	11,2	12,2	350
		E	252,7	280,0	325,6	307,3	250,7	213,6	145,8	118,1	95,0	114,8	164,3	214,1	2482,1
Lake Arthur Q4E01	1/1/26 to 30/9/67	P	19,1	33,8	10,4	40,6	19,1	29,2	41,7	10,4	15,5	11,9	5,1	4,3	241
		E	184,9	256,0	280,2	306,6	222,3	213,1	126,2	97,3	63,0	101,6	128,5	167,9	2147,6
Somerset East Q8E01	1/12/60 to 30/9/67	P	51,8	44,2	39,9	59,7	58,9	117,3	47,0	36,3	30,2	12,4	28,7	22,1	548,6
		E	180,8	200,2	276,9	254,8	195,3	106,8	120,7	100,8	98,3	122,7	153,2	170,9	2035,4
M E A N		P	30,9	38,7	30,5	52,9	40,2	67,0	41,4	21,6	13,3	12,4	25,2	13,7	582,2
		E	218,8	256,1	307,2	225,5	234,3	206,3	139,8	111,8	89,4	113,4	152,4	190,3	2245,3

Another factor which accounts for the loss of a percentage of the rainfall that is retained in a basin, is the transpiration by the vegetation. An important aspect of transpiration is the potential evapotranspiration which is explained by Louw and Krüger (1968). Van Rooyen (1977, p. 64) states that the figures for potential evapotranspiration are determined under conditions of water saturation and that where the soil moisture is decreased by $^2/3$ of the field capacity, the transpiration will decrease by 90 percent. This could mean that the actual transpiration may only be 10 percent of the potential evapotranspiration.

The potential evapotranspiration for both short and tall crops at Grootfontein and Somerset East is displayed in Table 2 - 5. Because of the low rainfall during the winter from April to September, the actual evapotranspiration for this period is taken at 10 percent of the potential value. During the summer, however, the potential values are easily reached.

Figures for the potential evapotranspiration are representative only for cultivated land and not for the veld which is sparsely covered by vegetation. In a semi-arid region such as the Great Fish River Basin, the cultivated land is limited almost entirely to the irrigation schemes along the river and according to Table 1 - 2 this area comprises only 18493 ha of the total area of 29376 km^2 above Blaawdrift. This, however, does not include the irrigation along the Cat River. Although not unimportant, it is difficult to believe that transpiration alone can account for the largest loss of rainwater in the area.

TABLE 2 - 5 Potential And Actual Evapotranspiration For Short And Tall Crops Determined At Grootfontein And Somerset East. (Data From Louw And Kruger 1968 in mm)

STATION		OCT	NOV	DEC	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	MEAN	SUMMER	WINTER	ANNUAL
								SHORT CROPS									
Groot-	Potential	5,45	6,30	7,01	6,77	6,13	4,76	3,82	2,95	2,48	2,60	3,39	4,51	4,68	1104,4	601,4	1705,8
Fontein	Actual	5,45	6,30	7,01	6,77	6,13	4,76	0,38	0,30	0,25	0,26	0,34	0,45	2,84	1104,4	60,6	1165,0
Somerset	Potential	4,91	5,57	7,03	6,17	5,36	4,45	3,78	3,20	3,03	3,21	3,92	4,47	4,59	1016,6	658,8	1675,4
East	Actual	4,91	5,57	7,03	6,17	5,36	4,45	0,38	0,32	0,30	0,32	0,39	0,45	2,97	1016,6	65,8	1082,4
								TALL CROPS									
Groot-	Potential	6,81	7,68	8,51	8,22	7,36	5,78	4,80	3,83	3,27	3,40	4,37	5,81	5,82	1345,6	775,8	2121,4
Fontein	Actual	6,81	7,68	8,51	8,22	7,36	5,78	0,48	0,38	0,33	0,34	0,44	0,58	3,90	1345,6	77,7	1423,3
Somerset	Potential	6,09	6,88	8,46	7,58	6,65	5,52	4,92	4,44	4,48	4,70	5,53	5,92	5,93	1250,1	914,5	2164,5
East	Actual	6,09	6,88	8,46	7,58	6,65	5,52	0,49	0,44	0,45	0,47	0,55	0,59	3,68	1250,1	91,2	1341,3

2. 2. 4 Infiltration

Whitmore (1961, p. 5) regards infiltration as a very crucial hydrological process, because it determines how much water enters the soil to sustain the plant growth and recharge to the groundwater table.

The rate of infiltration is greatly dependent on the nature of the soil. A fine-grained soil for instance possesses a great capillary force which exercises an initial positive influence on the infiltration rate. Soils rich in clay minerals such as montmorillonite, which expands by at least two thirds of its volume when wet, may have a diverse effect on the infiltration rate.

Other factors influencing the rate of infiltration are the type and abundance of vegetation, intensity of precipitation, temperature and the surface gradiënt.

The amount of water which eventually reaches the groundwater table is therefore dependent on the total infiltration less that amount retained in the soil to either be lost by evaporation or transpiration. The amount which is retained in the soil is dependent on the depth of the soil cover as well as the field capacity of the soil. The field capacity is the maximum amount of water which can be retained by the soil before being forced downwards by gravity.

Hounam (1971, p. 5) suggests that for a given time interval, the amount of rainfall, the infiltration and the surface runoff are in a state of equilibrium and may be represented by the following equation:

$$E = P - O - D - \Delta S \qquad 2 - 2$$

where E is evaporation

P is rainfall

O is surface runoff

D is underground drainage

ΔS is the change in the soil moisture content.

The value ΔS is of vital importance because before any of the rainfall can drain away, a certain amount of the rain is consumed to bring the soil to field capacity (Van Rooyen, 1977, p. 69). Once the field capacity is reached all excess water in the soil will perculate down to the groundwater table. It, however, depends on the degree to which evapotranspiration takes place, whether this excess water will reach the groundwater table or not and therefore the value of ΔS is incorporated by Hounam (1971) in the value for evapotranspiration.

Equation 2 - 2 can however be expressed as follows:

$$I = P - (E_p + E_t + O + D) \qquad 2 - 3$$

where I = amount of water reaching the groundwater table

P = precipitation

E_p = free water evaporation

E_t = evapotranspiration

O = surface runoff

D = subsurface drainage.

This equation is similar to that given by Lawrence (1975).

The problem which still arises in solving the above equation, is that in the area the annual free water evaporation as well as the annual evapotranspiration from the soil exceeds the annual precipitation and therefore a zero value for I is obtained. Only under exceptional conditions such as those which prevailed during 1974, when the precipitation exceeded even the free water evaporation (Fig. 2 - 1), could a large positive value for I be expected. During the above period the water levels in some of the bore-holes rose by at least 6 m. Many springs which had dried up for some time, started flowing again and in some cases seepage areas developed where they were the least expected. One such disastrous case occured in irrigation lands near Baroda (Fig. 2 - 4).

If during a normal year 5 percent of the total rainfall in the basin reaches the groundwater table, which occurs in rocks with an effective porosity of 5 percent, the water table should rise by 36 cm. In this case it is assumed that no groundwater seeps back into the river. It is, however, doubted whether even 5 percent of the annual rainfall reaches the groundwater table in the study area.
The groundwater is therefore replenished mainly during the periods of high rainfall, such as that encountered during 1974.

Fig. 2 - 4 Salt deposits resulting from saline groundwater seeping out of the soil near Baroda after excessive precipitation.

By assuming that 5 percent of the rainfall reaches the groundwater table in the Headbasin, equation 2 - 3 reveals that 93,8 percent of the annual precipitation is either retained in the soil or is lost to the atmosphere through evapotranspiration.

2. 3. VEGETATION

There appears to be a rather close relationship between the relief and the vegetation of the Great Fish River Basin.

Acocks (1975) describes many different veld types in the area, of which the most important are displayed in PLATE II.

The Interior Plateau is covered mainly by the Karroid Merxmeullera Mountain Veld, previously known as the Karroid Danthonia Mountain Veld (Acocks, 1975, p. 98). This veld type belongs to the Pure Grassveld Types and covers all the higher mountains of the False Karoo and Central Upper Karoo. Acocks (1975) states further that this veld type usually appears in regions which are too dry and / or too frosty for the development of any kind of forest.

The top of the Winterberg, which generally has a high rainfall and is often covered with snow during the winter, consists of the Highland Sour Veld which belongs to the Temperate and Transitional Forest and Scrub Types (Acocks, 1975, p. 82).

Grassveld, which is thickly sprinkeled with dwarf Acacia Karoo and occasionally changes into a dense clumpy shrub bushveld, appears along the Great Escarpment. This veld type belongs to the False Thornveld of the Eastern Province, which is classed among the False Bushveld Types by Acocks (1975, p. 50). Originally this thornveld could have belonged to the Eastern Province Grassveld or to the scrubforest which is marginal to the high forest of the mountains. The invasion of the thorn-bushclump is regarded as the reason for the reduction of grass cover, which in turn encourages erosion. Finally the deterioration of the grassveld results into the False Karroid Broken Veld which is an extremely poor substitute.

The lower part of the Marginal Region is represented by the

Fish River Scrub which is classed under the Valley Bushveld of the Karoo and Karroid Types. In the undamaged state, the veld consists of dense semi-succulent, thorny scrub about 2 m high. At present, however, the area has been opened up by over grazing, the result being an invasion of prickly pears and in some parts by Euphorbia bothae. Succulents and thorny plants are of great importance here.

Further to the west, the Marginal Region is covered by the vegetation of the False Karroid Broken Veld, which falls under the False Karoo Types. This veld type is, however, not only restricted to the Marginal Region, but extends all along the course of the Great Fish River and along some of the tributaries right into the Headbasin. According to Acocks (1975, p. 79) the following modes of origin for the False Karroid Broken Veld are suggested:

(i) At the foot of the Great Escarpment and up the Great Fish River Valley to Cradock an open grassy shrub savanna, which is marginal to the Spekboomveld and scrub of the lower mountain slopes, has been invaded by Central Lower Karoo and Karroid Broken Veld; the result of this being a destruction of the grass cover and soil erosion.

(ii) In the Headbasin north of Cradock the grassveld of the Dry Cymbopogon-Themeda-Veld Type has been invaded by the Central Lower Karoo, and at the same time the elements of the open, grassy Karroid Broken Veld scrub have spread. Together with the advancement of the above processes over-grazing and erosion of the grassveld has occured.

(iii) On the steep mountain slopes and in the Marginal Region the thinning out and destruction of the Valley Bushveld, Spekboomveld and Fish River Scrub has taken place. The grassveld of the latter veld types is devoured by overgrazing, erosion and the invasion by Central Lower Karoo. The destruction in recent years of the prickly pear by cochineal has given this veld type further chance to spread.

(iv) Along the foot of the Great Escarpment the beginnings of another method of the development of the False Karroid Broken Veld can be seen. Acacia karoo is invading from the south and the east, whilst the Central Lower Karoo and Central Upper Karoo are invaded from the west. Both are contributing towards the destruction of the grassveld.

(v) In the Middelburg area there appears to be signs that Acacia karoo is spreading into the False Upper Karoo, the result being the development of a False Karroid Broken Veld.

Quite a large part of the Headbasin is covered by False Upper Karoo Veld, which is the result of extreme erosion of the grassveld. This area differs from the rest of the False Karoo in the sense that a higher proportion of succulents are encountered. Pentzia incana rather than Pentzia globosa dominates this area.

According to Acocks (1975, p. 78) the False Karoo Types tend to be sparser than the genuine Karoo types; until the grassveld-soil has been eroded away, the Karoo types have no secure foothold. Only when the harder subsoil or the bare rock has been exposed will the full mixture of Karoo species become established.

A most striking aspect of the area is the correlation between the distribution of veld types and the quality of the groundwater. The groundwater may directly or indirectly be responsible for the deterioration of the veld types in the area. This aspect will be discussed later.

3. GEOLOGY

The stratigraphy and sedimentology of both the Cape and Karoo Sequences in the Eastern Cape Province have been described by Johnson (1966 and 1976). Kingsley (1977) has concentrated on the Ecca Group in the same area.
No geological map of the area has, however, been compiled by either author. The geological maps which were compiled by SOEKOR (1966) and the Department of Mines (1976) proved to be inadequate for the purposes of this study. A geological map (PLATE III) of the Great Fish River Basin was therefore compiled on a scale of 1 : 250 000. Some of the data from the above maps were used during compilation of PLATE III.

The area between 25^{o} 30'E and 26^{o} E was mapped in detail by the author and the occurrence of some of the lithological members was projected to the east as well as to the west of the area with limited field control.

Sedimentary rocks of the Beaufort Group, together with intrusions of dolerite represent by far the largest part of the area. Rocks of the Ecca Group and Dwyka Formation occur only south of the 33^{o} S latitude.

Vast areas, especially in the Headbasin and the low-lying Marginal Region are covered by recent deposits of alluvium and colluvium. The alluvium is concentrated close to the rivers, whilst the colluvium covers the larger part of the plains. The areas between Hofmeyr and Middelburg and around

Golden Valley are typical examples of colluvial plains.

Table 3 - 1 compares the stratigraphic sequence in the Great Fish River Basin with that presented by Johnson (1976, Folder 2) and Van Niekerk (1977, Table VI, p. 27).

It is interesting to note that the deposition of the Karoo Sequence in the area began during the Late Carboniferous, whilst the Early Triassic marks the end of the deposition of the Beaufort Group. The intrusion of dolerite occured during the Jurassic Period.

PLATE IV presents a stratigraphic column of the area. The sequence rests on quartzite and shale of the Witteberg Group of the Cape Sequence and is capped by sandstone of the Molteno Formation.

3. 1. SEDIMENTARY DEPOSITS

These deposits include diamictite of the Dwyka Formation, shale, "rhythmitite" and sandstone of the Ecca Group, mudrock and sandstone of the Beaufort Group as well as calcrete, alluvium and colluvium of Recent deposits. Each of the above units will be discussed in terms of their lithology, palaeontology, depositional environment, source and structure.

3. 1. 1 Dwyka Tillite Formation

TABLE 3 - 1 A Comparison Of The Stratigraphy Of The Great Fish River Basin With That Described In The Eastern Cape Province By Other Authors.

VAN NIEKERK (1977) GRAAFF-REINET

KAROO SEQUENCE	Beaufort Group	Platberg Fm.	? ? ? ? ?
KAROO SEQUENCE	Beaufort Group	Platberg Fm.	Steilkrans M.
KAROO SEQUENCE	Beaufort Group	Platberg Fm.	Letskraal M.
KAROO SEQUENCE	Beaufort Group	Apieskloof Fm.	Oudeberg Sandstone M.
KAROO SEQUENCE	Beaufort Group	Graaff-Reinet Fm.	

JOHNSON (1976) EASTERN PROVINCE

KAROO SEQUENCE			Molteno Fm.	
KAROO SEQUENCE	Beaufort Group	Tarkastad Subgroup	Burgersdorp Fm	
KAROO SEQUENCE	Beaufort Group	Tarkastad Subgroup	Katberg Fm.	
KAROO SEQUENCE	Beaufort Group	Adelaide Subgroup	Balfour Fm.	Zone 4 Palingkloof M
KAROO SEQUENCE	Beaufort Group	Adelaide Subgroup	Balfour Fm.	Zone 3
KAROO SEQUENCE	Beaufort Group	Adelaide Subgroup	Balfour Fm.	Zone 2 Daggaboersnek M
KAROO SEQUENCE	Beaufort Group	Adelaide Subgroup	Balfour Fm.	Zone 1
KAROO SEQUENCE	Beaufort Group	Adelaide Subgroup	Middleton Fm.	
KAROO SEQUENCE	Beaufort Group	Adelaide Subgroup	Koonap Fm.	
KAROO SEQUENCE	Ecca Group		Waterford Fm.	
KAROO SEQUENCE	Ecca Group		Fort Brown Shale	
KAROO SEQUENCE	Ecca Group		Ripon Fm.	
KAROO SEQUENCE	Ecca Group		Collingham Fm.	
KAROO SEQUENCE	Ecca Group		Whitehill Shale Fm.	
KAROO SEQUENCE	Ecca Group		Price Albert Fm.	
KAROO SEQUENCE			Dwyka Tillite Fm.	

OWN SUBDIVISION (GREAT FISH RIVER)

KAROO SEQUENCE			Molteno Fm.	
KAROO SEQUENCE	Beaufort Group	Tarkastad Subgroup	Burgersdorp Fm.	
KAROO SEQUENCE	Beaufort Group	Tarkastad Subgroup	Katberg Fm.	
KAROO SEQUENCE	Beaufort Group	Adelaide Subgroup	Balfour Fm.	Elandsberg M.
KAROO SEQUENCE	Beaufort Group	Adelaide Subgroup	Balfour Fm.	Barberskrans Sandstone M.
KAROO SEQUENCE	Beaufort Group	Adelaide Subgroup	Balfour Fm.	Daggaboersnek M.
KAROO SEQUENCE	Beaufort Group	Adelaide Subgroup	Balfour Fm.	Oudeberg Sandstone M.
KAROO SEQUENCE	Beaufort Group	Adelaide Subgroup	Middleton Fm.	
KAROO SEQUENCE			Koonap Fm.	
KAROO SEQUENCE	Ecca Group		Fort Brown Shale	
KAROO SEQUENCE	Ecca Group		Ripon Fm.	
KAROO SEQUENCE	Ecca Group		Collingham Fm.	
KAROO SEQUENCE	Ecca Group		Whitehill Shale Fm.	
KAROO SEQUENCE	Ecca Group		Prince Albert Fm.	
KAROO SEQUENCE			Dwyka Tillite Fm.	

3. 1. 1. 1 Lithology

This unit consists of deposits of glacial origin in which diamictite constitutes at least 90 percent of the rock-types. An insignificant amount of mudrock and sandstone also occur in this unit.

Because of the obvious glacial origin of the Dwyka, Johnson (1976, p. 200) has assigned the term "Dwyka Tillite" to the unit. Formally the term "Dwyka Tillite Formation" will be used.

In the area under consideration the mean thickness of the unit is about 680 m (PLATE IV).

The tillite is normally massive with erratics having a maximum size of about 2 m. According to Johnson (p. 204) the following average composition of the tillite was determined.

Quartz	19%
Feldspar	12,5%
Rock Fragments	4%
Accessory Minerals	1%
Matrix	63%
Secondary Material	0,5%

Q : F : R : : 54 : 35 : 11

Q is Quartz

F is Feldspar

R is Rock fragments

About 60 percent of the rock fragments in the tillite are composed of quartzite, while the rest were derived mainly from granitic material.

3. 1. 1. 2 Palaeontology

To date no fossils of significant importance have been found in this formation in the area (Johnson, 1976, p. 203).

3. 1. 1. 3 Depositional Environment

Du Toit (1954, p. 274) regards the tillite in the area to represent ice-rafted material which was released from floating ice into deep water. On the grounds of the fact that the massive tillite itself possesses a fabric of orientated elongated pebbles that can only have been induced by flowing ice, Stratten (1968, p. 32) feels that it represents material which accumulated as a terrestrial ground moraine.

Theron and Blignault (1973, p. 349) have recognised four depositional cycles in the Western Cape Province and these represent the advance and the retreat of the ice sheet. Each cycle commences with a thick massive tillite unit (melt-out till from a grounded ice shelf), overlain by stratified sediments (released from a floating ice shelf).

3. 1. 1. 4 Source

The heterogenous composition of the rock fragments in the tillite indicate that ice sheets from various sources were responsible for the transportation of the material.

Studies by Stratten (1968, p. 131 - 133) proved that four source areas contributed to the deposition of the Dwyka Tillite; a south-moving Transvaal Ice Sheet, a Natal Ice Sheet moving to the south-west, the Atlantic Ice Sheet moving to the east and a Southern Cape Ice Sheet moving to the north-west. It is most likely that the Atlantic Ice Sheet was responsible for the deposition of the tillite in the present area of study.

3. 1. 1. 5 Structure

The Cape Orogeny caused extreme folding of the Dwyka. Overfolding to the north of Kommadagga has resulted in the duplication of the strata. Extreme dips to the north of up to 60^{o} are encountered here.

3. 1. 2 Ecca Group

3. 1. 2. 1 Lithology

Resting conformably on the tillite is a sequence of 2340 m of shale and dark grey to grey sandstone. What was known as the Upper Dwyka Shale (Du Toit, 1954, p. 278) is now grouped with the Ecca Group by Johnson (1976, p. 207). Johnson regards it as wrong to group sediments of totally different origin with the tillite of the Dwyka, which is undoubtedly of glacial origin.

The following subdivision of the Ecca Group is suggested by Johnson (1976, p. 211):

Prince Albert Shale Formation

This unit consists of approximately 100 m of dark grey mudrock which rests directly on the tillite. The contact with the underlying tillite is conformable and Du Toit (1954, p. 278) regarded this unit, together with the overlying Whitehill Shale Formation (White Band), as part of the "Upper Dwyka Shale". According to Kingsley (1977, p. 54) the mudrocks consist entirely of shale.

Whitehill Shale Formation

The Whitehill Shale Formation is distinguished from the underlying and overlying shale formations by the prominent white colour displayed on weathered outcrops. A closer examination of the rock proves that the colour varies from

from light grey, grey, pale pink and even pale red to purple. According to Du Toit (1954, p. 277) the unweathered material of this formation is black and carbonaceous.

Thin layers or lenses of chert, approximately 15 cm thick and seldom longer than 1 m occur in the shale.

Kingsley (1977, p. 55) states that contorted lamination is the most conspicuous feature of the Whitehill Shale Formation. This contorted lamination occurs normally as small folds or faults within the formation, whilst the underlying and overlying formations reveal no distortion whatsoever. Such structures may be regarded as penecontemporaneous deformation (Reineck and Singh, 1973, p. 75 - 81).

In the area under consideration this formation is only 10 m thick.

Du Toit (1954, p. 278) and Johnson (1966, p. 31), as well as many other authors previously regarded the top of this formation as the base on which the Ecca Group was deposited. Johnson et al. (1975), however, decided to group this formation and the underlying Prince Albert Formation with the Ecca Group and therefore regard the top of the Dwyka Tillite as the base of the Ecca.

Collingham Formation

Kingsley (1977, p. 55) gives a thorough description of this formation. The formation is devided into a lower member consisting of light grey, silty, well laminated shales. The laminae contain many devitrified glass shards in a fine-grained matrix, thus suggesting material of volcanic origin.

In places carbonaceous limestone, not thicker than 30 cm, is encountered.

The upper part represents a rhythmitite consisting of grey to black shale, alternating with thin layers of yellow to orange, silty, pelitic beds (mudstone), which seldom exceed 4 cm in thickness. Several such yellow beds are reported in a single section by Lock and Wilson (1975, p. 171). Because these pelitic beds consist almost exclusively of volcanic glass shards in differing states of preservation, Lock and Wilson (1975) suggest the term "metabentonite" for them.

An interesting feature of the volcanic interval at the top of this formation is its geographical extent. Lock and Wilson traced it from east of Grahamstown to west of Laingsburg (i.e. the entire length of the fold-belt). The bulk composition of the yellow mudstones is given by the above authors as:

Quartz	30 - 40%
Orthoclase	10 - 25%
Plagioclase	15 - 20%
Illite	5 - 20%
Montmorillonite	trace

The illite is regarded as a decomposition product of montmorillonite.

A total thickness of 30 m is estimated for this formation in the area of study.

Ripon And Fort Brown Formations

Both Johnson (1976) and Kingsley (1977) were able to subdivide the sequence above the Collingham and below the Koonap Formation into two lithostratigraphic units, i.e. a lower Ripon and an upper Fort Brown Shale Formation.
In addition Johnson recognised a third lithostraphic unit, the Waterford Formation between the Fort Brown Shale and the Koonap Formation, although this unit is confined to the area west of longitude 26^{o} E. In the present study a subdivision into three units was not possible, although sequences similar to those described by Johnson (1976) and Kingsley (1977) are present. A thickness of 2200 m is estimated in the study area for the unit (PLATE IV).

According to Johnson (1976) an arenaceous sequence of about 1000 m, consisting of dark grey, fine-to very fine-grained, feldspathic sandstone (greywacke) with interbedded "varved" rhythmitite and mudrock (shale, mudstone and siltstone), is present above the Collingham Formation in the Eastern Cape Province. Because of its arenaceous character he distinguishes this sequence from his overlying argillaceous Fort Brown Shale Formation; the latter was originally described as the "Lower Ecca Stage" by Rossouw (1953), Mountain (1946) and Du Toit (1954). Johnson (1966, p. 35) originally proposed the name "Ecca Pass Formation", but later changed it to "Ripon" (Johnson, 1976, p. 213).

The average thickness of the individual sandstone units in the Rippon-type sequence in the present study are about 15 m and the lithosomes are normally tabular in shape. Mottled sandstones appear near the base of the sequence and are a common feature right into the higher Middleton Formation.

Graded bedding is common in the sandstone and ball-and-pillow structures are often encountered. Near the top of the sequence a fining-upward in the sandstone is common. Fining-upward cycles (average thickness of 30 m) are also reported near the top of the Ripon Formation as defined by Johnson (1976, p. 226) and Kingsley (1977, p. 113).

In addition Kingsley (1977, p. 59 and p. 113) describes thin limestone beds interbedded with the shale. Because these beds grade laterally into ordinary shale, Kingsley (1977) regards them as of secondary origin. The number of limestone beds appears, to increase toward the top of this sequence.

In the present study area a transition grading from sandstone to siltstone and finally to dark grey shale is present between the Ripon-type and Fort Brown-type sequences. The Fort Brown-type succession consists predominantly of pencil shale, "varved" rhythmitite with subordinate mudrock and sandstone.

The argillaceous Fort Brown sequence was formerly named the "Middle Ecca Shale" (Rossouw, 1953) or the "Middle Ecca Group" (Ryan, 1967, p. 19). Johnson (1966, p. 37) gave this unit the name "Fort Brown Shale Formation".

Kingsley (1977, p. 74 - 80 and p. 115) gives a detailed description of his Fort Brown Shale Formation in the Ecca Pass area, stating that the formation consists of several upward-coarsening cycles, starting with shale at the bottom and gradually grading into siltstone and in some cases finally into layers of sandstone. At least four such cycles, three of which end in sandstone, can be recognised in Kingsley's (1977, p. 115 - 116) description of the Carlisle Bridge Section. Calcareous laminae are found in the shale and towards the top of the formation limestone beds (some 20 cm thick) are frequently encountered. The sandstones at the top of each cycle shows wavy lamination as well as micro-cross-lamination. Sub-horizontal worm burrows are common in the siltstone and sandstone.

Johnson (1966, p. 37) reports that the shale becomes coarser grained towards the top of the formation and eventually grades into fairly coarse silt-shale in which a number of beds of massive, very fine-grained sandstone of about 2 m thick occur.

Fig. 3 - 1 is a simplified interpretation of the Fort Brown Shale Formation as described by Kingsley (1977, p. 74 - 80). It is quite clear that at the top of the Fort Brown Shale Formation a sequence occurs which shows fining-upward cycles

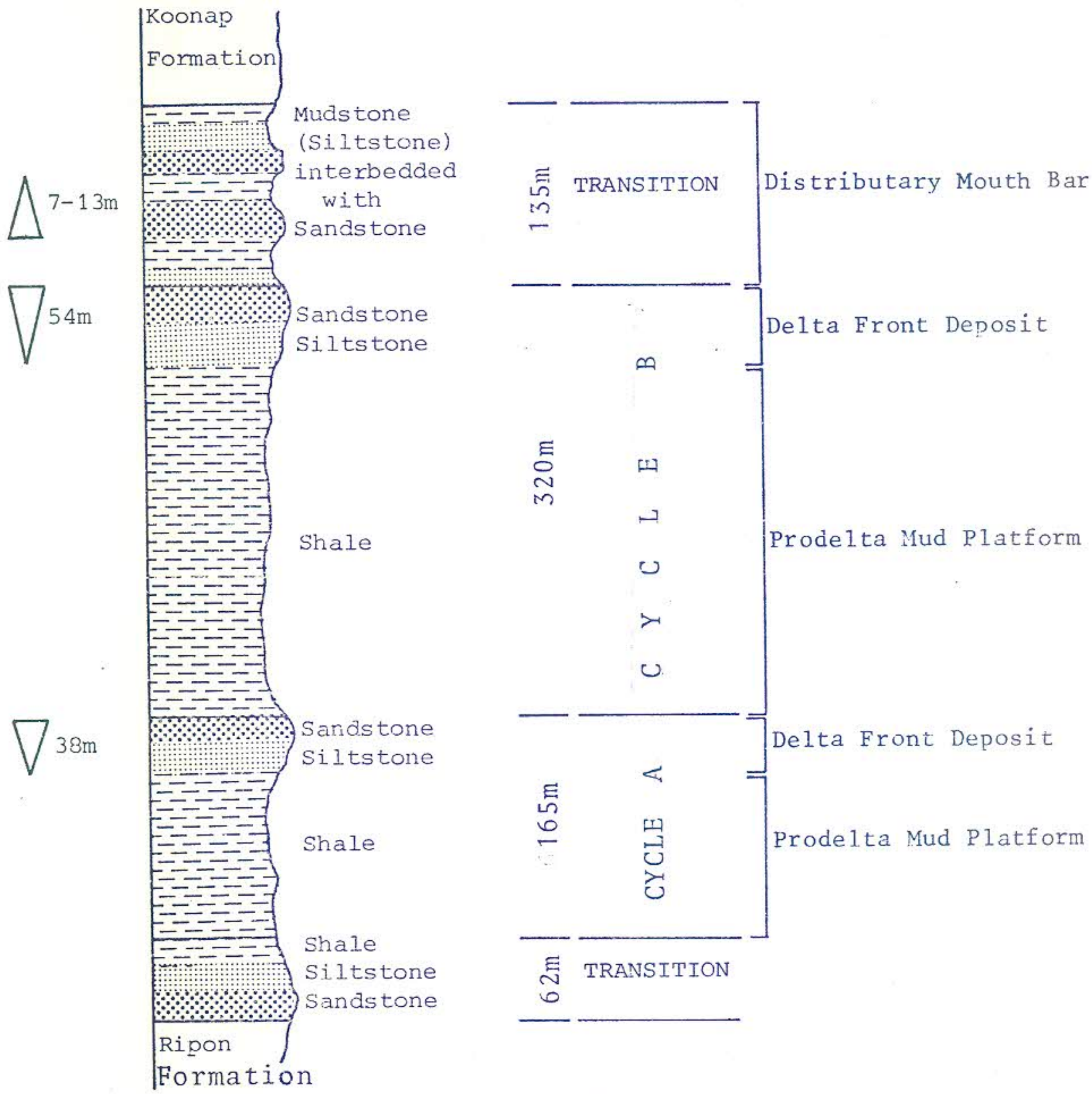


Fig. 3 - 1 A simplified section of the Fort Brown Shale Formation.

from sandstone to siltstone to shale. This sequence is regarded as the transition between the Fort Brown Shale and the overlying Koonap Formation.

Above the Fort Brown Shale Formation Johnson (1976, p. 211) describes a sequence of alternating grey (commonly speckled), massive lithofeldspathic sandstone, "varved" rhythmitite, shale and mudstone which he names the "Waterford Formation". He divides the formation into an arenaceous Main Member at the base, followed by the Middlewater Shale Member, above which lies the Transitional Member. He considers the Transitional Member as the transition from the pro-deltaic deposits of the Fort Brown Shale to the deltaic deposits of the overlying Koonap Formation.

Johnson (1966, p. 40) at first grouped part of this formation with the Koonap Formation, but after reconsidering the lithological features defining the Ecca and Beaufort Groups, the Waterford Formation was regarded as part of the Ecca, whilst the Koonap Formation was included in the Beaufort.

The lithology of Johnson's (1976) Waterford Formation corresponds well with the rocks of the "Upper Ecca" as described by Rossouw (1953).

According to Johnson (1976, p. 212) the Waterford Main Member does not extend farther east than longitude 26° E, in fact not farther east than the Port Elizabeth - Cookhouse railway line. This pinching-out effect leads the present author to believe that the Waterford Formation is merely part of an upward-coarsening cycle in the Fort Brown Shale Formation (Fig. 3 - 1) and therefore no justification exists for regarding this as a separate formation of the Ecca Group.

According to J.N.J. Visser (personal communication) the Waterford Main Member represents the delta front deposit of the top most cycle in the Fort Brown Shale as described by Kingsley (1977), while the Transition Member can be correlated with the transition at the top of the sequence (Fig. 3 - 1). The unusual thickness of Johnson's (1976) Main Member, in comparison to the relatively thin sandstone beds found elsewhere near the top of the Fort Brown Shale Formation, is attributed to the fact that the distributary channel involved, carried an exceptional amount of coarser material.

Oscillation ripple marks are common in the argillaceous rocks of the present study area, whilst slumping and ball-and-pillow structures occur in the sandstones. The sandstone lithosomes are tabular in shape and normally show massive bedding.

3. 1. 2. 2 Palaeontology

Very few fossil remains have been reported from the Ecca Group in this area. Those reported are mostly traces of fossils or unidentifiable plant remains.

Because of the extreme weathering of the Whitehill Shale very few remains of the reptile Mesosaurus, an index fossil of this formation, have been found. According to Kingsley (1977, p. 253), worm-burrow traces occur immediately above the Whitehill Shale and probably belong to the Nereites community, which is indicative of a relatively deep water environment.

Skolithos burrows, indicating a shallow water environment, were found by Kingsley (1977, p. 254) in sandstone of the Fort Brown Shale Formation. A single reptilian or amphibian fossilised bone was also found in one of the above sandstones.

3. 1. 2. 3 Depositional Environment

Kingsley (1977, p. 264) suggests a simple basin-fill sedimentation model for the deposition of the Ecca Group. Most of the material was deposited as turbidite fan complexes on the basin floor, on the slope and on the "shelf" of the basin. These turbidites were then covered by prograding, shallow-water deltaic complexes approaching from the south-east.

Pettijohn (1975, p. 552) remarks that large-scale (up to 100 m) coarsening-upward cycles, beginning with fine-grained dark shale, ending with coarse-grained sandstones, record the progradation or seaward growth of a shallow-water delta. The initial shales are regarded by Pettijohn (1975, p. 553) to be pro-delta clays which were deposited from suspension in quiet water below wave base. Periodically the pro-delta environment was invaded by turbidity flows descending from the delta margin area, bringing in fine sand and displaced debris to be deposited on the pro-delta clay as delta-front silts and sands.

Johnson (1976, p. 234) regards the absence of all current phenomena, the great lateral extent, the fine grained character and the extreme continuity of individual beds in the Prince Albert, Whitehill Shale and Collingham Formations as indicative of a deep-water environment.

The lower part of the Ripon Formation, according to Johnson (1976, p.235), represents a distal turbidite and the upper part a proximal turbidite sequence. It is believed that the turbidity currents were generated by the slumping of sediment over wide areas as a result of the approaching delta-front.

The main bulk of the Fort Brown Shale was deposited below, whilst the rhythmitites were deposited above the wave base (Johnson, 1967, p. 237). Reineck and Singh (1973, p. 323) regard storms as the cause of the alternating thin layers of sandstone and shale which are typical of the rhythmitites.

The arenaceous Main Member of Johnson's (1976) Waterford Formation is regarded by the present author merely as a large turbidity flow descending from the delta margin area and it therefore belongs to the same sequence of deposition as the Fort Brown Shale. In an area as vast as that in which the Fort Brown Shale was deposited, one would expect the development of several deltas. In some of these deltas more and coarser material was probably present, resulting in local lithological differences. The transition (fining-upward cycles) from Fort Brown-type to Koonap-type sedimentation (Fig. 3 - 1) probably represents distributary mouth bar deposits and indicates a transition from deltaic to fluvial deposition.

3. 1. 2. 4 Source

As stated previously a considerable proportion of the Ecca sediments have a predominantly volcanic source. Both Elliot and Watts (1974) and Martini (1974) support the above statement.

A source area of considerable areal extent and with a regional palaeoslope inclined towards the north-west, must have existed off the south-east coast (Johnson 1976, p. 239). One need merely consider the enormous amount of sedimentary material deposited in both the Ecca and Beaufort Groups to appreciate the extent of such a source area.

There appears to be very little difference in the mineralogy of the Ecca and Beaufort Groups. Only the relative amounts of the various minerals seem to vary. This fact proves that the same source area produced material for both groups.

Minerals encountered are quartz, feldspar (albite - oligoclase, orthoclase and microcline), rock fragments, mica (biotite and muscovite), matrix and accessory minerals.

The quartz is mainly of magmatic origin, although detrital quartz from pre-existing sedimentary rocks can be observed.

Plagioclase (albite-oligoclase) is the most abundant feldspar and appears to be relatively fresh, indicating a first cycle origin (Kingsley, 1977, p. 245). The plagioclase becomes more abundant in the Beaufort Group, whilst the amount of K-feldspar remains constant.

The rock fragments in sandstones of both the Ecca and Beaufort Groups are made up of volcanic rock, i.e. quartz-feldspar porphyry and pebbles of non-porphyritic "lava". This indicates a source area in which intermediate to felsic volcanics formed a prominent constituent (Johnson, 1976, p. 239). The presence of tuffaceous layers in the Collingham Formation is also indicative of contemporaneous volcanic activity.

Kingsley (1977, p. 250) concludes, from the heavy-mineral content of the sandstones, that the source rocks consisted mainly of low-grade metamorphic rocks, possibly of the greenschist facies. Sedimentary rocks must have been present, whilst granitic and silicic volcanic rocks delivered only minor, but significant, amounts of sedimentary material to the basin.

3. 1. 2. 5 Structure

The Ecca Group in the area has been subjected to folding during the Cape Orogeny. This folding resulted in east-west trending synclines and anticlines with an increased intensity of deformation to the south.

Prominent joints and fractures, also with an east-west trend have developed and can be associated with the folding.

3. 1. 3 Koonap Formation

Johnson (1976, p. 208) lists six differences which reflect a major change in the depositional environment from the Ecca Group to the Beaufort Group. The Ecca was deposited in a large body of water (probably marine), whilst the Beaufort sediments are typical of continental, fluviatile deposition. As no major unconformaty revealing a time-break between the deposition of the two groups can be determined a transitional environment should exist between the two. The Koonap Formation represents this transition. Both Johnson (1976)and Kingsley (1977), regard this formation as part of the Beaufort Group. Because such an environment is neither marine nor fluviatile in the strictest sense of the term, it is suggested that the Koonap Formation be regarded as an independant unit and not part of the Beaufort Group.

3. 1. 3. 1 Lithology

The Koonap Formation is distinguished from the underlying one by the conspicuous number of sandstone beds which grade upwards into fine-grained siltstone and mudstone.

The sandstone is greenish-grey, medium-to fine-grained and often mottled. Elliot and Watts (1974, p. 110) attribute this mottling to the presence of laumontite, which develops as a secondary zeolite from intermediate and felsic volcanic material.

These bodies of sandstone vary in thickness from 1 - 3 m. Micro-cross-lamination and flat-bedding are common features of the thinner sandstone lithosomes, while the thicker sandstone units (up to 7 m) show either massive bedding or large-scale trough-cross-bedding.

Greenish-grey, fine-grained mudstone is the most abundant rock-type in this formation while shale and siltstone occur in minor amounts. In PLATE IV (Section 1) the mudrock constitutes 73 percent of the sequence, a figure which corresponds well with the findings of Johnson (1976, p. 247). This section reveals the cyclic deposition in the formation and both major and minor cycles can be distinguished.
The major cycles normally begin with a relatively thick sandstone (up to 7 m) and grade upwards into finer siltstone and mudstone. A complete major cycle is normally 100 - 200 m thick and superimposed on it follows a minor cycle, 15 - 20 m thick. In both cycles an upward fining is discerned.

The Koonap Formation is 980 m thick and can be subdivided in two units on the ground of the sedimentary structures which occur in the sandstone lithosomes. In the lower cycles (both minor and major) rippledrift-lamination and ball-and-pillow structures are common, while trough-crossbedding only occurs in some of the units. The upper cycles show well developed micro-crosslamination and large-scale trough-crossbedding while ball-and-pillow slump structures are virtually absent.

The number of sandstone units increase towards the top, but their thickness seems to decrease.

Johnson (1966, p. 39) defined the Koonap Formation as a cyclic sequence of sandstone and mudstone occuring between the underlying Fort Brown Shale and the first "red" mudstone of the overlying Middleton Formation. This formation was, however, redefined by the same author (1976, p. 244) when he grouped the Waterford Formation with the Ecca Group and the Koonap with the Beaufort Group. The base of the formation is defined as the first thick (7 m) sandstone above the Fort Brown Shale.

Oxidising conditions, which are responsible for the red colour of the mudstones, may, however, prevail in a deltaic as well as in a fluviatile environment. The first "red" mudstone can therefore, not be used as the upper boundary of the Koonap Formation in terms of its present definition. In the present study the boundary between the two formations was therefore taken above the last occurrence of typical deltaic deposits, i.e. at the top of the last cycle containing tabular, horizontally laminated, upward-fining sandstone lithosomes. These lithosomes represent interdistributary delta-plain deposits, the last of which are to be seen in the road cutting just north of Middleton Station (Fig. 3 - 2). The first "red" mudstone was encountered a few kilometres south of Middleton, near the Sheldon turn-off. In the present investigation the contact between the Koonap and Middleton Formation is therefore placed much higher in the stratigraphic succession.

3. 1. 3. 2 Palaeontology

Practically no identifiable vertebrate fossils have been found in the formation in this area. Because of the lack of suitable outcrops, no particular attempt was made to search for fossils.

This unit may, however, represent part of the *Tapinocephalus* Zone (of Middle Permian age), which is described by Kitching (1977, p. 14).

Kingsley (1977) mentions the presence of possible *Glossopteris* remains, as well as the imprints of *Phyllotheca* stems.

3. 1. 3. 3 Depositional Environment

Kingsley (1977, Figure 115) suggests that the strata of the

Fig. 3 - 2 Tabular sandstone lithosomes of the Koonap Formation.

Koonap Formation represent delta-plain deposits. According to Reineck and Singh (1973, p. 269) flat bedding, ripple-drift-crosslamination and the tabular shape of thin sandstone lithosomes in minor cycles, however, suggest levee-splay and interdistributary delta-plain deposits. Trough-cross-bedding in the thicker sandstone units (major cycles) may be indicative of distributary channel deposits.

According to Kingsley (1977, p. 241) the major cycles are correlatable over a wide area, thus indicating an external tectonic control on their deposition. Atkinson (1962, p. 355) states that the minor cycles in such a deltaic environment could be attributed to one of the following factors: longterm climatic variations, shortterm meteorologic causes, tectonic movements or changes in the hydrodynamics of the area of deposition.

The ball-and-pillow slump structures and the scarce trough-crossbedding in the lower part of the Koonap Formation, and the abundance of trough-crossbedding, rippledrift-crosslamination and a decrease in thickness of the sandstone lithosomes in the upper part, indicate an upward-shallowing of the basin from an outer deltaic-plain to an inner deltaic-plain environment. Twenhofel (1950, p. 54 and 102) regards deltaic deposits as part of the transitional environment, which is further justification for separating the Koonap Formation from the Beaufort Group.

3. 1. 3. 4 Source

The source of the material deposited in this formation is the same as for the underlying Ecca Group.

The following average mineral assemblage was determined for the sandstones of the formation:

Q	F	R	Acc.	Cem	M	Q : F : R		
18	24	34,5	1	6,5	16	23,5	31,5	45

Q = quartz
F = feldspar
R = rock fragments
Acc.= accessory minerals
Cem.= cement
M = matrix

3. 1. 3. 5 Structure

Discernable, but inconspicuous gentle folds, with east-west trending axes, indicate a decreased influence of the Cape Orogeny.

Intense jointing and fracturing, also with an east-west trend, are very conspicuous.

3. 1. 4 Beaufort Group

A sequence 4500 m thick of alternating fine-grained, lithofeldspathic sandstone and mudrock lithosomes of varying thickness, follows conformably on the Koonap Formation. The sandstone / mudrock cycles have a fining-upward trend and display features which are generally considered to be characteristic of fluviatile deposits. Part of these cycles can be observed in PLATE IV(Sections 2, 3, 4 and 5).

The Beaufort Group has been subdivided on lithological grounds as well as on the colour of the mudstones into the Adelaide Subgroup (Koonap, Middleton and Balfour Formations) and the Tarkastad Subgroup (Katberg and Burgersdorp Formations) by Johnson (1966, p. 46 - 59 and 1976, p. 241 - 271). According to Johnson (1976) "red" mudstone is absent in the Adelaide Subgroup (excepting in the Middleton Formation) and abundant in the Tarkastad Subgroup.

In the present study the Middleton and Balfour Formations constitute the Adelaide Subgroup, as the Koonap Formation is considered to be a separate unit. Red mudstone is present in the Koonap and Middleton Formations, as well as in the overlying Tarkastad Subgroup. This is further proof that colour is an unreliable criterium for stratigraphic subdivision in the Karoo sequence.

3. 1. 4. 1 Lithology

3. 1. 4. 1. 1 Adelaide Subgroup

Middleton Formation

This formation was first described by Johnson (1966, p. 49) as the lower-most unit of the Beaufort Group, but he later (1976) altered this boundary (as discussed previously). According to Johnson (1976, p. 245) the road and railway cuttings north and south of Middleton Station display the typical and characteristic features of this formation.

In the present study, however, sandstone lithosomes answering to the type of deposition which is normally found in the Koonap Formation, were found north of Middleton. It would therefore appear that the name "Middleton Formation" is not quite suitable for this unit, and that the name "Golden Valley Formation" should be prefered. Because of the fact that the name "Middleton Formation" has already been accepted by the South African Committee For Stratigraphy (SACS, 1971) it is suggested that this formation be redefined as consisting of typical fluviatile strata in which no tabular-shaped sandstone lithosomes occur.

The formation consists mainly of a sequence of cyclic deposits commencing with a lenticular body of sandstone at the base and fines upward into greenish-grey mudstone. In places the mudstone is replaced by a greyish siltstone, while patches of "red" mudstone are encountered throughout the formation. The sandstone lithosomes consist mainly of fine-to medium-grained grey to light grey feldspathic sandstone. Motteling in the sandstone is very common.

A fluvial cycle commences with the deposition of sandstone. Most of these were deposited in channels scoured out of mudstone, thus revealing trough-crossbedding or massive bedding at the base, grading upward into rippledrift-crosslamination and flat lamination at the top. Clay-pellet conglomerate is present at the base of some of the sandstone units. The sandstone grades upward into a few centimetres of siltstone, followed by mudstone. Near the top of the mudstone calcareous lenses can sometimes be observed, providing that these were not destroyed by scouring, which defines the beginning of a new cycle.

As in the Koonap Formation, monor cycles with an average thickness of 15 m are superimposed on major cycles of about 100 m in thickness. The sandstone lithosomes of the minor cycles vary in thickness from about 1 - 2 m and pinch out laterally over a distance of a few hundred metres, whilst the thicker sandstones (10 - 15 m), which constitute the base of the major cycles, pinch out laterally over a slightly longer distance. Toward the top of the formation the sandstone bodies become extremely thick (PLATE IV, Section 2). One such body south of Cookhouse is 30 m thick, but pinches out laterally over a distance of 1 km.

Sandstone constitutes about 30 percent of the major cycles, but towards the top the amount increases to 40 percent.

Outcrops of the Middleton Formation are relatively poor as this formation is found mainly in the Marginal Region below the Great Escarpment.

The total thickness of this formation in the study area is about 1700 m, representing 47 percent of the thickness of the Adelaide Subgroup and 37 percent of the thickness of the Beaufort Group as a whole.

Balfour Formation

Johnson (1976, p. 246) defined the boundary between the Middleton Formation and the overlying Balfour Formation (both belonging to the Adelaide Subgroup) as a conformable, intertongued transition zone of approximately 100 m thick above which "red" mudstones are absent. In this study, however, the boundary between the two formations is taken at the base of the first sandstones which possesses a lateral extent of at least a few kilometres. This sandstone lithosome also defines the base of the Oudeberg Sandstone Member (180 m thick) which is correlated with Zone 1 of the Balfour Formation of Johnson (1976, p. 241). Following the Oudeberg Sandstone Member are the Daggaboersnek Member (1200 m), the Barberskrans Sandstone Member (190 m) and the Elandsberg Member (320 m), which are respectively correlated with the Zones 2, 3 and 4 of Johnson (1976).

The name "Balfour Formation" was first proposed by Johnson (1966, p. 49) after the village of Balfour around which rock-types typical of this formation occur.

The Oudeberg Sandstone Member was regarded by Keyser (1973, p. 8) as a single consistant sandstone body in the

Graaff-Reinet area above which Cistecephalus microrhinus fossils are found. This sandstone is only 9m thick in the area studied by Keyser (1973) and constitutes a bed at the top of a sequence of sandstones, which are consistent throughout his area. It is suggested, however that the name "Oudeberg Sandstone Member" be assigned to the sequence as a whole.

In the present study area sandstone constitutes about 70per-cent of this member, and the individual sandstone lithosomes have an average thickness of about 40m. These units are lenticular to tabular in shape and can be followed over relatively long distances. The sandstone bodies are normally massively bedded at the base, grading upward into rippledrift-crosslamination and flat bedding. Surfaces upon which the sandstones were deposited tend to be undulating and scoured channels are rare.

The sandstone is fine to medium-grained, while motteling can be observed in the lower lithosomes.

Fluviatile cycles, which have a fining-upward trend are normally 50-60m thick. The contact between the sandstone and the overlying mudstone is gradational and normally consists of siltstone of a few centimetres in thickness. The mudstone is greenish-grey to dark grey in colour.

In addition thin dark green argillaceous beds, approximately 2cm thick, are often present in the sandstone. These beds often occur immediately above massively bedded sandstone lithosomes and could possibly be of volcanic origin.

Fig. 3 - 3 presents a typical view of the Oudeberg Sandstone Member in the area around Cookhouse. The photograph was taken form the township of Uitkeer, which is a construction camp for the Fish-Sundays River Canal. In this area the unit forms part of the Great Escarpment, while farther to the east it occurs at the base of the escarpment.

Fig. 3 - 3 The Oudeberg Sandstone Member forming part of the Great Escarpment. View from Uitkeer.

The Daggaboersnek Member follows conformably on the Oudeberg Sandstone Member and consists of an argillaceous sequence of mudstone and subordinate sandstone of about 1200 m in thickness. This unit is well exposed at Daggaboersnek on the main road between Cookhouse and Cradock, from thence the

the name "Daggaboersnek Member" as suggested by Tordiffe (1974, p. 1). Johnson (1976, p. 241) describes this member as Zone 2 of the Balfour Formation and tentatively assigned the name "Daggaboersnek Member" to his sequence. Fig. 3 - 4 illustrates the main features of this member.

Fig. 3 - 4 A typical view of the Daggaboersnek Member illustrating the fine alternation of sandstone and mudstone layers. Daggaboersnek.

The Daggaboersnek Member is distinguished from the underlying strata by the abundance of mudstone, as well as the even stratification caused by thin sandstone lithosomes which appear at regular intervals. The sandstone is seldom thicker than 2 m and is tabular to subtabular in shape, persisting over relatively long distances. Rippledrift-crosslamination is common in the sandstone, while trough-crossbedding is rare because of the fact that the sandstone was deposited on a relatively even surface.

The mudstone is greenish-grey to grey and contains a considerable amount of carbonaceous material in places. Wave-ripplemarks with an east-north-east orientation are common in the mudrocks.

Fining-upward cycles, with an average thickness of 15 - 20 m, reveal a regular rhythmic alteration of sandstone and mudstone. The sandstone constitutes 11 percent of the total thickness.

This unit is located mainly in the Great Escarpment.

The Barberskrans Sandstone Member follows conformably on the Daggaboersnek Member and is an arenaceous sequence. The complete section is illustrated in PLATE IV (Section 3). The name for this member was derived from the road cutting at Barberskrans, 10 km south of Cradock.

This member has a surprisingly uniform thickness (190 m) throughout the area and consists mainly of a fine-grained, lithofeldspathic sandstone with interbedded greenish-grey mudstone. The sandstone features prominantly in the mountains around Mortimer (Fig. 3 - 5) and Johnson (1966, p. 52) suggests that this unit can possibly act as a marker horizon in the area. In the map compiled for SOEKOR (1966) this sandstone was mistakenly mapped as the base of the overlying Katberg Formation. Fig. 3 - 5 clearly shows the tabular shape of the unit. Trough-crossbedding in sandstone

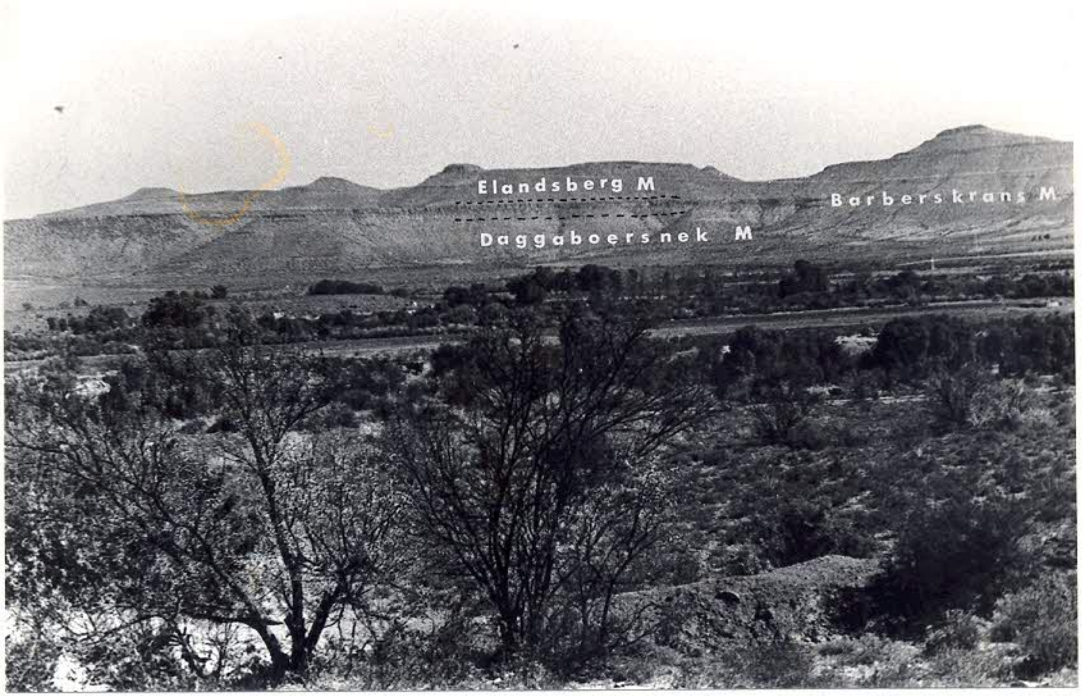


Fig. 3 - 5 Barberskrans Sandstone featuring prominantly in the mountains around Mortimer.

which was deposited in large channels scoured out of the underlying material, is quite common. In places, however, the sandstone is massive at the base, grading upward into rippledrift-crosslamination and flat bedding. Most of the overlying mudstone has been eroded away prior to the deposition of the following sandstone unit. This removal of the argillaceous material has resulted in the large fining-upward cycles (20 m) consisting mainly of sandstone, grading upward into thin (2 m) layers of mudstone. In places the mudstone has been removed completely and the sandstone of the following cycle is deposited on a flat or undulating surface of sandstone of the underlying cycle (Fig. 3 - 6).

Irregular and / or lenticular bodies of calcareous material are often encountered in this member (Fig. 3 - 6). Some of these bodies are several metres in length.

Fig. 3 - 6 Typical view of the Barberskrans Sandstone near Halesowen. Note the deposition of sandstone (b) on a flat surface of sandstone (a) of an older cycle. A thin veneer of mudstone is still present in places at the top of the older cycle. Note also the presence of irregular and / or lenticular bodies (c) of calcareous material.

Mudstone constitutes only a minor part of this member (15 percent). A slight thickening of the mudstone lithosomes can be observed toward the top of the unit (PLATE IV, Section 3).

To the north-west of Cradock this unit constitutes a large part of the low-lying areas of the Headbasin. Farther to the east, however, this sandstone member is located in the Great Escarpment.

The Elandsberg Member constitutes the top of the Balfour Formation and consists of a sequence of argillaceous cycles. Each individual cycle starts with a thin lenticular body of sandstone (3 - 5 m), which grades rapidly into a thick layer of mudstone (20 - 40 m).

In the bottom 200 m of the sequence the argillaceous material consists only of greenish-grey mudstone, while the top 120 m contains mostly "red" mudstone. Johnson (1976, p. 241) grouped the lower part with his Zone 3, i.e. the Barberskrans Sandstone Member of the present investigation, and named the upper part the "Palingkloof Member". Because of its argillaceous nature, it is incorrect to group the bottom part with the arenaceous Barberskrans Sandstone Member and neither is the colour of the mudrock a valid criterium for his subdivision.

Representative outcrops of this member are exposed at the foot of the Elandsberg on the farm Elandsberg (Cr. F. 3 - 1), to the east of Cradock (PLATE IV, Section 4). Sandstone becomes more abundant towards the top of the sequence, but generally constitutes a minor amount (22 percent) of the total thickness. The sandstone lithosomes are lenticular in shape and normally show trough-crossbedding at the bottom which grades upward into rippledrift-crosslamination and flat bedding.

Although the mudstone normally shows massive bedding, a feint lamination can often be detected at the bottom of mudstone cycles. Calcareous lenses and nodules are often present in the mudstone.

This member, together with the underlying Barberskrans Sandstone Member occurs in the low-relief area of the Headbasin to the north-west of Cradock. Farther to the east it is encountered at the base of the Interior Plateau or at the top of the Great Escarpment.

3. 1. 4. 1. 2 Tarkastad Subgroup

Katberg Formation

The Tarkastad Subgroup (Katberg and Burgersdorp Formations) is distinguished from the underlying Adelaide Subgroup by a greater abundance of sandstone and the predominance of "red" mudstone (Johnson 1976, p. 243). There is little difference in the lithologies of the Katberg and Burgersdorp Formations, except for an abundance of arenaceous material in the former. Johnson (1976, p. 247) expresses the view that north of 32^{o} S the distinguishing features between the two formations become so obscure, that they will have to be mapped as a single unit and named the "Tarkastad Formation", which will be the equivalent of the Tarkastad Subgroup in the south. The contact between the Katberg Formation and the underlying Elandsberg Member of the Balfour Formation presents no problem in the field.

Fig. 3 - 7 is a typical outcrop of the Katberg Formation south of 32° S. Sandstone constitutes almost 70 percent of the unit at this particular locality and 82 percent farther south-east, at Elandsberg (PLATE IV, Section 4). These figures do not correspond with the visual estimation of 30 - 35 percent by Johnson (1976, p. 248 - 249) for the same area.

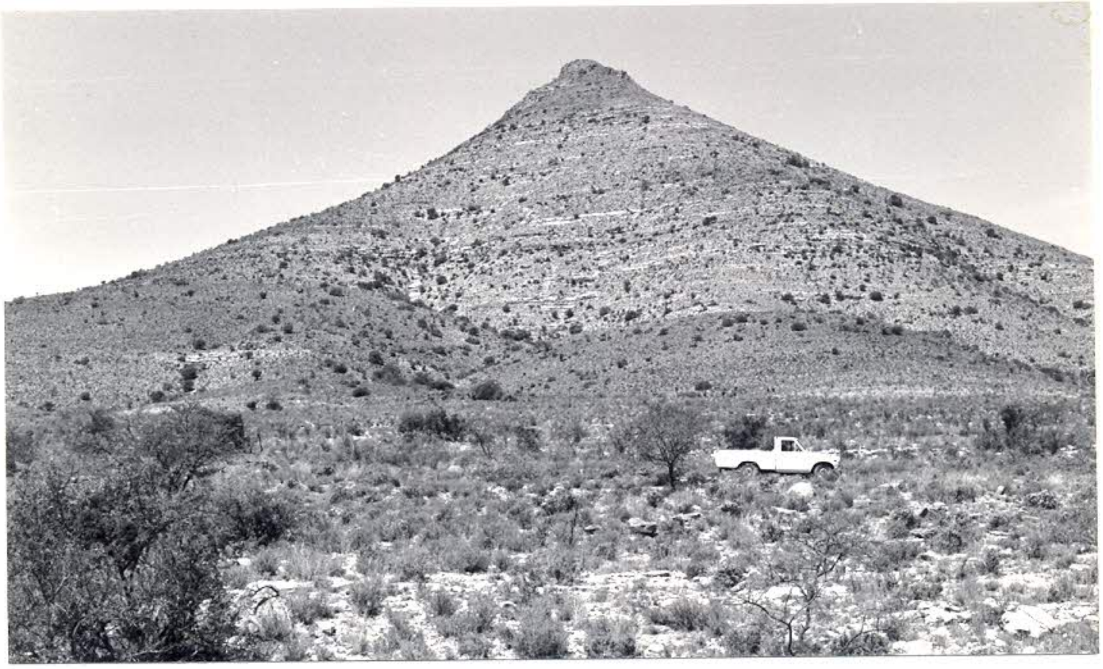


Fig. 3 - 7 A typical view of the Katberg Formation at Speelmanskop near Baroda

As in the lower formations, fining-upward, fluviatile cycles make up the entire sequence of 500 m of the Katberg Formation. The mudrock of a proceding cycle was often scoured away before the deposition of the sandstone of the following cycle. This results in the deposition of a sandstone lithosome directly upon the preceding one, with only a thin layer of mudstone, a few centimetres thick, occuring in places between the two sandstone units. In some cases the presence of mud clasts between two sandstone beds is the only indication that mudrock was ever deposited in the former cycle (Fig. 3 - 8).

Fig. 3 - 8 Mud clasts between successive sandstone cycles of the Katberg Formation. Road cutting south of Tarkastad.

The sandstone lithosomes are lenticular in shape and large trough-crossbedding is a common feature. The sandstone was sometimes deposited in deep channels scoured out of the underlying mudrock. Fig. 3 - 9 shows such a channel which was scoured right down to the sandstone of the lower cycle. Lenticular bodies of clay-pellet conglomerate are often present at the base of the sandstone lithosomes. The pinkish-grey colour of the sandstone is a distinguishing feature and contrasts with the greenish-grey colour of the same rock type

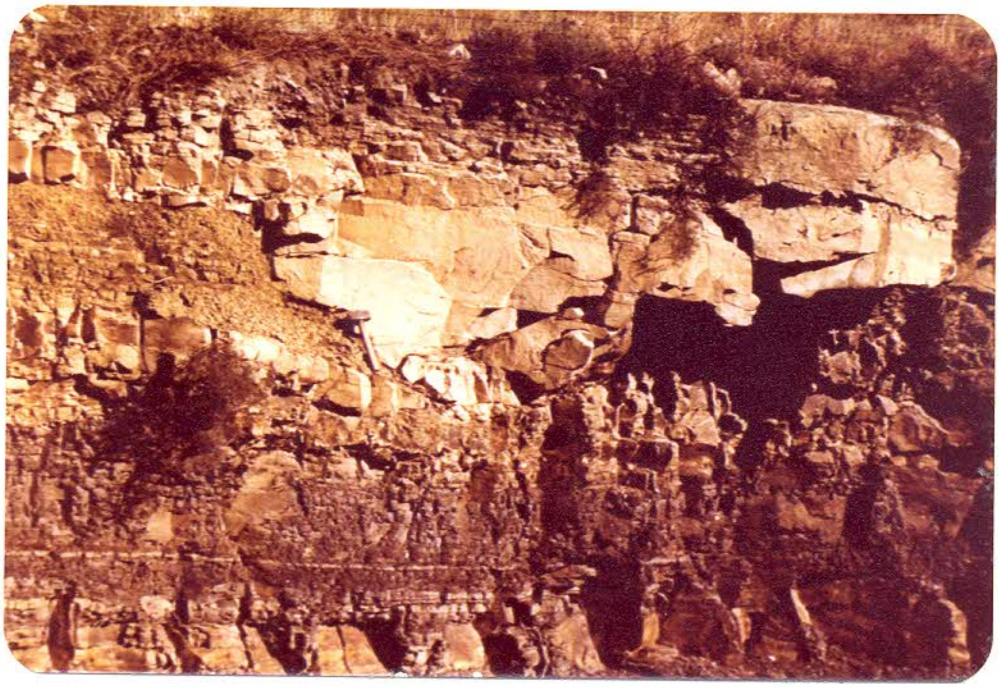


Fig. 3 - 9 Erosion channel in the Katberg Formation.
Road cutting between Tarkastad and Adelaide.

in the underlying Adelaide Subgroup. The sandstone is fine- to medium-grained and tends to become finer towards the north-west.

Maroon mudstone is the predominant mudrock of Katberg Formation, although dark red to purple varieties of siltstone and mudstone occur in places.

The contact between the Katberg and Burgersdorp Formations is gradational and makes the mapping of this boundary extremely difficult, especially in the low-lying areas of the Headbasin, which are often covered by colluvium.

Burgersdorp Formation

The only difference between the Katberg and Burgersdorp Formations is that sandstone is less abundant in the latter. Nine sections were measured in the Burgersdorp Formation around Steynsburg and sandstone constitutes an average of 35 percent.

According to Johnson (1976, p. 245) the name "Burghersdorp Beds" was first used by Du Toit in 1904. Later on these beds were included in what became known as the "Upper Beaufort Stage". Johnson (1966, p. 49) proposed the name "Queenstown Formation" for this unit, but the Karoo Working Group of the South African Committee for Stratigraphy has decided to retain the historical name "Burghersdorp" and merely change the spelling to "Burgersdorp".

A thickness of 450 m is estimated for this formation. The same cyclic deposition as in the under-lying Katberg Formation can be observed. Mudstone, however, features prominently.

The fluviatile cycles are better developed than in the underlying Katberg Formation. Numerous scour channels with clay-pellet conglomerate (50 cm) at the bottom, are encountered. The sandstone lithosomes begin with large trough-crossbedding, followed upward by rippledrift- cross-lamination, flat bedding and eventually by a thick bed of "red" mudstone. The thickness of individual cycles vary between 10 m and 60 m, with an average of 30 m.

The mudstone varies from maroon to purple-grey in colour. Near intrusions of dolerite a decolouration to greenish-grey is often observed. Thin calcareous lenses and nodules are commonly present with the mudstone.

3. 1. 4. 2 Palaeontology

The "Beaufort Series" has until recently been subdivided, on relatively vague lithological evidence into a "Lower", "Middle" and an "Upper Stage". It was possible at an early stage, however, to subdivide the Beaufort into a number of biostratigraphic zones because of the abundance of vertebrate remains. Du Toit (1954, p. 340 - 341) lists a number of vertebrate remains belonging to the Reptilia, Amphibia and Pisces Classes. Broom (1906) divided the "Beaufort Series" into six palaeontological zones, while Kitching (1977, p. 3) reduced the subdivision to five zones. Table 3 - 2 presents the biostratigraphic subdivision of the Beaufort as proposed by Broom (1906) and Kitching (1977).

Kitching (1977, p. 17) includes the Endothiodon Zone with the lower part of the Cistecephalus Zone and redefines the upper part of the latter as the Daptocephalus Zone. Keyser (1973, p. 6) appears to have used the same subdivision as Kitching (1977) but divided the Cistecephalus Zone further into the Cistecephalus microrhinus sub-zone and a Lower Sub-zone. The Procolophon Zone of Broom (1906) is grouped by Kitching (1977, p. 14) with the Lystrosaurus Zone.

TABLE 3 - 2. Biostratigraphic Subdivision Of The Beaufort As Proposed By Broom (1906) And Kitching (1977).

NEW NOMENCLATURE		OLD NOMENCLATURE	BIOSTRATIGRAPHIC ZONE	
Formation	Member	Stage	Broom 1906	Kitching 1977
Burgersdorp		Upper Stage	Cynognathus Procolophon	Cynognathus
Katberg		Middle Stage	Lystrosaurus	Lystrosaurus
Balfour	Elandsberg Barberskrans Daggaboersnek Oudeberg	Lower Stage	Cistecephalus	Daptocephalus
			Endothiodon	Cistecephalus
Middleton				
Koonap			Tapinocephalus	Tapinocephalus

Kitching (1977, p. 15) states that, due to scanty exposures, fossil remains are very rare, badly preserved and fragmentary in the Somerset East, Bedford and Fort Beaufort districts. The identifiable specimens which were found in this area include the genera Oudenodon and Gorgonops, as well as a number of small Pristerodon - like endothiodants. On these grounds Kitching (1977, p. 15) doubts whether the Tapinocephalus Zone is present in the area, and states that the main fossiliferous beds assigned to this zone were deposited in the south-western region of the main Karoo Basin.

The base of the Cistecephalus Zone is not very well defined in the area, but according to the description given by Keyser (1973, p. 7 - 12) this zone appears to start somewhere below the Oudeberg Sandstone Member, while the Cistecephalus microrhinus Sub-zone occurs in a thin unit at the base of the Daggaboersnek Member. Keyser (1973, p. 10) attributes the disappearance of the genus Cistecephalus to unusually heavy flooding, perhaps during the deposition of the Oudeberg Sandstone Member.

The Daptocephalus Zone occurs in the Daggaboersnek Member, Barberskrans Sandstone Member and in the lower part of the Elandsberg Member. Above this follows the Lystrosaurus Zone, which is largely associated with the Katberg Formation. The Cynognathus Zone is correlated with the Burgersdorp Formation.

Many imprints of plants are present in the Beaufort Group, but due to a scarcity of forms, they cannot be used for zoning (Du Toit, 1954, p. 292). Various species of Glossopteris are, however, characteristic of this group.

According to Romer (1970) it appears that the Adelaide Subgroup was deposited during the Middle to Late Permian, whilst the Tarkastad Subgroup was deposited during the Early Triassic.

3. 1. 4. 3 Depositional Environment

The fining-upward cycles of the Beaufort Group correspond well with the deposits in recent fluvial environments as described by Allen (1964, 1965a, 1965b and 1970) and Reinek and Singh (1973, p. 223 - 263). Such environments are marked by the presence of large cycles on which small cycles are superimposed. Each cycle invariably consists of a coarser grained sandstone unit at the base, which grades upward into a finer grained siltstone or mudstone body. The base of the sandstone is often marked by a thin basal conglomerate which contains intraformational clasts.

Allen (1965b, p. 127 - 128) divides alluvial sediments into two main groups, i.e. channel or substratum deposits and overbank or topstratum deposits. The former represents channel-floor, point-bar and channel-bar deposits and the latter pointbar-swale or abandoned braided-stream channel, levee, cravasse-splay or flood-basin deposits.

The channel deposits are coarser grained and display all the syndepositional structures like trough-crossbedding, flatbedding and micro-crosslamination. During extreme flooding the banks of the channel are overflowed and a large part of the load is deposited near the channel on the banks. The finer material is deposited away from the channel. Allen (1965b, p. 125) notes that point bars, channel bars and alluvial islands are the result of the lateral accretion of stream-bed loads during the sideways migration of the channels. The vertical accretion of suspended loads during overbank flow lead to the construction of levees, crevasse-splays and floodbasins on top of the channel deposits which have undergone latteral accretion. Reineck and Singh (1973, p. 246) state that levee sediments are made up of somewhat finer-grained material than their corresponding point-bar sediments. In most cases the mud layer is thicker than the sand layer.

The above description of a fluviatile environment fits the entire sequence of the Beaufort Group. It is, however, necessary to discuss the deposition of the Daggaboersnek Member and the Katberg Formation, as these two units do not seem to answer entirely to the fluviatile environment.

The regular bedding, abundance of thin, tabular sandstones, the presence of wave-ripple surfaces and the presence of dark "shales" containing occasional leaf impressions in the Daggaboersnek Member are features regarded by Johnson (1976, p. 270) to be indicative of a fairly extensive inland sea or lake. The dark carbonaceous shales, containing plant remains, point to the presence of marshes and swamps. This type of environment would be expected at a relatively advanced stage during the filling of the Karoo Basin.

Johnson (1976, p. 269) points out that the Katberg sandstone was undoubtly deposited in a braided-stream environment. His deduction is based on the relatively coarse-grained character of the sediments, the virtual absence of inter-bedded mudstone layers, the absence of micro-crosslamination and ripple marks, the low scatter of palaeocurrent data and the presence of a fan-shaped palaeocurrent pattern.

Miall (1977, p. 37), however, maintains that the variability in braided-river sediments calls for rigorous analysis, preferably using statistical techniques. The same author envisages several types of cycles in a braided-river environment:

(i) A flood cycle: a superimposition of beds formed at progressively decreasing energy levels.

(ii) A cycle due to lateral accretion: generated by side-or pointbar growth is possible, as in a meandering-river environment.

(iii) A cycle due to channel aggradation: this cycle would represent the fill of a channel or a local channel system. Waning energy levels would occur during sedimentation, followed by channel abandonment as a result of avulsion.

(iv) A cycle due to channel re-occupation: an abandoned, partly filled channel may be re-occupied by avulsion.

It is also possible that all four types of cycles may occur in a given braided-stream deposit and therefore all interpretations must be based on extremely careful field work and detailed local analyses. To date, insufficient work has been conducted on the Katberg Formation to reach Johnson's (1976) conclusion. One can merely accept the fact that the Katberg Formation was either deposited by a meandering-river system or a braided-river system. Miall (1977, p. 7 - 8) states that according to Schumm and Lichty (1963) a single major flood in 1914 changed the river morphology of the Cimarron River, Kansas, from a meandering, suspension-load stream to a broad, shallow-braided, bed-load stream. It is, however, possible that to the south-east the Katberg Sandstone was deposited by braided streams, which changed to a meandering river system farther to the north-west.

Another important characteristic of the sedimentary rocks of the Beaufort Group is the presence of "red" mudrock in the Middleton Formation, Elandsberg Member, Katberg Formation and Burgersdorp Formation. This red colouration is attributed by Reineck and Singh (1973, p. 132) to the presence of hematite and iron pigment as a matrix or coating on detrital grains. The "red" mudstones, therefore indicate an oxidizing environment, while the greenish-grey and black colours indicate a reducing environment.

It is, however, difficult to determine the exact climatic conditions under which such oxidation takes place. Reineck and Singh (1973, p. 132) point out that red-coloured sediments can originate either in desert or in humid, tropical climates. One may conclude that the Adelaide Subgroup was deposited largely under reducing conditions (except for part of the Middleton Formation), and that the Tarkastad Subgroup was deposited under oxidizing conditions. The Karoo Cycle of deposition terminates with eolian conditions (Beukes, 1969), i.e. the Clarence Sandstone Formation and, therefore, relatively arid conditions could have prevailed during the deposition of the Tarkastad Subgroup.

The deposition of the Beaufort Group can be regarded as a mere continuation of the filling of the Karoo Basin, a process which began with the deposition of the Dwyka Tillite (Late Carboniferous). The environment, however, changed from marine during the Dwyka to fluviatile during the Beaufort.

At least three fining-upward mega-cycles, each marking a period of major tectonic activity in both the provenance and the basin-area are recognised in the Beaufort Group. The first of these cycles started during the late Permian with a major uplift of the source and resulted in the rapid deposition of the Oudeberg Sandstone Member over a vast, but flat area. The massive nature (marked by rippledrift-crosslamination only at the top) of the sandstone lithosomes, is indicative of relatively rapid deposition (Reineck and

Singh, 1973, p. 113). As the tectonic activity subsided, finer material was transported into the basin, resulting in the upward-fining sequences of the Daggaboersnek Member. The second major tectonic cycle started with the deposition of the Barberskrans Sandstone Member and the third is represented by the Katberg Formation. This last event, at the beginning of the Triassic, must have been of considerable magnitude, judging from the amount of coarse clastics deposited. It is concluded that the Oudeberg Sandstone Member thus defines the first pulse of the Cape Orogeny and that subsequent pulses persisted into Molteno times.

3. 1. 4. 4 Source

The sedimentary material supplied to the Karoo Basin during Beaufort times had the same general provenance as those of the Koonap Formation and the Ecca Group. Due to the tectonic activity associated with the Cape Orogeny the basin, however, became cannibalistic and existing units of both the Cape and the lower part of the Karoo sequences in the south supplied material to the basin.

Table 3 - 3 lists the mineral composition of the sandstones of the Beaufort Group.

An increase in the quartz content and a decrease in the feldspar content is clearly illustrated. This same trend can be observed in Fig. 3 - 10 and 3 - 11 and in PLATE V.

TABLE 3 - 3 Mineral Composition Of Sandstone In The Beaufort Group

Unit	Q	F	R	Acc.	Cem.	M	Ratio Q	: F	: R
Burgersdorp	42	9	34	0,5	0,5	14	49	11	40
Katberg	36	16	37	1	1	9	40	18	42
Barberskrans	18	29	39	2	2	10	21	34	45
Oudeberg	20	29	40	0,5	2,5	8	22	33	45
Middleton	19	29	35	1	2	14	23	35	42

Q = quartz

F = feldspar

R = rock fragments

Acc. = accessory minerals

Cem. = cement

M = matrix

This trend is attributed to the recycling of existing sedimentary material in the Cape-Karoo basin and is due largely to the effect of the Cape Orogeny.

Johnson (1976, Table XV1 - 6, p. 267) lists a number of rock-fragment types in the sandstones of the Beaufort Group. Felsite fragments, consisting of a fine-grained quartz-feldspar mosaic, are the most abundant and are followed in order of abundance by micaceous, schistose and chert fragments. This points to the fact that volcanic, metamorphic, granitic and older sedimentary rocks constituted the source.

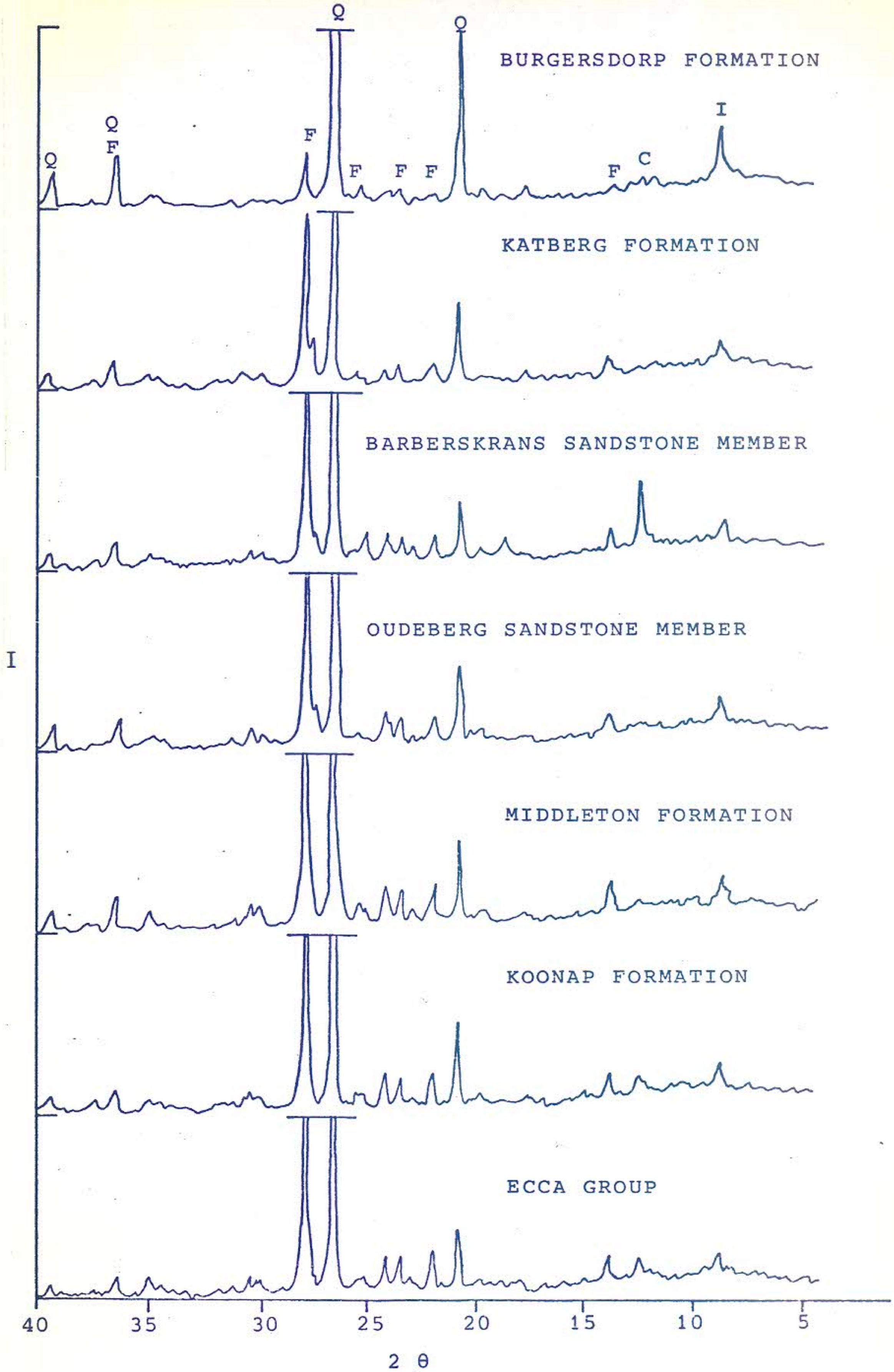


Fig. 3 - 10 X-ray diffractograms of sandstone from different stratigraphic units of the Karoo sequence in the Great Fish River Basin. (Q =quartz, F = feldspar, I = illite, C = chlorite).

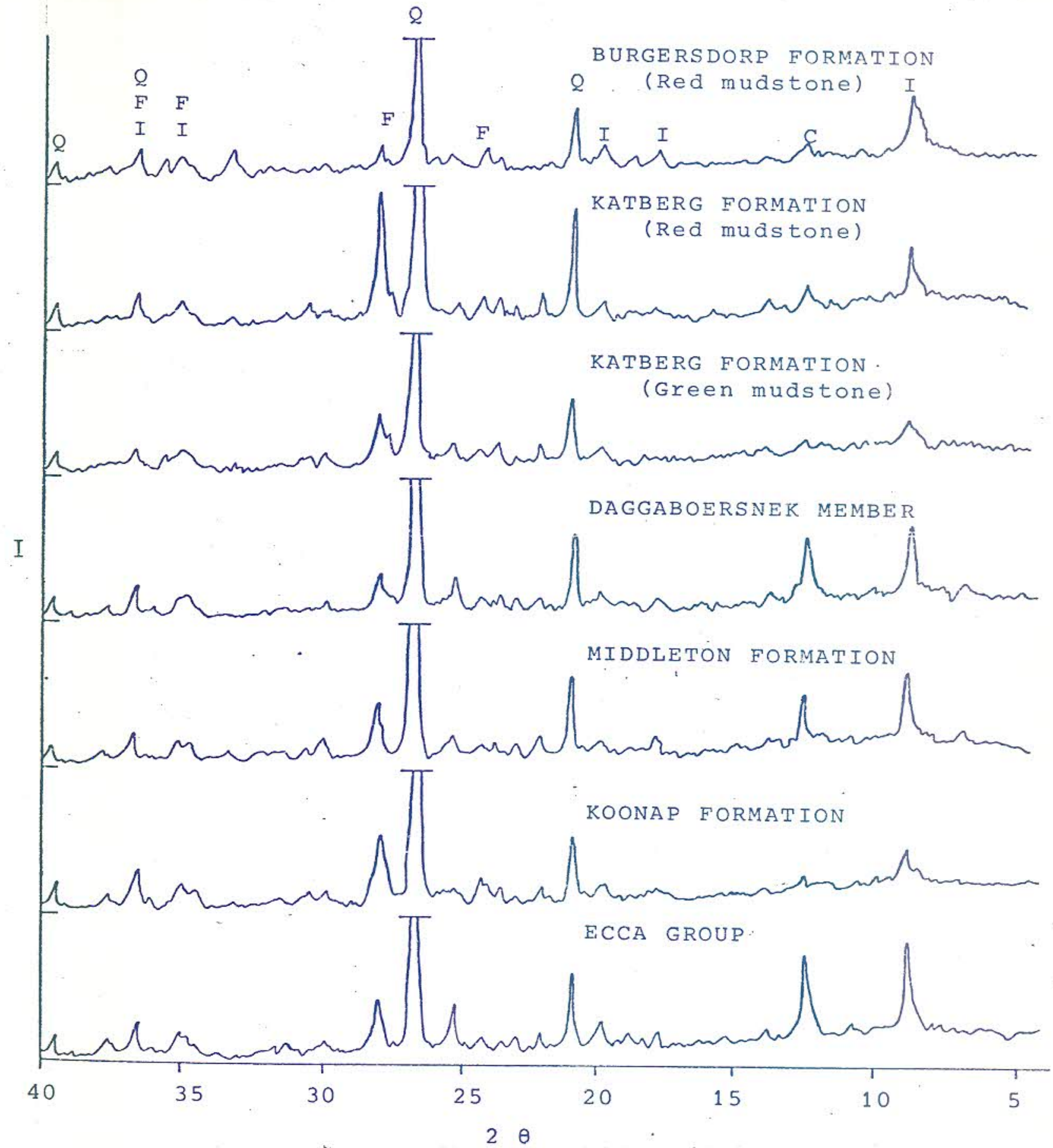


Fig. 3 - 11 X-ray diffractograms of mudstone from different stratigraphic units of the Karoo sequence in the Great Fish River Basin. (Q = quartz, F = feldspar, I = illite, C = chlorite).

Fig. 3 - 10 and 3 - 11 are respectively X-ray diffractograms of sandstone and mudstone/shale from the different stratigraphic units in the area. The matrix of the sandstones (Fig. 3 - 10) consists mainly of illite showing various degrees of crystallinity. The sandstones of the Koonap Formation and of the Ecca Group appear to contain a normal amount of chlorite, which is attributed by Rowsell and De Swardt (1976, p. 91,111) to diagenetic processes active during the deep burial of these sediments.
The anomalous amount of chlorite in the sample of the Barberskrans sandstone may be attributed to the metamorphic influence of a nearby dolerite intrusion and is not the result of anchimetamorphism, which produces chlorite during deep burial.

The mudstones of the area contain a fair amount of quartz (15 - 25 percent PLATE V and Fig. 3 - 11) in a fine-grained matrix of illite and chlorite. A minor amount of feldspar, which is normally altered to some extent to illite, also occurs in these rock types. Fig. 3 - 11 shows a gradual decrease in the chlorite content of the mudstones towards the top of the Karoo sequence. It is interesting to note that alteration of feldspar to kaolinite was not observed.

The origin of the illite and chlorite in the argillaceous rocks is of great importance, especially where the quality of the groundwater is concerned. According to Kübler (1966), only illite, chlorite and occasionally pyrophyllite are present in rock-types which have undergone advanced diagenesis and early metamorphism. Aspects of the diagenesis of the sedimentary rocks of the Karoo sequence are discussed in detail in the following chapter.

Although the sedimentary material of the Beaufort and the Ecca Groups and of the Koonap Formation was derived from the same source, diagenetic processes have altered the clay mineralogy so that slight differences can be observed in the different stratigraphic units.

3. 1. 4. 5 Structure

The structures observed in the Beaufort Group are confined mainly to sin- and post- depositional sedimentary features.

The regional dip of the strata varies from horizontal to less than 5^{o} to the north-east. Higher up in the sequence the dip tends to change towards the east.

A few prominent joints, with an east-west trend, are encountered south of Cradock. Severe jointing can be observed in the Middleton Formation.

Exceptional tilting of the strata has occured locally near intrusions of dolerite.

3. 1. 5 Molteno Formation

A minor outcrop-area of this formation is found around

Bamboesberg east of Steynsburg. No particular attention was paid to this arenaceous unit.

3. 1. 6 Recent Deposits

Vast areas of the Headbasin and some areas of the Marginal Region are covered by relatively deep deposits of soil. Areas of particular interest are the colluvial plains between Middelburg and Steynsburg, around Hofmeyr, Visrivier Station and Mortimer. In the Marginal Region soil covered plains occur to the north of Cookhouse and around Golden Valley.

Most of the soils are derived from the weathering and erosion of the surrounding rock-types. These soils mainly represent colluvial fan deposits, having been transported merely from the mountain area to the plains below. The deposits normally form pediments close to the mountains and may reach depths of up to 50 m.

Alluvial soil is scarce and is confined mainly along river courses. Due to the ephemeral character of most streams, little over-bank deposition takes place. At the confluence of the larger tributaries with the Great Fish River a fair amount of alluvial material is deposited. Of particular interest are the alluvial deposits at the confluences of the Brak and Tarka Rivers with the Great Fish River.

These soils are normally drained by a network of water channels which eventually converge into one major channel.

Poor farm management often results in catastrophic soil erosion as can be seen in Fig. 3 - 12.

Fig. 3 - 12 Typical example of soil erosion in the Headbasin of the Great Fish River. View south of Steynsburg.

The soils are typical of a semi-arid region. Clay minerals (mostly montmorillonite) are predominant throughout the profile and because of the sparse vegetation no true A-horizon has developed. Because of the high clay-mineral content of the soils, they are virtually impervious to penetration of surface water. The soils fit the description of "Brown Soils" as described by Levinson (1974, p. 106).

Levinson (1974, p. 94) states that alkaline soils are characteristically found in dry or semi-arid regions, where water is insufficient to drain completely through to the water table. This results in a "closed" chemical system, because most of the dissolved material is retained in the soil. The lack of water and sparse vegetation from which organic acids may be derived results in less leaching and the precipitation of a calcium carbonate layer, called caliche. According to Levinson (1974, p. 94) caliche is the most striking result of illuviation in semi-arid regions. It often occurs as nodules at various depths, depending on the rainfall. Hawkes and Webb (1962, p. 111) point out that the depth to the upper limit of the calcareous horizon increases with the intensity of the annual rainfall, while the lower limit indicates the maximum depth to which rain water percolates before being dissipated by evaporation or by transpiration by plants. Calcrete or caliche occurs at or near the surface of most of the soils in the area.

Around Halesowen, south of Cradock, the calcrete is quite thick (10 m) and was deposited by circulating groundwater, rather than by penetrating rain water.

3. 2. KAROO DOLERITE

3. 2. 1 Origin And Mode Of Intrusion

During the Triassic, deposition of the Molteno Formation (600 m), Elliot Formation (500 m) and the Clarens Formation (250 m) followed the Beaufort Group (Johnson, 1976).

The close of the depositional cycle is marked by the injection during the Jurassic of basaltic magma into the sedimentary rocks in the form of sills and dykes and the eruption of vast flows of basaltic lava, which resulted in the 1200 m thick basalts of the Drakensberg Group. During the Jurassic the Cape Fold Belt and adjacent area to the north were still subjected to Compressive forces, resulting in the absence of dolerites in this area.

Almost the entire sequence of sedimentary rocks in the Great Fish River Basin north of Middleton is, however, riddled by a complex network of dolerites giving rise to a host of intrusive forms such as concordant sills, transgressive sheets, curved sheets, dykes, bell-jar intrusions and laccoliths.

Du Toit (1920, p. 28 - 36) is of the opinion that only after the lava flows of the Drakensberg Group had accumulated to a thickness of a few thousand metres, did the effusive phase cease and only then did the hypabyssal injection of the Karoo sequence proceed. During this phase the sedimentary rock layers were extensively ruptured and the magma permeated along transverse fissures, fissile bedding planes and irregular fractures, giving rise to the host of intrusive forms mentioned above. Contrary to this suggestion, Lombaard (1952, p. 185) is of the opinion that initially the magma rose along fissures and spread from them in the form of relatively few, large concordant sheets in the lower parts of the Beaufort Group. The magma proceeded from these sills along conical, vertical and inclined fractures and soon reached heights where gasses could escape, thus forming vents. Lombaard (1952, p. 185) goes on to state that the passages to the above vents were effectively plugged by the freezing of the magma from the roots of the vents downward. This plugging resulted in the revival of high vapour pressures at relatively low levels from where magma, finding its upward passages effectively sealed off again, spread laterally to form many smaller sills.

Botha and Hodgson (1976, p. 190) state that the undulating sheets are confined mainly to areas with an original thick sedimentary cover; in the upper part of the Karoo sequence dykes are the normal mode of intrusion.

Rhodes and Krohn (1972, p. 20) proved geochemically that shallow fractionation of the basaltic magma occured in the central Karoo Basin, while high-pressure fractionation occured in areas marginal to the basin.

The mode of intrusion in the Great Fish River Basin changes from south to north. Transgressive sheets with an east-west strike and a dip to the north appear in the Middleton Formation in the south (PLATE III). The first concordant to slightly transgressive sills appear above the Oudeberg Sandstone Member. Although these sills are termed "concordant" local undulations and variations in thickness of individual bodies can be observed. Not more than four of these sills are present and most of them are located in the strata between the Oudeberg Sandstone Member and the Katberg Sandstone Formation (PLATE III).

Transgressive intrusions of dolerite with annular outcrops and inward dips are associated especially with the Katberg and Burgersdorp Formations. The average diameter of these conical bodies is about 20 km. The bodies are invariably dissected by later dykes of dolerite, while smaller ring-shaped structures are often superimposed on the larger ones. In some cases the rings are arcuate and incomplete. Lombaard (1952, p. 183) notes that when followed laterally or vertically along the horizontal sediments, these basin-shaped intrusions may pass upwards into concordant or inclined sheets or into vertical dykes. Lombaard (1952) is of the opinion that they closely resemble true cone-sheets, from which they differ mainly in not being associated with ring-dykes and in the absence of two or more concentrically arranged bodies. An arrangement of incomplete concentric bodies, however, does occur in the area immediately north of Lake Arthur.

The dykes in the area vary from a few centimetres to several metres in thickness, while their average length is approximately 1 km. Dykes of several kilometres in length are, however, present in the area. One such dyke runs for 140 km from Middelburg, past Cradock and disappears in a concordant sill in the mountains to the north of Cookhouse. This dyke can clearly be seen when crossing over the Witkransnek, about 30 km south of Middelburg. About 2400 dykes totaling 3100 km in length are encountered in the area.

Fig. 3 - 13 illustrates the orientation of the dolerite dykes in the area. A significant change in orientation can

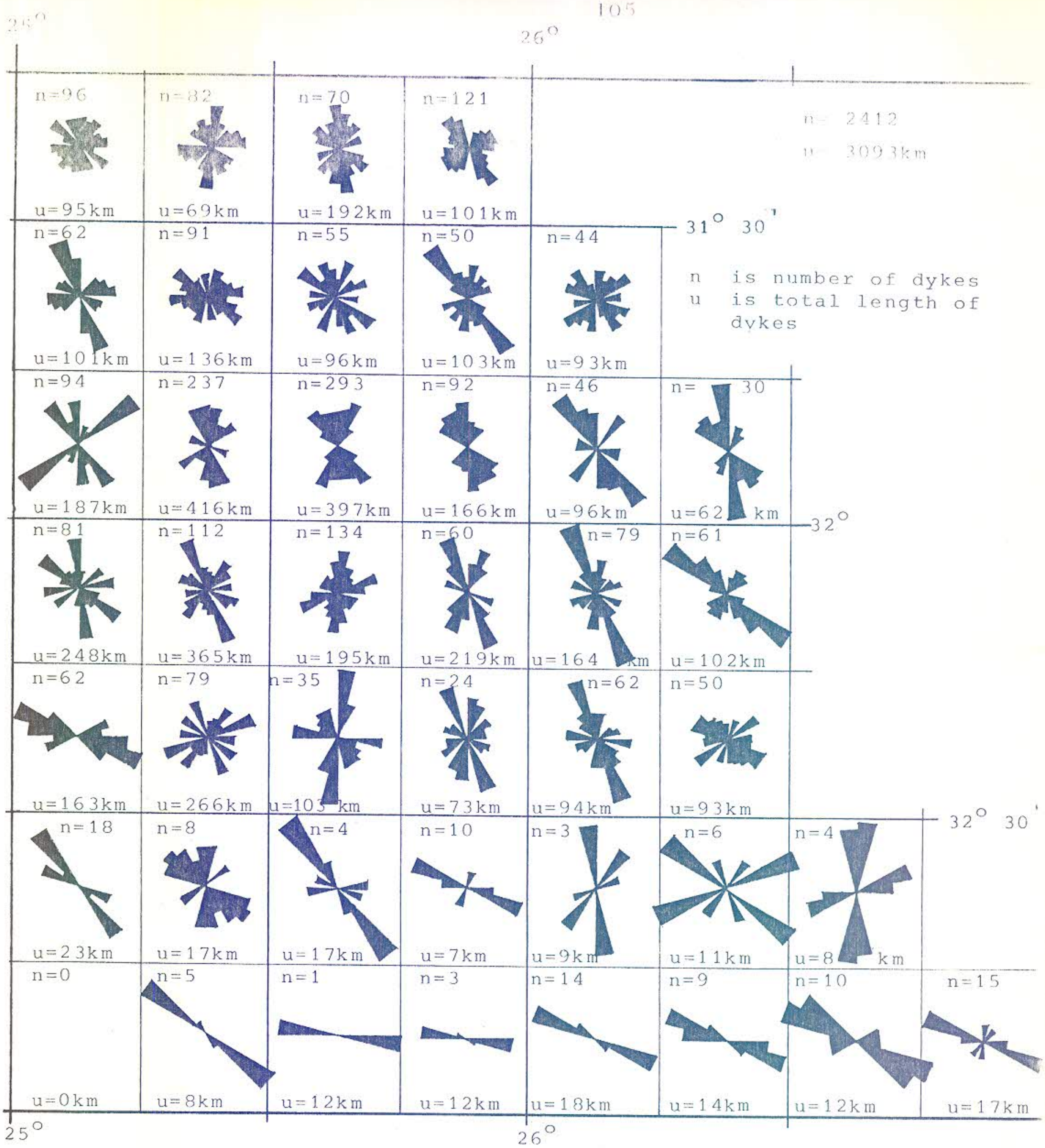


Fig. 3 - 13 Diagram showing the orientation of dolerite dykes in the Great Fish River Basin.

be observed from south to north:

(i) No dykes are present south of latitude 33° S.

(ii) The predominant direction between 32° 45'S and 33° S is approximately 290°; which is almost parallel to the fold axes in the Cape Fold Belt of the area under discussion.

(iii) The dominant direction between 32° 15'S and 32° 45'S is approximately 315°, although a minor trend between 65° and 90° makes its appearance.

(iv) Between 31° 45'S and 32° 15'S the dominant trend is 345° although a north-easterly trend becomes prominent.

(v) North of 31° 45'S the spread becomes more random although a north-westerly and a north-easterly trend can still be observed.

Greef (1968, Fig. 7) shows that the dominant direction of fractures in the Southern African sub-continent is either north-west or north-east or both. The dominant direction of strike of dolerite dykes in the Great Fish River Basin also reveals this trend. It is believed, however, that both the folding in the south, as well as the thickness of sediment cover had an influence on the fracture pattern in the area under discussion.

3. 2. 2 Petrography

The dolerite sills are less than 100 m thick and this did not promote a pronounced gravity differentiation. Limited differentiation did, however, take place in the thicker sills, thus developing rocks like picrite near the base. Such a differentiated sill is described by Walker and Poldervaart (1942, p. 58) near Middelburg.

Most sills and dykes reveal chilled selvages of a few centimetres in contact with the surrounding sedimentary rocks. The material in the selvage is usually fine grained and can be regarded as "contact basalt" (Maske, 1966, p. 33). This chilled phase is composed mainly of a fine-grained plagioclase and pyroxene, with minor amounts of biotite, apatite and opaque minerals. The feldspar forms a network of minute laths which are often randomly embedded in pyroxene. Occasional large plagioclase phenocrysts in glomeroporphyritic aggregates give the rock a porphyritic appearance.

Most of the dolerite sills in the area, however, resemble the Hanover-type which is described by Walker and Poldervaart (1942, p. 58). Other types of dolerite encountered in the area are the Perdekloof/Blaauwkrans-type described by McLaren (1974, p. 166) and quartz-dolerite. The following is a brief description of the types of dolerite encountered:

Hanover-type. This is the most common type present. The rock is a medium- to coarse-grained, ophitic, olivine dolerite which is characterised by the presence of large phenocrysts (2 - 4 mm) of orthopyroxene (Fs_{20}). Smaller crystals of Ougite ($2V_z = 60^o$, $Z \wedge c = 48 - 54^o$) often contain cores of pigeonite.

Plagioclase (An_{57}) is either ophitically intergrown with pyroxene or is present as seriated laths which constitute the interstitial material.

The olivine (Fa_{70}) is subhedral to anhedral with rounded edges and has much the same grain size as the plagioclase, i.e. 0,8 mm.

Magnetite is invariably present as skeletal crystals, showing signs of resorption.

Perdekloof/Blaauwkrans-type. This is a fine- to medium-grained type and contains no orthopyroxene.

The clinopyroxene (augite) consists of large ($>$ 2 mm), tabular phenocrysts, often with a poikilitic core of magnesium-rich pigeonite. Very rarely iron-rich pigeonite constitutes the outer edges of the augite. The clino-pyroxene is ophitically to sub-ophitically intergrown with plagioclase.

Plagioclase once again constitutes the interstitial material. In many cases the small ($<$ 0,2 mm) plagioclase laths seem to have grown from a central point, thus revealing a radiating pattern. According to Maske (1966, p. 24) this texture is indicative of fast cooling and is more common in the chilled zones where the "contact basalts" occur. Plagioclase constitutes more than 50 percent of this rock-type.

Occasionally large tabular crystals of plagioclase occur as clusters, thus presenting the rock with a glomeroporphyritic texture.

Olivine is present in varying amounts. When present, this mineral is extremely altered and appears as irregular crystals.

Picritic Dolerite and Olivine Hyperite. A picrite, in which olivine constitutes more than half of the rock, is described by Walker and Poldervaart (1942, p. 58) in a locality 12 km north of Middelburg. The basal part of most of the thicker sills consists of a medium-grained dolerite with an olivine (Fa_{18}) content of 15 to 25 percent (picritic dolerite) or an olivine-hyperite containing 20 - 35 percent olivine.

Some of the dykes of dolerite in the area contain up to 30 percent olivine (Fa_{20}).

The euhedral to subhedral phenocrysts of olivine are embedded in a finer matrix of plagioclase and pyroxene. In many cases the olivine shows alteration to green serpentinitic material and magnetite along parallel fractures which often cross through the adjoining pyroxene. Rounded, poikilitic olivine in large tabular crystals of pyroxene are common.

The pyroxene, mainly augite, is tabular and is ophitically to subophitically intergrown with plagioclase. Orthopyroxene, which is typically non-ophitic, is present in subordinate amounts.

The plagioclase is lath-shaped and seriate with an average length of 0,4 mm.

Quartz Dolerite. Walker and Poldervaart (1942), Lombaard (1952), Mountain (1960) and Maske (1966) have described felsic rocks which are associated with the Karoo dolerites. Mountain (1960) presents evidence of sediments which were mobilized by the intrusive basaltic magma and furthermore discusses some xenoliths of sedimentary rocks, dolerite pegmatite, coarse granophyre and albitic veins and inclusions which are associated with dolerite. Maske (1966, p. 83) demonstrates the differentiation trend in the Ingeli intrusion; crystal fractionation resulted in a continuous gradation from picrite, through gabbro, to quartz monzonite.

Leucocratic dolerite, which has intruded a large sill of normal dolerite in the form of a transgressive sheet (which eventually changes upward into a concordant sill), is located in the area around Speelmanskop, about 15 km to the east of Visrivier Sation (Fig. 3 - 14). The average thickness of this body is about 50 m where it forms the sill and about 20 m where it forms the transgressive sheet.

The sill has a slight dip ($< 5^{o}$) to the south and follows

Alluvium

Colluvium

Dolerite Dyke

Leucocratic dolerite

Dolerite

Katberg Sandstone Formation
Balfour Formation

Road

River

Farm Boundary

Fig. 3-14 Geological map showing the distribution of leucocratic dolerite in a sill of normal dolerite around Speelmanskop.

the regional dip of the surrounding dolerite sill.
In some places this body appears to pinch out abruptly in the dolerite.

The rock is medium grained and has a more leucocratic appearance than the surrounding normal dolerite. There is, however, a tendancy for this rock to become lighter toward its contact with the overlying dolerite.

All the contacts with the surrounding dolerite are sharp, showing no gradational features. At the contact with the underlying dolerite a large number of xenoliths of normal dolerite (approximately 4 cm X 2 cm) are included in the leucocratic phase (Fig. 3 - 15). The number of inclusions decreases upward and no inclusions are present in the top half of the sill. These xenoliths, are proof of the intrusive nature of the leucocratic dolerite.

Fig. 3 - 15 Inclusions of normal dolerite in leucocratic dolerite at Speelmanskop.

In the lower half of the sill, plagioclase (An_{50-55}) is the dominant mineral and consists of seriated laths and tabular crystals. The larger tabular crystals often show zoning which is absent in the laths. Alteration of this mineral to saussurite is common and toward the top of the sill the plagioclase is altered to such an extent that it is hardly recognizable.

Augite is relatively abundant as large tabular phenocrysts in the lower parts; augite, rimmed by biotite and chlorite becomes progressively more abundant toward the top. In the top half of the sill only biotite and chlorite are present.

The most striking feature of this rock is the myrmekitic intergrowths of plagioclase and quartz, which becomes more abundant toward the top(Fig. 3 - 16). Staining tests as

ppl x 80

Fig. 3 - 16 Micrographic intergrowths of quartz and plagioclase in Quartz dolerite near Speelmanskop.

[illegible] (1974, p. 18), together with the
[illegible] vestigations, proved K - feldspar to be
[illegible] ples which were examined. Terms such as
[illegible] ntain, 1960, p. 145), "granophyre"
[illegible] rt, 1942; Lombaard, 1952 and Mountain,
[illegible] zonite" (Maske, 1966) may therefore not be
[illegible] fic rock-type. The most appropriate
[illegible] ite".

[illegible] kite toward the top is accompanied by
[illegible] "free" quartz; right at the top of the
[illegible] as abundant as plagioclase, while biotite
[illegible] accessory amounts.

[illegible] rystalization of this intrusive,leucocratic
[illegible] e occurrence of lenses of medium- to
[illegible] -plagioclase material near or at the
[illegible] Round nodules (1,5 cm in diameter),
[illegible] orite and zeolites, give this rock-type
[illegible] earance. Prospectors mistook this for
[illegible] several pits were sunk along the
[illegible] rly 1920's.
[illegible] mineralisation (chalcocite and
[illegible] ever, present along the horizon.
[illegible] rs more readily than the normal
[illegible] cterized by the presence of calcrete

[illegible] several dolerite samples from the area
[illegible] le 3 - 4. The CIPW - norms, which are

described by Hutchison (1974, p. 18), together with the normal microscopic investigations, proved K - feldspar to be absent in all the samples which were examined. Terms such as "micropegmatite" (Mountain, 1960, p. 145), "granophyre" (Walker and Poldervaart, 1942; Lombaard, 1952 and Mountain, 1960) and "quartz monzonite" (Maske, 1966) may therefore not be applied to this specific rock-type. The most appropriate term is "quartz dolerite".

The increase in myrmikite toward the top is accompanied by the crystalization of "free" quartz; right at the top of the sill quartz becomes as abundant as plagioclase, while biotite and chlorite occur in accessory amounts.

The final stage of crystalization of this intrusive,leucocratic phase is marked by the occurrence of lenses of medium- to coarse-grained quartz-plagioclase material near or at the overlying dolerite. Round nodules (1,5 cm in diameter), containing mainly chlorite and zeolites, give this rock-type a "conglomeratic" appearance. Prospectors mistook this for gold-bearing reef and several pits were sunk along the horizon during the early 1920's.
Very low-grade copper mineralisation (chalcocite and chrysocolla) are, however, present along the horizon.
This rock type weathers more readily than the normal dolerite and is characterized by the presence of calcrete at the surface.

Chemical analyses of several dolerite samples from the area are presented in Table 3 - 4. The CIPW - norms, which are presented in the above table, were calculated according to the method after Johannsen (1931) as modified by Kelsey (1965) in Hutchison (1974, p. 414 - 417).

There is a close correspondence between the analyses of normal dolerite (Samples 1 and 2) at Speelmanskop, the average composition of Karoo dolerite and the upper chilled basalt from Ingeli as presented by Maske (1966, Table V). Rhodes and Krohn (1972, p. 12) have, however, indicated that there is a difference between the chemical composition of dolerite from the central Karoo Basin and that from the basin margins. This contradicts the findings of Walker and Poldervaart (1949) who state that the basaltic magma in the northern part of the basin have a lower SiO_2 and Al_2O_3 and a higher TiO_2 and Fe_2O_3 content than the magma in the south. Rhodes and Krohn (1972, p. 14) comes to the conclusion that:

(i) Rocks from the central Karoo Basin are significantly higher in SiO_2, Al_2O_3, MgO and CaO and lower in Fe_2O_3, Na_2O, K_2O, TiO_2 and P_2O_5 with respect to rocks from the basin margins.

(ii) Rocks from the basin margin north of 26°S are similar to those from the basin margin south of 26°S except for SiO_2, TiO_2 and Fe_2O_3, which show significant differences.

(iii) Rocks from the central basin are significantly higher in Al_2O_3, MgO and CaO and lower in TiO_2, Na_2O, K_2O and P_2O_5 with respect to rocks from the basin margin south of 26°S".

TABLE 3 - 4 Chemical Analyses Of Karoo Dolerite From The Great Fish River Basin.

	1	2	3	4	5	6
SiO_2	51,02	52,00	55,01	54,12	58,15	65,31
TiO_2	0,78	0,61	0,81	0,69	0,81	0,69
Al_2O_3	14,83	14,55	13,68	13,04	14,90	5,33
Fe_2O_3	2,27	2,11	2,31	2,19	2,31	2,19
FeO	7,88	7,71	7,17	7,53	5,22	3,06
MnO	0,18	0,17	0,16	0,17	0,12	0,08
MgO	6,87	6,49	7,50	9,35	3,65	1,93
CaO	11,25	9,29	7,83	7,94	5,99	2,10
Na_2O	1,86	1,91	1,77	1,62	1,63	1,46
K_2O	0,66	0,89	1,35	0,98	2,04	3,45
P_2O_5	0,10	0,11	0,14	0,15	0,16	0,20
LOI	0,37	0,55	0,54	0,34	3,06	3,91
	98,07	96,39	98,27	98,12	98,04	99,68

C.I.P.W. - NORMS

	1	2	3	4	5	6
Salic						
Q	3,71	6,55	10,04	8,29	20,36	35,51
Or	3,90	5,23	7,96	5,57	12,02	20,37
Ab	15,73	16,15	14,95	13,63	13,79	12,38
An	30,16	28,52	25,43	25,59	27,32	9,13
C	-	-	-	-	-	5,85
femic						
Ap	0,24	0,37	0,30	0,34	0,37	0,47
Il	1,46	1,15	1,53	1,37	1,55	1,32
Mt	3,29	3,06	3,33	3,24	3,36	3,17
Di	20,41	13,71	10,26	10,67	1,12	
Hy	18,73	21,01	24,23	29,83	15,30	7,58
LOI	0,37	0,55	0,54	0,34	3,06	3,91
	98,00	96,20	98,57	98,87	98,25	99,69

Analyst. Department Of Geology, UOFS

1. = Normal dolerite below sill of quartz dolerite.
2. = Inclusions of normal dolerite in quartz dolerite.
3. = Quartz dolerite near bottom contact.
4. = Bulk sample of quartz dolerite with inclusions of normal dolerite
5. = Quartz dolerite near the top of the sill.
6. = Quartz dolerite at a higher elevation in an up-dip direction.

The above differences are attributed by Rhodes and Krohn (1972, p. 21) to low-pressure fractionation of the magma in the central basin, compared with high-pressure fractionation in areas marginal to the basin.

Analyses 3, 4, 5 and 6 (Table 3 - 4) represent the quartz dolerite sill at Speelmanskop.

There is a pronounced increase in SiO_2, K_2O and Al_2O_3 from the base to the roof of the sill, as well as in the up-dip direction. The MgO, CaO and FeO decreases in the same direction. This trend corresponds well with the general increase in quartz and disappearance of pyroxene toward the top of the sill.

Of particular interest is analysis 6 (Table 3 - 4) in which corundum appears in the norm. This rock is located at the highest point in the inclined sill. Maske (1966, p. 79) suggests that the peraluminous character of the Ingeli quartz monzonite is related to the replacement of primary ferromagnesian minerals by "chlorophaeite" during the deuteric stage. All the analyses of the quartz monzonite at Ingeli (Maske, Table 1, analyses 12, 13 and 14) produce normative corundum. Scholtz (1936, p. 96) suggests that this is indicative of the assimilation of sedimentary material.

Scholtz (1936, p. 146), by comparing chemical analyses of hornfels situated at different distances below the floor of the Intsizwa Sheet, deduced that alkalis were transfered from the intrusive magma into the adjacent wall rock.

This deduction is supported by Walker and Poldervaart (1949, p. 679 - 681) who also present evidence of the transfusion of Na- and K- bearing emanations from the highly fractionated residue of the dolerite magma into the surrounding sediments. This is accompanied by a simultaneous metasomatic transfer from Al from the sediments to the magma. They suggest, from chemical data, that metasomatic processes could have been responsible for the generation of quartz dolerite.

Maske (1966, p. 79 - 80), however, points out that it is not unusual for residual magmas to contain alumina in excess of that required to form feldspar, and therefore concludes that the felsic quartz monzonite phase at Ingeli (which has a similar chemistry as sample 6, Table 3 - 4) could have been derived from the original magma by strictly magmatic processes.

The quartz dolerite at Speelmanskop could have been the result of a metasomatic, a syntectic or a magmatic differentiation process, but the fact remains that, because of the intrusive nature, none of these could have occured in situ. The leucocratic material must have been generated at a greater depth. If the quartz dolerite is the product of a metasomatic or a syntectic process, the sedimentary rocks must have melted to a certain extent before being intruded into the surrounding dolerite sill. Walker and Poldervaart (1949, p. 675) are of the opinion that pure melting of sediments by the dolerite magma is rare. Furthermore, the amount of dolerite magma required to produce a body the size and extent of that at Speelmanskop would be far greater than that of any dolerite sill encountered in the area. It is therefore concluded that this leucocratic sill is probably the result of magmatic differentiation lower down in the sequence, possibly at a late magmatic stage of the relatively thick concordant sills present in the Daggaboersnek Member. It is interesting to note that this rock-type occurs in the zone in which conical sheets are present. Limited differentiation within the leucocratic body itself, both from floor to roof and in an "up-dip" direction, must have occurred as well.

Fig. 3 - 17 illustrates the MgO-FeO- (Na_2O + K_2O) variation in this sill, compared with variations in similar rocks associated with Karoo dolerite. The Speelmanskop quartz dolerite closely resembles the middle and the late stages of crystallization of the Ingeli magma (Maske, 1966, Figure 12). There is a marked enrichment in alkalis together with a slight increase in FeO from the bottom to the top of the sill. Sample 6, however, which was obtained from the highest elevation in the sill shows a rather sharp decline in FeO and MgO in relation to the alkalis.

All the chemical analyses of the felsic rocks described by Lombaard (1952, Table 111) and Mountain (1960, p. 142) fall in the alkali-rich region of the diagram.

3. 2. 3 Effect On Wallrocks

The adjacent sedimentary rocks show signs of induration and

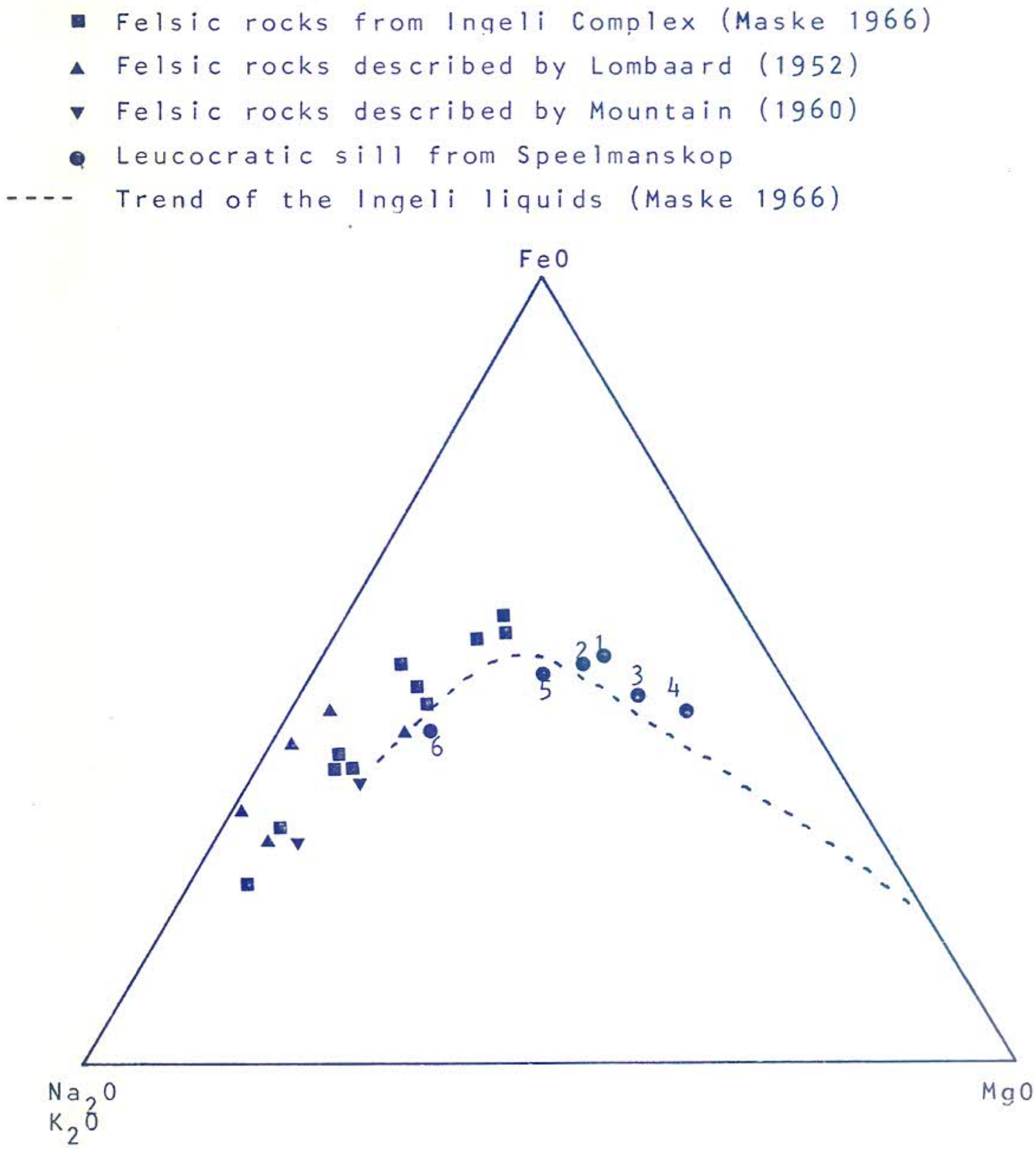


Fig. 3 - 17 Variation diagram of the Speelmanskop sill and felsic rocks associated with Karoo dolerite from other areas.

in some cases the mudrocks were altered to crystaline porphyroblastic hornfels. The porphyroblasts are normally nodular (5 - 10 mm in size) and contain mostly quartz, cordierite, biotite, magnetite and pyroxene. This effect increases with the thickness or width of the intrusion.

In some cases Walker and Poldervaart (1949, p. 675) found that the dolerite magma actually mobilized some of the sedimentary rock which then intruded the dolerite body itself, as well as the surrounding rock, as rheomorphic veins.

One of the most important effects of the intrusions, is the fracturing of the immediate adjacent rocks, thus increasing the porosity and permeability. Most of the bore-holes for groundwater are therefore located in the contact zone of dolerite intrusions. Some of the dykes have created a passage for deep-percolating groundwater to reach the surface, e.g. the warm sulphuretted spring north of Cradock.

4. GEOCHEMISTRY

4. 1, INTRODUCTION

There is a perpetual interaction in the geospheres between the lithosphere, hydrosphere, atmosphere and biosphere. This interaction, which results in the transfer of matter from one sphere to another, is brought about by the tendancy to reach a state of equilibrium. Should equilibrium, however, be reached, it is soon disturbed by the addition of foreign substances or the removal of substances which are essential to maintain that equilibrium. One of the most important results of the above interaction is the weathering of the lithosphere due to changes in the chemical environment in which the various rock-types find themselves. Minerals, formed under magmatic, hydrothermal, metamorphic or sedimentary conditions, become potentially unstable when exposed to the present atmospheric conditions. The result is that these minerals are attacked largely by water (hydrosphere), oxygen and carbon dioxide (atmosphere) and to a lesser extent by elements of the biosphere.

Apart from rock-weathering, other geochemical reactions which have an effect on the chemistry of the hydrosphere (surface and groundwater) are the diagenetic processes which proceeded the deposition of the sediments in the Karoo Basin and the adsorption and ion exchange during the interaction of the surface and groundwater with the surrounding rocks.

4. 2. CHEMICAL WEATHERING

The processes involved in the weathering of rocks and minerals of the Earth's crust are discussed by authors such as Lukashev (1958), Hawkes and Webb (1962), Krauskopf (1967), Loughnan (1969) and Levinson (1974). According to Loughnan (1969, p. 28) three simultaneous processes appear to be involved in the weathering of the silicate minerals:

(i) The breakdown of the parent mineral structures with the concomitant release of cations and silica. The released silica may be reduced to the monomeric form, that is, a molecularly dispersed state, or it may persist in a polymerized form, that is much of the original structure is retained.

(ii) The removal in solution of some of the "released" constituents.

(iii) The reconstitution of the residue with components from the atmosphere such as water, oxygen and carbon dioxide, to form new minerals which are in stable or metastable equilibrium with the environment.

The following aspects of rock weathering are regarded as important factors controlling the chemical quality of the groundwater in the Great Fish River Basin and will therefore be discussed briefly:

(i) Some Reactions Of Chemical Weathering.

Chemical weathering of rocks is merely one aspect of the entire weathering phenomenon which also entails mechanical processes such as expansion of water on freezing, growth of plant roots and swelling of minerals due to hydration.

Another point which is stressed by Krauskopf (1967, p. 100) is that due to the slowness of the weathering reactions, which normally occur at atmospheric temperatures and pressures, the knowledge of weathering is largely qualitative rather than quantitative. It is therefore possible to decipher what happens chemically in the decay of a rock, but there are no means of predicting accurately what the state of the rock will be at a particular time in the future. Krauskopf (1967, p. 99) has observed that all chemical reactions related to chemical weathering involve four relatively simple processes: ionization, addition of water and carbon dioxide, hydrolysis and oxidation. The dissolving of soluble minerals and the addition of water to form hydrates are among the simplest of the weathering reactions (Krauskopf, 1967, p. 107). Such reactions are confined mainly to the evaporites and carbonate rocks and will not be discussed here in any great detail.

Oxidation has a great influence on the mobility of certain cations such as iron and titanium, which are capable of existing in more than one valence state. Loughnan (1969, p. 41 - 42 and Fig. 15) illustrates the mobility of iron as ferric hydroxide, which becomes soluble in water only at pH-values smaller than 3 and is therefore normally retained in the weathering environment. Ferrous hydroxide, however, is soluble in water up to pH-values of 9 and is therefore often removed from the environment by natural waters which have a pH of between 4 and 9.

The redox potential (Eh) in a weathering environment is controlled by the accessibility of atmospheric oxygen and the presence or absence of organic matter (Loughnan, 1969, p. 43). Oxidation tends to proceed spontaneously in the airated zone above the groundwater table. Fluctuations in the level of the groundwater table can therefore often be responsible for the accumulation or mobilization of certain cations. It can therefore safely be deduced that the Eh of an environment is dependent on the prevailing climate and topography of the area. (Loughnan, 1969, p. 43).

According to Krauskopf (1967, p. 113), the weathering of silicates is primarily a process of hydrolysis. He defines hydrolysis as the reaction between water and ions of a weak acid or a weak base. In the natural state at normal atmospheric pressure and temperatures of 25°C water tends to dissociate as follows:

$$H_2O = H^+ + OH^- \qquad Kw = (H^+)(OH^-) = 10^{-14}$$

Where the water comes into contact with the surfaces of the silicate minerals by penetrating through pores and cleavages in the rock and micro openings in the minerals, the hydrogen ions which are produced are small enough to penetrate into the crystal lattice. The high charge-to-radius ratio of these H^+ - ions which is greater than for any other ion, has a marked disrupting effect on the charge balance within the lattice (Loughnan, 1969, p. 31) In order to restore the charge balance, various cations and polyhedra belonging to the crystal lattice are released to the surrounding water. According to Loughnan (1969, p. 29) the released polyhedra may form amorphous colloids which with ageing become oriented into the structures of the secondary minerals such as clays and oxides.

The weathering of forsterite is described by Krauskopf (1967, p. 113) to illustrate a simple example of hydrolysis:

$$Mg_2SiO_4 + 4H^+ + 4OH^- \longrightarrow 2Mg^{++} + 4OH^- + H_4SiO_4$$

The extent to which the above reaction will proceed to the right is clearly a function of the $H^{\pm}$ concentration (pH) of the water. Water near the surface often contains more H^+ - ions because of CO_2 which it dissolves. The amount of dissolved CO_2 in natural waters is furthermore a function of the climate e.g. more CO_2 is dissolved under cold climatic conditions than under hot conditions. Therefore, by incorporating carbonic acid in the equation, Krauskopf (1967, p. 113) suggests the following reaction:

$$Mg_2SiO_4 + 4H_2CO_3 \longrightarrow 2Mg^{++} + 4HCO_3^- + H_4SiO_4$$

The hydrolysis reaction becomes more complex where silicates containing several cations are involved. In such cases the different cations go into solution at different rates, with the result that the silicate grains are coated with an outer shell which protects the interior of the grain to further attack.

During the weathering of aluminum silicates some clay mineral or another is produced as one of the residual products, thus presenting further problems. De Vore (1959) points out that decomposition of the feldspar structure releases chains which have a certain degree of stability and retain the original Si - Al ordering of the tetrahedra. The released chains can polymerize directly into tetrahedral sheets if they originate from the (100) and(010) surfaces of the feldspar. Clay minerals like illite, montmorillonite and chlorite form when the above sheets are bound by cations such as Al^{3+}, Mg^{++}, Fe^{++}, Fe^{3+} and K^{+}. Where the parent minerals break down to individual tetrahedra, kaolinite is formed. The breakdown of Al-silicates therefore occurs in a series of steps and Krauskopf (1967, p. 114) suggests that feldspar first breaks down to gibbsite $[Al(OH)_3]$, setting silica free as dissolved silicic acid, which later reacts with the gibbsite to form kaolinite. The following simplified hydrolysis reaction involving K-feldspar is suggested by Krauskopf (1967, p. 114):

$$\underset{\text{K- feldpar}}{4\,KAlSi_3O_8} + 22H_2O \longrightarrow 4K^+ + 4OH^- + \underset{\text{kaolinite}}{Al_4Si_4O_{10}(OH)_8} + 8H_4SiO_4$$

According to Krauskopf (1969, p. 115) any solution in contact with silicate minerals cannot long remain appreciably acid, and if contact is continued, the solution must eventually become alkaline. This increase in pH (which seldom exceeds a value of 9) is normally confined to closed systems where the released components of weathering are retained in the environment. Such conditions are often encountered in arid and semiarid environments.

If the system is open and the soluble products are removed at a rate equal or greater than their release from the parent mineral, the residue becomes more acid (Loughnan, 1969, p. 39). Loughnan (1969, p. 39) therefore reaches the conclusion that the pH of the environment is not an independent variable in chemical weathering but rather is a function of several interrelated factors: (a) the composition and structure of the parent minerals, (b) the rate of leaching of the bases and (c) the nature and, in particular, the cation exchange capacity of the residual mineral products. The cations which are readily lost to the environment by leaching are Ca^{++}, Mg^{++} and Na^{+}, whilst K^{+} is readily lost, but the rate may be retarded through fixation in the illite structure. Fe^{3+}, Si^{4+} and Al^{3+} are relatively immobile and are mostly retained in the environment.

The ambient pH, however, has a large influence on the nature of the residual and soluble end-product of chemical weathering.

Lukashev (1970, p. 89) illustrates the transformation of muscovite under different pH conditions as follows:

$$\underset{\text{muscovite}}{(OH)_2KAl_2(Alsi_3O_{10})} \xrightarrow{pH = 9,5} \underset{\text{illite}}{(OH)_2\ K_{1-n}\ (H_2O)nAl_2(AlSi_3O_{10})} \xrightarrow{pH = 9,5 - 7,8\ (?)}$$

$$\underset{\text{beidellite}}{n(H_2O)(OH)_2Al_2(AlSi_3O_{11})} \xrightarrow{pH = 8,5 - 7,5} \underset{\text{montmorillonite}}{nH_2O(OH)_2Al_2(Si_4O_{10})} \xrightarrow{pH = 8,5 - 7,0}$$

$$\underset{\text{halloysite}}{nH_2O(OH)_4Al_2(Si_2O_5)} \xrightarrow{pH = 7 - 6} \underset{\text{metahalloysite}}{H_2O(OH)_4Al_2(Si_2O_5)} \xrightarrow{pH = 6 - 5\ (?)}$$

$$\xrightarrow{pH = 5 - 3,5} \underset{\text{kaolinite}}{(OH)_4\ Al_2\ (Si_2O_5)}$$

Crystal structures obviously play an important role in rock-weathering and Goldich (1938, Table 18, p. 56) presented a sequence of resistance to weathering for the common rock-forming minerals, which coincides closely with the Bowen sequence for the crystallization of minerals from a silicate melt. This sequence appears to be dependent on the number of Si - O - Si bonds in the mineral structure and is zero in the case of olivine (nesosilicate) and 4 in the case of quartz (tectosilicate), which is the most resistant to weathering. In the case of olivine, which consists of isolated silica tetrahedra bonded together by magnesium and ferrous ions, percolating water easily leaches the above ions from the margins and fractured surfaces of the crystals. This loss of the cationic bridges releases the silica tetrahedra and fresh surfaces are thereby exposed for further attack by the percolating waters (Loughnan, 1969, p. 60).

In spite of the highly mobile potassium and sodium ions which form essential constituents of the alkali feldspars, these minerals are relatively resistant to breakdown because of the framework structure of the silica tetrahedra. According to Loughnan (1969, p. 60) the Al - O - Si bonds in the feldspar structure form the weakest links in the chains and therefore the greater the substitution of aluminum for silicon in the tetrahedra, the greater the number of weak links. This fact accounts for the higher susceptibility of Ca^{++} - rich plagioclase to weathering than Na^{+} - rich plagioclase.

One can therefore conclude that the rate of weathering of silicate minerals by means of hydrolysis is greatly dependent on the crystal structure of the primary minerals, as well as on the prevailing pH of the weathering environment. The pH in turn is dependent on the rate of leaching, which is a function of the topography and climate.

(ii) Environmental Factors Influencing Chemical Weathering.

A state of equilibrium is soon reached between the parent material and the weathering agents, should the products of the weathering not be removed from the environment through leaching.

The equilibrium may furthermore be disrupted by the introduction of foreign constituents from another environment. These constituents may either be individual ions, polymerized silica or both, which then react with the soluble products of weathering. Such reactions may produce a pecipitate which results in the removal of constituents essential for maintaining the state of equilibrium and therefore the weathering reactions may proceed further.

In the previous section it was suggested that the most important agents of rock-weathering are water and air; the water to partake in the hydolysis reactions, as well as to leach the soluble products, whilst the air produces oxygen for the oxidizing reactions and CO_2 to produce H_2CO_3. The influence of these agents is, however, dependent on certain environmental factors.

Levinson (1974, p. 87) regards climate, biological activity, parent material, topography and time as the more important environmental factors which influence chemical weathering. In this study the influence only of climate, topography and parent material will be discussed briefly. The biological activity is mainly dependent on the climate, whilst the time taken for weathering to reach a mature stage, is a function of all the above factors combined.

Climate. Both Loughnan (1969, p. 67) and Levinson (1974, p. 87) confirm that climate is one of the major environmental factors which control chemical weathering.

The annual rainfall in an area controls the amount of water which is available for the chemical reactions and for the leaching of the soluble products from the weathering environment. Temperature on the other hand controls the reaction rate, the amount of CO_2 which can dissolve in the meteoric water and the rate of evaporation. Apart from this, climate also controls the vegetation of an area, which in turn is a biogenic weathering agent.

The influence of climate on rock-weathering in arid regions is explained by Loughnan (1969. p. 67) as follows: " where evaporation exceeds rainfall, water may penetrate the rocks, but during the ensuing long dry spell, it is returned to the surface and ultimately becomes lost through evaporation. As a result, soluble constituents of the rocks are not removed and reactions are slowed down accordingly. Moreover, not only is vegetation scant but, in addition, plant debris is quickly destroyed by the oxidizing atmosphere brought about by high temperatures and the considerable depth to the water table. Consequently, such areas are characterized by an abundance of unaltered or partly altered parent minerals, the presence of salts such as gypsum and carbonates, alkaline pH values (7,5 - 9,5) and a general paucity of organic matter."

In such environments montmorillonite, illite, chlorite and mixed layers of these minerals are regarded to form as the typical secondary products.

It is clear that the semi-arid climate of the Great Fish

River Basin has produced similar conditions as above. The soil in the area is predominantly montmorillonitic and in places large deposits of calcrete and gypsum are encountered. The groundwater is highly saline, with pH values between 7 and 8,3.

Topography. Topography mainly controls, (a) the rate of surface runoff of rainwater and hence the rate of moisture intake by the parent rock, (b) the rate of subsurface drainage and therefore the rate of leaching of the soluble constituents, and (c) the rate of erosion of the weathered products and thereby the rate of exposure of fresh mineral surfaces. (Loughnan, 1969, p. 71).

The various geomorphological features of the area are discussed in Chapter 2. Because of the flat-lying nature of the Headbasin only 2 percent of the annual precipitation in this area is lost by surface runoff. This means that 98 percent of the rainfall is available to infiltrate the soil and the rock strata. Much of this remaining water is, however, returned to the atmosphere through evapotranspiration. Movement of the meteoric water that eventually does infiltrate, tends to be sluggish because of the low topography and therefore the soluble products released by the hydrolyzing reactions persist in the environmental waters.

The runoff from the Great Escarpment on the other hand is about 4 percent of the annual rainfall, i.e. double that from the Headbasin. Leaching of the soluble products is higher here, whilst accumulation of dissolved substances can be expected in the lower regions at the foot of the escarpment.

Parent material. Although the influence of the parent material on weathering is a geological factor, it can hardly be separated from the environmental factors. Apart from the chemical reactivity of the parent material, such features as texture, porosity, permeability and the nature of the rock itself can exert a considerable influence on the rate of leaching.

Porous rocks for example are chemically more readily attacked than dense ones because water is able to penetrate more easily. Fine-grained minerals are more susceptible to chemical attack than the coarse-grained varieties. The reason for this is that, in the case of the coarser grains, a protective residual coating develops around the grain, thus preventing any further chemical reaction between the unaltered surface of the grain and the water. In the case of the fine-grained rocks the grains are completely altered before the protective layer is developed.

The mudrock and sandstone in the area have such a dense and compact nature that the penetration of water into the rocks is greatly inhibited. Very little of the annual rainwater thus penetrates the rock strata and as only 2 percent is lost to the environment by surface run-off, one may deduce that most of the annual precipitation in the area returns to the atmosphere by evapotranspiration.

It is, however, true that the soils in the area can absorb more water than the rocks. Where a thick cover of soil is encountered, the amount of groundwater increases considerably. The water is, however, often very saline with Na^+ and Cl^- the dominant soluble ions.

A factor which greatly inhibits the penetration of the rainwater into the soil, is the high montmorillonite content. As soon as this clay mineral absorbs sufficient water it expands to such an extent that the entire soil unit becomes dense, thus preventing any further penetration of water.

The percolation of groundwater in the area is largely limited to joint and fracture zones in the sedimentary rocks. Many of these fractures are the result of dolerite intrusions. These fracture zones, however, appear to be incapable of sufficiently leaching the entire environment.

The dolerite in the area is also more prone to chemical weathering than the sedimentary rocks. The reason is that most of the constituents in the sedimentary rocks have already withstood a least one cycle of chemical weathering. It is therefore unlikely that these rock-types will be in any state of great disequilibrium with the weathering environment. Minor chemical adjustments are, however, expected.

(iii) Palaeoweathering In The Provenance-area Of The Karoo Sequence And Weathering During Transport Of This Material.

The presence of soluble and insoluble components associated with the Karoo Sequence started with the evolution of the depositional basin. Water reacted with the material present in the basin as well as with the material surrounding the basin. Certain components were thus dissolved, whilst residual secondary products were formed. These reactions continued throughout the geological history of the Karoo Basin, i.e. the deposition of the clastic material, the diagenesis of the sedimentary deposits, the intrusion of the dolerite magma and finally the influence of the present environmental conditions in different regions within the basin.

The chemical quality of the initial (palaeo) water in the Karoo Basin depended on the degree of chemical weathering in the provenance. The degree of chemical weathering, as pointed out previously, is greatly dependent on the ambient environmental factors such as the type and nature of the source rocks, the climatic conditions and the topography.

The source rocks of the sequence in the Great Fish River Basin consisted mainly of intermediate volcanic rocks (andesite), felsic volcanic rocks (dacite, rhyolite, quartz-feldspar porphyry), low-grade metamorphic rocks of the greenschist facies, granitic rocks and pre-existing sedimentary rocks. There is sufficient evidence that during the deposition of the lower part of the sequence, i.e. the Ecca Group and the Middleton Formation, a fair amount of volcanic activity commenced in the provenance.

Loughnan (1969, p. 75 - 114) discusses the chemical weathering of various rock-types and points out that most of the primary minerals which are encountered in the provenance of the Karoo Sequence decompose during weathering to form secondary clay minerals, whilst the alkali and alkali-earth cations such as Na^+, K^+ and Mg^{++}, Ca^{++} respectively go into solution. In almost all cases the rate of leaching determines whether a residue of illite and montmorillonite or a residue of kaolinite and halloysite will be the likely end-product (Loughnan, 1969, p. 95).

The minerals in the intermediate magmatic rocks weather more readily than those from the felsic types. Plagioclase from andesite (andesine), which is more calcic than albite from the felsic rocks, tends to weather more readily. The pyroxene and amphibole from the andesite will likewise decompose more readily than the K-feldspar in the felsic rocks. Quartz which is common in the felsic but rare in the andesitic rocks is very stable and tends to form part of the unweathered detrital residue.

During the initial period of evolution relatively cold conditions must have prevailed over the entire provenance of the Karoo Sequence as indicated by the presence of the Dwyka Tillite Formation at the bottom of the sequence. Under such conditions mechanical weathering will prevail and is confirmed by the large amount of detrital material in the tillite.

Following the ice-period the provenance must have been subjected to a moderate to high rainfall. The vast amount of water, however, was not given enough time to chemically react with the surrounding rock-types, but rather stripped the partly weathered material and transported it to the basin. Most of the sedimentary rocks of the Karoo Sequence consist of clastic deposits which show a poor degree of sorting, thus revealing rapid transport and deposition.

Apart from the clastic nature and the poor sorting of the sediments, the sequence becomes thinner towards the north, therefore, suggesting a wedge-like deposit. Such a wedge indicates a provenance of fairly high relief not very far to the south of the basin. The magnitude of the provenance is expressed by the enormous amount of detrital material which was deposited in the southern part of the basin. The high relief again promoted the stripping of the weathered material at a far greater rate than that at which chemical weathering could occur.

One can therefore conclude that, in spite of the fact that the provenance contained rock-types which possess the potential to be decomposed by chemical weathering, mechanical weathering prevailed because of the existing environmental conditions. Chemical decompositon of this material, however, became more pronounced in the lower lying areas at the foot of the provenance, as well as within the basin itself. In these parts leaching was rather limited and secondary minerals such as montmorillonite and illite therefore developed instead of kaolinite and halloysite. Because of the closed nature of the basin none of the constituents which remained in solution, could be removed entirely from the environment and therefore ions such as Na^+ and Cl^- started to accumulate in the water.

One must, however, bear in mind that the conditions which prevailed over the southern provenance did not necessarily have to prevail over the northern and north-western provenance of the basin. Here conditions could have fovoured chemical weathering rather than mechanical weathering, thus encouraging the concentration of soluble constituents in the basin by means of leaching. This will account for the presence of kaolinite in the northern part of the Karoo Basin.

(iv) Post-depositional Weathering.

During the weathering cycle one of the end results is the development of a soil profile. Such soil profiles may be residual as a result of the in situ weathering of the underlying rock strata or the profile may develop as a result of the transport of the weathered material from the higher regions by means of (a) sheet-wash during periodic showers to form colluvium or by (b) river channels to be deposited as alluvium in the river bed itself or on the banks thereof (Levinson, 1974, p. 90).
Whether the soil is of residual or of transported origin the vertical profile starts developing the moment sufficient material is accumulated. An A horizon, characterized by the depletion of soluble colloidal and mineral matter, develops

as a result of eluviation (leaching), whilst a B horizon, in which the above components are accumulated, develops lower down as a result of illuviation. Levinson (1974, p. 94) states that a well-drained soil in a humid area tends to present an "open" chemical system which results in the development of acid soils, whereas alkaline soils tend to develop in dry or semi-arid environments, where water is insufficient to drain completely through to the water table. The poor leaching in such semi-arid regions often causes the precipitation of a calcium carbonate layer, called caliche in the soil.

The origin of caliche is described by Jenny (1950) as follows: "If a uniform parent material containing some calcium carbonate is assumed, the formation of the lime horizon may be vizualized as the consequence of calcium carbonate-bicarbonate equilibria which are regulated by the carbon dioxide pressure of soil air." At the surface sufficient CO_2 is present in the air and is produced by plants to reduce the pH of the natural waters to a value of 5,7 (Krauskopf, 1967, p. 40). Under these conditions $CaCO_3$ can be dissolved and transported to the B horizon where the CO_2 pressure of the soil is reduced and the calcium carbonate is precipitated.

In the light of the above discussion the nature of the soils in the Great Fish River Basin can now be briefly examined. It is, however, obvious that such a complex subject as soils, which integrates aspects of geology, chemistry, microbiology and other sciences, is far too detailed to discuss here in more than a cursory manner.

Soils of the area. Vast expanses of the Headbasin and the Marginal Region are covered by soil, which in places reaches a thickness of up to 50 m. Most of these soils are derived from the weathering and erosion of the surrounding escarpment as well as from the various inselberge which occur within the above areas. Geomorphological features such as pediments on the slopes of the mountainous areas and bahadas in the lower lying areas are common as a result of the transport and deposition of the weathered material. Because most of the material appears to have been transported only over short distances, mainly by gravitational movement and by sheet-wash during episodic floods, the term "colluvium" is attributed to these soils.

The rivers and their tributaries have, however, scoured deep channels into the colluvium, thus re-working the material and depositing it as alluvium in the river bed during the normal flow and on the river banks during exceptional floods.

Residual soils develop only in the higher parts of the mountainous areas and on the Interior Plateau.

Owing to the semi-arid climatic conditions of the area, leaching of the upper horizon of the soil is poor and therefore the development of the A horizon is greatly inhibited. Eluviation is further restricted by the sparse vegetation in the basin. Where irrigation has, however, been applied over a relatively long period, leaching in the A horizon has occured only to develop a calcium-rich layer one or two metres lower down, which is almost impervious to circulating groundwater.

Conditions are generally suitable for the development of a thick B horizon, which is normally enriched in the "free" - salts of Ca^{++}, Mg^{++}, and Na^{+}.

A very striking example of illuviation, which has occured laterally rather than in a vertical profile, is the calcrete and gypsum deposits south of Cradock, which is described by Visser et al. (1963, p. 55 - 57).

The soils are generally chestnut-brown in colour and contain a high percentage of clay minerals. In the areas where "red" mudstone of the Katberg and Burgersdorp Formations is the predominant parent rock-type, the soils adopt a maroon-red colour.

Montmorillonite, illite and other mixed-layer clay minerals make up the bulk composition of the soils. The ratios of these clay minerals vary greatly in the soils throughout the area. This fact is confirmed by P.A.L. Le Roux of the Department of Soil Science, University of the OFS (personal communication) who found that laterally and vertically the soils often vary considerably in their swelling characteristics. In some horizons the swelling characteristics are extremely high, whilst an overlying or underlying horizon may contain clay minerals with virtually no swelling characteristics at all.

Quartz invariably froms an integral part of the mineral composition of the soils, but never exceeds the clay mineral content.

The caliche layer consists of pea-sized calcium carbonate nodules which are imbedded in a reddish-brown, clay-rich soil with extreme swelling characteristics. The nodules may, however, range up to a metre or more in length or the lime may occur as disseminated flecks and particles in the soil.

As far as the evolution of the clay minerals in soil is concerned, Millot (1970, p. 354) states that clay minerals which are encountered within the residual products of weathering are born, evolve, and die within the following three principal stages of the geochemical cycle: the zone of weathering, the zone of sedimentation and the zone of diagenesis. Within the above three zones the following types of clay minerals can be found: (a) inherited clay minerals, which are detritally transported from a preceeding environment, (b) clay minerals transformed by the degradation of the inherited particles, (c) clay minerals transformed by the aggradation of degraded particles, and (d) neoformed clay minerals, which are built entirely from solution. The nature of the environment thus determines which clay mineral will evolve and to what extent.

The variation in the types of clay minerals encountered in the study area can be explained by the fact that in certain areas conditions are suitable for the degrading of inherited minerals, whereas in other areas (where leaching is limited) inherited minerals may occur with neoformed minerals as well as with degraded minerals which are transformed by aggradation. The origin of the clay minerals in the soils, however, is determined by the weathering of the dolerite and the sedimentary rocks in the area.

Weathering of Sedimentary rocks. When considering the weathering of the sedimentary rocks, one must always bear in mind that the minerals which constitute such rocks have already endured at least one cycle of some form of weathering, sedimentation and diagenesis. The difference between the existing atmospheric conditions and the last of one of the above processes to which the rock was exposed, will determine the degree of disequilibrium which exists between such a rock and the ambient environment.

There appears to be very little difference between the present alkaline conditions (caused by the semi-arid climate and poor drainage in the basin) and the palaeo-depositional conditions of the Karoo strata in the area. It has already been suggested that the Karoo strata were deposited in a confined basin under brakish conditions and under climatic conditions which become progressively arid with time. The Beaufort Group was furthermore deposited under fluviatile conditions, which bring the palaeo-environmental conditions of deposition even closer to the present weathering environment.

If the depositional environment was alkaline, the diagenetic environment was even more so, the only difference here being the exclusion of oxygen and carbon dioxide.

One can therefore attribute the weathering of the sedimentary rocks in the area to a disequilibrium caused by the introduction of "fresh" rainwater, oxygen and CO_2, as well as to a limited leaching from the higher lying areas.

Mudrock constitutes the largest part of the sequence and consists mainly of illite and chlorite of diagenetic origin and lesser amounts of quartz, feldspar and other accessory minerals (Section 3. 1. 4. 4 and Fig. 3 - 11). The quartz and feldspar, as well as the accessory minerals, have obviously withstood the previous weathering cycles although the feldspar is sometimes altered to illite.

Aluminous illites according to Loughnan (1969, p. 105) weather in much the same manner as muscovite. Their finer grain size and lower degree of crystallinity, however, render them more vulnerable to leaching solutions. Where leaching is vigorous, potassium is removed from the mineral and the highly charged residual structure is free to expand in a similar manner as montmorillonite.

The only parts of the basin where leaching is vigorous enough to remove K^+ from the illite structure are the escarpment areas, the slopes of the <u>inselberge</u> within the low lying areas and the road cuttings. As soon as sufficient K^+ is leached, the rock tends to crumble into sub-cubic fragments of about 1,5 cms. The loose material is then easily removed by sheet-erosion and accumulates at the foot of the hills to form either pediments or bahadas, where further weathering can proceed.

Loughnan (1969, p. 106) is of the opinion that degraded illite similar to the above, is particularly vulnerable to desilification and may even be converted to kaolinite. Leaching in the basin is, however, so limited that kaolinite is unable to develop. The degraded mineral therefore tends to stabilize by absorbing certain soluble cations which results in the forming of a 14Å mineral that resembles vermiculite in that it shows little expansion on treatment with polar liquids, yet collapses to a 10Å mineral after heating to 450°C (Loughnan, 1969, p. 106). Where the sorbed cation is predominantly Mg^{++}, montmorillonite will evolve.

Where the connecting links between the layers of chlorite are attacked by the weathering agents a chlorite-vermiculite mixed layer mineral and with increased weathering, vermiculite is formed (Millot, 1970, p. 108). Loughnan, (1969, p. 106) states that a similar process takes place during the weathering of illite. Millot (1970) has shown that, under extreme conditions, vermiculite, can be transformed to vermiculite-montmorillonite and even further to montmorillonite. The following sequence of evolution of illite and chlorite is suggested by Millot (1970, p. 108).

illite → illite-vermiculite

vermiculite → vermiculite-montmorillonite → montmorillonite

chlorite → chlorite-vermiculite

It is, therefore, obvious that a mudrock containing substantial amounts of illite and chlorite can weather to form a residue consisting of any combination of the above products. The abundance of any one product will depend greatly on the weathering conditions in that particular area.

Millot (1970, p. 359) is of the opinion that the geochemical evolution of silicates in the hydrosphere is governed in its direction and its intensity by the activity of the environment of lessivage (environment in which water escapes by means of percolation, thus carrying material away in solution) or by the activity of the confined environment (an environment in which water is lost by evaporation rather than by percolation).

Both environments can be recognised in the Great Fish River Basin. An environment of lessivage exists in the area of high relief surrounding the Headbasin, in the Great Escarpment and on the slopes of the inselberge within the basin. Apart from the loss of interlayer cations, mostly K^+, but also Ca^{++}, Mg^{++} and H^+ from the illite and chlorite structures in the lessivage environment, ions of the octahedral sheets tend to migrate between the layers in an attempt to equilibrate the lattice, but these ions (mostly Al^{3+} but also Mg^{++}, Fe^{++} and Fe^{3+}) are also removed by lessivage. Degradation of the illite and chlorite proceeds further when the aluminum which replaces Si^{4+} in the tetrahedral sheets tends to move into the octahedral positions. This generally gives rise to the vermiculite and montmorillonite structures as described previously.

The lower lying confined environment greatly impede the percolation of groundwater, which results in an accumulation of soluble salts leached from the environment of lessivage. In this semi-confined environment the degraded residual product (which is transported from the higher regions) comes once again into contact with the soluble product (which was previously separated by leaching).

According to Millot (1970, p. 361) the inherited silicates, whether layered or not, that have resisted lessivage in the desaturated environment have no reason to evolve in an environment richer in cations. These minerals therefore remain inert and are buried in a state of continuing stability. Minerals such as quartz, feldspar and most of the accessory heavy minerals will behave in this manner.

The degraded minerals of the desaturated environment will, however, undergo transformation by aggradation to some extent in the confined environment. Loughnan (1969, p. 20) for example notes that degraded illite (which is merely stripped or partially stripped of K^+) expands in a manner similar to montmorillonite but, by the addition of sufficient K^+, the structure collapses irreversibly to 10Å and illite is regenerated. The same author (p. 19), however, states that the K^+ may partially be replaced by Ca^{++}, Mg^{++} and H^+, whilst Millot (1970, p. 8) is of the opinion that the H^+ ion combines with the water molecules to form H_3O^+ (oxonium) ions which are also capable of partially replacing the K^+ ions. Sodium may also replace the K^+, but to a far lesser extent than the other cations mentioned.

Should the illite be degraded to montmorillonite and the montmorillonite is saturated with K^+, only a partial collapse takes place while much of the expandable material remains (Loughnan, 1969, p. 20). The number of interlayer cations which can be absorbed by the montmorillonite structure is greatly dependent on the charge deficit which is caused by the substitution by Al^{3+} for tetrahedral Si^{4+} and by Mg^{++} for octahedral Al^{3+} (Deer et al., 1963, p. 227). These cations need not necessarily be K^+, but are frequently Na^+ or Ca^{++} (Millot, 1970, p. 11).

The degree of degradation of the clay minerals within the environment of lessivage, therefore, appears to determine the type of mineral that will evolve in the lower lying confined environment, i.e. aggradation of partly degraded illite and partial adsorption of various cations by montmorillonite.

Aluminum and Si^{4+} are capable of being dissolved by natural waters in the environment of lessivage and transported to the lower lying areas. Such solutions are possible only under alkaline conditions. The environment permitting these soluble products can regroup in the lower areas to precipitate from the solution as neoformed minerals. Millot (1970, p. 362) states that neoformed montmorillonites are often formed in calcic and hydromorphic soils. Such neoformations occur as growths by addition of ions to exsisting nuclii.

The silicate minerals (other than the clay minerals), which occur in the sedimentary rocks of this area, although fairly resistant to weathering because of their survival of the previous cycle of weathering, tend to decompose at a faster rate than the clay minerals. Feldspar (which is an important component of both the mudrock and the sandstone) for example is partly hydrolysed in the environment of lessivage, with the result that the structure on the surface of every mineral is partly degraded. Even these minerals on weathering produce degraded residues, soluble products and particles which are resistant to weathering.

Quartz, although relatively resistant to weathering, becomes more soluble with an increase in alkalinity (Fig. 4 - 1) and can therefore become an important source of silica for the neoformation of clay minerals in the semi-confined environment.

One would expect the sandstone to weather more readily than the mudrock on account of its higher porosity, permeability

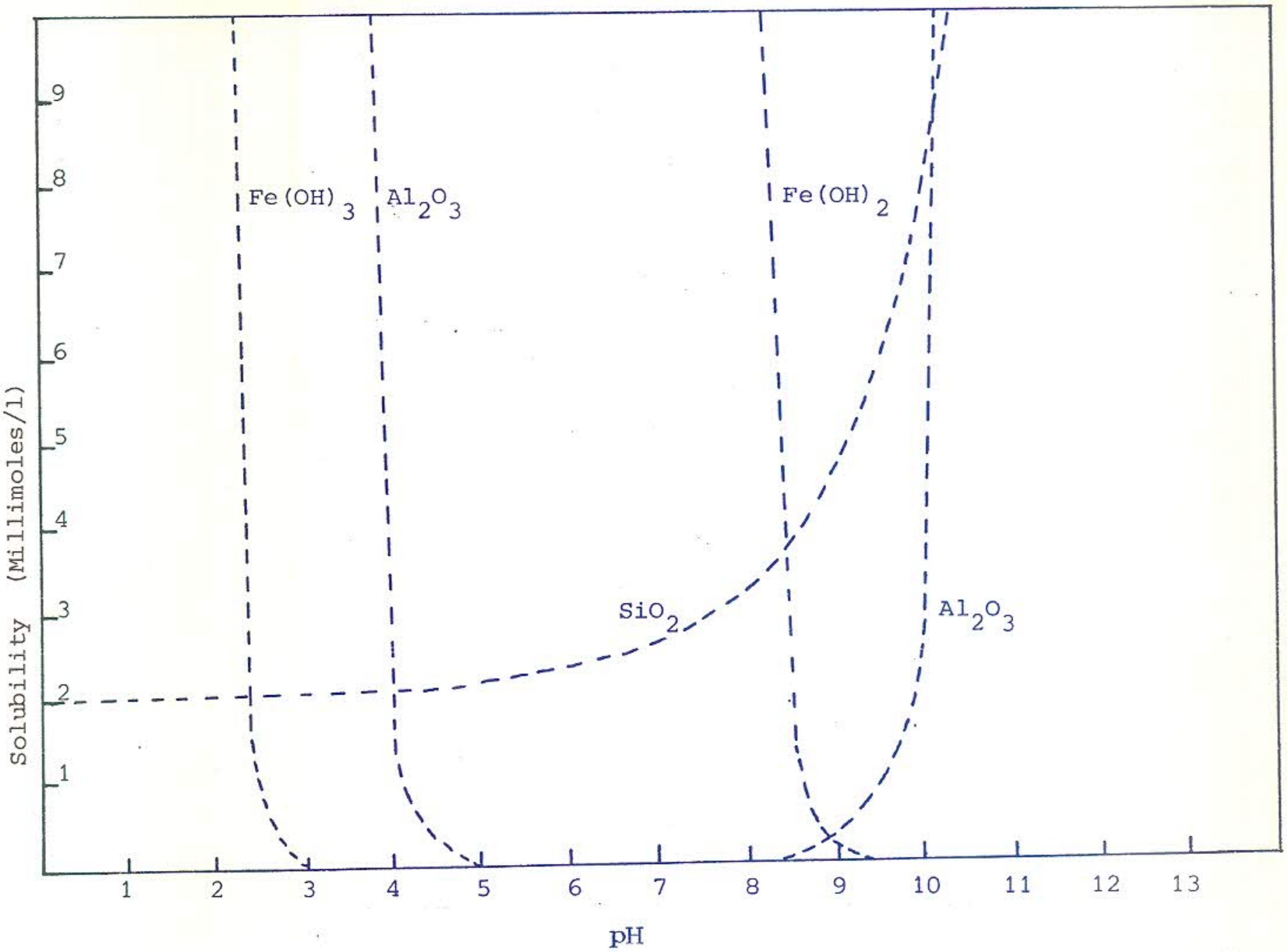


Fig. 4 - 1 Solubility in relation to pH for $Fe(OH)_3$, $Fe(OH)_2$, Al_2O_3 and SiO_2 released by chemical weathering. (After Loughnan, 1969, p. 32, Fig. 15).

and feldspar content. The semi-arid conditions, however, cause the moisture in the sandstone to evaporate rapidly near the surface, resulting in the precipitation of salts and dissolved silica. Most of the sandstones in the area of lessivage are coated with such a siliceous shell which prevents any further penetration of water and therefore retards the weathering of the sandstone.

The mudrock on the other hand weathers more readily because of the swelling character which the illite adopts when it loses its interlayer potassium.

The calcareous bodies which are often interbedded in the mudrock and sandstone remain relatively stable in an environment where lime tends to be deposited as calcrete.

It is, therefore, clear that the minerals in the sedimentary rocks in the area possess most of the ions which are normally encountered in the natural waters. On weathering, these ions are readily released to solution, but some of them are retained in the residual weathering product by adsorption, whilst others are even accomodated within the structures of aggraded and neoformed clay minerals.

Weathering of dolerite. Because of its igneous origin, the dolerite is more vulnerable to chemical weathering than the sedimentary rocks. The mafic minerals (olivine and pyroxene) and the calcic plagioclase (andesine-labradorite) are readily hydrolized. During the hydrolizing reactions cations such as Na^{+}, Ca^{++}, Mg^{++} and Fe^{++} are released to solution, whilst the residual structure is partly dissolved and partly reformed to some clay mineral.

Several samples of dolerite, ranging from unweathered to extremely weathered, were analysed and the average results of these analyses, after recalculation to 100 percent, are presented in Table 4 - 1.

TABLE 4 - 1 Calculation Of Gains And Losses During The Weathering Of Dolerite In The Great Fish River Basin.

	AVERAGE FRESH DOLERITE	AVERAGE WEATHERED DOLERITE	IN CASE OF SiO_2 STABLE	IN CASE OF Fe_2O_3 STABLE	LOSS IF SiO_2 STABLE	LOSS IF Fe_2O_3 STABLE	LOSS (IN%) IF SiO_2 STABLE	LOSS (IN%) IF Fe_2O_3 STABLE
WEATHERING INDEX	80	65						
SiO_2	51,69	55,76	51,69	48,29	0	-3,40	0	-6,58
Al_2O_3	14,83	12,70	11,77	11,00	-3,06	-3,83	-20,63	-25,83
Fe_2O_3	10,99	12,69	11,76	10,99	+0,77	0	+7,01	0
Mg O	7,55	4,40	4,08	3,81	-3,47	-3,74	-45,96	-49,54
Ca O	10,41	7,75	7,18	6,71	-3,23	-3,70	-31,03	-35,54
K_2 O	0,56	0,81	0,75	0,70	+0,19	+0,14	+33,93	+25,00
Na_2O	2,22	2,54	2,35	2,20	+0,13	-0,02	+5,86	+0,90
TiO_2	0,90	1,35	1,25	1,17	+0,35	+0,27	+38,89	+30,00
Mn O	0,18	0,18	0,17	0,16	-0,01	-0,02	-5,56	-11,11
P_2O_5	0,16	0,22	0,20	0,19	+0,04	+0,03	+25,00	+18,75
H_2O	0,51	1,60	1,48	1,39	+0,97	+0,88	+190,20	+172,55
Total	100	100	92,68	86,61	-7,32	-13,39		

Analyst: NPRL of the CSIR;

The weathering index, as indicated in Table 4 - 1, is calculated according to the equation:

$$WI = \frac{(Na)a}{0,35} + \frac{(Mg)a}{0,9} + \frac{(K)a}{0,25} + \frac{(Ca)a}{0,7} \times 100$$

WI = weathering index

(X)a = the atomic proportion of element X defined as atomic percentage divided by atomic weight.

The denominators in the function represent the bond strength between the element and oxygen in each case.

This equation is regarded by Parker (1970, p. 502) as a reasonably reliable method of comparing the degree of weathering of silicate rocks. The reliability of the method lies in the fact that the individual mobilities of only the most mobile major elements, based on their bond strength, are taken into account. Silica is disregarded on account of its irregular movement within a profile.

Note must be taken of the difference of 15 between the weathering index of the fresh dolerite and that of the weathered dolerite.

Calculation of the rest of the data in Table 4 - 1 is according to the description by Krauskopf (1967, p. 103). A few points in the calculations, however, need explaining:

(i) There is a decrease in Al_2O_3, which means that it cannot be regarded to have remained immobile.

(ii) The apparent increase in SiO_2 turns into a decrease when compared with the increase of Fe_2O_3. This is in accordance with the decrease in Al_2O_3, as both substances become more soluble in an alkaline environment.

(iii) Under alkaline conditions ferrous iron oxidizes readily to ferric iron and becomes insoluble, thus losing its mobility.

(iv) Some of the other elements, which appear to be immobile, have such low concentrations that no significant deductions can be made concerning them.

The solubility of the elements involved above, compared with an increase in pH is presented in Fig. 4 - 1.

It will be noticed that although Al_2O_3 is insoluble within the normal pH ranges of natural waters, it becomes more soluble when the pH exceeds 9. SiO_2 also becomes more soluble with an increase in pH and therefore it appears to be impossible for SiO_2 to remain immobile in an alkaline environment while Al_2O_3 is removed.

In the same alkaline environment $Fe(OH)_3$ is completely insoluble whilst $Fe(OH)_2$ oxidizes to form $Fe(OH)_3$. One can therefore readily assume that in an alkaline environment where Al_2O_3 and SiO_2 are easily removed by solution, Fe_2O_3

remains immobile.

When Fe_2O_3 is immobile, equal percentages of SiO_2, Al_2O_3, Mg O and Ca O are lost (Table 4 - 1). The total percentage loss is 14,71 percent, while the gains reduce the loss to 13,39 percent. The loss of 14, 71 percent is comparable with the discrepancy of 15 between the weathering index of the fresh rock and that of the weathered rock.

From the above discussion one may therefore conclude that apart from the removal of Mg^{++}, Ca^{++} and Na^{+} from the environment of lessivage during the weathering of dolerite, a substantial amount of SiO_2 and Al_2O_3 is also removed, should the environment be alkaline. In the confined area these elements are regrouped to promote the neoformation of montmörillonite.

The lack of sufficient rainwater in the area, however, retards the weathering of dolerite to a great extent. No deep weathering of this rock-type occurs and it is for this reason that dolerite weathers positively in relation to the sedimentary rocks. This is often illustrated by ridges of dolerite and the capping of most of the hills and mountains in the area by dolerite.

4. 3. DIAGENESIS IN THE KAROO STRATA

4. 3. 1 Introduction

Diagenesis commenced soon after deposition and is still active at present. This process is regarded by many to comprise all those transformations which start at the deposition of the sediment and continue throughout the history of the sedimentary rock until the metamorphic stage is reached (Read and Watson, 1962; Fairbridge, 1967; Müller, 1967; Von Engelhardt, 1967). Much controversy, however, appears to exist concerning the definition or delimitation of the term "diagenesis" and furthermore each researcher in this field has used his own terminology to characterize the phenomena encountered (Larsen and Chilingar, 1967, p. 523). Fairbridge (1967, p. 32) terms the various stages syndiagenesis, anadiagenesis and epidiagenesis whilst Müller (1967, p. 130) uses the terms "pre-burial, shallow-burial and deep-burial stages". The description of the terms given by Fairbridge (op. cit.) do not correspond with those given by Müller (op. cit.), although there appears to be some similarity between the zone of anadiagenesis and the deep-burial stage. Dapples (1967, p. 97 - 99) confuses the matter even more by using the terms "redoxomorphic, locomorphic and phyllomorphic" to describe the stages of diagenesis in a sandstone, the latter corresponding with the deep-burial stage.

Because all diagenetic reactions involve some interaction between the solid phase of the sediment and the pore fluid or formation waters (Pettijohn <u>et al</u>., 1973, p. 412), it is essential that some aspects of diagenesis of the Karoo strata are considered in this study. Not only do such reactions influence the physical properties of the sediments involved,

i.e. their porosity and permeability, but also the chemical composition of the interstitial water is affected.
At some stage or another, this water may mix with circulating meteoric water.

4. 3. 2 Physical Changes

Rowsell and De Swardt (1976, p. 118 - 121) consider the sediments in the southern part of the Karoo Basin to be in a state of very strong compaction, which indicates the former presence of a very thick overburden. The clay mineralogy of the sedimentary rocks and the irregular increase in density or porosity from the higher to the lower rock strata verifies this assumption. The mudstone in the Beaufort Group for example has an average porosity of 0,7 percent, whilst that of the Dwyka Tillite Formation is 0,4 percent (Rowsell and De Swardt, p. 118). These authors (p. 126) estimate that a minimum 3000 m of material was removed from the present surface in most of the Southern Karoo. In the case of the Headbasin it is probable that even more material was removed.

On account of the alteration of illite into chlorite and because of the high crystallinity of these clay minerals in the lower parts of the stratigraphic succession, Rowsell and De Swardt (1976, p. 116 and 126) conclude that at least the lower formations have already reached the stage of anchimetamorphism, where temperatures are estimated to have ranged between 270° C and 300° C. The sedimentary rocks of the Koonap Formation and the Beaufort Group appear to be less affected.

Where dolerite has intruded the sedimentary rocks, the generation of metamorphic chlorite in the sedimentary rock can clearly be observed in thin sections.

The initial porosity and therefore the water content of argillaceous muds before compaction is very much higher than that of sands (Müller, 1967, p. 135). During compaction of such sediments, caused by deposition of overlying material, the porosity is greatly reduced, whilst most of the interstitial water is displaced upward. Such displacement of formation water occurs at a relatively early stage in the sedimentary cycle. Müller (1967) points out that the rate of compaction of argillaceous sediments increases to a depth of 500 m and then decreases with a further depth of burial. It must, however, be borne in mind that not all the formation water is removed from the sediment by compaction and that at least some of it is trapped in the rock to form connate water.

In most cases the water which is displaced by compaction tends to flow upward into the overlying sediments. Von Engelhardt (1967, p. 514), however, proved that where the bedding is not perfectly horizontal, the water tends to collect in overlying sandstone bodies (which normally are not as prone to compaction as the argillaceous rocks) and can then flow via such a body towards the margins of the basin. Such a flow is illustrated by Von Engelhardt

(1967, p. 514) where in a sandstone body 10 m thick and with a dip of only 10', a single flow line would run within the sandstone for 3000 m before reaching the top surface and entering into the overlying strata.

Clay-rich sandstone, normally has a porosity of between 5 and 10 percent when exposed to moderate conditions of compaction. The average porosity and permeablility of similar rocks in the Southern Karoo is, however, 2 percent and 1 md respectively (Rowsell and De Swardt, 1976, p. 117). One may therefore deduce that even the sandstones in the Karoo Basin were greatly affected by compaction.

Müller (1967, p. 168) observed that once a sediment had reached a certain stage of compaction, the process becomes irreversible even when the load is removed by the erosion of the overburden. In this case one would expect very little difference between the porosity of the deeply buried rock and that of an exposed rock belonging to the same stratigraphic horizon. Du Toit (1915, p. 171), however, reported a decrease in the mean porosity of sandstones from the higher to the lower stratigraphic horizons of the Karoo Sequence. The values given by Du Toit (op. cit.) are higher than those of Rowsell and De Swardt, i.e. 2,9 percent for the Lower Beaufort Beds, 4,9 percent for the higher horizons of the Lower Beaufort Beds, 5,2 percent for the Middle Beaufort Beds and 5,5 percent for the Upper Beaufort Beds. Olivier (1972, p. 200) also records higher porosity values for the sedimentary rocks dissected by the Orange-Fish Tunnel, i.e. 2,2 to 8,2 percent for the various types of mudstone and 3,0 to 7,8 percent for the siltstone and sandstone.

If the state of compaction remains unaffected by a reduction in load, once a certain limit is reached, then another reason must be found for the higher porosity encountered in samples located close to the surface. Du Toit (1915, p. 174) noted that much of the cement material, as well as the weathering products of the feldspar in the sandstone, can easily be removed by percolating meteoric waters, thus creating a higher porosity. The decrease in porosity of the lower strata may thus be due to poor leaching conditions in the lower lying environment in which these strata are encountered.

Enslin (1956, p. xxi) regards the unweathered sedimentary rocks of the Karoo Sequence, because of their low porosity and permeability, as a poor potential for groundwater. The tectonic activity prior to and associated with the intrusion of dolerite into the Karoo Sequence, however, greatly affected the movement of the formation water which was partially entrapped during the diagenetic process. These fractures provided permeable passage-ways for the connate water to migrate upwards and where the fractures extend to the surface, mixing of connate and meteoric water can take place. Kent (1949, p. 252) notes that most of the thermal springs in the Karoo Sequence rise alongside dolerite dykes.

4. 3. 3 Chemical Changes

The clay minerals in the mudrocks of the study area consist of varying amounts of illite and chlorite, with minor amounts of mixed-layer clay minerals. No montmorillonite or kaolinite is present in any of these rocks (Fig. 3 - 11).

In the northern part of the Karoo Basin Rowsell and De Swardt (1976, p. 111) report substantial amounts of both montmorillonite and kaolinite in the argillaceous rocks and conclude that the presence of mixed-layer clay minerals in the southern part of the basin indicates a diagenetic origin and not a detrital origin for the illite and chlorite encountered here.

The transformation of montmorillonite into illite occurs during the early stages of diagenesis as the result of a simple fixation of K^+ from the interstitial water (Müller, 1967, p. 155). During the deep-burial stage montmorillonite is completely transformed into illite (Müller, 1967, p. 156).

Chlorite, which is transformed from montmorillonite, kaolinite and illite, poses a further problem in that the chemical structure is altered from a dioctahedral sheet to a trioctahedral sheet. Such a reaction appears to be endothermic and Müller (1967, p. 158) points out that chlorite forms in large amounts only under low-temperature metamorphism. The deduction by Rowsell and De Swardt (1976, p. 116) that most of the lower Karoo Sequence reached the anchimetamorphic stage, therefore appears to be correct.

During the above transformation some interaction between the pore water and the clay minerals occurs. Pettijohn et al. (1973, p. 431) state that: "Though there is substantial evidence that there is a relationship between depth of burial and illite-chlorite diagenesis it is not certain that it is a case of simple dependence on temperature and pressure, although that must be a large part of it. Depth of burial is also linked, in most sedimentary basins, with increase in concentration and change in composition of pore waters: pore water chemistry undoubtedly has an important interaction with clay diagenesis, though the water composition may be more the result of clay diagenesis than the cause."

Montmorillonite is transformed into illite by the fixation of K^+ from the interstitial solution, whilst SiO_2, H_2O, Na^+, Ca^{++}, Mg^{++} and Fe^{++} are released to the solution (Pettijohn et al., 1973, p. 431). Kaolinite is transformed into illite by the fixation of K^+ and SiO_2, while Al_2O_3 and H_2O are released. This latter transformation does not appear to have been prominent in the area.

According to Pettijohn et al.(1973, p. 431) montmorillonite may also be transformed into chlorite by the fixation of Fe^{++} and Mg^{++} from the solution, whilst the same solution is enriched in SiO_2, H_2O, Na^+ and Ca^{++} which are released by the montmorillonite. Rowsell and De Swardt (1976, p. 116) have, however, indicated that most of the chlorite in the Karoo Sequence was generated from illite at the boundary between anchi- and epimetamorphism. In this case the illite releases iron and magnesium which are then available for the neoformation of chlorite while the illite itself becomes more aluminous (Boltenhagen, 1969).

Because of the above transformation one would expect an increase in concentration in the interstitial waters of those components released to the solution. Müller (1967, p. 155), however, found that the ratio Na^+/Cl^-, K^+/Cl^-, Mg^{++}/Cl^-, HCO_3^-/Cl^- and $SO_4^=/Cl^-$ are lower in the pore waters than in normal sea water, whilst the ratio Ca^{++}/Cl^- is higher. The reason for this difference is explained by Fairbridge (1967, p. 65) as the effect of "natural chromatography" when the interstitial waters are squeezed out of a compacting sediment to flow through a clay-rich mudstone which acts as a porous membrane. In the case of a montmorillonite-rich mud, which has strong negative charges, the passage of the anions are mechanically restricted by repulsion. This mechanism of retention of certain ions by a clay membrane is supported by Degens and Chilinger (1967, p. 495) who conclude that the charge which is permanently attached to the clay membrane and the charge on the ions in the solution, rather than the size of the ions in the electrolyte, determine the amount of salts retained in the remaining pore waters. The water molecules for example are able to pass through the membrane and therefore the filtrate has a lower salt level than the original solution.

As far as the movement of the cations through such a membrane is concerned, Blatt et al. (1972, p. 341) point out that the controlling mechanism is not pore size, but the fact that these ions move through the membrane by "site hopping" on the clay surfaces. The ease of replaceability during such "site hopping" is Li>Na>K>Rb and therefore the concentration pattern in the remaining brine is the reverse of the above sequence.

The filtration of formation waters through a porous clay membrane can therefore, account for the concentration of salts in the pore waters of deeply buried zones of a sedimentary sequence. Fairbridge (1967, p. 65) notes that such deep basin brines may achieve salinities of up to six times that of ordinary sea water. In the case of the Karoo Sequence, with all its mudrock, such concentrations are quite possible especially in the Dwyka Formation and in the Ecca Group.

The enrichment of Ca^{++} in the remaining waters, as suggested previously, is explained by the fact that the divalent cations are bound to the clay surfaces more strongly than the univalent ones. According to Blatt et al. (1972, p. 342) the carbonate equilibria in the remaining pore waters is affected by such calcium enrichment, therefore increasing the ion activity product (Ca^{2+}) (CO_3^{2-}) which encourages the precipitation of calcite cement. Fig. 3 - 6 illustrates irregular calcareous concretions in the Barberskrans Sandstone Member, which could possibly have formed under such diagenetic conditions.

Fairbridge (1967, p. 66) notes, that apart from mechanical filtering, the rising solutions lead to various reactions, the most obvious being cementation and decementation. In most cases these cements are $CaCO_3$ or SiO_2. Pettijohn *et al.* (1973, p. 402) describes a mechanism of diffusion whereby quartz in a sandstone is replaced by calcite as a result of reactions between the pore water and the mineral grains.

Blatt *et al.* (1972, p. 342) report the following chemical variation in formation waters with depth in the Donetz Basin: "calcium-bicarbonate water near the surface, superseded successively at a depth of sodium-bicarbonate water, sodium sulfate water, sodium-bicarbonate water, sodium-bicarbonate-chloride water and possibly saline sodium-chloride water". A great variety of precipitates may therefore be formed by mixing of these waters as a consequence of faulting, jointing or dissolution of earlier formed cements. A similar trend is noticed in the groundwater of the Karoo Sequence in the Great Fish River Basin and will be delt with in the following chapter.

4. 4. INTERACTION OF GROUNDWATER WITH MUDROCK

4. 4. 1 Introduction

The fact that the Karoo Sequence in the Great Fish River Basin consists largely of mudrock, which in turn is composed mainly of clay minerals like illite and chlorite, influences the groundwater quality in the area. Apart from the influence of the mudrock on the connate and formation water during compaction and diagenesis, the clay minerals within these rocks possess the ability of sorbing cations and anions on their surfaces and within their structures, thus altering the chemical composition of the solutions with which they come into contact.

According to Grim (1962, p. 30) both illite and chlorite have a cation-exchange capacity of 10 - 40 me/100g, which is caused mainly by broken bonds around the edges of the silica-alumina tetrahedra and which therefore increase with a decrease in particle size. Another cause, which is commonly found in montmorillonite and less common in illite and chlorite, is the substitution within the clay mineral lattice of Al^{3+} for Si^{4+} in the tetrahedral sheet, thus resulting in unbalanced charges, which are satisfied by adsorbed cations (Grim, 1962, p. 31).

Krauskopf (1967, p. 162) states that no simple rules can be applied to the behavior of ions toward adsorbents because adsorption depends on various factors such as the nature of the adsorbent, the temperature and the kind and amounts of other ions present in the solution. The relative adsorbability of two ions to the same adsorbent, however, depends to some extent on such properties as ion sizes, their charges and their ability to form covalent bonds.

The difference in adsorbability of two cations is illustrated by Krauskopf (1967, p. 162) where Ca^{++} in dilute solution replaces Na^{+} in a Na-zeolite during the "softening" of water. In order to remove the newly adsorbed Ca^{++} from the zeolite, a much more concentrated solution of Na^{+} is required. This illustration proves that when natural waters containing certain cations come into contact with clay minerals with adsorbed cations, cation-exchange may occur, thus altering the chemical composition of that particular water. As will be illustrated later, all the mudrocks in the area contain vast amounts of adsorbed Ca^{++} and only minor amounts of Na^{+}, whilst the reverse is observed in the chemistry of the groundwater. Sayles and Mangelsdorf (1977, p. 959) point out that the Ca^{++}, which is adsorbed by continental clays may only be replaced by Na^{+} in a concentration equivalent to that of sea water.

Garrels and Christ (1965, p. 272) state that the cation-exchange reactions to some extent follow the Law of Mass Action and that where two univalent cations are involved in the reaction, the following equation may be satisfied:

$$\frac{(A^{+})}{(B^{+})} = K_{A\bar{B}} \quad \frac{(Ax)}{(Bx)}^{\quad n}$$

where

(A^{+}) and (B^{+}) = the activities of the cations in solution

(Ax) and (Bx) = the concentrations of the cations in the exchanger

n = an exponent

$K_{A\bar{B}}$ = the exchange constant

Where multivalent cations are involved in the exchange reactions, e.g.

$$Z_B A^{+Z_A} + Z_A B^{+Z_B} R_{Z_B} = Z_A B^{+Z_B} + Z_B A^{+Z_A} R_{Z_A}$$

Golubev and Garibyants (1971, p. 50) found that the use of the Law of Mass Action in ion-exchange, without consideration of the activity - coefficients, leads to the expression:

$$Kc = \frac{C_B^{Z_A} \; q^{Z_B}AR}{C_A^{Z_B} \; q^{Z_A}BR}$$

where

Z_A and Z_B = the valences of the ions A and B

C_A and C_B = the concentrations of the ions in solution

qAR and qBR = the concentrations of the ions in the resin

Kc = the equilibrium coefficient.

According to the above authors, however, the equilibrium coefficient can possibly be considered a constant value only during the exchange of univalent ions from dilute solutions. Where several cations of various valences are involved, as is often the case with natural waters, the equation becomes even more complex and almost impossible.

The determination of the adsorbed cations Na^+, K^+, Ca^{++} and Mg^{++}, as well as the cation-exchange capacity (CEC) of the mudrock in the area, was conducted as follows:

(i) The sample is pulverized to - 200 #.

(ii) Five grams of the sample is flushed several times with de-ionized water, which is kept for the determination of "free" cations.

(iii) After flushing the sample is dried and weighed.

(iv) The dry sample is then immersed in a 1N solution of $CH_3CO_2NH_4$ (ammonium acetate) at pH = 7 and shaken for two hours. During this process NH_4^+ from the ammonium acetate replaces the adsorbed cations in the sample.

(v) The residual solution is separated from the sample by centrifugen and kept for analysis of the exchangeable cations.

(vi) The residual sample with the adsorbed NH_4^+ is then immersed in a 1N solution of CH_3CO_2Na (sodium acetate) at pH = 7 in order to replace the NH_4^+ with Na^+.

(vii) The sample now should contain only adsorbed Na^+ in all the possible adsorbing positions.

(viii) By re-immersing the sample in a 1N solution of $CH_3CO_2NH_4$, this Na^+ is again replaced by NH_4^+ and the residual solution now containing the exchanged Na^+ is analysed for this ion to determine the CEC of the sample.

The results of the above analyses on mudrock samples from the various strata in the area are presented in Tables 4 - 2 to 4 - 6.

Table 4 - 2 Exchangeable Cations Of The Mudrock In The Dwyka Tillite Formation, Ecca Group And Koonap Formation.

SAMPLE	CO-ORDINATES		EXCANGEABLE CATIONS					
No	Lat. S	Long. E	Na^+ me/100g	K^+ me/100g	Ca^{++} me/100g	Mg^{++} me/100g	Total me/100g	CEC me/100g
			KOONAP FORMATION					
15 - 1	33° 00′	25° 30′	0,28	2,17	4,16	2,37	8,98	9,51
15 - 2	33° 00′	25° 37½′	0,28	1,78	2,97	1,85	6,88	6,60
15 - 3	33° 00′	25° 45′	0,35	2,26	10,69	1,73	15,03	7,57
15 - 4	33° 00′	25° 52½′	0,24	0,48	4,75	3,79	9,26	9,51
15 - 5	33° 00′	26° 00′	0,71	1,69	3,56	2,85	8,81	8,74
15 - 6	33° 00′	26° 07½′	0,33	1,09	26,73	2,54	30,69	8,16
			ECCA GROUP					
16 - 1	33° 07½′	25° 37½′	0,32	1,54	27,33	2,20	31,39	9,13
16 - 4	33° 07½′	26° 00′	0,45	0,35	4,75	2,46	8,01	8,54
16 - 5	33° 07½′	26° 07½′	0,36	0,91	33,86	1,42	36,55	6,41
16 - 6	33° 07½′	26° 15′	0,36	1,11	17,82	1,38	20,67	5,63
16 - 7	33° 07½′	26° 22½′	0,26	0,30	5,34	2,98	8,88	9,51
16 - 8	33° 07½′	26° 30′	0,30	0,43	5,94	2,16	8,83	9,90
16 - 9	33° 07½′	26° 37½′	0,27	1,19	16,63	1,08	19,17	7,19
16 - 10	33° 07½′	26° 45′	0,27	0,80	6,54	1,42	9,03	6,60
16 - 11	33° 07½′	26° 52½′	0,58	0,17	4,16	3,67	8,58	7,96
			DWYKA TILLITE FORMATION					
16 - 2	33° 07½′	25° 45′	0,47	1,09	8,91	1,90	12,37	4,66
16 - 3	33° 07½′	25° 52½′	0,64	0,30	2,38	2,37	5,69	5,82

TABLE 4 - 3 Exchangeable Cations Of The Mudrock In The Middleton Formation.

SAMPLE	CO-ORDINATES		EXCHANGEABLE CATIONS					
No	Lat. S	Long. E	Na^+ me/100g	K^+ me/100g	Ca^{++} me/100g	Mg^{++} me/100g	Total me/100g	CEC me/100g
12 - 4	32° 37½′	25° 45′	0,3	0,87	3,24	1,60	6,01	5,39
12 - 5	32° 37½′	25° 52½′	0,3	0,75	34,80	1,36	37,21	10,27
13 - 2	32° 45′	25° 30′	1,20	1,47	4,18	3,46	10,27	10,35
13 - 3	32° 45′	25° 37½′	0,51	1,21	5,74	3,03	10,49	12,40
13 - 4	32° 45′	25° 45′	0,8	1,94	14,10	0,72	17,56	7,26
13 - 5	32° 45′	25° 52½′	0,53	0,81	6,79	5,10	13,23	13,79
13 - 6	32° 45′	26° 00′	0,50	1,62	7,57	3,76	13,45	13,40
13 - 7	32° 45′	26° 07½′	0,67	2,00	2,87	3,95	9,49	9,70
14 - 1	32° 52½′	25° 22½′	0,74	1,64	18,27	2,60	23,25	10,35
14 - 2	32° 52½′	25° 30′	0,65	1,64	18,27	2,60	23,16	10,44
14 - 3	32° 52½′	25° 37½′	1,37	1,13	7,83	3,85	14,18	11,53
14 - 4	32° 52½′	25° 45′	0,65	1,68	2,09	1,97	6,39	7,35
14 - 5	32° 52½′	25° 52½′	1,11	1,21	3,65	2,96	8,93	8,61
14 - 7	32° 52½′	26° 07½′	0,53	1,60	23,75	1,22	27,10	7,00
14 - 8	32° 52½′	26° 15′	0,57	1,17	5,22	2,40	9,36	10,61
14 - 9	32° 52½′	26° 22½′	0,32	0,32	6,26	4,28	11,18	11,70
14 - 10	32° 52½′	26° 30′	1,02	1,32	6,00	2,47	10,81	8,87
14 - 11	32° 52½′	26° 37½′	0,43	1,15	5,94	3,41	10,93	8,74
15 - 7	33° 00′	26° 15′	0,25	1,46	4,16	2,54	8,41	9,51
15 - 8	33° 00′	26° 22½′	0,38	1,65	2,97	1,81	6,81	6,41
15 - 9	33° 00′	26° 30′	0,28	2,35	5,94	1,29	9,86	9,13
15 - 10	33° 00′	26° 37½′	0,74	0,67	11,29	2,46	15,16	8,16
15 - 11	33° 00′	26° 45′	0,62	0,37	5,34	2,29	8,62	9,51
15 - 12	33° 00′	26° 52½′	0,35	0,78	5,34	2,37	8,84	9,71

TABLE 4 - 4 Exchangeable Cations Of The Mudrock In The Balfour Formation.

SAMPLE	CO-ORDINATES		EXCHANGEABLE CATIONS					
No	Lat. S	Long. E	Na^{+} me/100g	K^{+} me/100g	Ca^{++} me/100g	Mg^{++} me/100g	Total me/100g	CEC me/100g
			ELANDSBERG MEMBER					
3 - 1	31° 30′	24° 45′	0,2	0,38	8,98	0,74	10,30	7,90
3 - 3	31° 30′	25° 00′	0,2	1,00	2,50	2,47	6,17	8,30
4 - 1	31° 37½′	24° 45′	0,2	0,32	5,45	2,91	8,88	11,09
4 - 2	31° 37½′	24° 52½′	0,2	0,76	8,45	6,47	15,88	18,05
4 - 3	31° 37½′	25° 00′	0,2	0,73	13,96	2,91	17,80	14,24
4 - 5	31° 37½′	25° 15′	0,1	0,76	13,36	1,62	15,84	15,12
5 - 4	31° 45′	25° 15′	0,1	0,32	24,55	2,22	27,19	10,55
5 - 5	31° 45′	25° 22½′	0,2	0,79	4.74	3,46	9,19	9,35
6 - 1	31° 52½′	25° 00′	0,2	0,14	5,45	3,14	8,93	10,00
6 - 5	31° 52½′	25° 30′	0,1	0,76	9,00	3,33	13,19	15,44
7 - 1	32° 00′	25° 00′	0,1	0,25	4,36	3,23	7,94	10,53
7 - 2	32° 00′	25° 07½′	0,1	0,32	10,09	3,42	13,93	14,68
7 - 3	32° 00′	25° 15′	0,1	0,73	48,00	2,13	50,96	12.51
7 - 4	32° 00′	25° 22½′	0,2	0.25	4.91	3,60	8,96	10,66
8 - 2	32° 07½′	25° 07½′	0,2	0,37	18,82	3,84	23,23	15,22
8 - 3	32° 07½′	25° 15′	0,2	0,07	7,63	3,00	10,90	9.46
9 - 1	32° 15′	25° 00′	0,1	0,30	7,63	1,94	9,97	10,00
9 - 2	32° 15′	25° 07½′	0,2	0.37	3,82	3,00	7,39	9,24
9 - 5	32° 15′	25° 30′	0,3	1,04	12,10	1,64	15,08	6,79
9 - 8	32° 15′	25° 52½′	0,4	0,58	4,86	3,32	9,16	10,18
9 - 9	32° 15′	26° 00′	0,6	0,31	6,74	4,69	12,34	15,88
10 - 7	32° 22½′	26° 00′	0,5	0,17	10,79	5,99	17,45	16,53
10 - 8	32° 22½′	26° 07½′	0,5	0,15	10,79	5,99	17,43	15,31
			BARBERSKRANS SANDSTONE MEMBER					
6 - 3	31° 52½′	25° 15′	0,1	0,41	5,73	2,96	9,20	10,87
6 - 4	31° 52½′	25° 22½′	0,1	0,78	13,09	-2,63	16,60	11,42
8 - 5	32° 07½′	25° 30′	0,2	0,46	7,37	3,14	11,17	12,29
8 - 6	32° 07½′	25° 37½′	0,2	1,28	3,00	2.73	7,21	9.79
11 - 7	32° 30′	26° 07½′	0,3	0,62	15.91	1,40	18,23	8,44
11 - 10	32° 30′	26° 30′	0,4	0,06	5.93	4,71	10,90	9,83
			DAGGABOERSNEK MEMBER					
9 - 6	32° 15′	25° 37½′	0,3	0,53	6,47	3,60	10,90	12,01
9 - 7	32° 15′	25° 45′	0,3	1,13	7.01	3,43	11,87	14,62
10 - 1	32° 22½′	25° 15′	0,2	0,17	8,90	5,50	14,77	13,57
10 - 3	32° 22½′	25° 30′	0,2	0,17	8,90	4,23	13,50	19,49
10 - 4	32° 22½′	25° 37½′	0,3	0,15	8,36	6,51	15,32	14,62
10 - 5	32° 22½′	25° 45′	0,4	1,09	44,24	1,75	47,48	12,35
10 - 6	32° 22½′	25° 52½′	0,2	0,17	27,50	1,64	29,52	15,05
11 - 1	32° 30′	25° 22½′	0,2	1,34	34,80	1,75	38,09	11,14
11 - 2	32° 30′	25° 30′	0,3	0,51	8,63	4,63	14,07	15,48
11 - 4	32° 30′	25° 45′	0,6	0,49	9,46	1,68	12,23	7,74
11 - 5	32° 30′	25° 52½′	0,2	2,15	5,39	2,07	9,81	8,61
11 - 6	32° 30′	26° 00′	0,2	0,15	6,47	2,76	9,58	10,09
11 - 8	32° 30′	26° 15′	0,7	0.17	11,06	6,19	18,12	19,49
11 - 9	32° 30′	26° 22½′	0,5	1,53	1,89	1,68	5,60	6,70
12 - 3	32° 37½′	25° 37½′	0,2	0,21	8,90	4,39	13,70	15,31
12 - 11	32° 37½′	26° 37½′	0,7	0,34	12,01	3.52	16,57	10,61
			OUDEBERG SANDSTONE MEMBER					
12 - 1	32° 37½′	25° 22½′	0,3	0,51	6,47	3,60	10,88	12,18
12 - 6	32° 37½′	26° 00′	1,0	0,49	20,88	2,90	25,27	11,53
12 - 7	32° 37½′	26° 07½′	0,8	0,15	8,87	6,45	16,27	14,88
12 - 8	32° 37½′	26° 15′	0,3	0,34	24,28	1,61	26,53	14,14
12 - 9	32° 37½′	26° 22½′	0,7	0,53	6,53	5,63	13,39	13,88
13 - 8	32° 45′	26° 15′	0,42	0,58	43,85	2,90	47,75	13,88
13 - 9	32° 45′	26° 22½′	0,76	1,43	24,54	2,73	29,46	9,79
13 - 10	32° 45′	26° 30′	0,58	0,68	55,33	2,04	58,63	10,70
13 - 11	32° 45′	26° 37½′	0,39	1,07	59,51	2,37	63,34	15,57

TABLE 4 - 5 Exchangeable Cations Of The Mudrock In The Katberg Formation.

SAMPLE	CO-ORDINATES		EXCHANGEABLE CATIONS					
No	Lat. S	Long. E	Na^+ me/100g	K^+ me/100g	Ca^{++} me/100g	Mg^{++} me/100g	Total me/100g	CEC me/100g
1 - 1	31° 15'	25° 00'	0,2	0,5	5,2	1,0	6,90	6,3
1 - 3	31° 15'	25° 15'	0,1	0,5	4,0	1,6	6,20	8,3
1 - 4	31° 15'	25° 22½'	0,1	0,4	8,0	2,1	10,60	10,9
2 - 1	31° 22½'	24° 52½'	0,3	0,2	4,2	2,6	7,30	10,4
2 - 2	31° 22½'	25° 00'	0,2	0,6	5,2	1,4	7,40	7,4
2 - 3	31° 22½'	25° 07½'	0,1	0,4	12,97	2,06	15,51	13,2
2 - 6	31° 22½'	25° 30'	0,2	0,74	16,97	0,74	18,65	7,9
3 - 6	31° 30'	25° 22½'	0,2	0,87	22,46	3,95	27,48	14,5
3 - 7	31° 30'	25° 30'	0,1	0,43	36,93	0,74	38,20	14,4
4 - 4	31° 37½'	25° 07½'	0,2	0,41	5,29	1,81	7,71	7,0
4 - 6	31° 37½'	25° 22½'	0,1	0,90	24,20	4,03	29,23	13,5
4 - 7	31° 37½'	25° 30'	0,2	0,41	3,99	1,65	6,25	8,16
4 - 8	31° 37½'	25° 37½'	0,1	0,50	12,00	2,20	14,80	9,90
4 - 9	31° 37½'	25° 45'	0,2	0,46	14,72	2,30	17,68	14,40
5 - 2	31° 45'	25° 00'	0,1	0,54	15,72	0,66	17,02	6,74
5 - 6	31° 45'	25° 30'	0,2	0,18	8,18	5,54	14,10	15,55
7 - 6	32° 00'	25° 37½'	0,2	0,02	2,99	2,06	5,27	6,96
7 - 7	32° 00'	25° 45'	0,2	0,54	20,71	0,90	22,35	11,09
7 - 8	32° 00'	25° 52½'	0,2	0,90	24,70	1,89	27,69	11,53
7 - 9	32° 00'	26° 00'	0,3	0,02	5,24	6,50	12,06	12,61
7 - 10	32° 00'	26° 07½'	0,1	0,80	2,80	1,50	5,20	7,50
8 - 7	32° 07½'	25° 45'	0,3	1,61	8,98	1,15	12,04	8,48
8 - 8	32° 07½'	25° 52½'	0,2	1,51	12,72	1,56	14,99	13,70
8 - 10	32° 07½'	26° 07½'	0,2	0,36	13,72	2,30	16,58	13,81
8 - 11	32° 07½'	26° 15'	0,2	0,90	3,24	0,74	5,08	7,50
8 - 12	32° 07½'	26° 22½'	0,2	1,15	3,24	1,23	5,82	8,05
9 - 10	32° 15'	26° 07½'	0,2	0,01	11,48	3,46	15,15	15,22
9 - 11	32° 15'	26° 15'	0,2	1,37	9,55	1,62	12,74	7,93
9 - 12	32° 15'	26° 22½'	0,1	0,18	7,37	4,16	11,81	13,16
10- 11	32° 22½'	26° 30'	0,2	0,41	17,73	1,89	20,23	12,07

TABLE 4 - 6 Exchangeable Cations Of The Mudrock In The Burgersdorp Formation.

SAMPLE	CO-ORDINATES		EXCHANGEABLE CATIONS					
No.	Lat. S	Long. E	Na^+ me/100g	K^+ me/100g	Ca^{++} me/100g	Mg^{++} me/100g	Total me/100g	CEC me/100g
1 - 5	31° 15′	25° 30′	0,3	0,3	3,0	2,1	5,70	9,0
1 - 6	31° 15′	25° 37½′	0,3	0,9	3,5	2,1	6,80	9,2
1 - 8	31° 15′	25° 52½′	0,1	0,5	23,0	1,1	24,70	11,1
1 - 9	31° 15′	26° 00′	0,3	0,2	1,7	4,2	6,40	9,2
2 - 7	31° 22½′	25° 37½′	0,5	0,2	24,0	1,3	26,00	7,8
2 - 8	31° 22½′	25° 45′	0,3	0,5	6,2	2,6	9,60	13,9
2 - 9	31° 22½′	25° 52½′	0,1	0,6	11,0	1,0	12,70	9,5
3 - 10	31° 30′	25° 52½′	0,2	0,8	5,7	4,1	10,80	12,8
4 - 10	31° 37½′	25° 52½′	0,2	0,4	5,2	6,1	11,90	14,4
4 - 11	31° 37½′	26° 00′	0,3	0,1	6,2	3,2	9,80	12,2
5 - 9	31° 45′	25° 52½′	0,1	0,6	3,5	3,9	8,10	9,9
5 - 10	31° 45′	26° 00′	0,4	0,5	9,0	12,3	22,23	16,0
5 - 12	31° 45′	26° 15′	0,50	0,20	2,20	9,90	12,80	16,6
6 - 10	31° 52½′	26° 07½′	0,1	0,5	4,0	2,2	6,80	8,9
6 - 11	31° 52½′	26° 15′	0,1	0,5	1,7	1,6	3,90	7,1
7 - 11	32° 00′	26° 15′	0,1	1,3	44,9	1,2	47,50	6,6

4. 4. 2 Cation-adsorption Of Karoo Mudrock

It is interesting to note that the mean total adsorbed cations of the mudrock from all the strata in the area exceeds the corresponding mean CEC (Fig. 4 - 2), but that the lowest value of total adsorbed cations falls either within, or is lower than the CEC - range for any particular stratigraphic unit. Especially in the Barberskrans Member and in the Katberg and Burgersdorp Formations total adsorbed cation-values lower than the lowest CEC-values were recorded. The highest CEC-values for the mudrock of the Middleton Formation, Daggaboersnek, Barberskrans and Elandsberg Members, as well as that for the Katberg and Burgersdorp Formations exceed the corresponding mean total adsorbed cations. One may therefore accept that the mean total adsorbed cations of the above strata fall within the normal CEC-range.

At a first glance, one would be inclined to attribute the discrepancy between the mean total adsorbed cations and the mean CEC to an excess of "free" Ca^{++} in the sample. Fig. 4 - 2 for example shows that in many cases, i.e. in the Dwyka Formation, Ecca Group, Koonap Formation, Oudeberg Member, Daggaboersnek Member and Katberg Formation, the mean exchanged Ca^{++} exceeds the mean CEC. The water, which was used to rinse the sample before commencing with the cation-exchange reactions, however, proved to contain very little and often indeterminable amounts of Ca^{++} or any other cation. The presence of "free", unadsorbed Ca^{++} was therefore ruled out. Calcium present in the form of calcite can also be ruled out as this mineral was detected only in a few samples

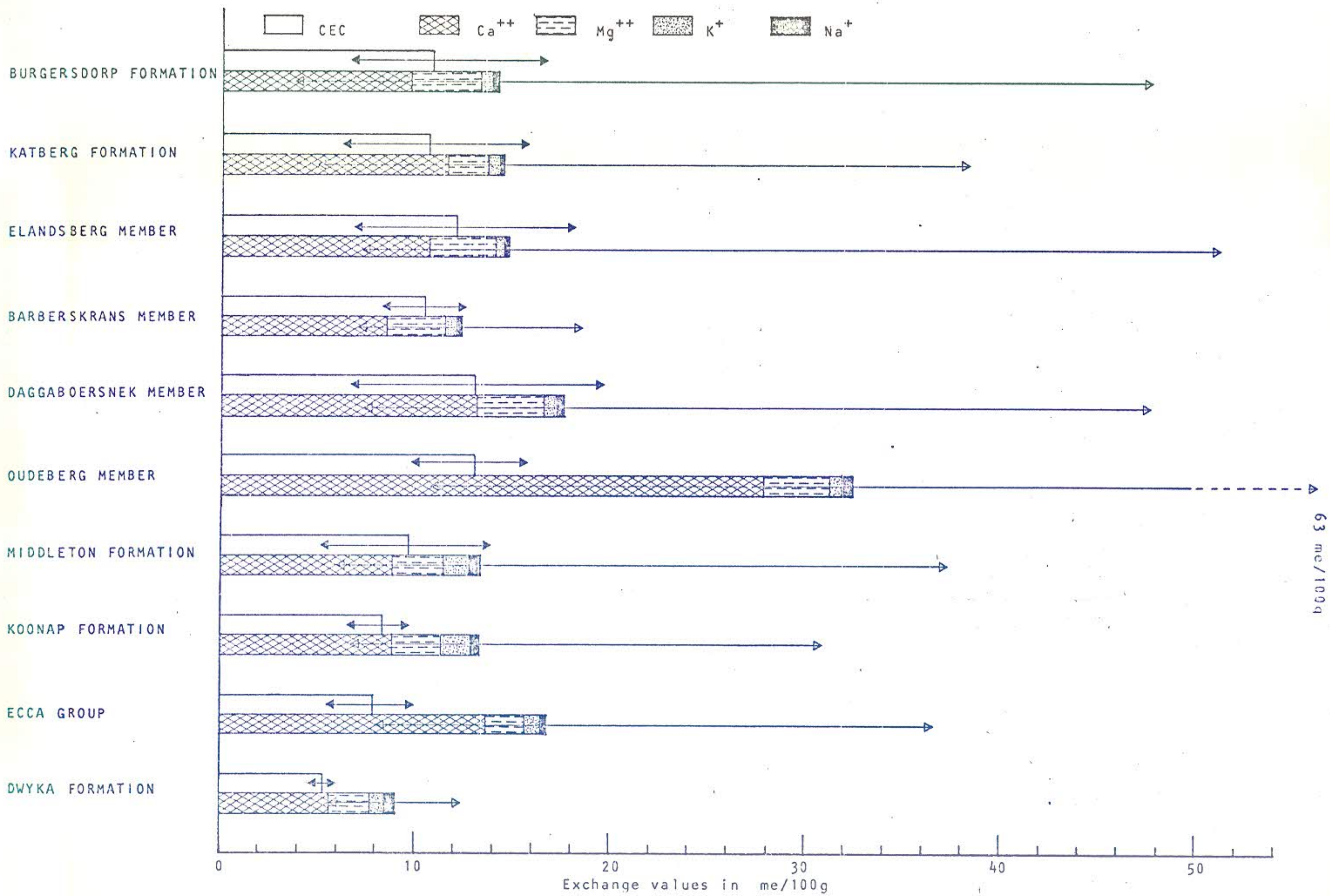


Fig. 4 - 2 Bar diagrams showing the mean cation-exchange properties of the mudrock in the study area.

by X-ray diffraction techniques. Samples with exceptionally high Ca^{++}- values were especially checked by this technique to make sure that no Ca^{++} was derived from calcite in the sample.

Similar discrepancies, which could not be explained, are recorded by Sayles and Mangelsdorf (1977, p. 955). In the case of the present investigation the discrepancy appears to lie in the method used to determine the CEC. The possibility exists that the 1N solution of CH_3CO_2Na was not adequate to replace all the NH_4^+ in the exchange positions, thus presenting a lower CEC value. Although these CEC-values can not be compared with the total adsorbed cations, they can nevertheless be used to compare one stratigraphic unit with another.

In Fig. 4 - 2 the mean CEC of the mudrock appears to increase gradually from the Dwyka Formation to the Oudeberg Member, where it reaches a maximum for the whole of the sequence. From the Oudeberg Member to the Burgersdorp Formation a gradual, but erratic, decline in the mean CEC can be observed.

It is also interesting to note that maximum CEC - values of the mudrock in the various strata also increase from the Dwyka Formation, but in this case the highest value is recorded in the mudrock of the Daggaboersnek Member.
From the Oudeberg Member upward higher maximum CEC - values are recorded in the mudrock of the argillaceous units, i.e. in the Daggaboersnek Member, Elandsberg Member and Burgersdorp Formation, than in the mudrock of the more arenaceous units, i.e. in the Oudeberg Member, Barberskrans

Member and Katberg Formation. This phenomenon may be attributed to a higher quartz-content in the mudrock of the arenaceous units.

In the argillaceous units the maximum CEC shows a decline from the Daggaboersnek Member to the Burgersdorp Formation.

Calcium is by far the dominating adsorbed cation in all the mudrock of the sequence in the area (Fig. 4 - 2). Differences in the mean total adsorbed cations of the various strata in the area can be attributed to the differences in the mean Ca^{++}- content as in each case the $Mg^{++} + K^{+} + Na^{+}$ - values appear to remain fairly constant (Fig. 4 - 2).

Extremely high Ca^{++} - values are recorded in the mudrock of the Oudeberg Sandstone Member. This may be due partly to the enrichment of calcium in the formation waters of the Oudeberg Sandstone Member during the compaction process, as explained by Blatt <u>et al</u>. (1972, p. 341 - 342). Should this be the case, it means that the Oudeberg Sandstone Member marks a major diagenetic event in the history of the Karoo Sequence in the area.

Adsorbed sodium appears to reach a maximum in the mudrock of the Middleton Formation and Oudeberg Sandstone Member, while it remains fairly constant in the rest of the sequence. Sodium, however, constitutes a very minor part of the total adsorbed cations (Fig. 4 - 2).

Potassium, although also a minor constituent of the total adsorbed cations, is more abundant than Na^+ and appears more prominently in the mudrock of the Koonap and Middleton Formations.

Magnesium remains rather constant throughout the sequence and is more abundant than Na^+ or K^+.

The ratio of adsorbed $Na^+ + K^+ : Ca^{++} : Mg^{++}$ in the various samples of mudrock from the different formations in the study area are presented in Fig. 4 - 3. The prominence of Ca^{++} is again reflected in most of the diagrams. It is clear that the mudrock of the Ecca Group contains exchangeable Ca^{++} in excess of 60 percent of the total adsorbed cations. The mudrock of the overlying Koonap Formation, however, appears to have less Ca^{++}, i.e. less than 50 percent of the adsorbed cations. This decrease is apparently related to a relative increase in K^+, as discussed previously.

Although most of the samples from the Middleton Formation contain exchangeable Ca^{++} in excess of 50 percent of the total adsorbed cations, the trend is similar to that in the Koonap Formation, i.e. a higher $Na^+ + K^+$ percentage due to higher K^+- values. In this case the $Na^+ + K^+$ - values of only three samples are below 10 percent of the total adsorbed cation content, while a quarter of the samples analysed, contain adsorbed Ca^{++} in amounts less than 50 percent and adsorbed $Na^+ + K^+$ in amounts more than 20 percent of the total exchangeable cations. The situation in the Balfour Formation is similar to that in the Ecca Group. Here most of the samples contain exchangeable Ca^{++} in amounts exceeding

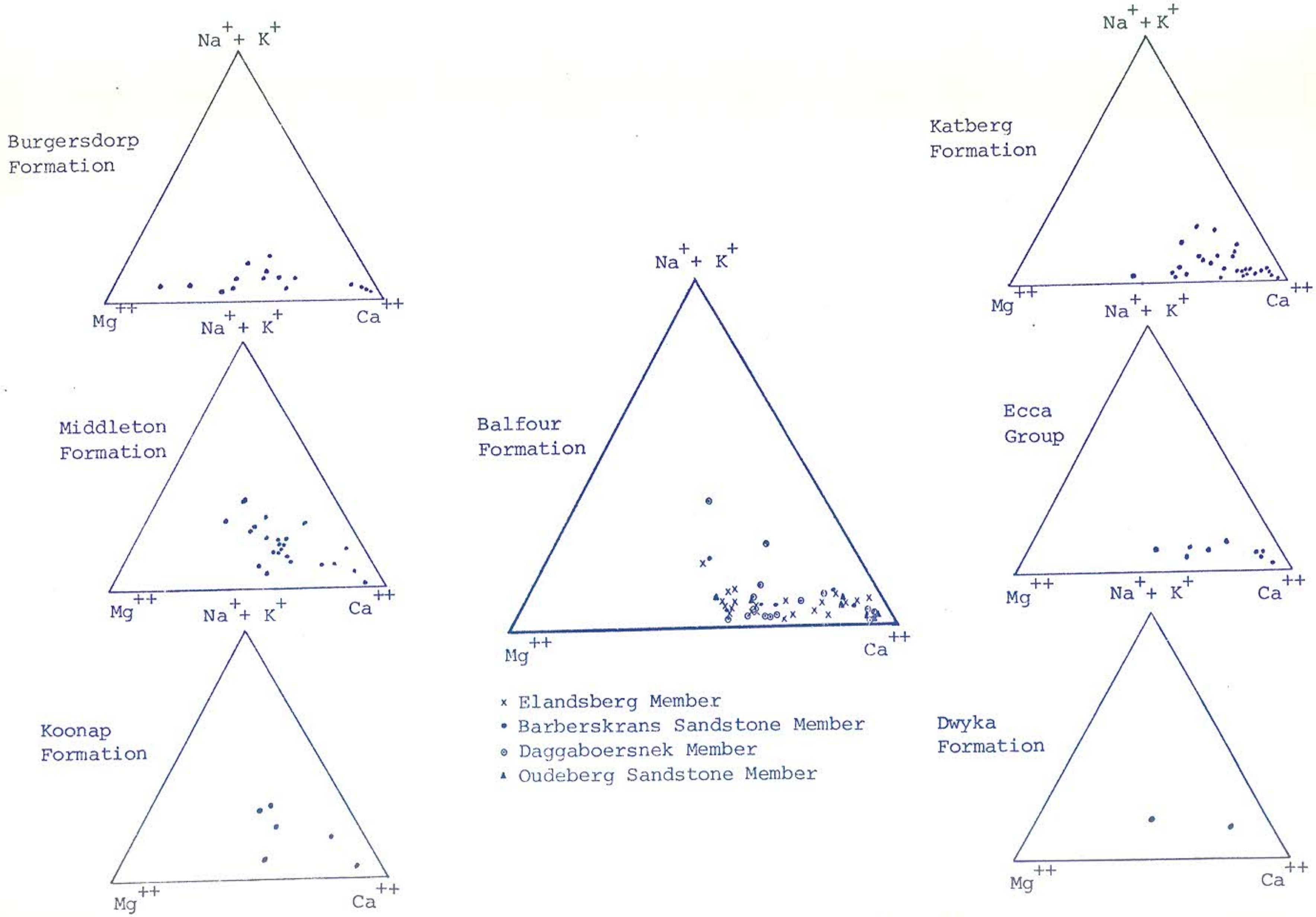


Fig. 4 - 3 Diagrams showing the ratio of adsorbed $Na^+ + K^+:Ca^{++}:Mg^{++}$ in the mudrock of the study area.

50 percent and $Na^+ + K^+$ in amounts less than 15 percent of the total adsorbed cations. Of special interest is the fact that most of the samples from the Oudeberg Member contain exchangeable Ca^{++} in amounts in excess of 80 percent of the total adsorbed cations. The Katberg Formation exhibits the same trend as the underlying Balfour Formation, while in the Burgersdorp Formation the mudrock appears to contain slightly higher percentages of exchangeable Mg^{++}.

To conclude: The Oudeberg Sandstone Member appears to be a very prominent geochemical marker in the hydrogeochemical history of the Karoo Sequence in the area. During some geological event, possibly compaction during the diagenetic stages, vast amounts of Ca^{++} were introduced into these rocks while Na^+ was taken into solution. Field observations prove that of all the sandstone units in the Karoo Sequence around Graaff-Reinet, the Oudeberg Sandstone Member underwent the most severe processes of diagenesis in which especially silicification played an important role.

4. 4. 3 Trends In The Cation-exchange Properties Of Karoo Mudrock In The Great Fish River Basin

The samples presented in Tables 4 - 2 to 4 - 6 were collected at selected localities on a grid of $7\frac{1}{2}'$ in order to obtain a good regional coverage of all the mudrock in the area. Moving-average isochemical contours of the data obtained from the cation-exchange analyses of these samples are presented in PLATES VI to XVI.

Special note must be made of the fact that the exchangeable cations are expressed in terms of mg/kg and not as me/100g, which is the accepted conventional notation. This is done in order to compare these figures with the ion concentrations in the groundwater, which are expressed as mg/l.

4. 4. 3. 1 Sodium

It has already been indicated that Na^+ constitutes only a very minor part of the total exchangeable cations in the mudrock of the area. In spite of this, however, PLATES VI and VII respectively show a definite trend in both the exchangeable and the percentage exchangeable sodium in the Great Fish River Basin.

Over a large part of the Headbasin, especially the catchments of the Klein Brak and Paul's Rivers, as well as the Teviot Basin, the exchangeable Na^+ in the mudrock is lower than 50 mg/kg (0,22 me/100g). Values of up to 100 mg/kg (0,43 me/100g) are, however, encountered in the central part of the Groot Brak River catchment.

The sodium becomes more abundant farther south and in most of the area below the Great Escarpment the mudrock contains more than 100mg/kg of exchangeable Na^+. High values are especially encountered over most of the Koonap River catchment and in the lower catchment of the Little Fish River.

It seems to be more than just coincidence that the higher Na^+- values occur either within or below the Oudeberg Sandstone Member. PLATE VII defines this change even better than PLATE VI, as a pronounced zone of Na^+ lower than 2 percent can be observed to the north of an imaginary line joining Somerset East, Bedford, Adelaide and Fort Beaufort, i.e. the strike of the Oudeberg Sandstone Member. To the south of this line higher Na^+ - percentages are observed.

The cause of this phenomenon is at this stage, however, not very clear. As will be discussed later, the groundwater in the area to the south of this line is extremely rich in sodium and could therefore have some influence on the adsorbability of the Na^+ relative to Ca^{++} in the mudrock. On the other hand, leaching in the low-lying area below the Great Escarpment is relatively poor, thus promoting the concentration of cations in the mudrock of such areas. Should the latter be the controlling factor, a general build-up of total exchangeable cations in this area should be encountered. This, however, does not appear to be the case, as revealed by PLATE XV.

The most likely explanation for the increase in exchangeable Na^+ in the mudrock of the southern part of the area, possibly lies in the fact that the sediments of the Ecca Group were deposited in a shallow marine environment. During the compaction of these sediments, the saline water would migrate upward in the strata, thus introducing Na^+-rich water into the overlying rocks which was deposited in a fluviatile environment. In order to satisfy the Law of Mass Action some of the other adsorbed cations would be exchanged for Na^+ only until equilibrium is reached. Because of the limited Na^+- adsorption these interstitial waters would still maintain a predominantly Na^+- character, while the surrounding mudrock adopts a slightly higher content of adsorbed Na^+. Such saline waters, however, do not appear to have migrated higher in the sequence than the Oudeberg Sandstone Member.

4. 4. 3. 2 Potassium

Potassium is more strongly adsorbed by clay minerals than Na^+ and therefore occurs in greater exchangeable quantities in the mudrock throughout the area. Although the smaller of two ions with the same valence should be more firmly held to a surface, Krauskopf (1967, p. 162) has shown the reverse in the case of K^+ and Na^+. This phenomenon is explained by Golubev and Garibyants (1971, p. 49) as a matter of increase in hydration or solvation with decrease in radius and this leads to a decline in electrostatic interaction with the charged surface of the adsorbent.

Potassium, although occuring in higher quantities than Na^+, appears to have the same trend in that higher values are encountered in the south of the area (PLATE VIII). Values exceeding 300mg/kg (0,77 me/100g) do, however, occur around Middelburg, Cradock and Tarkastad, but are of limited extent.

The influence of the Oudeberg Sandstone Member on the percentage exchangeable K^+ in the mudrock is clearly illustrated in PLATE IX. As with sodium, in and below the Oudeberg Sandstone Member high K^+ - percentages (>10) are encountered, whilst above this member the average percentage is smaller than 5 with higher values occuring only in limited areas around Middelburg, Cradock and Tarkastad.

4. 4. 3. 3 Calcium

It has already been pointed out that Ca^{++}, occurs as the most highly adsorbed cation in the mudrock of the area. As is noticed in PLATE X, there are very few areas in which the mudrock contains less than 1000 mg/kg (4,99 me/100g) of exchangeable Ca^{++}.

There appears to be no particular trend in the amount of exchangeable Ca^{++} in the area, although an area of relatively high (> 3000 mg/kg) Ca^{++}- values is present in the northern part of the Headbasin. Smaller areas of high exchangeable Ca^{++} are rather erratically scattered throughout the area.

According to PLATE XI the largest part of the area contains mudrock with exchangeable Ca^{++} exceeding 60 percent of the total exchangeable cations. Only in the headbasin of the Tarka River, to the north of Tarkastad, and in part of the area below the Great Escarpment do mudrocks occur which contain Ca^{++} in amounts less than 50 percent of the total exchangeable cations.

The ratio Ca^{++}/Na^+ shows a gradual decrease toward the south

(PLATE XII). In the central part of the Headbasin, i.e. in the area around Visrivier Station and farther north, the exchangeable Ca^{++} in the mudrock exceeds the exchangeable Na^{+} by a factor of 100, whilst below the Great Escarpment, this factor decreases to values lower than 20. This trend appears to be partly due to the higher Na^{+}- values observed in the Oudeberg Sandstone Member and in the Middleton Formation (Fig. 4 - 2).

The Ecca Group lies in the area of lowest Ca^{++}/Na^{+} - ratio. As this stratigraphic unit was deposited under shallow marine conditions, this appears to be the main cause for this low ratio. Sayles and Mangelsdorf (1977, p. 959) point out that the net reaction between fluvial clays and seawater is primarily an exchange of seawater Na^{+} for bound Ca^{++}. As previously pointed out, however, the Na^{+} concentration of this water was not sufficient to replace any large amounts of Ca^{++}.

4. 4. 3. 4 Magnesium

Magnesium is more readily adsorbed by the clay minerals than Na^{+} and K^{+}, but in relation to adsorbed Ca^{++} remains a minor constituent of the total adsorbed cations.

According to Fig. 4 - 2 the mean exchangeable Mg^{++} remains fairly constant, ranging between 2 and 4 me/100g. These values, however, appear to increase upward from the mudrock of the Dwyka Formation and reach a maximum in the Middleton and Balfour Formations. Lower values are again encountered within the mudrock of the Katberg Formation, whist a higher mean occurs in the Burgersdorp Formation, i.e. 3,68 me/100g.

The above trend accounts for the apparent irregular distribution of exchangeable Mg^{++} in the mudrock of the area. The areas of low exchangeable Mg^{++} to the east of Middelburg, to the east of Baroda and Cradock and around Tarkastad (PLATE XIII) appear to coincide with the outcrops of the Katberg Formation, whilst the area of low exchangeable Mg^{++} in the south coincides with outcrops of the Dwyka Formation, Ecca Group and Koonap Formation.

PLATE XIV shows a rather uniform distribution of the percentage exchangeable magnesium throughout the area. A rather poor correlation between PLATE XIV and PLATE XIII can, however, be observed in the central and northern parts of the area, but not in the south. Here Mg^{++} appears to constitute a higher percentage of the total exchangeable cations.

4. 4. 3. 5 Total Exchangeable Cations And CEC

No meaningful deductions can be made from the distribution of total exchangeable cations in the mudrock. Apart from the few areas in the north, as well as in the south, which show values exceeding 5000 mg/kg of total exchangeable cations, most of the mudrock have values ranging between 1500 and 5000 mg/kg (PLATE XV).

The distribution of the cation exchange capacity (CEC) of the mudrock, however, shows a reasonable trend of declining values toward the south (PLATE XVI). This trend can also be observed in Fig. 4 - 2.

In PLATE XVI the high CEC-values ($>$ 16 me/100g) of the mudrock of the Daggaboersnek Member relative to the low CEC-values of the mudrock of the Barberskrans Member ($<$ 10 me/100g) are clearly illustrated. The areas of low CEC to the west and east of Cradock and around Tarkastad represent mudrock from the Barberskrans and Elandsberg Members and from the Katberg Formation. To the south of these areas high values, mostly representing mudrock of the Daggaboersnek Member, are recorded.

There is, however, no correlation between the distribution of the total exchangeable cations and the CEC of the mudrock in the area.

5. GROUNDWATER

Before considering the chemical quality of the groundwater in the area, a brief discussion of the hydrologic properties of the water-bearing strata, as well as of the replenishment and movement of the groundwater is necessary. The proper understanding of these aspects is important, especially since the influence of groundwater on surface water is to be examined.

5. 1. HYDROLOGIC PROPERTIES OF THE WATERBEARING STRATA

Wilke (1961, p. 617) points out that the principal hydrologic properties of a geological unit are its porosity, specific yield and permeability. The soils, the sedimentary rocks and the dolerite will be discussed separately in order to obtain an idea of the relative importance of each regarding the distribution of subsurface waters.

5. 1. 1 Soils

Bond (1946, p. 9) suggests that the amount and nature of the soil and subsoil in an area play an important role as far as the infiltration of meteoric water is concerned. The nature of the soils is discussed in Chapter 3 and only the relevant aspects will be repeated here. The soils are restricted mainly to the low-lying areas of the Headbasin and of the

Marginal Region, while fairly deep alluvial deposits are encountered along the banks of the Great Fish River. Most of the soils in the low-lying areas occur as pediments, which change over into bahadas farther away from the source. The bahadas usually represent large flat plains consisting of relatively deep soil. Because of the sparse vegetation, these soils are often eroded by deep water channels. Apart from the pediments and bahadas in the low-lying areas, talus and rubble occur on the slopes of the mountains. These deposits are, however, of local extent.

The soil in the area consist mainly of clay minerals such as montmorillonite and illite which are derived mainly from the weathering of the mudrock. Loughnan (1969, p. 20) points out that montmorillonites, which are dried at temperatures in excess of 200°C, have an interlayer distance ranging between 9,1 and 10,0 Å, but on saturation with water the structure may expand to 10 times this value. The expanding nature of the soil therefore presents a rather serious problem as far as the infiltration of meteoric water is concerned. Initially the dry soil absorbs quite a vast amount of the rain water, but the moment the clay minerals in the soil have adsorbed sufficient water they swell to such an extent that insufficient pores are available for any more water to pass through and infiltration ceases. Wilke (1961, p. 618) mentions the fact that even although a loamy soil has a high porosity and specific retention, its low permeability prevents the retained water to infiltrate to the aquifers of the underlying solid rock.

The soils of the area can generally be regarded as poor aquifers because of the low permeability and low specific yield. It was pointed out in Chapter 2 that more than 90 percent of the annual precipitation is retained in the area. The possibility that very little of this water actually reaches the groundwater level via the soil horizons, means that the replenishment of the groundwater occurs via some other means.

It is suggested by Wilke (1961, p. 618) that talus and rubble, due to their high porosity and permeability allow water an easy downward passage to augment the underground water supplies. This loose, unconsolidated rock debris is often found on slopes between two successive plateaux or on the flanks of mountains and its extent and thickness is therefore rather limited.

Infiltration of meteoric water may also occur along the contact between the soil and the underlying rock. This zone often consists of coarse-grained rock fragments and gravel with a relatively high porosity and permeability. Surface outcrops of such zones are, however, rather limited but their cummulative effect may have a considerable influence. These zones are often covered by talus and rubble, in which case an ideal situation for the infiltration of meteoric water exists.

5. 1. 2 Sedimentary Rocks

Bond (1946, p. 10) points out that water-yielding rock strata can be subdivided into:-

(a) Porous or permeable rocks which hold water throughout their mass.

(b) Rocks practically impervious in mass, but holding water in joints and cleavages, fissures and other openings such as faults and shatter belts.

5. 1. 2. 1 Primary Hydrologic Properties

The primary hydrologic properties of the sedimentary rocks are those inherent to the rock itself and have not been modified by external or secondary processes like faulting, folding, jointing, weathering etc. Although these properties are of minor importance, they nevertheless require further discussion.

Sandstone has always been regarded as an important water-bearing rock in the area. Rowsell and De Swardt (1976, p. 117) however, record that the porosity of a number of sandstone samples from the Beaufort Group is below 2 percent, while the permeability ranges between 0 and 2 md. They conclude that the extremely poor porosity and permeability of the Beaufort sandstones make them unattractive as potential reservoirs. The porosity of a sandstone sample, probably from the Barberskrans Sandstone Member, was determined by Wyberg (1932, p. 118) as 4 percent. Water absorption tests on this sample indicate fine, almost disconnected textural pores and a few isolated structural cavities, whilst staining tests show even permeability and very strong secondary silicification. Du Toit (1915, p. 171) records porosities of 4,9 percent for sandstones from the higher horizons of the "Lower Beaufort" in the area. Frommurze (1953, p. 65) suggests, however, that the determinations made by Wyberg (1932) and Du Toit (1915) were probably carried out on weathered and partially weathered samples and that the unweathered rock might have an even lower porosity. This suggestion is clearly supported by the findings of Rowsell and De Swardt (1976).

The mudrock in the area has an even lower porosity, i.e., 0,4 to 0,7 percent and very little change in both the density and the porosity with depth can be observed (Rowsell and De Swardt, 1976, p. 118).

The extremely low specific yield of the rocks of the Karoo Sequence appears to be a general phenomenon as Van Wyk (1960) points out that field permeability tests on Karoo rocks in Natal, proved that "none of the rocks in the area is permeable enough to yield, in a 6" borehole an amount even approaching 100 g.p.h." Wilke (1961, p. 620), therefore, concludes "that the primary hydrologic properties of the sedimentary rocks are of minor importance regarding the storage and supply of underground water".

5. 1. 2. 2 Secondary Hydrologic Properties

In spite of the low porosity and permeability of the sedimentary rocks, Frommurze (1937, p. 168) recorded 1241 boreholes in the study area, which on average yielded 109,7 m^3 of water per 24 hours of pumping. This proves that at least some porous and permeable zones do exist in these rocks.

The geological changes which occured after solidification and diagenesis of the sedimentary rocks and which constitute the more important secondary hydrologic properties are jointing, fracturing, the intrusion of dolerite and weathering; diagenetic processes are excluded.

Joints And Fractures. Wilke (1961, p. 621) states that investigations on water levels and general water-level conditions in boreholes drilled in the Karoo sediments of the Fraserburg area showed that the holes tap confined water. From pumping tests and other observations he concludes that the groundwater occurs mainly in interconnected fissures, joints and fractures in the rock. The following observations were made by Wilke (1961):.

(i) A number of random pump tests were carried out in the Fraserburg district. After pumping at a high rate for several minutes, draw-downs reaching several metres were registered, but, as soon as the pump was stopped, the water level returned to its original rest level within a few seconds.

(ii) During drilling operations, the drill-holes remained comparatively dry until a water-bearing joint or fissure was struck, on which water entered the hole and rose a few metres to a point above the position where the water was struck.

(iii) On various occasions, the water level in two neighbouring holes was measured and found to differ by several metres. On the farm "Tweesusters", the water level in two holes, 6 m apart, was found to differ by as much as 12 m.

Similar observations concerning pumping tests and ground-water-level measurements were made in the present area of study. Unfortunately no data on actual drilling operations are available.

It is clear that the water levels in most of the bore-holes in the study area represent a pressure or piezometric surface rather than an actual water table. Fig 5 - 1, however, shows that regionally the groundwater levels represent a surface closely resembling that of the surface topography. Wilke (1961, p. 622) attributes this phenomenon to the fact that all excess water, after replenishment, is forced vertically and laterally into available openings until eventually a condition of equilibrium is reached. The upper surface of the zone of saturation thus conforms to the general ground level. Most of the joints and other water-bearing structures are thus regionally interconnected in such a manner that the groundwater level has an average depth of 15 m over most of the area.

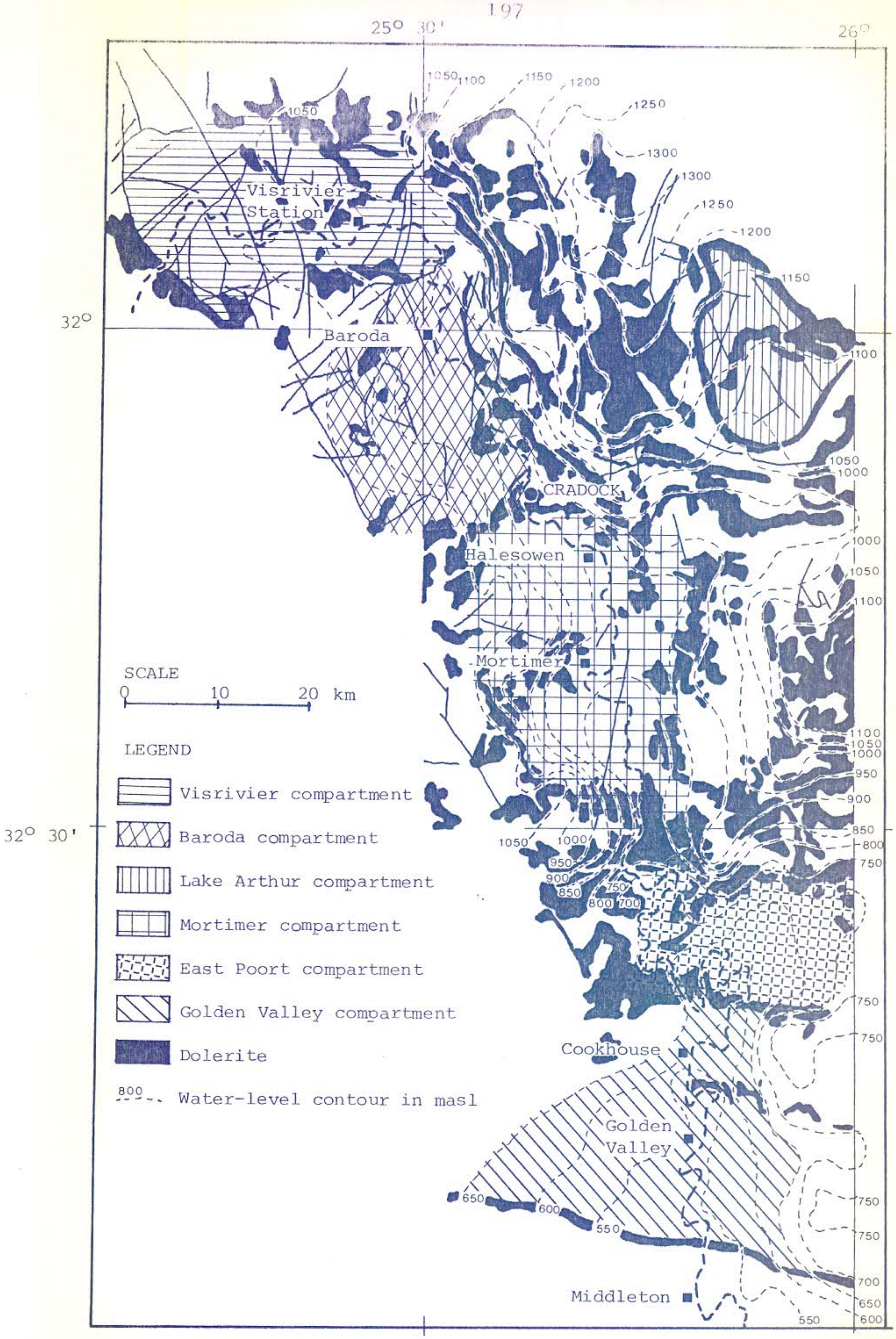


Fig. 5-1 Groundwater-level contours in the Great Fish River Basin illustrating the influence of dolerite on the formation of groundwater compartments.

Joints and fractures in the sandstone appear to hold more water than those in mudrock. The main reason for this is that such openings in mudrock are frequently sealed by clayey or calcareous material, thus preventing the circulation of groundwater.

The importance of any particular stratigraphic unit as an aquifer therefore depends largely on the amount of sandstone present. Taking the same approach as Wilke (1961, p. 623) one cubic metre of sandstone containing only two vertical joints 1 mm wide, will store 2×10^{-3} m^3 of water if the sandstone is immersed in water. If the joints and fractures are open to a depth of 100 m, the volume of water stored in the sandstone of each stratigraphic unit can be calculated. This will be determined by the percentage sandstone in that particular unit, as well as the proportion of such a unit in the total sequence.

The approximate area of the Great Fish River Basin is 25000 km^2. The total volume of rock up to a depth of 100 m is therefore 25×10^{11} m^3. The top 15 m of rock are, however, above the groundwater level and therefore only $21{,}25 \times 10^{11}$ m^3, contain sandstone with joints and fractures carrying 2×10^{-3} m^3 water per m^3 of rock. The volume of groundwater carried by the various stratigraphic units is therefore:

$$V = (21{,}25 \times 10^{11})\ (2 \times 10^{-3})\ (\frac{A}{100})\ (\frac{B}{100})\ m^3$$

V = total volume of water stored in joints and fractures of the sandstone in the unit

A = percentage which the unit constitutes of the whole stratigraphic column.

B = percentage sandstone in the unit.

The values for A are obtained from the stratigraphic column (PLATE IV), whilst the values for B are derived from the detailed sections measured in the area. The probable volume of water in each unit of the Beaufort Group were claculated and the result is presented in Table 5 - 1.

The total volume of 757,78 x 10^6 m^3 of groundwater which could be stored in joints and fractures in the sandstone of the Beaufort Group would cover the area of 25000 km^2 with a layer of water 30 mm deep. Other secondary structures, which are to be discussed in the following sections, may increase the above volume quite considerably.

Viljoen (1972, p. 1) reports a relatively low natural flow of water in the Great Fish River, i.e. 0,113 to 0,283 cumecs. If such a flow is attributed to the normal seepage of groundwater to the river, then the annual depletion of the groundwater reserves by this means varies between 3,563568 x 10^6 and 8,924688 x 10^6 m^3. These volumes respectively represent 0,47 percent and 1,18 percent of the 757,78 x 10^6 m^3 of groundwater which is probably stored in joints and fractures in the sandstone of the Beaufort Group in the area. Should the above water not be replenished at all, it would theoretically take over 100 years to deplete the calculated volume of water by seepage alone.

TABLE 5 - 1 Probable Volume Of Groundwater Present In Joints And Fractures In Sandstone Of The Beaufort Group In The Great Fish River Basin.

UNIT	PERCENTAGE OF STRATIGRAPHIC COLUMN	PERCENTAGE OF SANDSTONE	VOLUME IN m^3
Burgersdorp Fm	5	35	$74,38 \times 10^6$
Katberg Fm	6	76	$193,80 \times 10^6$
Elandsberg M	4	22	$37,40 \times 10^6$
Barberskrans M	2	85	$72,25 \times 10^6$
Daggaboersnek M	14	11	$65,45 \times 10^6$
Oudeberg Sandstone M	2	70	$59,50 \times 10^6$
Middleton Fm	20	30	$255,00 \times 10^6$
		Total =	$757,78 \times 10^6$

Intrusion Of Dolerite. Most of the water-yielding boreholes in the area are located in or near intrusions of dolerite. The reason for this is that the contact zones are almost always fractured and jointed, thus presenting highly permeable conduits to circulating groundwater. The sedimentary material in the fractured zone is invariably indurated so that the joints and fractures even in the mudrock are relatively open compared to those occuring away from the dolerite intrusion.

Such indurated and fractured zones are almost always linked with the major water-bearing fissures described in the previous section (Wilke, 1961, p. 624). Depending on the amount and extent of dolerite intrusion, the water-bearing capacity of the various stratigraphic units will therefore be considerably higher than the figures mentioned in Table 5 - 1. The extent to which the dolerite increases the water-bearing capacity of the sedimentary rocks is difficult to evaluate as this is dependent on the width of the fractured zone, as well as the degree of induration. Some relatively thick dolerite bodies appear to have had very little influence on the surrounding rocks, whilst in other cases almost insignificant bodies appear to have greatly indurated and fractured the surrounding sedimentary rocks.

If the dykes of dolerite cross the general flow of the groundwater, these bodies may act as impervious barriers, thus daming the water, often to such an extent that it seeps to the surface immediately above the dyke. A striking example of such a dyke occurs on the farm "Selection" near Baroda, halfway between the railway line and the national road. The dyke is here covered by colluvium and after the excessive rains of 1974, water started seeping out of the soil thus turning the cultivated lands in the immediate vacinity into a marsh.

Weathering. Because of the low porosity and permeability of the sedimentary rocks, weathering is confined mainly to the surface and to joints and fractures. It has, however, already been emphasized that due to the sedimentary nature of the rocks, as well as the prevailing semi-arid climatic conditions no large-scale chemical weathering occurs in the area. Near surface weathering proceeds mainly along joints and fractures, thus widening the structures and increasing their water-bearing capacity. Weathering of the sedimentary rock itself also improves infiltration of meteoric water, especially in the areas where no soil cover is present.

5. 1. 3 Igneous Rocks

Due to unfavourable conditions for chemical weathering the sills and dykes of dolerite themselves can be regarded as poor aquifers for groundwater. The dolerite is seldom sufficiently weathered to hold significant quantities of groundwater in the weathered zone. Groundwater is, however, often encountered in joints and fractures in the dolerite. Where such joints are interconnected, considerable amounts of groundwater may be present. Wilke (1961, p. 627) reports that a borehole in jointed dolerite in the Fraserburg area yielded a supply of 180 m^3/day.

5. 2. ORIGIN AND MOVEMENT OF GROUNDWATER

Before discussing the chemical quality of the groundwater, an understanding of the origin and movement of the groundwater is necessary. The origin of the water determines the concentration of dissolved substances already present in the groundwater prior to entering the environment, as well as the amount of water available. This in turn determines the extent to which chemical interaction between the new environment and the introduced water will take place.

The rate of movement of groundwater through rock strata in turn determines the time available for chemical reactions between the water and the surrounding rocks. It has already been pointed out that leaching of the more soluble substances occurs where there is good circulation of groundwater, whilst an accumulation of substances occurs in a stagnant environment.

5. 2. 1 Origin Of Groundwater

Groundwater may originate from magma penetrating the strata, connate water trapped in the sedimentary rocks after solidification and diagenesis, or from meteoric water infiltrating from the surface. According to Du Toit (1954, p. 548) the supply of groundwater in the Karoo Sequence is maintained by meteoric infiltration and is usually of a local origin. Frommurze (1953, p. 61) comes to the same conclusion and states further that the available quantities of groundwater of connate origin are rather limited, whilst no evidence has yet been produced to prove the existance of groundwater from a plutonic source.

The migration of connate groundwater during the compaction and disgenesis of the lower strata of the Karoo Sequence has already been discussed. One can therefore not rule out the possibility that at least some of the groundwater in the Adelaide Subgroup is of connate origin.

Factors controlling the infiltration of meteoric water are the amount and distribution of rainfall, the climate, the thickness and nature of the soils, the porosity and permeability of the underlying rocks, the structure of the sedimentary strata, the distribution and orientation of joints, faults and intrusive dykes, the geomorphology of the area and the vegetation in the area.

Although most of the groundwater in the area originates from rainwater, Du Toit (1954, p. 549) points out that the yields are to a great extent independent of the magnitude of the seasonal precipitation. The other factors mentioned above must therefore play an important part in concentrating the rainwater in the underlying strata.

Part of the rainwater which accumulates on the surface penetrates the overlying soil cover and eventually reaches the groundwater level. Most of this water is, however, lost to the atmosphere through evapotranspiration before it even reaches the groundwater level and therefore the nature and depth of the soil, climate, vegetation, topography and structures such as joints, faults, dykes and bedding planes play a major role in controlling the rate of infiltration. It is suggested in Chapter 2 that less than 5 percent of the annual precipitation ever reaches the groundwater level. Only during excessive precipitation, when the evaporation rate remains considerably low for some time, will the groundwater level show a marked elevation.

5. 2. 2 Movement Of Groundwater

The groundwater levels in several hundred boreholes in the immediate area around the Great Fish River were measured in order to determine the regional slope of the water table. This survey was conducted over a relatively short period so that marked fluctuations due to climatic conditions could be eliminated.

Contours of the groundwater level above mean sea level are illustrated in Fig. 5 - 1, showing a definite slope toward the river, suggesting groundwater seepage into the river at suitable localities. There is also a gradual regional slope toward the south, indicating topographic control on the movement of groundwater. The influence of topography on the chemical quality of the groundwater will be discussed later.

In sedimentary strata groundwater generally flows along bedding planes in a down-dip direction. In the study area the dip of the strata is, however, opposite to the slope of the landscape and is, furthermore, so slight that the regional direction of the groundwater flow is almost entirely controlled by the topography.

Groundwater will flow readily along joints and fracture zones and the orientation of these structures therefore control the local direction of flow. The dolerite dykes and therefore the fracture zones along the dykes, are orientated mainly in a north-south direction (Fig. 3 - 13). This together with the topography, will encourage the groundwater to flow in a southerly direction. The orientation of many of the dolerite intrusions, however, deviates from the north-south direction. The maze of dolerite intrusions will therefore deviate or arrest the groundwater flow along its general southerly course.

Fig. 5 - 2 is a schematic illustration of groundwater in joints and fractures in Karoo strata intruded by a dyke of dolerite, thus causing a higher water level on the "up-flow" side. The formation of groundwater compartments, as illustrated in Fig. 5 - 1, is implied by this mechanism.

It must be stressed, however, that these compartments do not represent areas in which the bulk of the sedimentary rock is saturated with water; only the joints and fractures in the rock are filled to a slightly elevated level.

Where joints, fractures and dolerites occur in low-lying areas like river beds, springs or diffuse effluent seepage may occur. Many of the permanent water pools in the Fish River represent such seepages of groundwater. Due to the fact that the yields of these seepages are dependent on climatic conditions, Wilke (1961, 632) suggests that they should rather be regarded as intermittent springs.

5. 3. CHEMISTRY OF GROUNDWATER

5. 3. 1 Sampling Procedure And Chemical Analyses

Hem (1970, p. 69) is doubtful whether a single water sample

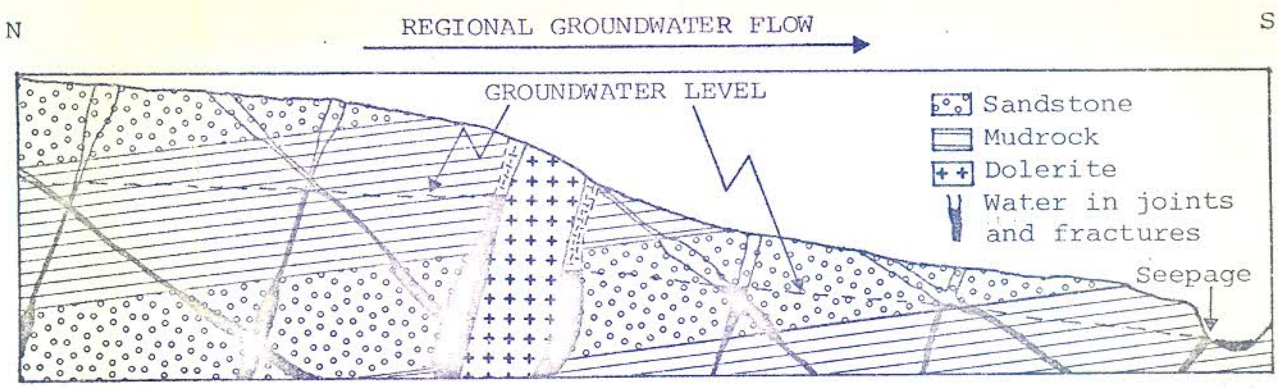


Fig. 5 - 2 Schematic illustration of groundwater in joints and fractures. An impervious dolerite dyke inhibits the free flow of groundwater.

from a borehole represents the chemical composition of all the groundwater in the vertical section at that point; it is, however, a useful indication of the average composition. Several lines of approach were followed in sampling the groundwater during this study. Numerous boreholes have been drilled in the Great Fish River Basin and sampling of all holes proved an impossible task. Samples were therefore collected from the nearest borehole(s) on a grid of approximately 12 km x 12 km. The chemical analyses of these samples are presented in Tables 5 - 2 to 5 - 6.

It will be noticed that the samples are grouped according to the geological strata in which that particular groundwater occurs, as one of the aims was to establish the influence of the various geological strata on the quality of the groundwater.

Of the major groundwater compartments shown in Fig. 5 - 1, only the Baroda and Golden Valley Compartments were investigated in detail (Fig. 5 - 1) and all available boreholes were sampled here. The groundwater quality of these compartments (Table 6 - 2 and 6 - 3) are discussed

TABLE 5-2 Inorganic Quality Of Groundwater Samples From The Burgersdorp Formation.

SAMPLE	CO-ORDINATES		CATIONS					ANIONS				Cations + anions mg/1	pH
No	Lat. S	Long. E	Na^+ mg/1	K^+ mg/1	Mg^{++} mg/1	Ca^{++} mg/1	Total mg/1	Cl′ mg/1	(HCO_3′) mg/1	SO''_4 mg/1	Total mg/1		
1 - 5	31° 15′	25° 30′	26,3	0,6	25,0	212,4	264,3	11	866,0	4,0	881	1145	7,7
1 - 6	31° 15′	25° 37½′	25,0	0,3	29,6	46,1	101,0	18	311,0	9,0	338	439	8,0
1 - 9	31° 15′	26° 00′	24,0	0,6	25,6	80,2	130,4	22	296,0	31,0	349	479	7,6
2 - 8	31° 22½′	25° 45′	29,4	1,5	38,4	24,0	93,3	12	323,0	5,0	340	433	7,6
2 - 9	31° 22½′	25° 52½′	29,3	2,5	31,4	38,1	101,3	24	305,0	4,0	333	434	8,0
2 -10	31° 22½′	26° 00′	67,0	1,7	30,8	48,1	147,6	38	378,0	28,0	444	591	7,6
3 - 9	31° 30′	25° 45′	100,0	0,1	45,4	4,2	149,7	43	397,0	28,0	468	618	8,2
3 -10	31° 30′	25° 52½′	36,4	6,0	43,2	18,0	103,6	34	279,0	32,0	345	449	8,3
3 -11	31° 30′	26° 00′	70,0	6,0	34,4	38,1	148,5	44	264,0	114,0	420	569	8,0
4 -10	31° 37½′	25° 52½′	200,0	2,0	36,8	6,3	245,1	206	372,0	7,0	585	830	7,8
4 -11	31° 37½′	26° 00′	200,0	2,0	6,0	42,1	250,1	91	451,0	71,0	613	863	8,0
5 - 8	31° 45′	25° 45′	35,4	2,3	62,0	5,0	104,7	82	214,0	55,3	351,3	456	7,8
5 - 9	31° 45′	25° 52½′	270,0	4,8	45,0	20,0	339,8	206	525,0	101,6	832,6	1173	7,7
5 -10	31° 45′	26° 00′	150,0	8,3	41,0	17,0	216,3	106	403,0	72,0	581,0	797	7,9
5 -11	31° 45′	26° 07½′	49,5	1,6	36,0	38,0	125,1	26	366,0	18,0	410,0	535	7,7
5 -12	31° 45	26° 15′	79,5	1,4	35,0	46,0	161,9	76	323,0	61,8	460,8	623	7,8
6 -10	31° 52½′	26° 07½′	40,5	1,6	33,0	16,0	91,10	45	220,0	21,9	286,9	378	8,1
6 -11	31° 52½′	26° 15′	32,5	0,5	10,0	28,0	71,0	26	171,0	1,04	198,0	269	7,8
7 -11	32° 00′	26° 15′	108,0	14,0	60,0	65,0	247,0	210	348,0	74,6	632,6	879	7,6

TABLE 5-3 Inorganic Quality Of Groundwater Samples From The Katberg Formation.

SAMPLE	CO-ORDINATES		CATIONS					ANIONS					
No	Lat. S	Long. E	Na^+ mg/l	K^+ mg/l	Mg^{++} mg/l	Ca^{++} mg/l	Total mg/l	Cl′ mg/l	(HCO_3′) mg/l	SO″$_4$ mg/l	Total mg/l	Cations + anions mg/l	pH
1 - 1	31° 15′	25° 00	29,2	0,4	18,2	56,1	103,9	27	262	28	317	421	7,7
1 - 2	31° 15′	25° 07½′	26,0	0,7	15,5	26,1	68,3	18	171	23	212	280	8,0
1 - 3	31° 15′	25° 15′	20,6	1,0	19,2	44,1	84,9	20	226	19	265	350	7,7
1 - 4	31° 15′	25° 22½′	71,0	0,6	6,6	28,1	106,3	31	244	8	263	369	8,1
2 - 1	31° 22½′	24° 52½′	20,3	0,5	38,8	19,1	78,7	28	232	19	279	358	7,8
2 - 2	31° 22½′	25° 00′	32,1	0,7	22,0	46,1	100,9	20	281	15	316	417	7,5
2 - 3	31° 22½′	25° 07½′	63,6	0,8	31,6	24,6	120,6	64	262	28	354	475	7,8
2 - 4	31° 22½′	25° 15′	57,5	1,0	44,0	16,4	118,9	50	293	36	379	498	7,9
2 - 5	31° 22½′	25° 22½′	18,9	2,0	39,2	28,1	88,2	21	277	7	305	393	7,6
2 - 6	31° 22½′	25° 30′	61,5	4,2	35,2	40,1	141,0	44	318	59	421	562	7,8
3 - 6	31° 30′	25° 22½′	40,4	2,9	40,6	27,1	111,0	92	210	22	324	435	7,6
3 - 7	31° 30′	25° 30′	65,5	0,2	28,6	16,8	111,1	67	159	67	293	404	7,9
3 - 8	31° 30′	25° 37½′	300,0	17,0	14,8	126,3	458,1	431	311	180	922	1380	7,8
4 - 4	31° 37½′	25° 07½′	42,6	1,4	37,4	28,1	109,5	53	251	39	343	453	8,2
4 - 6	31° 37½′	25° 22½′	40,4	2,9	40,6	27,1	111,0	92	210	22	324	435	7,6
4 - 7	31° 37½′	25° 30′	94,3	5,0	20,6	37,2	157,1	106	234	48	388	545	7,6
4 - 8	31° 37½′	25° 37½′	114,0	6,9	42,2	35,8	198,9	145	312	57	514	713	7,7
5 - 2	31° 45′	25° 00′	27,0	1,2	39,2	7,5	74,9	35	226	5	266	341	7,8
5 - 3	31° 45′	25° 07½′	21,0	1,1	24,0	36,0	82,1	18	256	3	277	359	7,5
5 - 6	31° 45′	25° 30′	232,0	4,7	45,0	43,0	324,0	220	397	154	771	1095	7,8
5 - 7	31° 45′	25° 37½′	124,0	4,5	76,0	28,0	232,5	160	393	98	651	882	7,5
6 - 7	31° 52½′	25° 45	36,0	1,8	73,0	14,0	124,8	60	378	18	456	581	7,9
7 - 7	32° 00′	25° 45	103,5	3,0	52,0	30,0	188,5	125	305	88	518	707	7,6
7 - 8	32° 00′	25° 52½′	87,0	5,7	48,0	54,0	194,7	103	360	80	543	738	7,7
7 -10	32° 00′	26° 07½′	31,4	1,6	47,0	22,0	102,0	22	348	4	374	476	7,9
8 - 8	32° 07½′	25° 52½′	92,0	3,3	42,0	8,0	145,3	100	232	68	400	545	8,0
8 -10	32° 07½′	26° 07½′	80,0	0,3	30,0	38,0	148,3	75	317	30	422	569	8,2
8 -11	32° 07½′	26° 15′	48,2	1,4	21,0	64,0	134,6	42	323	23	388	522	7,8
8 -12	32° 07½′	26° 22½′	57,5	1,0	33,0	42,0	133,5	61	299	37	397	531	8,0
9 -11	32° 15′	26° 15′	17,8	1,0	10,0	18,0	46,8	21	122	1	144	191	7,7
9 -12	32° 15′	26° 22½′	45,5	2,6	30,0	66,0	144,1	57	329	36	422	567	7,5

TABLE 5 - 4 Inorganic Quality Of Groundwater Samples From The Balfour Formation.

SAMPLE	CO-ORDINATES		CATIONS					ANIONS				Cations + Anions mg/l	pH
No	Lat. S	Long. E	Na^+ mg/l	K^+ mg/l	Mg^{++} mg/l	Ca^{++} mg/l	Total mg/l	Cl' mg/l	(HCO_3') mg/l	SO''_4 mg/l	Total mg/l		
					ELANDSBERG MEMBER								
3 - 1	31° 30'	24° 45'	26,8	1,0	35,4	8,7	71,9	57	153	14	224	296	7,7
3 - 3	31° 30'	25° 00'	129,0	2,4	16,4	54,1	201,9	111	281	95	487	639	7,8
4 - 1	31° 37½'	24° 45'	22,9	0,8	41,6	26,1	91,4	17	287	28	332	423	7,8
4 - 3	31° 37½'	25° 00'	61,5	1,6	25,3	86,2	174,6	39	435	41	515	690	7,7
4 - 5	31° 37½'	25° 15'	125,0	1,4	38,9	22,9	188,2	100	342	66	508	696	7,7
5 - 4	31° 45'	25° 15'	11,5	1,1	25,0	12,0	49,6	9	177	1,2	187	237	7,6
5 - 5	31° 45'	25° 22½'	117,2	2,0	15,0	43,0	177,2	108	293	33,4	434	611	7,6
6 - 1	31° 52½'	25° 00'	31,0	0,7	42,0	6,0	79,7	35	220	25,7	281	361	7,9
6 - 2	31° 52½'	25° 07½'	154,0	3,1	27,0	44,0	203,8	123	342	100,4	565	769	7,7
6 - 5	31° 52½'	25° 30'	200,0	3,9	29,0	34,0	266,9	240	177	154,4	571	838	7,5
7 - 1	32° 00'	25° 00'	69,0	0,7	18,0	17,0	104,7	58	189	30,9	278	383	7,9
8 - 1	32° 07½'	25° 00'	29,5	0,7	26,0	42,0	98,2	25	293	1,3	319	417	8,0
8 - 2	32° 07½'	25° 07½'	36,0	2,0	18,0	16,0	72,0	27	189	1,6	218	289	7,8
8 - 3	32° 07½'	25° 15'	55,0	1,6	38,0	38,0	132,6	49	311	47,6	408	540	7,9
9 - 2	32° 15'	25° 07½'	14,1	0,6	9,0	16,0	39,7	20	98	1,9	120	160	8,0
10 - 7	32° 22½'	26° 00'	50,5	1,0	34,0	34,0	119,5	40	317	17,0	374	494	7,9
10 - 8	32° 22½'	26° 07½'	50,5	1,0	34,0	34,0	119,5	42	320	16,0	378	498	7,9
					BARBERSKRANS MEMBER								
6 - 3	31° 52½'	25° 15'	43,6	2,8	83,0	22,0	153,4	163	311	12,9	487	640	7,5
6 - 4	31° 52½'	25° 22½'	58,0	5,0	22,0	56,0	141,0	68	262	52,8	383	523	7,8
11 - 7	32° 30'	26° 07½	88,5	1,2	37,0	7,5	134,2	55	336	9,0	400	534	8,1
					DAGGABOERSNEK MEMBER								
9 - 6	32° 15'	25° 37½'	1694	9,0	314	1112	3129	4875	61	869	5805	8934	7,6
9 - 7	32° 15'	25° 45'	853	3,9	94	265	1216	1638	153	447	2238	3454	6,9
10 - 2	32° 22½'	25° 22½'	31,4	1,0	24	42	98,4	38	201	54	293	391	7,8
10 - 4	32° 22½'	25° 37½'	110,0	1,0	14	14	139,0	41	288	40	369	508	7,8
10 - 5	32° 22½'	25° 45'	284,0	5,0	58	3	350,0	298	384	127	809	1159	8,1
10 - 6	32° 22½'	25° 52½'	73,0	0,8	27	22	123,0	44	288	26	358	481	7,8
11 - 3	32° 30'	25° 37½'	134,0	1,2	50	28	213,2	120	470	19	609	822	7,8
11 - 4	32° 30'	25° 45'	1035,0	4,0	4	215	1258,0	1960	0	44	2004	3262	5,1
11 - 8	32° 30'	26° 15'	49,0	1,6	23	26	99,6	47	250	1	298	398	7,8
					OUDEBERG SANDSTONE MEMBER								
13 - 8	32° 45'	26° 15'	238	3,0	47	36	324	376	313	23	712	1036	7,8
13 -10	32° 45'	26° 30'	692	3,4	123	42	860,4	1090	438	216	1744	2604	7,5

TABLE 5-5 Inorganic Quality Of Groundwater Samples From The Middleton Formation

SAMPLE	CO-ORDINATES		CATIONS					ANIONS					
No	Lat. S	Long. E	Na^+ mg/1	K^+ mg/1	Mg^{++} mg/1	Ca^{++} mg/1	Total mg/1	Cl′ mg/1	(HCO_3′) mg/1	SO″$_4$ mg/1	Total mg/1	Cations + Anions mg/1	pH
12 - 5	32° 37½′	25° 52½′	400,0	3,9	110,6	34,1	548,6	677	287	201,7	1166	1715	8,0
13 - 2	32° 45′	25° 30′	254,0	12,0	51,0	21,0	338,0	387	249	75,0	711	1049	7,8
13 - 4	32° 45′	25° 45′	680,0	9,0	117,0	58,0	864,0	1029	544	209,0	1782	2646	7,6
13 - 5	32° 45′	25° 52½′	1505,0	6,0	171,0	49,0	1731,0	2581	414	121,0	3116	4847	7,8
14 - 6	32° 52½′	26° 00′	386,0	17,0	50,0	46,0	499,0	575	354	82,0	1011	1510	7,6
14 - 7	32° 52½′	26° 07½′	460,0	11,5	73,0	24,0	568,5	675	368	118,0	1161	1730	7,8
14 - 8	32° 52½′	26° 15′	235,0	12,0	58,0	64,0	369,0	334	476	112,0	922	1291	7,6
14 - 10	32° 52½′	26° 30′	434,0	10,0	56,0	12,0	512,0	548	451	69,0	1068	1580	7,9
14 - 11	32° 52½′	26° 37½′	485,0	7,4	71,0	32,0	595,4	720	384	99,0	1203	1798	7,8

TABLE 5-6 Inorganic Quality Of Groundwater Samples From The Koonap Formation, Ecca Group And Dwyka Formation.

SAMPLE	CO-ORDINATES		CATIONS					ANIONS						
No	Lat. S	Long. E	Na^+ mg/1	K^+ mg/1	Mg^{++} mg/1	Ca^{++} mg/1	Total mg/1	Cl′ mg/1	(HCO_3') mg/1	SO_4'' mg/1	Total mg/1	Cations + Anions mg/1	pH
					KOONAP FORMATION								
15 - 1	33° 00′	25° 30′	324,0	10,0	63,0	62,0	459	475	476	66	1017	1476	7,7
15 - 4	33° 00′	25° 52½′	152,0	16,0	33,0	36,0	237	210	293	42	545	782	7,7
					ECCA GROUP								
16 - 4	33° 07½′	26° 00′	515,0	5,8	79,0	74,0	674	770	464	163	1397	2071	7,7
16 - 6	33° 07½′	26° 15′	411,0	12,0	85,0	96,0	604	635	537	154	1326	1930	7,7
16 - 8	33° 07½′	26° 30′	581,6	23,0	147,0	78,0	830	1250	106	280	1636	2466	7,6
					DWYKA FORMATION								
16 - 9	30° 07½′	26° 37½′	4660,0	60,0	830,0	29,0	5579	7742	683	2123	10548	16127	8,3

in the next chapter. Groundwater samples were also collected at various localities along the Great Fish River (Table 6 - 1) and these are also discussed in the next chapter. The aim of the above two investigations was to determine the influence of groundwater on the quality of base flow in the river. The present chapter deals with regional variations in groundwater quality.

It is pointed out by Hem (1970, p. 71) that the movement of groundwater in an aquifer is usually slow enough so that samples taken monthly may easily reveal changes in the quality with time. Such samples from borehole B93 near Baroda were taken over the period November 1973 to September 1974. The analytical results of these samples are presented in Table 6 - 2.

Although natural waters can contain numerous ions in varying quantities, Chebotarev (1955a, p. 41) states that the types of soluble salts in such waters remain largely unchangeable; he, therefore, only uses the ions Na^{+}, K^{+}, Ca^{++}, Mg^{++}, Cl^{-}, HCO_3^{-}, $CO_3^{=}$ and $SO_4^{=}$ to determine and express the possible chemical reactions and properties of water. Assuming that the anions are independent ingredients, while the cations are dependent variables Chebotarev (1955a, p. 41) recognises the following three major categories of natural waters, which practically cover all the variety of the chemical compositions of subterranean waters: (a) bicarbonate waters, (b) sulphate waters and (c) chloride waters.

Because Ca^{++} and HCO_3^{-}-concentrations may exchange as a result of $CaCO_3$ precipitation in stored water, especially when stored for some time, two 500 ml polythene containers were filled with the water from each particular sampling point. A few drops of dilute HCl were then added to one sample in order to prevent any $CaCO_3$ precipitation, whilst the other sample was used for determining the concentrations of the rest of the ions present. All samples were collected from boreholes on which windmills are erected. Care was, however, taken to ensure that the mill had been pumping water for at least 15 minutes before the sample was obtained. A portable pH - meter was used to determine the hydrogen ion concentration of the samples in the field.

The cations mentioned above were determined by means of atomic-absorption spectrophotometry and flame emission spectrography, whilst the anions were determined by means of wet chemical analyses. The analyses of $CO_3^{=}$ in all the samples tabulated were so low, that the results were incorporated into the HCO_3^- concentration and expressed as total alkalinity.

5. 3. 2 Origin Of Major Chemical Components In Groundwater

Chebotarev (1955b, p. 210 - 211) discusses the metamorphic cycle of natural waters and concludes that chloride brines, which are formed even in "fresh-water" sediments, can be attributed to the "metamorphism" of meteoric water. He also states that the "salinity distribution of subterranean waters obeys a definite hydrological and geochemical law which can be formulated as the cycle of metamorphism of natural waters in the crust of weathering".

The main environments in which the chemical quality of groundwater in the Great Fish River Basin is affected, are illustrated in Fig. 5 - 3.

5. 3. 2. 1 Evaporation, Condensation And Precipitation Of Meteoric Water

It is a well known fact that water evaporating from the sea may contain as much as 1 mg/1 of $SO_4^=$ and of Cl^- (Hem, 1970, p. 164 - 176). As the water vapour, however, moves inland the above concentrations become less due to an initial fall-out close to the shore. It is therefore unlikely that much $SO_4^=$ and Cl^- is introduced to the area by these means.

During the condensation and precipitation of meteoric water, CO_2 and O_2 from the atmosphere are dissolved in the water, thus increasing the ability of the water as a weathering agent. As a result of the solution of CO_2 in the rainwater, the HCO_3^- and H^+ - contents are increased considerably. Krauskopf (1967, p. 14, 36 - 37) points out that sufficient CO_2 from the atmosphere can dissolve in water to form a 10^{-5} M solution of H_2CO_3, which according to his calculations may reduce the pH to 5,7. Such natural waters are therefore better weathering agents.

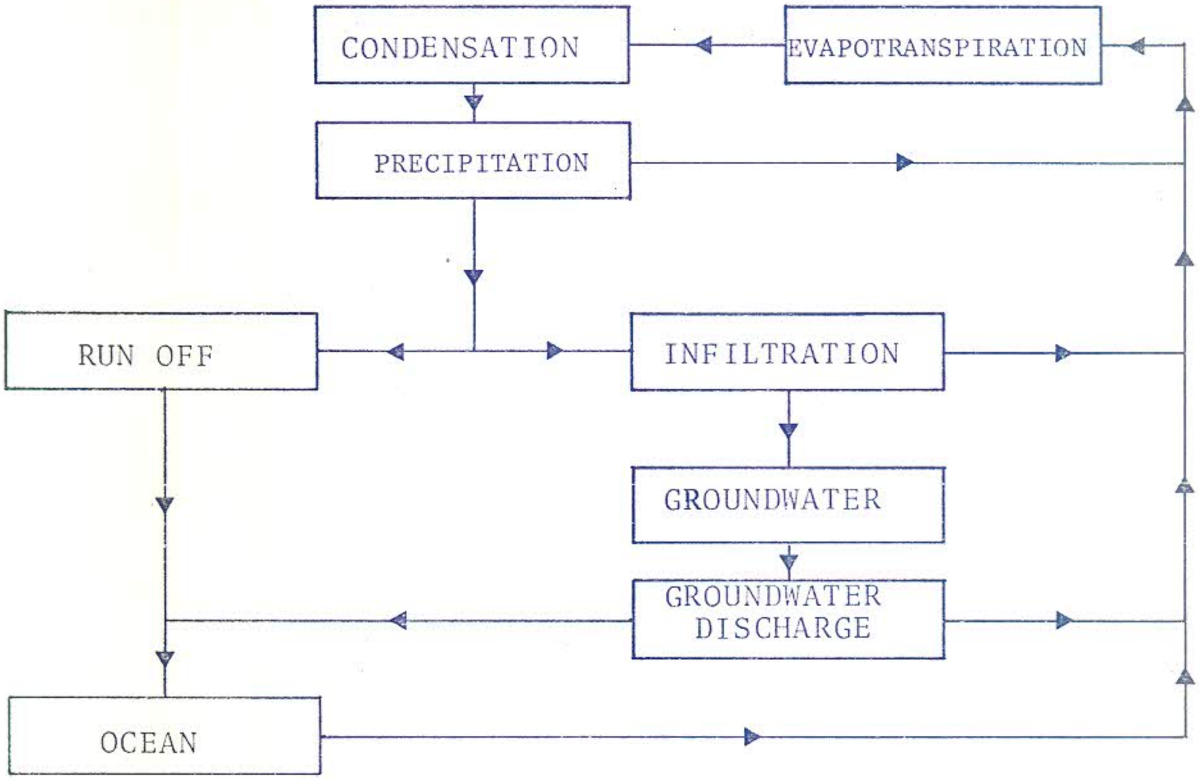


Fig. 5 - 3 Diagram showing the various environments in which the chemical quality of groundwater in the Great Fish River Basin may be affected.

5. 3. 2. 2 Infiltration Of Meteoric Water

It has previously been pointed out that only a very small percentage(approximately 3 percent) of the total annual precipitation in the area is lost by runoff. The remaining 97 percent either evaporates immediately, infiltrates the soil where most of it is lost through evapotranspiration or is retained in the soil and infiltrates the rock strata to eventually reach the water-saturated zone below the groundwater level.

More CO_2 is present in the soils as a result of the respiration of micro-organisms and plant roots. In such environments the CO_2 - concentration may be even several hundred times greater than that of the atmosphere (Lawrence, 1975, p. 107). During the infiltration of meteoric water chemical reactions between the water and the soil and rock may occur, resulting in the solution of Na^+, K^+, Ca^{++} and Mg^{++}. Other reactions, which occur in the unsaturated zone above the groundwater level, are the precipitation of colloidal iron, silica and carbonates while cation exchange is common. Most of the dissolved substances are, however, concentrated in or near the upper horizons of the soil as a result of evapotranspiration.

It is therefore quite understandable why the small volume of meteoric water which eventually reaches the groundwater level has a reasonably high salinity.

5. 3. 2. 3 Saturated Zone Below The Groundwater Level

Once the groundwater reaches the saturated zone, reactions involving cation exchange predominate as this environment contains less oxygen, which promotes the oxidizing reactions higher up. Adsorbed Na^+ is normally replaced by Ca^{++}, therefore increasing the Na^+ concentration of the groundwater.

The solution or precipitation of minerals may also take place in the saturated zone, depending on whether the solubility product of that particular mineral is reached or not.

Garrels and Christ (1965, p. 83 - 85) explain a method according to which one can determine the degree of calcium carbonate saturation of a water sample by merely taking the normal chemical analysis into account. The molal concentration of Ca^{++} obtained from the chemical analysis is compared with the molal concentration of Ca^{++} which should be in equilibrium with the amount of HCO_3^- in the sample. One may therefore calculate the degree of $CaCO_3$ saturation by means of the following expression:

$$\frac{m\ Ca^{++}\ \text{analy.} - m\ Ca^{++}\ \text{equil.}}{m\ Ca^{++}\ \text{equil}} \times 100$$

where m Ca^{++} analy. and m Ca^{++} equil. are respectively the analytical molal concentration of Ca^{++} in the sample and the molal concentration of Ca^{++} which is calculated to be in equilibrium with the analytical HCO_3^- of the sample.

The mean $CaCO_3$ saturation of the groundwater in the various sedimentary strata of the area is presented in Fig. 5 - 4. It will be noticed that in all cases the groundwater is to a lesser or greater extent oversaturated with $CaCO_3$ and where the environment is suitable precipitation is bound to occur.

Thus far a fair account of the origin of most of the cations in the groundwater, as well as the origin of HCO_3^-, has been presented. The origin of Cl^- concentrations, however, presents a slight problem as Lawrence (1975, p. 112) points out that the chloride concentration in almost all minerals,

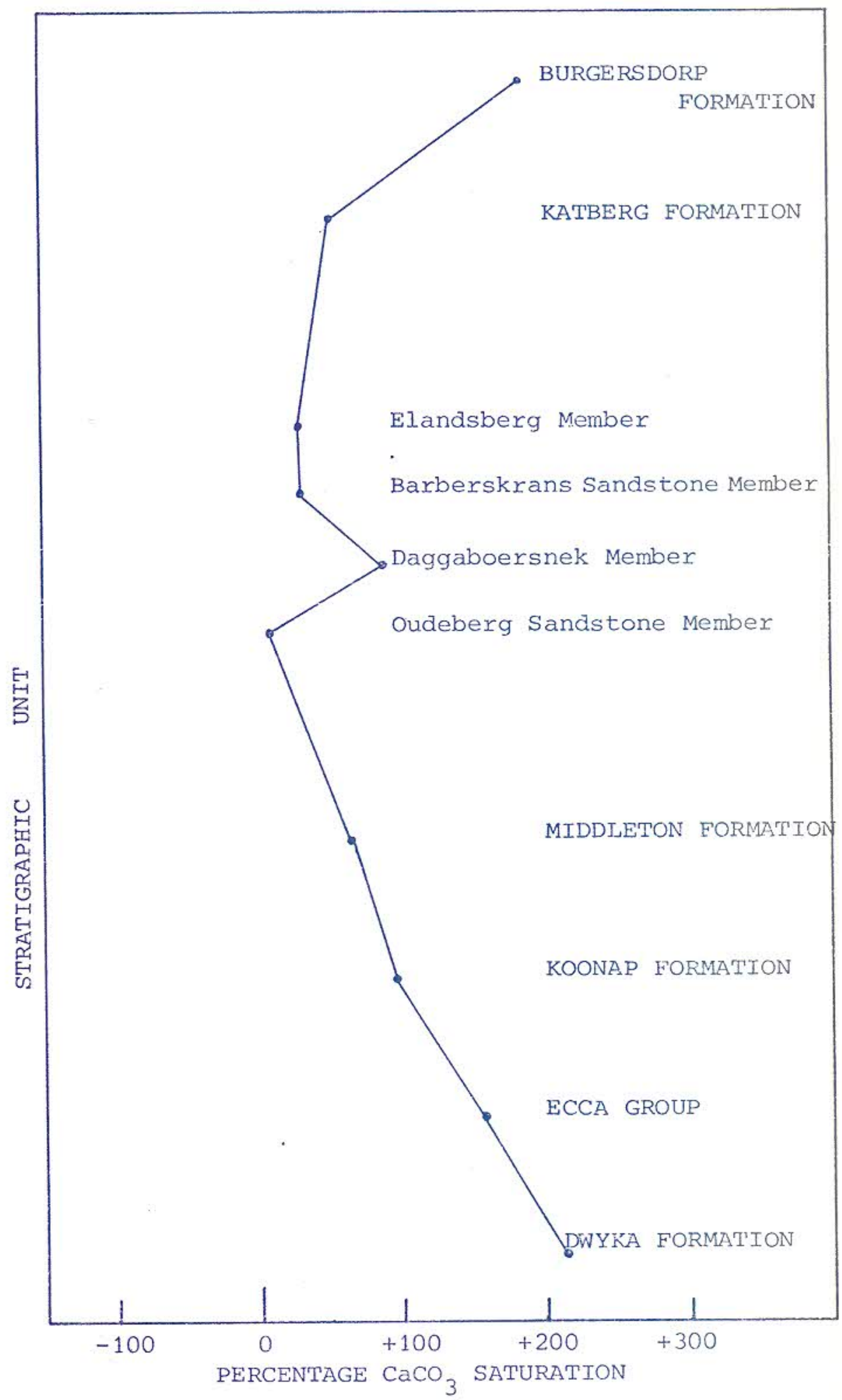


Fig. 5 - 4 Comparison of the mean $CaCO_3$ saturation in the groundwater of the various strata in the Great Fish River Basin.

excepting some of the evaporites, is so low that mineral dissolution cannot be regarded as a source for this element. He also states that igneous emanations are at present a very minor source of chloride and that the release of sodium chloride-rich water from fluid inclusions in the primary minerals during weathering is a very minor source of Cl^-. Evaporites are practically absent in the study area and cannot be regarded as a major source of Cl^-.

In spite of the above facts, Chebotarev (1955b, p. 203) states that a common evolution in the groundwater chemistry along the regional flow path is:

$$HCO_3^- \rightarrow HCO_3^- + Cl^- \rightarrow Cl^- + HCO_3^- \rightarrow Cl^- + SO_4^=$$

$$\text{or} \quad SO_4^= + Cl^- \rightarrow Cl^-$$

He attributes this change to the difference in mobility of the various ions involved and to the physical chemical processes which occur in the subsurface reservoir.

Lawrence (1975, p. 110 - 112) supports the above model and states that although the chloride at present in the earth's crust originated from "degassing" at an early stage in the geological history of the earth, the Cl^- is now mainly present in solutions rather than in the solid phase.

The main source of Cl^- in the natural waters of the study area is therefore attributed probably to the mixing of connate water in the sedimentary strata with the younger meteoric water.

In the case of the Karoo Sequence in the area the sedimentary strata from the Ecca Group downward were deposited under marine conditions with the result that connate water rich in Cl^- was trapped in the sedimentary rocks. Some alteration of the original water must, however, have taken place during diagenesis of the sediments as explained in the previous chapter.

Another mechanism which is responsible for the concentration of especially Cl^- in certain sedimentary rocks, is the ultra-filtration of the connate water as it is forced upward through the mudrock layers which act as semi-permeable membranes during the compaction of the sediments. This mechanism is discussed in Chapter 4.

Lawrence (1975, p. 109 - 110), however, emphasizes the fact that the amount of replacement of connate water by meteoric water is dependent on the following three main factors:

(i) The length of time that groundwater flow has been established in the basin.

(ii) The velocity of flow in the basin, which depends on the hydrolic conductivity and anisotropy of hydrostratigraphic units, the spatial relationship of these units, the topography and the climate.

(iii) The size of the basin in relation to factors (i) and (ii).

Other chemical reactions which are regarded by Lawrence (1975, p.106, 109) to occur below the groundwater level are the sulphate reduction by bacteria together with the generation of CO_2 by the breakdown of organic material. He states that the growth of sulphate-reducing bacteria requires an anaerobic environment where an adequate supply of organic matter is available and suggests the following generalized exothermic reaction:

$$SO_4^{=} + (CH_2O) \longrightarrow CO_2 + H_2S + H_2O,$$

where CH_2O represents a number of organic compounds. This may be the type of reaction which is responsible for many of the sulphuretted waters encountered in the area. According to Kent (1949, p. 253) the thermal spring at Cradock contains $CO_3^{=}$ and HCO_3^{-} as the main anions and therefore could be the result of such a reaction. The oxidation of pyrite in the sedimentary rocks must, however, not be disregarded as a possible source of H_2S in some of the groundwater.

5. 3. 2. 4 Points Of Groundwater Discharge

It has already been pointed out that the discharge of groundwater occurs mainly in the rivers or low-lying areas. The groundwater is here once again exposed to evapotranspiration and therefore the salinity increases. When the solubility product of certain ions in solution is exceeded, salts of these ions are precipitated, thus resulting in the white deposits often encountered in the area.

The high salinity of the groundwater, which is discharged into the Great Fish River, may be diluted by the runoff water after rains or by irrigation water which is periodically let into the river. During periods when neither rain or irrigation water flows down the river a gradual build-up of saline water occurs.

5. 3. 3 Influence Of Topography On Groundwater Quality

Although there are many factors which control the accumulation of salts in groundwater, topography appears to play one of the more important roles in the case of the Great Fish River Basin. Lukashev (1970, p. 258) points out that evaporation and its ratio to drainage are the principal factors influencing salt accumulation in groundwater. Topography in turn controls (a) the rate of surface runoff and therefore the amount of moisture available for chemical reactions and (b) the rate of groundwater movement which consequently controls the rate of removal of the soluble products (Levinson, 1974, p. 88).

The topography of the area is clearly illustrated in PLATE I and was discussed in some detail in Chapter 2. 1. Of the four major geomorphological environments, both the Headbasin and the Marginal Region represent areas in which the groundwater movement is retarded, thus promoting the accumulation of dissolved substances.

Contour maps which were compiled from the chemical data in Tables 5 - 2 to 5 - 6 are presented in PLATES XVII to XXXIII. The relationship between the groundwater chemistry and the topography is illustrated by comparing PLATE I with the above.

5. 3. 3. 1 Total Dissolved Solids

Groundwater with relatively low TDS (< 400 mg/l) is encountered only in the Interior Plateau, as well as in the mountainous regions within the Headbasin (PLATE XVII). In the Interior Plateau groundwater with a moderate concentration of TDS (400 - 600 mg/l) is also encountered. This groundwater, however, extends from the above plateau to the outer perimeter of the Headbasin.

The poorly-drained central parts of the Headbasin contain groundwater with a relatively high salinity (600 - 1000 mg/l). It will also be noted that the Teviot Basin, in which several salt pans are encountered, occurs within this zone.

Along the Great Escarpment the groundwater appears to contain a surprisingly high salinity (600 - 1000 mg/l). This may be due to the fact that the argillaceous Daggaboersnek Member occurs mainly in this area. It was pointed out in the previous chapters that this sedimentary unit was probably deposited in a brackish lake environment. The saline groundwater in the Great Escarpment appears therefore to be the result of geological processes rather than the result of topographic control.

Groundwater with extreme high salinity (> 1000 mg/l) is often encountered in the Marginal Region below the Great Escarpment (PLATE XVII). The source of this high salinity may be attributed partly to the gradual build-up of dissolved substances as the groundwater progresses along its course down the basin and also to the presence of saline connate water from the marine deposits of the Dwyka Formation and the Ecca Group.

Apart from a few local environments where the groundwater quality is affected by factors other than topography, the general trend in the Great Fish River Basin is for the groundwater to become more saline as it moves from the higher lying areas, where most of the recharge occurs, to the lower lying areas where more stagnant conditions are encountered.

According to Lukashev (1970, p. 258) the salinity of groundwater increases with the intensity of evaporation and therefore a high salinity is associated with shallow groundwater levels. The following stages in the formation of saline groundwater with increasing salinity are suggested (Lukashev 1970, p. 258):

(i) Silicate-alkali water of low salinity (1-3g/l) and appreciable prevalence of HCO_3^-, $CO_3^=$, Mg^{++}, Na^+ and silicate ions. Saturation of the water with respect to these ions causes precipitation of carbonates and silicates.

(ii) Calcium-carbonate sulfate water with a salinity of 3 - 5g/1. Saturation of the water causes precipitation of calcium carbonates and sulfates together with silicon compounds.

(iii) Chloride-sulfate water with a salinity of 5 - 20g/1, occasionally 70-100g/1. A characteristic feature of this stage is the continous coprecipitation of silicon compounds, $CaCO_3$, $CaSO_4$, $MgCO_3$ and parially Na_2SO_4.

(iv) The most highly saline sulfate-chloride water with salinities exceeding 35 - 50g/1 occasionally 100 - 200 g/1. This stage is characterized by the saturation of the solution in silicates, Ca- and Mg- carbonates and Ca- and Na- sulphates, all of which precipitate. The high salinity and high Na- content of this water is important, since the Na^+ is believed to exchange Ca^{++} and Mg^{++} in the surrounding rocks with the resultant accumulation of large amounts of Ca^{++} and especially Mg^{++} in the water.

Although concentrations in the Great Fish River Basin are not as high as those mentioned above, the same trend in the groundwater quality is observed. In the lower lying areas the groundwater levels are much shallower than those in the areas with higher relief and therefore concentration of salts are encouraged by evapotranspiration.

5. 3. 3. 2 The pH

The pH of the groundwater appears to decrease only slightly from the higher lying areas of recharge to the lower lying parts of the Headbasin and the Marginal Region (PLATE XVIII). This phenomenon is attributed to the hydrolysis reactions which occur during the solution of $CaCO_3$ in the newly recharged water. According to Hem (1970, p. 90) the following generalized reaction occurs during the solution of $CaCO_3$ in the water:

$$CaCO_3 + H_2O = Ca^{++} + HCO_3^- + OH^-.$$

According to the above reaction Ca^{++}, HCO_3^- and OH^- - concentrations in the groundwater will increase as a result of $CaCO_3$ solution. The low pH of the initial meteoric water, as a result of the solution of atmospheric and biogenic CO_2 is therefore increased.

PLATES XXII and XXVIII reveal that in the areas of recharge, i.e. the Interior Plateau, Ca^{++} and HCO_3^- respectively constitute the major proportions of the dissolved cations and anions in the groundwater. This fact therefore supports the above reaction.

The lower pH values which occur in the central part of the Headbasin and in the Marginal Region are apparently associated with the relatively stagnant groundwater conditions.

5. 3. 3. 3 Sodium

It is reported by Hem (1970, p. 145) that once sodium is brought into solution, it tends to remain there because of the fact that there are no important precipitation reactions which can maintain low sodium concentrations in natural waters. The only way sodium may be retained in the solid phase is by adsorption on clay minerals in the soil or in sedimentary rocks. Because of the preferential adsorption of Ca^{++} above Na^{+}, only small amounts of sodium are, however, retained this way in the study area (Tables 4 - 2 to 4 - 6).

As the groundwater progresses into the basin from the area of recharge, the Ca^{++}, which was initially dissolved, is either precipitated as $CaCO_3$ or is adsorbed in exchange for Na^{+}. The sodium concentration of the groundwater therefore increases along the regional flow direction. This trend is clearly observed in PLATES XIX and XX where low Na^{+} - values are encountered in the groundwater of the Interior Plateau, whilst the concentrations increase toward the centre of the Headbasin.

The high concentration of sodium in the Marginal Region may, however, be partly due to the presence of Na^{+} - rich connate water from the Dwyka Formation and from the Ecca Group.

Another striking fact concerning the increase in sodium in the groundwater of the area, is its direct relationship to the increase in salinity of the water (PLATE XVII and XIX).

5. 3. 3. 4 Calcium

Equilibria involving carbonates are according to Hem 1970,

p. 132) the major factor in limiting the solubility of calcium in most natural water. Ca^{++}, because of its position in the so-called lyotropic series (Golubev and Garibyants, 1971, p. 49) is, however, easily adsorbed especially by clay minerals and therefore has a limited mobility in relation to Na^{+}.

It is therefore obvious that as soon as the groundwater containing the dissolved Ca^{++} leaves the area of recharge, the calcium starts exchanging adsorbed Na^{+} from the surrounding material through which the water flows. The Na^{+} - concentration increases with an equivalent decrease in Ca^{++}.

Calcium is furthermore removed from the solution as the groundwater penetrates environments which are suitable for $CaCO_3$ precipitation.

PLATE XXI displays a trend in the calcium concentration which may be associated with the trend of increased salinity in the direction of regional flow of the groundwater.

It is interesting to note that in PLATE XXI the highest Ca^{++}-concentrations are encountered in the area around Cradock and Halesowen, where relatively pronounced deposits of calcrete and gypsum are found.

The percentage Ca^{++} of the total cations in solution, however, reveals clearly a general pattern of limited mobility for this ion (PLATE XXII). Although the percentage of Ca^{++} in solution seldom exceeds 50 percent of the total cations, the highest percentages, i.e. $>$ 30 percent are encountered in the Interior Plateau and gradually decreases toward the centre of the Headbasin and the Marginal Region where values lower than 10 percent are encountered.

5. 3. 3. 5 Magnesium

Although Hem (1970, p. 143) is of the opinion that waters in which magnesium is the predominant cation are somewhat unusual, areas occur in the Headbasin where Mg^{++} constitutes more than 50 percent of the total cations in the groundwater (PLATE XXIV). It is quite clear that the high percentages of Mg^{++} occur in the outer perimeter of the Headbasin, slightly farther along the flow-path of the groundwater than the calcium. This means that the magnesium is more mobile than the calcium, but less mobile than the sodium, as lower percentages of Mg^{++} are also encountered in the central parts of the Headbasin where Na^{+} is the dominant cation.

The concentration of magnesium in the groundwater is generally very low, but tends to increase with an increase in total dissolved solids. (PLATE XXXIII and XVII).

As far as the soluble cations in the groundwater of the Great Fish River Basin are concerned, one may conclude that without any exception all tend to increase in concentration with an increase in salinity. Calcium, however, is prominently the least mobile cation and dominates only in the Interior Plateau, while Mg^{++} is only slightly more mobile and Na^+ is the most mobile. This trend is clearly expressed in PLATE XXXI where the ratio of Na^+ to $Ca^{++} + Mg^{++}$ is greater than one mainly in the central part of the Headbasin and all the way down the river into the Marginal Region.

5. 3. 3. 6 Chloride

A brief discussion on the geochemical behaviour of chloride is necessary before actually investigating the concentration of this ion in the groundwater of the various geomorphological environments.

Chloride ions do not significantly enter into oxidation or reduction reactions, they form no important solute complexes with other ions (especially in dilute aqueous solutions), do not form salts of low solubility, are not significantly adsorbed on mineral surfaces and play few vital biochemical roles (Hem, 1970, p. 172). It may be deduced therefore that the circulation of Cl^- in the hydrological cycle is largely through physical rather than chemical processes.

The chloride ions therefore tend to remain in solution and are able to withstand most of those processes which reduce the concentration of other ions. Chloride concentrates in groundwater close to the surface as a result of evapotranspiration, but can also concentrate in deepseated connate waters which filter through fine-grained argillaceous sediments during compaction (Chapter 4).

There is a marked increase in the actual concentration of chloride, as well as in the percentage chloride of the total anions, in the groundwater as it proceeds towards the centre of the Headbasin (PLATES XXV and XXVI). This trend can also be observed toward the Marginal Region, but of great interest is the narrowing immediately south of Mortimer. Such a narrowing may be attributed to the presence of some form of hydrological barrier in the area and is displayed by most of the other soluble constituents. This area marks the lower end of the Headbasin where the river cuts through the Great Escarpment.

A similar trend as that of Na^+ is revealed by the Cl^-, which agrees with the statement by Hem (1970, p. 175) that "The most common type of water in which chloride is the dominant anion, is one in which sodium is the predominant cation."

The higher concentration of chloride in the groundwater of the Marginal Region may partly be attributed to the retention of Cl^- during the upward migration of the oceanic connate water as a result of compaction of the sediment.

5. 3. 3. 7 Alkalinity

Hem (1970, p. 158) is of the opinion that the alkalinity of many streams is caused mainly by the solution of CO_2 of the atmosphere rather than from the rocks of a drainage basin. Where this is the case low Ca^{++} or Mg^{++} - concentrations must, however, be encountered as these ions are produced mainly by the $CaCO_3$ - hydrolysis reaction.

It is therefore clear that the alkalinity in groundwater may originate from two sources, i.e.:

(i) The hydrolysis of $CaCO_3$:

$$CaCO_3 + H_2O \rightleftharpoons Ca^{++} + HCO_3^- + OH^-,$$

In this case the Ca^{++}-concentration, alkalinity and pH are high.

(ii) The solution of CO_2 from the atmosphere or from biological action:

$$H_2O + CO_2 \rightleftharpoons H_2CO_3$$

$$H_2CO_3 \rightleftharpoons H^+ + HCO_3^- \qquad K = 10^{-6,4}$$

$$HCO_3^- \rightleftharpoons H^+ + CO_3^= \qquad K = 10^{-10,3}$$

In this case no Ca^{++} is produced and although the alkalinity increases, the pH clearly decreases.

In the Great Fish River Basin there appears to be quite a good correlation between the alkalinity (PLATE XXXVII) and the pH (PLATE XXXVII). The most striking coincedence is that in the Interior Plateau regions the pH is high with a relatively high alkalinity, thus suggesting the hydrolysis of $CaCO_3$. In the Marginal Region high alkalinity concentrations are encountered, whilst the pH appears to be relatively low. This suggests that most of the HCO_3^- in this area is due to the mere solution of CO_2 in the groundwater.

Whilst the actual alkalinity of the groundwater in the area appears to occur in irregular concentrations (PLATE XXVII), the percentage alkalinity of the total dissolved anions reveal a regular trend in which the alkalinity appears along the flow direction of the groundwater. The highest alkalinity percentages (> 70 percent) occur in the Interior Plateau, whilst percentages exceeding 50 percent also occur over a large section of the outer perimeter of the Headbasin. High alkalinity percentages are also encountered in the groundwater of the Great Escarpment. Only a small section of the central part of the Headbasin and the whole of the Marginal Region contain groundwater in which the alkalinity is less than 50 percent of the total anions in solution.

This trend is very similar to that of the calcium(PLATE XXII). It may therefore be deduced that the concentration of Ca^{++} in the groundwater of the area is controlled mainly by the equilibria of the carbonates in the water, or that the alkalinity is controlled mainly by the solution of $CaCO_3$ in the water.

5. 3. 3. 8 Sulphate

The sulphate ion is chemically stable in aerated water and forms salts of low solubility with only a few metals (Hem, 1970, p. 164). He also states that most shales and fine-grained sediments, which are freshly raised above sea level, are exposed to the natural processes of weathering which bring about oxidation within the aerated zone right down to zones below the water table. During such processes sulphate is produced which may be transported away from the source. The rate at which the sulphate is removed, depends on the runoff rate and is therefore greatly dependent on environmental factors such as climate and topography.

In semi-arid environments such as encountered in large areas of the Great Fish River Basin, the supply of solutes is relatively large in proportion to the water volume in which it can be carried away and therefore where the subsurface drainage is poor, an accumulation of dissolved solids occurs with $SO_4^=$ as one of the main constituents.

PLATE XXIX proves that there is a pronounced migration of $SO_4^=$ from the high lying Interior Plateau to the lower lying central parts of the Headbasin. There also appears to be a further migration of sulphate from the Great Escarpment to the Marginal Region.

Sulphate, however, constitutes only a small percentage of the total anions in the groundwater and only in a limited portion of the Headbasin does this ion ever exceed 15 percent of the total dissolved anions (PLATE XXX).

One may therefore conclude that the topography of the Great Fish River Basin plays an important role in controling the migration and accumulation of ions in the groundwater of the area. It has been shown that Ca^{++} and HCO_3^-, for example, dominate the groundwater in the areas close to where it is recharged by meteoric water, i.e. in the Interior Plateau region. The more mobile ions (Na^+ and Cl^-) on the other hand are concentrated in the areas where movement of groundwater is limited, i.e. the central parts of the

Headbasin and the Marginal Region. Magnesium and sulphate, which have intermediate mobilities, often tend to concentrate in the groundwater somewhere between the two extremes mentioned. This trend is especially noticed with Mg^{++}, whilst the $SO_4^{=}$ has a slightly greater mobility under the prevailing conditions and is therefore often encountered in areas where high Na^{+} and Cl^{-}-concentrations occur in the groundwater.

PLATES XXXII and XXXIII, which respectively represent the ratios

$$Cl^{-}/HCO_3^{-} \quad \text{and} \quad \frac{Na^{+} + Cl^{-} + SO_4^{=}}{Ca^{++} + Mg^{++} + HCO_3^{-}}$$

in the groundwater of the area also illustrate the control of topography on the concentrations of these ions. The ratios are determined by comparing the concentrations in milliequilivalents per litre.

5. 3. 4 Influence Of Geology On Groundwater Quality

Most of the geological factors influencing groundwater quality have already been discussed either directly or indirectly. Rock weathering, as a result of the chemical reaction between circulating groundwater and the minerals constituting the surrounding rocks, may be regarded as one of the main geological factors influencing the quality of the water. During such reactions some of the elements contained in the rock-forming minerals are released to solution, thereby changing the chemical composition of the water. Chebotarev (1955a, p. 36) pointed out, however, that the proportion and type of soluble matter which is taken up from the rock-material depends on:

(i) The type of geological formations
(ii) The structural features of the area
(iii) The temperature of the water
(iv) The salinity concentration and the abundance of particular ions and compounds in the water
(v) The amount of water moving through the particular rock-type
(vi) The velocity with which the water flows through the rock.

Hem (1970, p. 287) points out that the main cations which occur in groundwater are derived mainly from the solution of minerals during chemical weathering, whilst the anions may be derived mainly from nonlithologic sources.

The dolerite in the study area, because of its igneous origin, is more susceptible to chemical attack by the prevailing atmospheric agents than are the sedimentary rocks. Cations such as Ca^{++}, Na^{+} and Mg^{++} are released to solution as a result of the weathering of plagioclase and pyroxene, which constitute the main mineral assemblage of the dolerite.

The sedimentary rocks, however, also contain resistant primary minerals which may undergo weathering under the present environmental conditions. It must, however, be born in mind that such minerals have already endured at least one and in some cases two cycles of weathering and will therefore not be readily decomposed by the weathering agents.

The precipitation of cementing materials or the solution thereof in the sedimentary rocks cause a marked change in the quality of the groundwater. Such cementing materials may include calicum carbonate, silica and ferric hyroxide or ferrous carbonate.

Ion-exchange is another factor which may be responsible for a major change in the chemical quality of the groundwater in an area. As a result of the "Law of Mass Action" sediments which are deposited in a marine environment will naturally have Na^+ as the main adsorbed cation. Once such sediments are exposed to the atmosphere where meteoric water rich in Ca^{++} can percolate through them, the adsorbed Na^+ is exchanged for the divalent Ca^{++}, thus rendering the remaining groundwater rich in Na^+.

It was pointed out in Chapter 3 that most of the rocks of the Middleton and Balfour Formations were deposited under reducing conditions. Disseminated pyrite may therefore occur within these deposits, but will on exposure to the atmosphere oxidise to hematite or other ferrous hydroxides, whilst sulphate is released to solution. Many oxidised remnants of pyrite are found in these sediments. Biological reactions responsible for the reduction of the sulphate concentration in groundwater have already been discussed.

Water is not easily transmitted through the sedimentary rocks of the study area because of their low porosity and permeability. Especially in the case of the fine-grained mudrock, connate water, which was entraped during the deposition of the sediment, is able to remain there for a considerable length of time.

Such waters are found to be highly saline (Hem, 1970, p. 300) and may have a great influence on the groundwater quality of the area. This is especially true where marine deposits are encountered.

Chebotarev (1955b, p. 201) points out that whilst groundwater is not saturated, it cannot remained unchanged in chemical composition as long as it is in contact with rocks containing soluble material.

In order to compare the quality of the groundwater in the various stratigraphic units of the area, the data in Tables 5 - 2 to 5 - 6 were plotted on Piper-Palmer trilinear diagrams (Fig. 5 - 5 to 5 - 9), while in Fig. 5 - 10 histograms of the major ions in the groundwater of the various strata are compared. The mean concentration of the ions in the groundwater from the various strata of the area is presented by Fig. 5 - 11.

5. 3. 4. 1 Groundwater Quality Of The Dwyka Tillite Formation

Due to the very steep dip of this unit, outcrops are limited

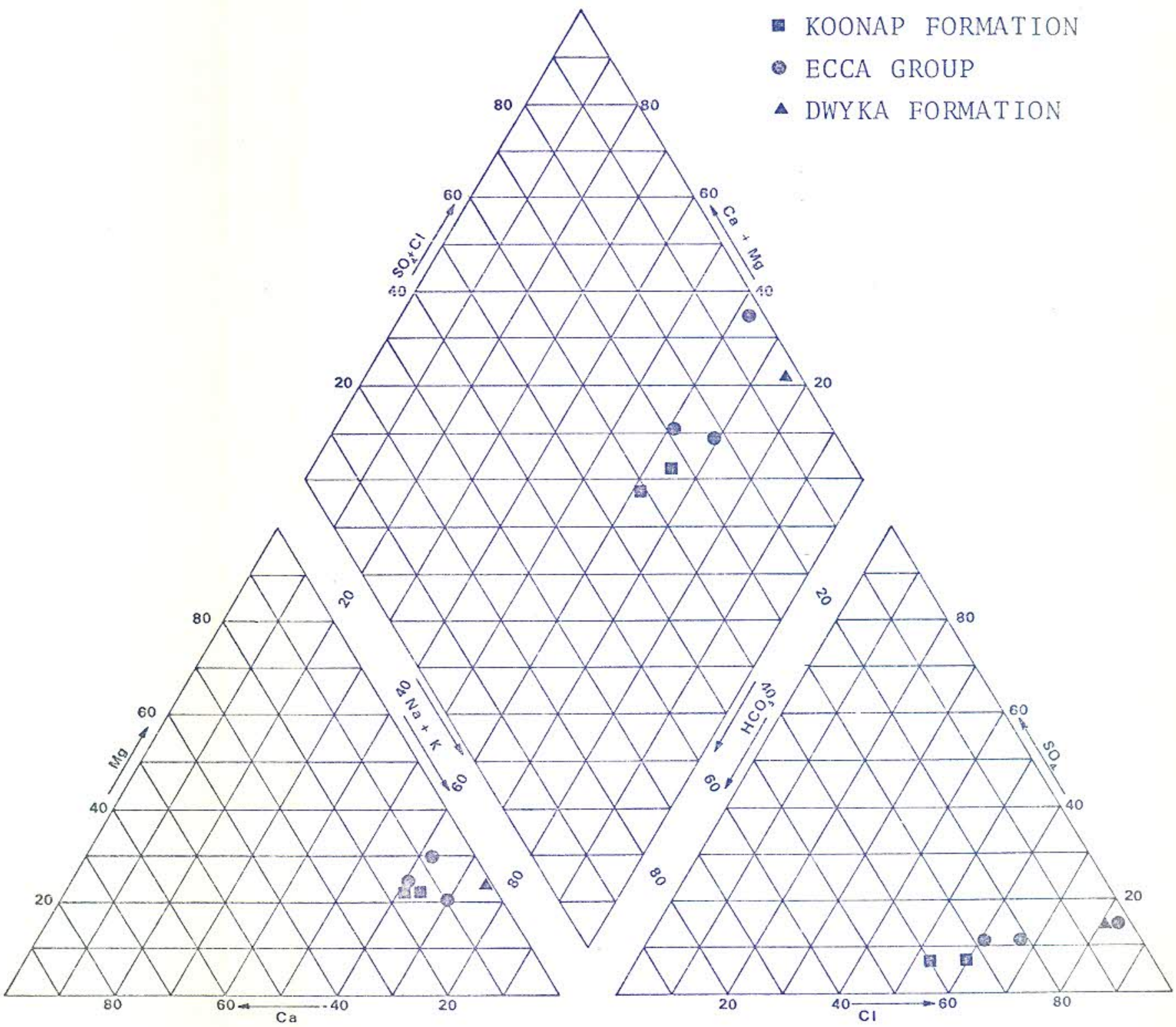


Fig. 5 - 5 Percentages of the major ions (epm) in groundwater associated with the Dwyka Tillite Formation, Ecca Group and Koonap Formation.

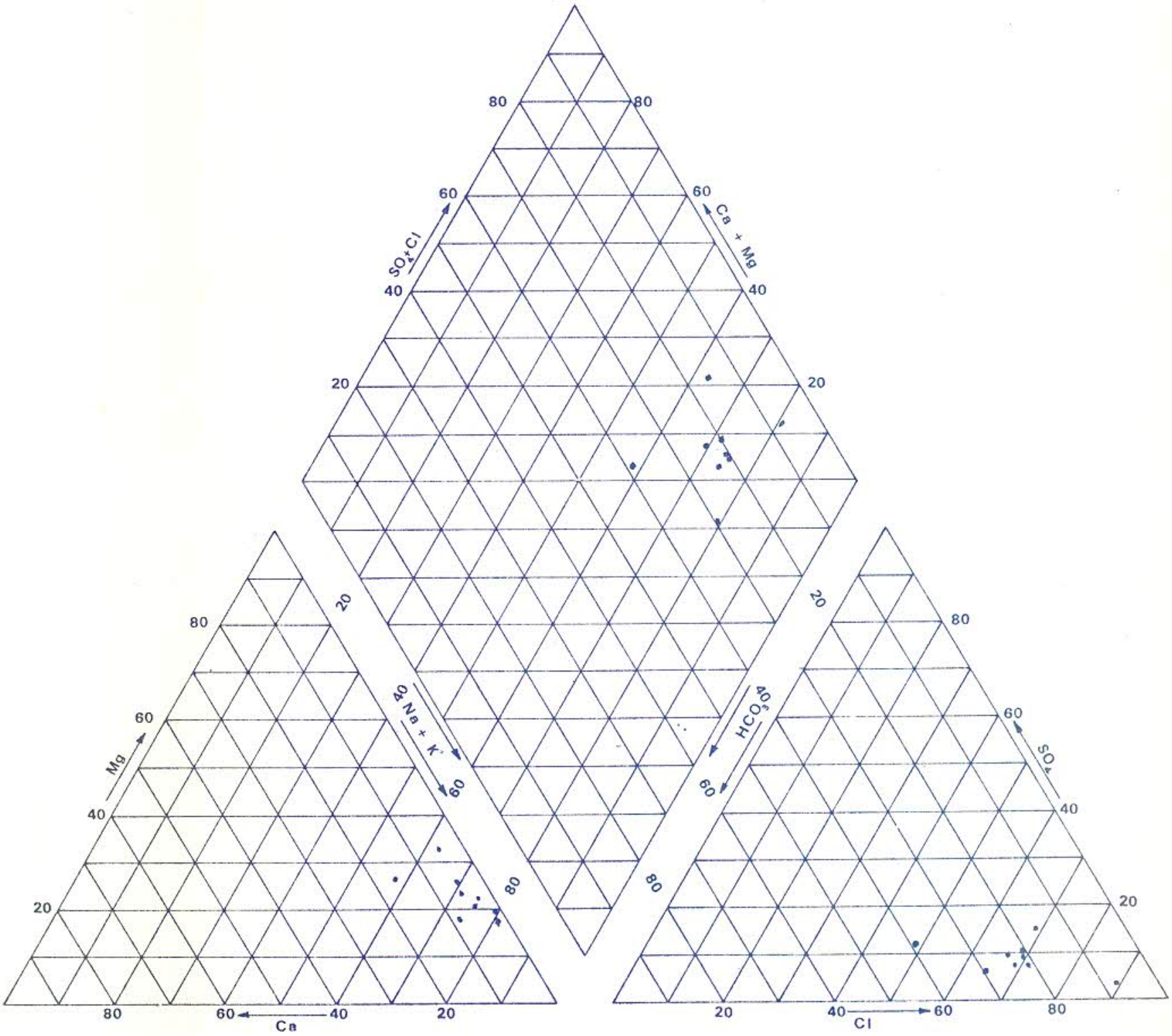


Fig. 5 - 6 Precentages of the major ions (epm) in groundwater associated with the Middleton Formation.

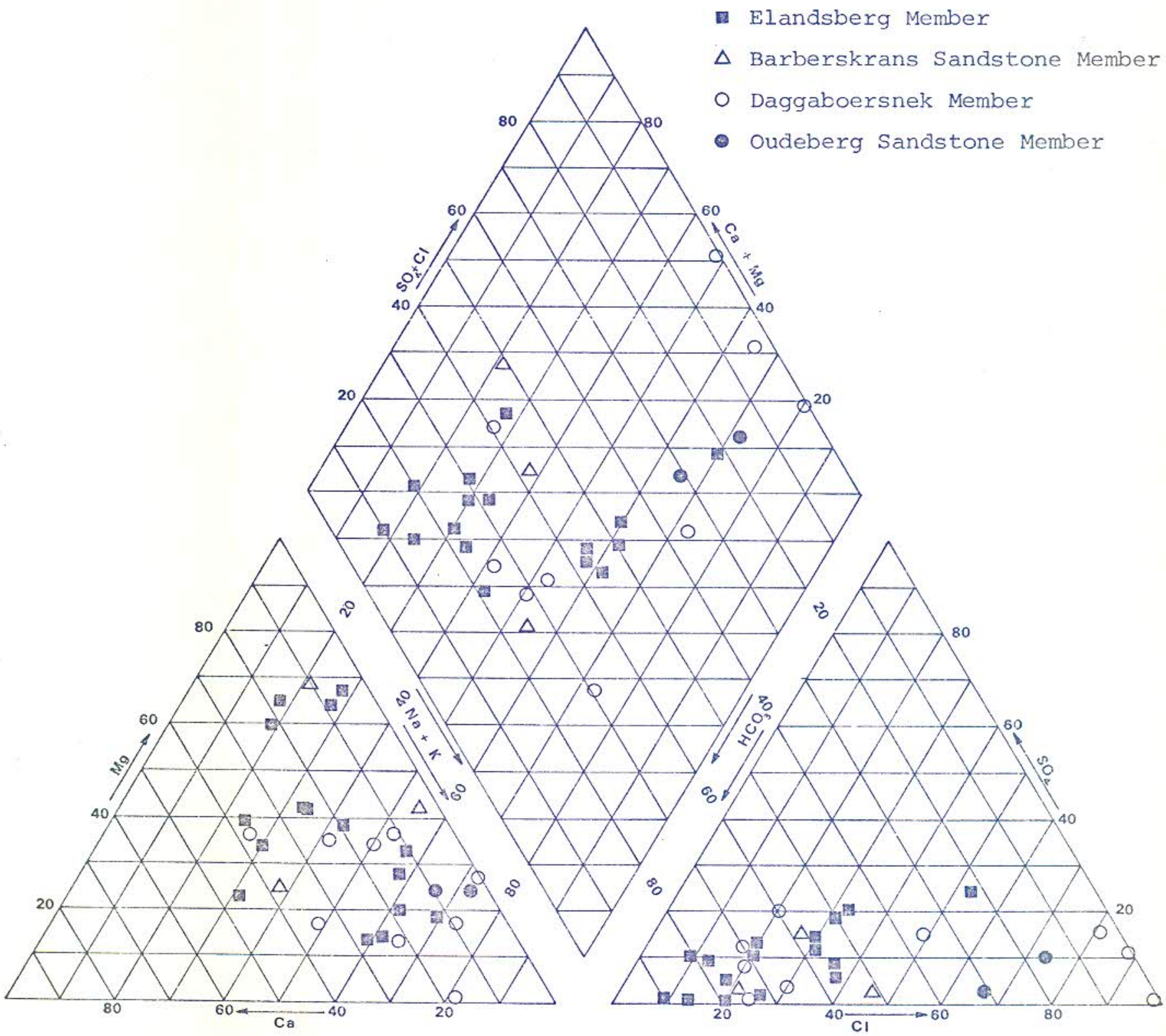


Fig. 5 - 7 Percentages of the major ions (epm) in groundwater associated with the Balfour Formation.

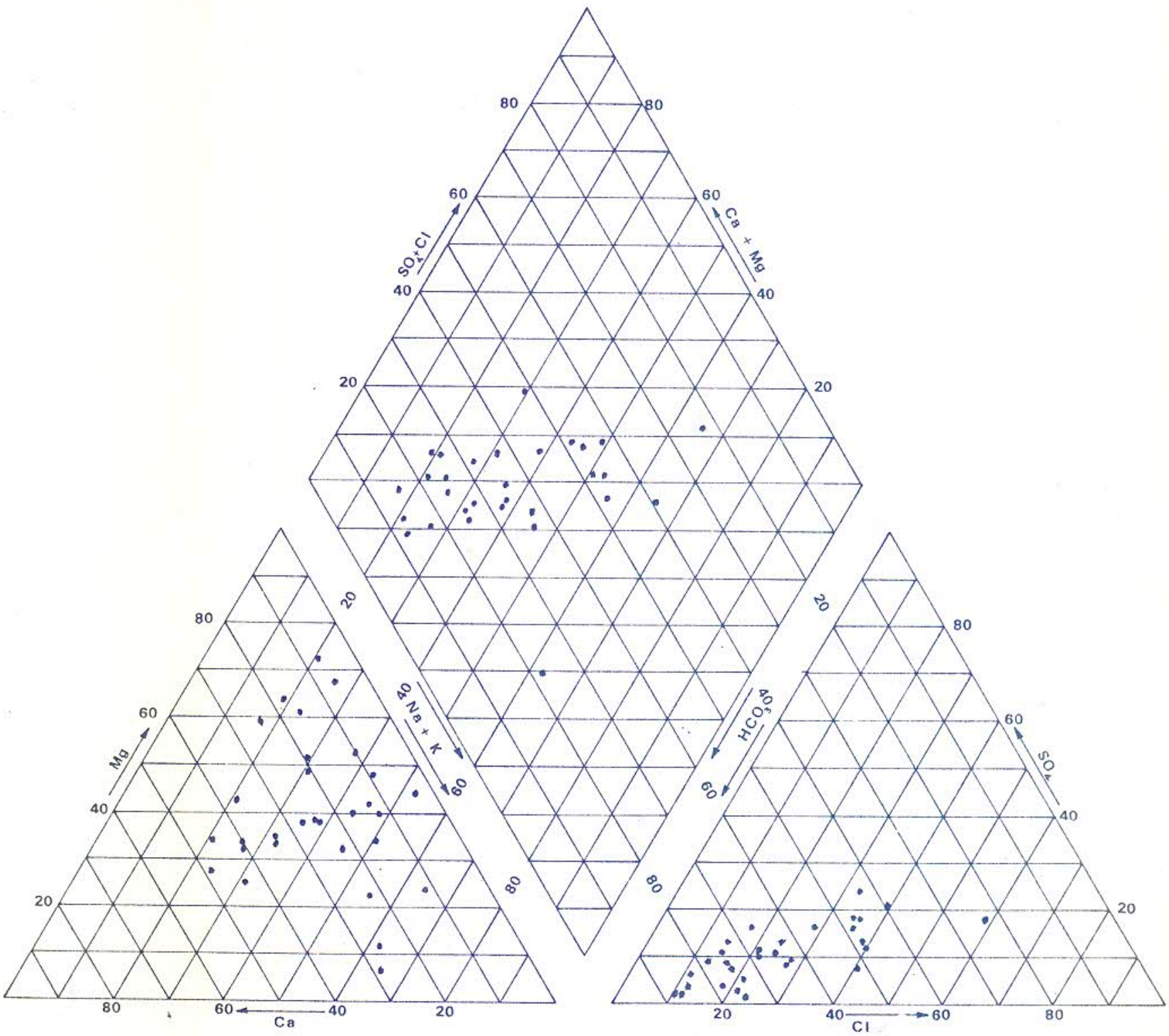


Fig. 5 - 8 Percentages of the major ions (epm) in groundwater associated with the Katberg Formation.

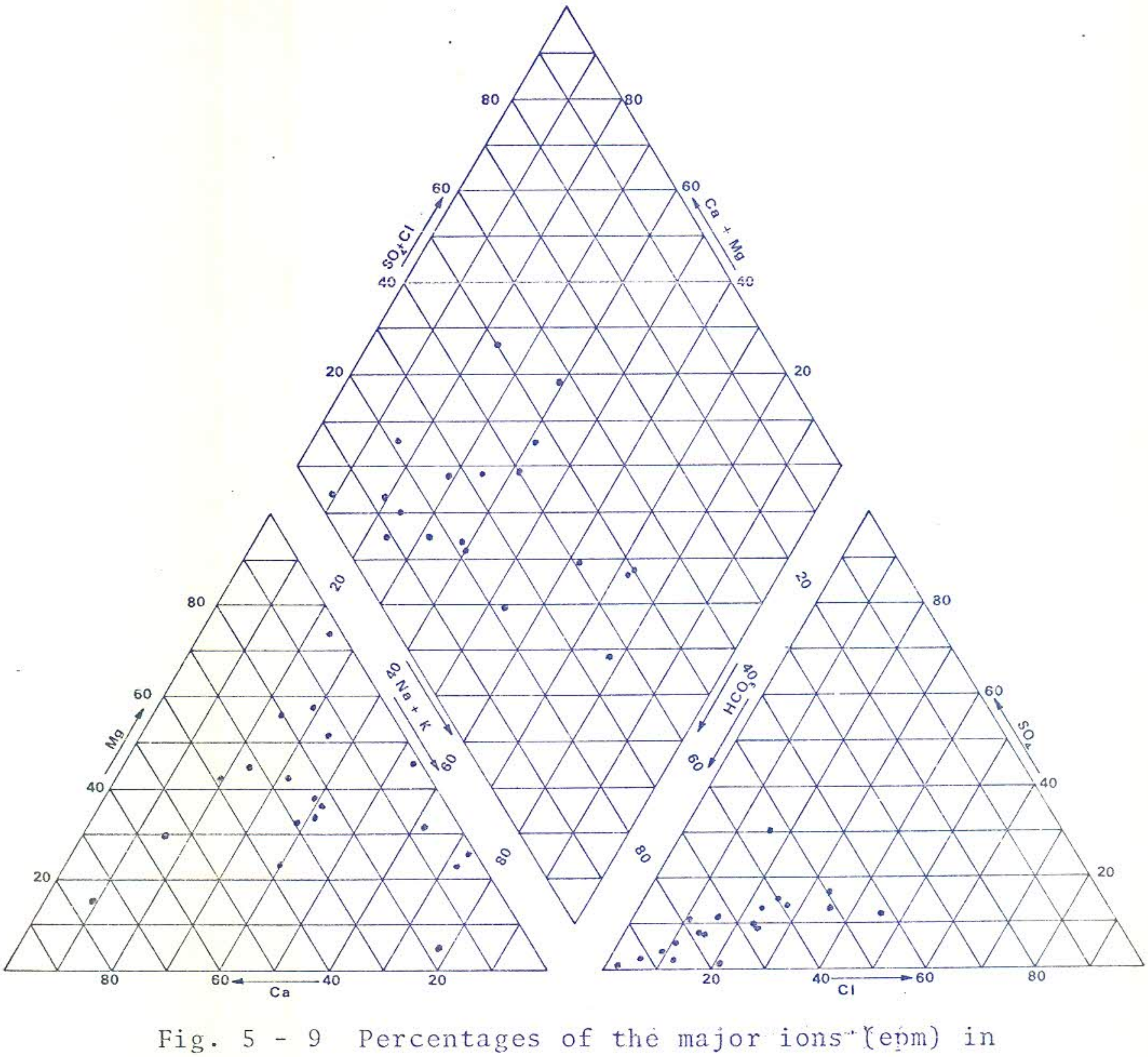


Fig. 5 - 9 Percentages of the major ions (epm) in groundwater associated with the Burgersdorp Formation.

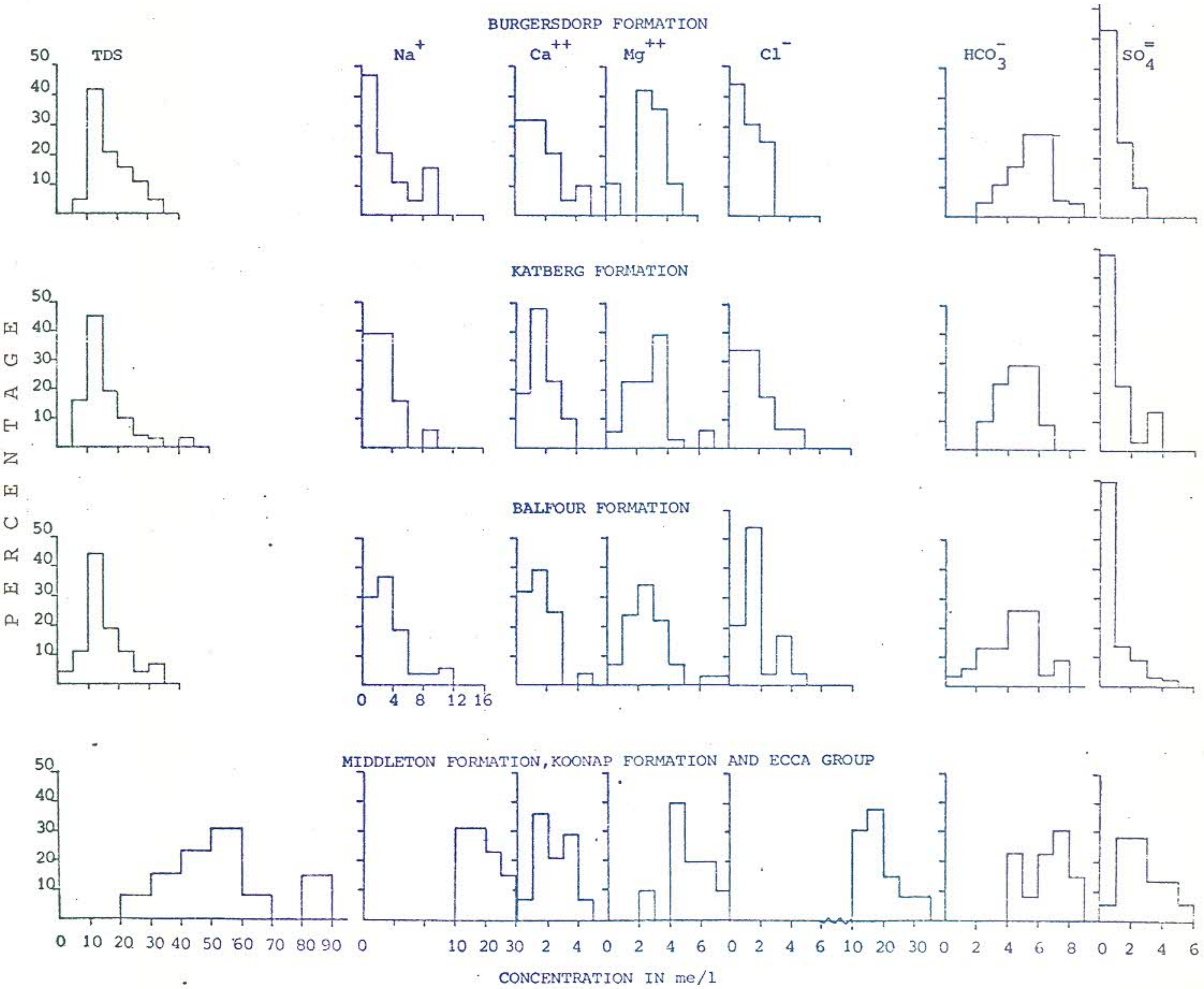


Fig. 5 - 10 Histograms of the major ions in the groundwater from the different stratigraphic units in the Great Fish River Basin.

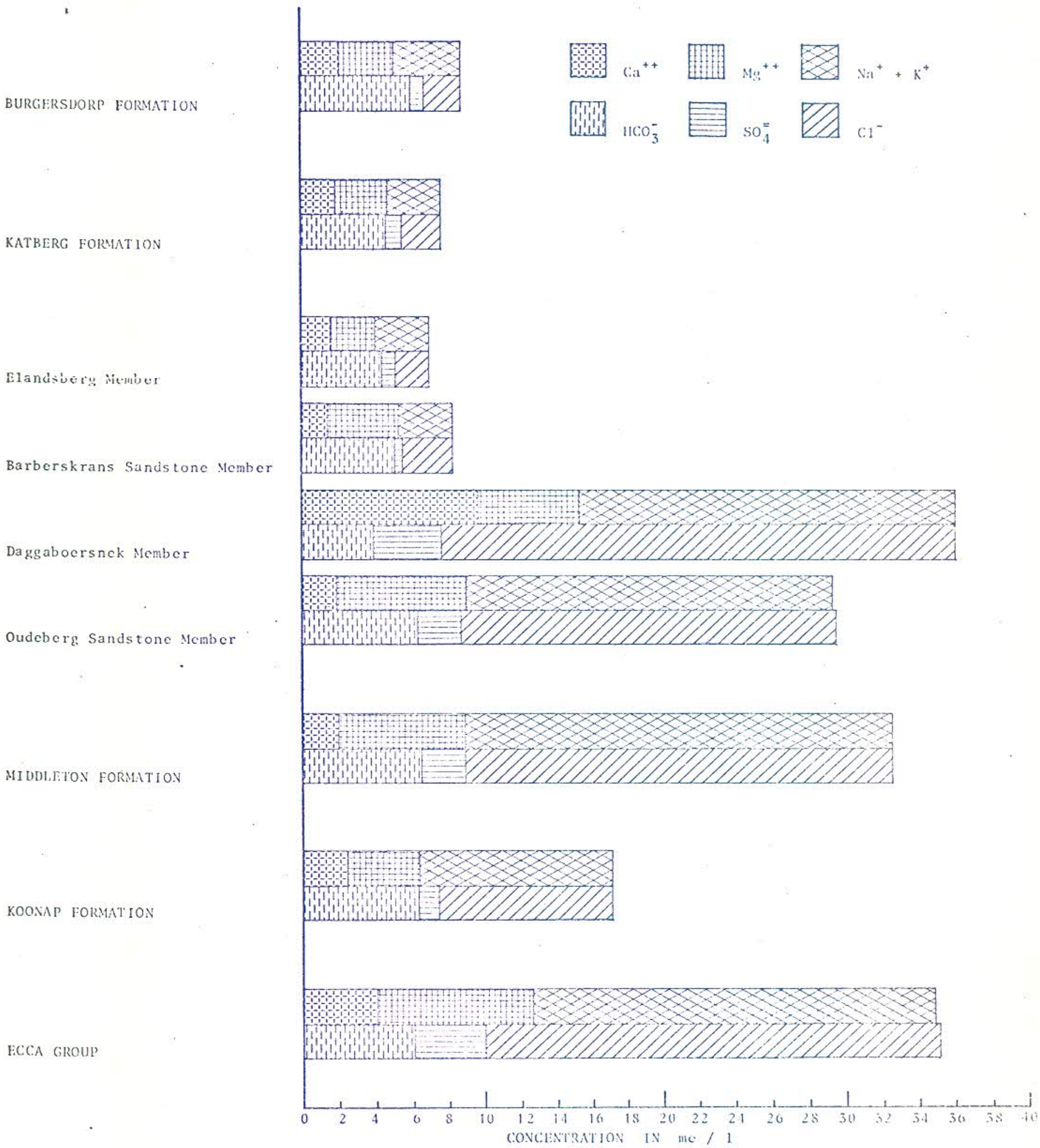


Fig. 5 - 11 Comparison of the mean concetration of the major ions in the groundwater from the various stratigraphic units in the Great Fish River.

to only a narrow zone in the south of the study area, with the result that only one groundwater sample on the present sampling grid was obtained. Boreholes in this formation are furthermore relatively sparse, because of the high salinity of the groundwater encountered therein.

The analytical data of this sample were, however, plotted on Fig. 5 - 5 in order to compare it with the results obtained from groundwater samples of the overlying strata. This sample compares well with results obtained from seawater (Johnson, 1974, p. 1).

The total dissolved solids in the sample are extremely high in comparison with the rest of the groundwater in the area, while Na^+ and Cl^- constitute the main ions. Groundwater quality of this nature may, however, be attributed to several factors such as the mobility of the chemical elements, the hydrodynamic conditions in the subsurface reservoir, as well as the local geochemical environment of the area (Chebotarev, 1955b, p. 201). It is therefore deduced that the following factors are the main cause of the high salinity of the groundwater in the Dwyka Tillite Formation of the area:

(i) The influence of saline connate water of marine origin.

(ii) The poor groundwater drainage in the tillite due to its low porosity and permeability and to the fact that this unit occurs in the Marginal Region which has low relief.

(iii) The weathering of rock fragments in the tillite.

Bond (1946, p. 111) points out that the Dwyka tillite contains more fresh and decomposable primary rock material than any other strata in the Karoo Sequence.

5. 3. 4. 2 Groundwater Quality Of The Ecca Group

As in the case of the Dwyka Formation, outcrops of this unit are limited to only a narrow zone in the south of the study area and only three groundwater samples were obtained.

The salinity of these samples are generally slightly lower than in the previous one, although Na^+ and Cl^- still remain the dominant ions in the water. According to Fig. 5 - 5 the groundwater of the Ecca Group has a higher HCO_3^- - percentage than the groundwater of the underlying Dwyka Formation. Because of the insufficient samples from these strata, this trend cannot be accepted with absolute certainty. By comparing these results with data from the overlying Koonap Formation there, however, appears to be no doubt that the chloride-content decreases in the groundwater toward the upper strata. Fig. 5 - 11 shows that the upward increase in HCO_3^- - percentage is not necessarily due to an increase in the HCO_3^- - concentration, but rather due to a decrease in the Cl^- - concentration.

The significance of this trend is that this is the only bit of proof that Cl^- - filtration occured during the upward migration of the connate water as a result of the compaction of the underlying sediments.

It may therefore be deduced that the groundwater quality of the Ecca Group is largely affected by the mixing of connate water of marine origin. The chloride-content of the water was, however, reduced by means of clay-membrane filtration as a result of compaction during the diagenetic stages.

The influence of topography on the quality of this water must, however, never be ruled out completely, as outcrops of this group occur within the Marginal Region where drainage is poor and stagnant conditions may prevail.

5. 3. 4. 3 Groundwater Quality Of The Koonap Formation

Only two groundwater samples from this formation were analysed.

According to Fig. 5 - 11 the mean concentration of total dissolved substances in the groundwater of this unit appear to be considerably lower than that in the groundwater of the underlying Ecca Group. Should connate water be regarded as the main factor influencing the quality of the groundwater in all three the stratigraphic units mentioned thus far, the lower salinity concentrations in the Koonap Formation must therefore clearly indicate a different depositional environment. This conclusion corresponds with geological observations which were discussed previously.

The relative ratios of the major cations in the groundwater of the Dwyka Formation, the Ecca Group and the Koonap Formation remain constant between the limits: $Na^+ + K^+ =$

60 - 80 percent, Ca^{++} = 0 - 20 percent and Mg^{++} = 20 - 30 percent, although the concentration of these ions vary greatly. This fact indicates that chemical weathering which is mainly responsible for the concentration of cations in the groundwater, is not the main factor influencing the groundwater quality in these units.

5. 3. 4. 4 Groundwater Quality Of The Middleton Formation

The mean salinity concentration of the groundwater in the Middleton Formation is much higher than that in the underlying Koonap Formation (Fig. 5 - 11). This increase is due mainly to an increase in the Na^{+} and Cl^{-}- concentration and compares well with the concentrations in the groundwater of the Ecca Group.

It must, however, be noted that due to insufficient samples from the Koonap Formation no accurate comparison between the two formations can be made. According to Fig. 5 - 6 the groundwater in the Middleton Formation is dominated by Na^{+} and Cl^{-}, which coincides with the "stagnant waters" described by Johnson (1974, p. 4). These samples differ from those of the underlying strata in that slightly higher values of Na^{+} (70 - 80 percent), together with lower values of Ca^{++} (0 - 10 percent) are present. The Mg^{++} percentages remain the same. This phenomenon is attributed to the exchange of Na^{+}, adsorbed by the sediment, for Ca^{++} in the groundwater which is preferentially adsorbed because of its divalent charge.

The anion-percentages of the groundwater in this formation do not vary greatly. There, however, appears to be a higher percentage of Cl^- in these waters than in the underlying Koonap Formation.

It is therefore deduced that factors other than the upward migration and anion-filtration of the connate water, which is assumed for the underlying strata, were responsible for the groundwater quality of this unit. Should the process of Cl^- - filtration have continued from the Koonap to the Middleton Formation, lower chloride percentages than those observed in Fig. 5 - 6 would be encountered. This does not imply that the upward migration of connate water did not continue into the Middleton Formation, but rather that other factors responsible for the concentration of Cl^- become more prominent in this unit.

The higher Cl^- - content of the groundwater of this formation is therefore attributed merely to a change from a bicarbonate character to a chloride character (Johnson, 1974, p. 2) as the water gradually progresses through the basin from the environment of recharge to the environment of discharge. Further indications of this is the decrease in the Ca^{++} in relation to the Na^+ - percentages, as discussed previously.

The influence of topography on the groundwater quality is expected to be uniform throughout the strata which has thus far been discussed, because of the fact that outcrops of the various units occur within the Marginal Region.

All differences encountered in this area may therefore be attributed largely to various geological factors. The exception, however, is the Middleton Formation which lies closest to the Great Escarpment from which percolating groundwaters change in chemical composition as they reach the low-lying parts.

Data obtained from the chemical analyses of the groundwater samples from the Ecca Group and from the Koonap and Middleton Formations were grouped together in order to compile the histograms presented in Fig. 5 - 10. This was done because, no sensible pattern could be obtained for each individual unit (because of the few samples available). According to this figure the salinity concentration of the groundwater of these lower units is much higher than that of the units higher up in the sequence. This is due mainly to higher Na^+ and Cl^--concentrations of which the equivalent concentration fall within the same range, i.e. 10 - 30 me/l. Magnesium concentration remains relatively constant. The higher $SO_4^=$ -concentrations encountered in these strata may be attributed to the oxidation of pyrite which formed during reducing conditions of deposition.

5. 3. 4. 5 Groundwater Quality Of The Balfour Formation

Bond (1946, p. 132) remarked on the high salinity of the groundwater arising in the lower beds of the Beaufort Group and also on the chloride character of these waters. He states, however, that no appreciable difference in quality between the groundwater of the Ecca Group and the Lower Beaufort is present.

According to Fig. 5 - 11 the chloride-rich water with high salinity occurs only as high as the Daggaboersnek Member. The high percentages of sodium and chloride in the groundwater of the Oudeberg Sandstone and Daggaboersnek Members are also illustrated in Fig. 5 - 7. The fact that the samples from the Oudeberg Sandstone Member and the Middleton Formation plot in the same area on the diamond of Fig. 5 - 6 and 5 - 7 proves that the same factors were responsible for the groundwater quality of these waters.

The variation in groundwater quality of the Daggaboersnek Member requires some discussion. According to Fig. 5 - 7 the samples vary from groundwater rich in Ca^{++}, Mg^{++} and HCO_3^- to water containing mainly Na^+ and HCO_3^- and finally to water with a sodium-chloride character. This trend corresponds with the normal chemical cycle of groundwater as described by Johnson (1974, p. 3). It therefore appears that the Daggaboersnek Member contains recent recharged, dynamic underflow, as well as stagnant groundwater. The high salinity of some of these samples, may, however, suggest saline connate water entrapped in the sedimentary rocks of this member. This is in accordance with deposition in a brack, lacustrine environment (Chapter 3).

In the Barberskrans Sandstone Member the groundwater is much less saline than in the underlying strata and contains mainly Mg^{++}, Ca^{++} and HCO_3^- as the major ions (Fig. 5 - 7 and 5 - 11).

Such characteristics, according to Johnson (1974, p. 4), represent recently recharged groundwater. Another observation (Fig. 5 - 11) is that the mean $SO_4^=$ - concentration in this water is much lower than that of the underlying strata. This suggests that the reducing conditions, which are responsible for producing pyrite in the sediments, did not prevail during the deposition of this unit.

The groundwater of the Elandsberg Member is mainly of the magnesium-calcium-bicarbonate type, although a few samples reveal a sodium-bicarbonate character (Fig. 5 - 7). According to Fig. 5 - 11 the mean salinity of this water is also relatively low and one may consider it to also represent recently recharged meteoric water.

The groundwater of the Balfour Formation may therefore be subdivided into two groups:

(i) The highly saline water of the Oudeberg Sandstone and Daggaboersnek Members, with a predominantly sodium-chloride character and corresponding well with the stagnant groundwater of the underlying Middleton Formation.

(ii) The relatively fresh water of the Barberskrans Sandstone and Elandsberg Members, with a predominantly magnesium-calcium-bicarbonate character, representing recently recharged, meteoric water.

Apart from the main division between Cl^- - and HCO_3^- - water (Fig. 5 - 7), there appears to be very little variation in the anion percentages in the groundwater of the Balfour Formation. The cation percentages, however, vary considerably i.e. from Na^{+} to Mg^{++}- rich water, while none of the samples contain more than 50 percent Ca^{++} of the total cations. This observation is contrary to Bond's (1946, p. 135) statement that there is almost invariably more calcium than magnesium in the groundwater of the Lower Beaufort Beds.

If the conclusion of Hem (1970, p. 287) is correct that the cation concentration in groundwaters is controlled mainly by the solution of minerals in rocks, then the great variation in the cation percentages of these waters must be attributed to chemical weathering of the rocks in the area. In the Great Escarpment and around the outer margin of the Headbasin, where most of the outcrops of this formation occur, conditions are favourable for such chemical weathering.

5. 3. 4. 6 Groundwater Quality Of The Katberg Formation

The groundwater of the Katberg Formation, like that of the underlying Barberskrans Sandstone and Elandsberg Members, has a low salinity, with magnesium, calcium and bicarbonate as the predominant ions (Fig. 5 - 8 and 5 - 11). Because of this character the water represents recently recharged, meteoric water.

The anion percentages in this water varies even less than in the underlying units and only two samples contain chloride in quanities exceeding 60 percent of the total anions (Fig.

5 - 8).

There is, however, a large variation in the cation percentages, although Ca^{++} never exceeds 50 percent of the total. Chemical weathering of the rocks may therefore also be attributed to this phenomenon.

5. 3. 4. 7 Groundwater Quality Of The Burgersdorp Formation

Although there appears to be very little difference between the groundwater quality of this formation and that of the underlying Katberg Formation, Fig. 5 - 9 reveals slightly higher bicarbonate percentages. Most of the samples also plot in the lower triangle of the diamond which, according to Johnson (1974, p. 1), suggests groundwater of a dynamic basin environment. Groundwater which plots in the upper triangle represents a static environment.

It is interesting to note that quite a few samples have a $NaHCO_3$-character indicating groundwater of a dynamic underflow regime. Bond (1946, p. 135) also found that many of the groundwater samples from the upper horizons of the Beaufort are rich in Na^+ and HCO_3^- and attributes this fact either to the weathering of feldspar in the sandstone of this strata or to the weathering of dolerite in the area. He, however, eventually disregards the possibility of the weathering of dolerite as the cause of the high $NaHCO_3$-content in the water, on the grounds that groundwater around such bodies contain much higher Mg^{++} than Ca^{++}. According to Bond's (1946) analyses the Ca^{++} in the groundwater is much higher than the Mg^{++} - concentration. In the present study it was, however, found that in most cases magnesium prevailed over calcium and therefore the weathering of dolerite as a source of soluble Na^{+} cannot be ruled out. The lower calcium percentages may be attributed to the fact that this cation is more readily adsorbed by clay minerals in the sediments than are magnesium or sodium and calcium is thus removed from the solution.

One may therefore conclude that the groundwater in the sequence below the Barberskrans Sandstone Member has a predominantly sodium-chloride character together with a high salinity. This phenomenon is caused partly by the upward migration of saline connate water during the compaction of the sediments (vide the groundwaters of the Dwyka Formation, Ecca Group and Koonap Formation), while saline groundwater containing sodium and chloride as the major ions in the Middleton Formation and lower members of the Balfour Formation is the result of normal percolating meteoric waters reaching stagnant conditions. Chemical weathering of the sedimentary rocks in the lower stratigraphic units is at a minimum,hence the small variation in cation percentages.

In the upper parts of the sequence chemical weathering does occur to some extent and therefore produces the great variation in the cation percentages. The groundwater is relatively fresh and contains mainly magnesium, calcium and bicarbonate, which is an indication of recently recharged, meteoric water.

Although most of the above variations can be observed in Fig. 5 - 10, the decreasing trend of $SO_4^=$ in the groundwater from the lower to the upper stratigraphic units needs special mention. This trend is especially noticed in the decrease in maximum concentrations of the above ion, i.e. from 6 to 5 to 4 to 3 me/1. The stratigraphic units below the Katberg Formation were deposited mainly under reducing conditions, thus enabling pyrite to develop, which under later oxidising conditions is able to produce $SO_4^=$. The Katberg and Burgersdorp Formations were, however, deposited under oxidising conditions, which do not permit the formation of pyrite.

It must be stressed, however, that although the geology plays an important role in controling the quality of the groundwater in the study area, this factor still remains subordinate to the influence of topography.

5. 3. 5 Influence Of Time On Groundwater Quality

According to Hem (1970, p. 41) alternating dry and wet seasons favour weathering reactions which produce considerably large amounts of soluble inorganic matter during certain seasons of the year. It has already been pointed out that the rainfall in the Great Fish River Basin occurs mainly over the months December to April and that although evapo-transpiration always exceeds the precipitation, it is highest over the same period (Fig. 2 - 1). Only during exceptional rainfall of 1974 did the precipitation exceed the evapotranspiration over the period from February to the beginning of April.

In order to determine the influence of the seasonal fluctuations on the groundwater quality of the area, borehole B93 (Table 5 - 7) near Baroda was selected. Due to unforeseen circumstances, the samples were collected at rather irregular intervals, which makes any deductions related to seasonal fluctuations almost impossible.

According to Fig. 5 - 12, the salinity of the groundwater does appear to decrease from November to May and then starts to increase slightly. This observation can, however, not be accepted without a certain amount of uncertainty. Fortunately most of the samples were collected over the period when extreme rainfall conditions prevailed over the area and therefore the effect of such conditions on the groundwater quality can be observed. Within $1\frac{1}{2}$ months after the floods of March 1974 the salinity concentration of the groundwater rose to a maximum (Fig. 5 - 12). This increase in salinity was followed by a dilution of the groundwater, which resulted in a decrease in salinity over a further period of one month, after which the normal seasonal fluctuation appears to have occured.

According to Fig. 5 - 12 the increase in salinity appears to have been caused by an increase in the concentration of Na^+ and Cl^- in the groundwater. This increase is believed to be due to the solution of salts from the aerated zone above the groundwater level as the first meteoric waters

TABLE 5 - 7 Change In Chemical Quality Of The Groundwater In Borehole B 93 * With Time

DATE	Na^+	K^+	Ca^{++}	Mg^{++}	Cl'	HCO_3'	SO_4''	TDS
15-11-1973	320	3	150	96	475	355	445	1844
26- 2-1974	260	4	120	70	460	362	280	1556
28- 3-1974	290	0	110	70	430	309	345	1554
11- 4-1974	420	0	110	100	696	272	410	2008
17- 5-1974	215	0	73	51	300	271	200	1110
9- 9-1974	216	4	114	58	370	309	220	1291

*Locality of borehole B93: Latitude S 32° 00,4′ and Longitude E 25° 30,1′

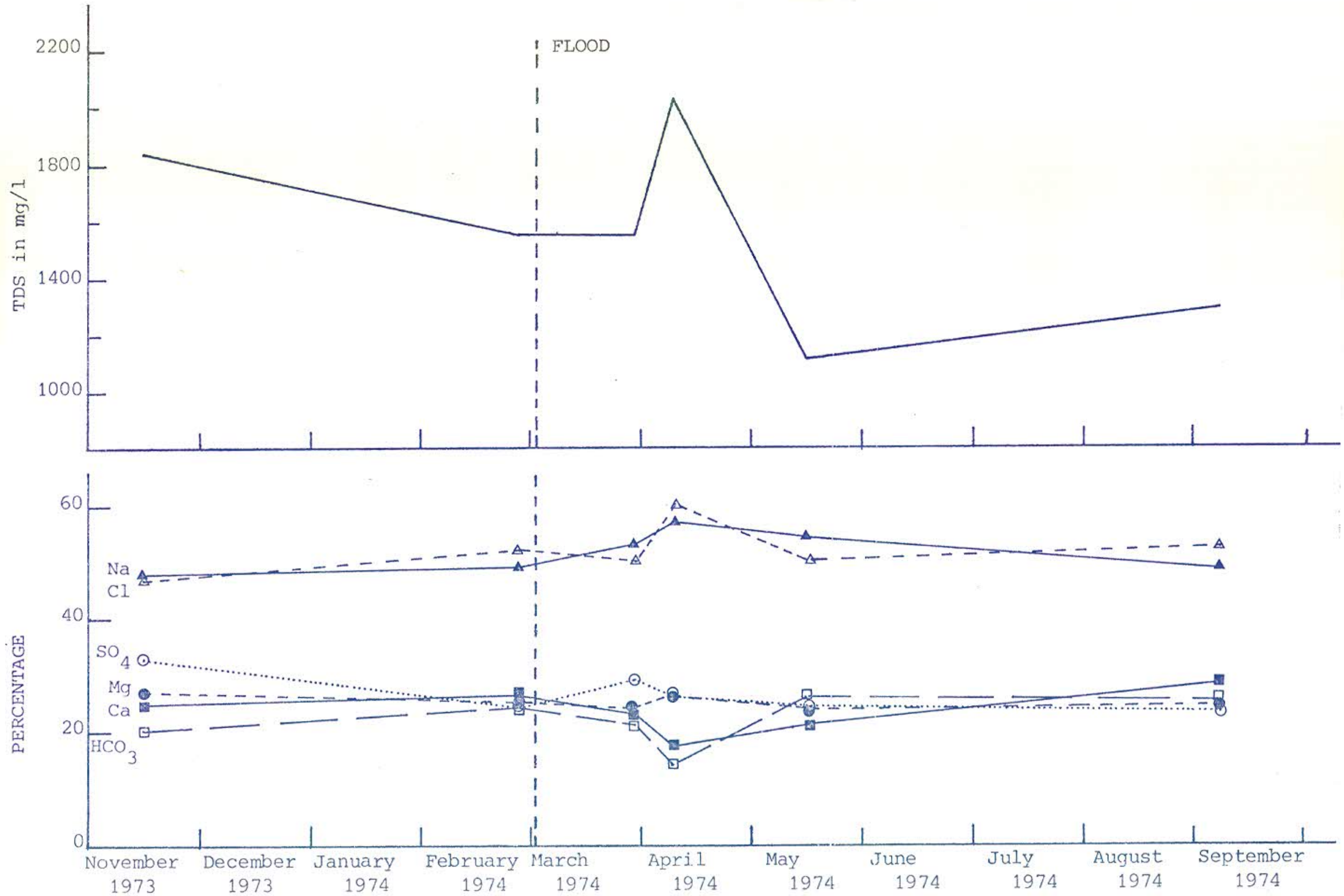


Fig. 5 - 12 Change in chemical quality of the groundwater of borehole B93 with time.

started to migrate down through this zone toward the groundwater level. Prior to this event very little meteoric water reached the groundwater level as this water was soon removed to the atmosphere by high evapotranspiration, resulting in the accumulation of soluble salts in the soil. Once these salts were leached due to high infiltration, relatively fresh meteoric water followed, which in turn was able to dilute the saline groundwater.

The importance of this observation is that, after every period of excessive rainfall, salts in the soil are transported down to the groundwater level. This saline groundwater then reaches the river via the various points of discharge and can therefore present a hazard to the irrigation along the river. It is reported by P.T. Viljoen (personal communication) that soon after the above floods the water in Lake Mentz on the Sundays River became so contaminated with Na^+ and Cl^- that it was unsuitable for irrigating the orange orchards in the Kirkwood area.

6. SURFACE WATER

6. 1. QUALITY OF SEEPAGE WATER IN THE GREAT FISH RIVER

It was pointed out previously that any water applied to an area, either by normal precipitation or by irrigation, will follow the various courses mentioned in the hydrological cycle and that only a small percentage of this water will infiltrate down to the groundwater level, where it may enter the river-system as seepage flow. Viljoen and Liebenberg (1974a) therefore conducted an intensive sampling program at various weirs down the river (Fig. 1 - 1) in order to determine the change in the chemical quality of this seepage water. Their program consisted of taking water samples daily at the various selected weirs in the river and also measuring the rate of flow in the water. Arrangements were also made to have all the water flowing in the river diverted at each weir so that only the water seeping into the river between the two weirs could be sampled. This exercise was conducted over a period of two months during the winter when irrigation ceased.

The results of the program by Viljoen and Liebenberg (1974a) are summarized in Fig. 6 - 1. A gradual increase in the average flow from 4,5 Ml/d to 7 Ml/d can be observed between Katkop and Mortimer weirs. The fairly steep rise at Mortimer may, however, be attributed to an additional seepage flow from the Tarka River.

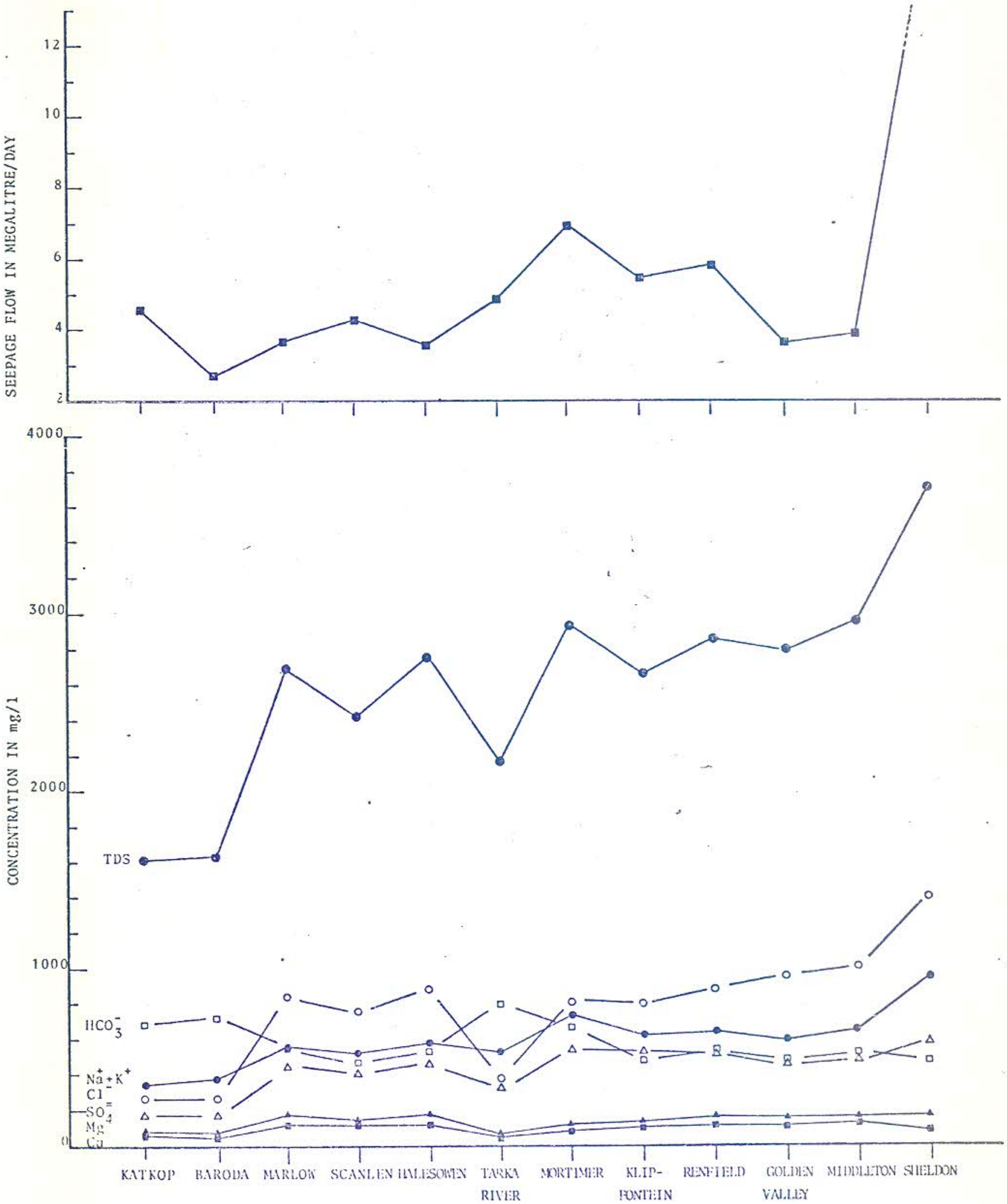


Fig. 6 - 1 Variation in the chemical quality of seepage water at consecutive weirs along the Great Fish River.

From Mortimer to Middleton the flow rate decreases considerably to 3,8 Ml/d and then rises to over 15 Ml/d at Sheldon. The flow rates at the various sampling stations are of great importance because once the concentration of dissolved substances has been determined, the actual salt load carried by the water can be determined.

According to Fig. 6 - 1 the total salinity of the seepage water remains unchanged between the Katkop and Baroda weirs, but then increases rapidly from there to Marlow. From here the salinity increases gradually to Middleton, with only a few fluctuations in between. A major fluctuation is encountered in the Tarka River, which appears to produce seepage water with a lower salinity than that of the Great Fish River. It is also important to note that in spite of the increased flow of seepage water between Middleton and Sheldon, there appears to be a rapid increase in the total salinity of the water. This in turn means that the total salt load over this section of the river is extremely high. Viljoen and Liebenberg (1974a, p. 4) compared the salt load passing each weir with the area of land irrigated above such a weir and found the salt load at the Katkop weir to be 6,3 kg/ha/d whilst the salt load at Sheldon weir is 19,4 kg/ha/d.

Sodium is clearly the cation which constitutes most of the total dissolved solids and shows a gradual increase downstream. It is interesting to note the slight decrease in the Na^+ - concentration in the Tarka River and the sharp increase between Middleton and Sheldon. According to Viljoen and Liebenberg (1974a, p. 4) the sodium-load is

1,3 kg/ha/d at Katkop weir while it increases to 5,0 kg/ha/d at Sheldon. The other two cations Ca^{++} and Mg^{++} play a rather insignificant role in the salinity of the water, although Mg^{++} appears in slightly higher concentrations than the Ca^{++}. This trend was also observed in the groundwater of the area and was discussed in the previous chapter.

The anions play a very significant role in contributing towards the amount of total dissolved solids in the seepage water. Between Katkop and Baroda weirs bicarbonate dominates over chloride, but farther downstream toward Marlow, the water is dominated by Cl^-. In the Tarka River, however, the seepage water contains bicarbonate as the major anion, whilst a marked decrease in the Cl^- can be observed.

Farther downstream a progressive increase in the Cl^--concentration, with a corresponding decrease in the HCO_3^-, is noticed. Of significant importance is the sudden increase in the Cl^-- concentration between Middleton and Sheldon weirs. The $SO_4^=$ - concentration is generally the lowest of the anions, but downstream from Klipfontein weir its concentration is very much the same as that of the bicarbonate.

It may therefore be concluded that a progressive downstream deterioration in the quality of the seepage water in the Great Fish River exists. This deterioration is caused mainly by an increase in the salinity of the water, which in turn is due to a progressive increase in sodium and chloride-concentrations in spite of an increase in the flow-rate of the water. The load of salts carried down the river by seepage water therefore increases progressively downstream and may have an adverse effect on the quality of the irrigation water in the lower areas.

It has already been pointed out that the groundwater becomes progressively saline down the Great Fish River Basin and that this salinity is caused mainly by an increase in the concentration of sodium and chloride. There can, therefore, be little doubt that the quality of the seepage water is closely related to the quality of the groundwater in the area.

Viljoen and Liebenberg (1974a, p. 4) point out that there is some correlation between the amount of seepage-flow at a weir and the irrigated area within the particular section above the weir. The fact that the general nature of the soils in the area does not permit a fast enough infiltration of water before it is returned to the atmosphere by evapotranspiration, makes it difficult to believe that a return-flow of irrigation water seeping through the soil is solely responsible for the seepage-flow in the river.

6. 2. INFLUENCE OF GROUNDWATER ON THE QUALITY OF SEEPAGE WATER IN THE GREAT FISH RIVER

Samples of groundwater at various localities along the Great Fish River were collected in order to determine the relationship between the groundwater quality and the quality of the seepage water. In addition a number of samples from both the Baroda and the Golden Valley Groundwater Compartments (Fig. 5 - 1) were studied. The purpose was to compare the influence of a compartment in the Headbasin with that of a compartment in the Marginal Region on the seepage water.

6. 2. 1 Variation In Groundwater Quality Down The Great Fish River

The mean analyses of a number of groundwater samples taken at various localities down the Great Fish River are presented in Table 6 - 1 and are displayed graphically in Fig. 6 - 2.

A slight decrease in the salinity concentration of the groundwater occurs between Visrivier Station and Baroda. From here to Golden Valley a gradual increase in the salinity can be observed, whilst extremely high salinities are encountered at Halesowen and Cookhouse.

Sodium has the highest concentration of all the major cations encountered in the groundwater. The concentration of this ion shows a similar trend to that of the salinity in the downstream direction. On account of this fact and also because the other two cations Ca^{++} and Mg^{++} occur only in minor concentrations (showing very little variation down the river), one may deduce that the salinity of the groundwater along the river is greatly dependent on the sodium concentration.

As far as the relationship between the anion-concentration and the salinity is concerned, the chloride-concentration

TABLE 6 - 1 Mean Chemical Composition Of Groundwater At Various Localities Along The Great Fish River

LOCALITY	Na^+	K^+	Ca^{++}	Mg^{++}	Cl'	HCO_3'	SO_4''	TDS
Visrivier	215	4	14	33	103	525	72	966
Baroda	110	4	62	46	85	342	96	745
Hales Owen	850	4	265	94	1638	153	447	3451
Mortimer	410	12	52	46	628	244	120	1512
East Poort	398	4	34	111	677	287	202	1713
Cookhouse	813	8	52	114	1322	433	163	2905
Golden Valley	591	6	182	104	1042	418	375	2718

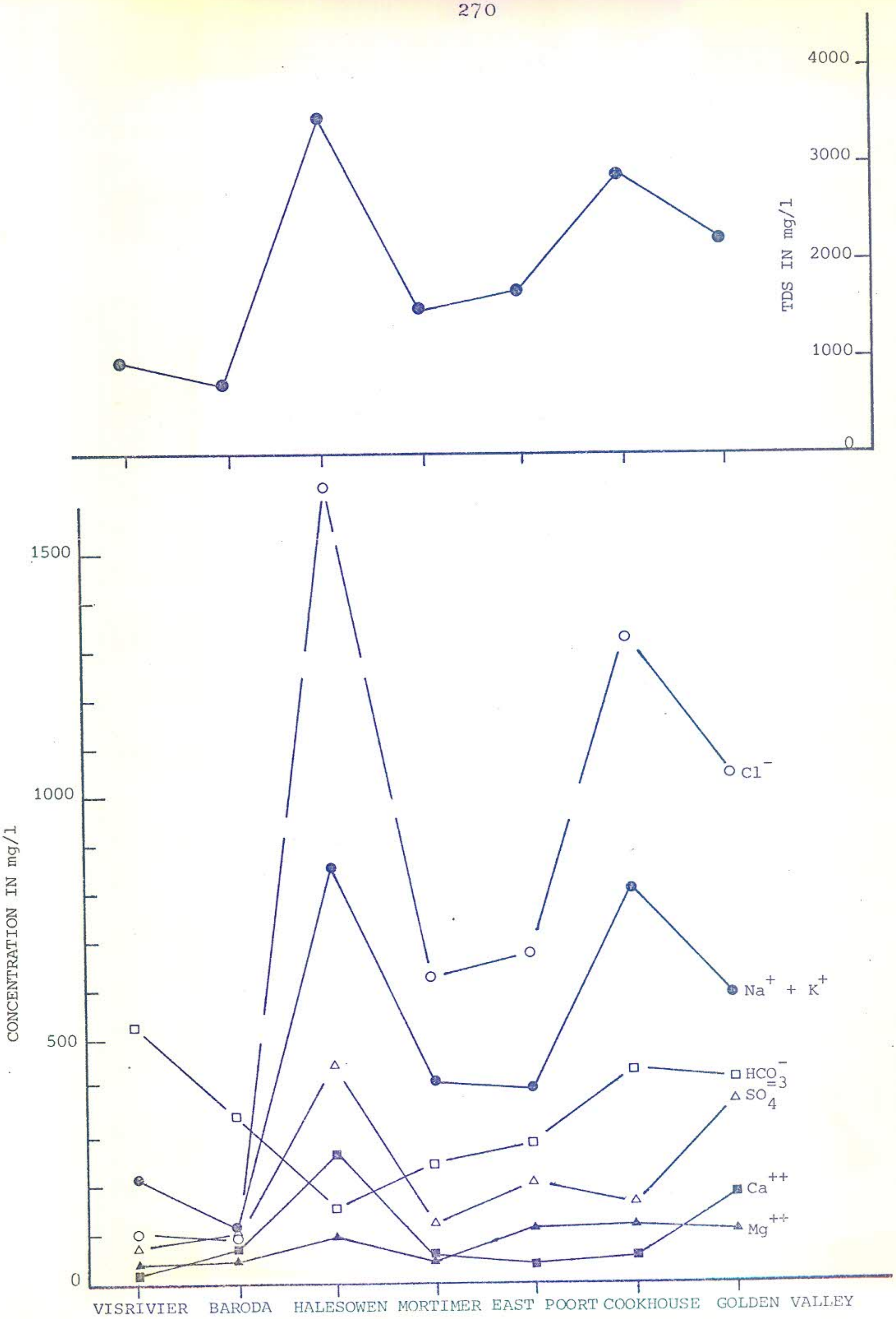


Fig. 6 - 2 Change in groundwater quality down the Great Fish River.

follows the salinity trend even more closely than the sodium. In this case, however, the groundwater between Visrivier Station and Baroda contains a higher concentration of bicarbonate than chloride. This situation soon changes downstream of Baroda, where the water adopts a prominent chloride character and retains it all the way down to Golden Valley.

It is interesting to note that, whereas all the other ions show a distinct increase in concentration from Baroda to Halesowen, the HCO_3^- - concentration shows a marked decrease. This decrease is attributed to the precipitation of $CaCO_3$ in the form of calcrete in the vacinity, as discussed previously. The higher concentration of $SO_4^=$ in the groundwater at this locality may also account for the occurrence of gypsum in the area.

The following similarities between the groundwater quality and the quality of the seepage water down the Great Fish River are observed by comparing Fig. 6 - 1 and 6 - 2:-

(i) Upstream of Baroda both types of water have a distinctly bicarbonate character, which changes to a distinctly chloride character father downstream.

(ii) Upstream of Baroda both types of water have a relatively low salinity, which shows a reasonable increase farther downstream.

(iii) In both types of water sodium is the dominant cation and it's concentration closely follows the trend of the salinity.

(iv) In both types of water the chloride-concentration follows the trend of the salinity even more closely than the sodium-concentration.

Because of these similarities the influence of groundwater on seepage water is clearly illustrated. The seepage is therefore due mainly to groundwater discharge. This does, however, not imply that seepage from the irrigation water, as suggested by Viljoen and Liebenberg (1974a, p. 4), does not occur. In sandy soils near the river direct seepage of the irrigation water to the river is bound to occur, but in the clay-rich soils farther away from the river, the water will have to infiltrate down to the groundwater level before it can ever hope to reach the river. This infiltration rate is extremely low because of the low permeability of the soils, as well as the high evapotranspiration conditions in the area.

6. 2. 2 Influence Of Major Groundwater Compartments On Seepage Water

6. 2. 2. 1 Baroda Compartment

This compartment is the result of a dolerite dyke cutting across the Great Fish River at the sulphur baths, immediately north of Cradock. Although the term "groundwater compartment" is used, one must bare in mind that due to the hydrological features of the area, such a compartment may consist of a number of sub-compartments, each with its own unique chemical character.

A total of 42 groundwater samples were collected in the area between Katkop weir and the sulphur baths, the chemical results of which are presented in Table 6 - 2. Fig. 6 - 3 is a graphical illustration of these results.

The salinity of the samples varies between 10 and 80 me/1, with more than 60 percent having a total salinity of between 10 and 30 me/1. According to the histogram the mean salinity concentration is 29,9 me/1, with a standard deviation of 15,1 me/1.

A very limited variation in the cation prcentages can be observed and most of the samples plot in the area where every cation constitutes less than 50 percent of the total. The result is that calcium and magnesium together exceed the sodium percentage. There are, however, a few samples in which Na^+ constitutes more than 50 percent of the total cations.

The anion percentages show a greater variation than the cations and range mainly from bicarbonate-rich to chloride-rich waters. A slight increase in the sulphate percentage occurs with the increase in the chloride percentage. In the diamond it will be noticed that most of the samples plot either in the Ca^{++}-Mg^{++}- HCO_3^- area, representing recently recharged groundwater, or in the Ca^{++}- Mg^{++}- Cl^- area, representing groundwater of a stagnant environment. Only very few samples have a Na^+-

TABLE 6 - 2 Inorganic Quality Of Groundwater Samples From The Baroda Compartment.

SAMPLE	CO-ORDINATES		CATIONS					ANIONS				TDS
No	Lat. S	Long. E	Na^+ mg/l	K^+ mg/l	Mg^{++} mg/l	Ca^{++} mg/l	Total mg/l	Cl^- mg/l	HCO_3^- mg/l	$SO_4^=$ mg/l	Total mg/l	
B 1	32° 00,6'	25° 30,3'	150	2	35	89	276	220	261	150	631	907
B 2	32° 00,2'	25° 30,6'	100	3	33	39	175	103	262	115	480	655
B 3	32° 01,5'	25° 31,3'	220	1	5	180	406	550	52	120	722	1128
B 4	32° 01,4'	25° 32,1'	340	5	57	70	472	390	461	240	1091	1563
B 5	32° 02,2'	25° 32,8'	510	4	140	110	764	830	359	445	1634	2398
B 7	32° 05,9'	25° 36'	130	3	155	77	365	280	256	120	656	1021
B 8	32° 00,8'	25° 33,2'	90	3	27	65	185	53	400	41	494	679
B 10	32° 00'	25° 30,9'	100	5	30	45	180	97	287	85	469	649
B 23	32° 05,5'	25° 33,8'	110	5	60	66	241	165	393	115	673	914
B 24	32° 03,9'	25° 34,2'	90	5	19	40	154	85	185	120	390	544
B 26	32° 05,2'	25° 35,6'	76	3	92	43	214	77	599	63	739	953
B 27	32° 03,1'	25° 34,6'	94	4	39	48	185	17	509	21	547	732
B 28	32° 01,6'	25° 35'	34	1	17	43	95	8	270	5	283	378
B 29	32° 02,6'	25° 34,2'	110	3	41	45	199	112	210	185	506	706
B 30	32° 00,3'	25° 32,8'	370	5	270	350	995	1160	183	900	1248	2243
B 31	32° 01,7'	25° 30,9'	220	4	81	93	398	395	196	290	881	1279
B 32	32° 02,4'	25° 30,2'	240	2	71	110	423	245	516	290	1051	1474
B 33	32° 02,3'	25° 31,2'	200	3	77	96	376	305	301	265	871	1247
B 34	32° 02,6'	25° 32,9'	54	7	33	59	153	37	373	37	447	600
B 57	32° 06,2'	25° 36,8'	60	2	42	60	164	75	351	39	465	629
B 80	32° 04,8'	25° 31,7'	72	2	32	39	145	45	306	53	404	549
B 81	32° 01,5'	25° 32,7'	250	6	56	87	399	325	393	210	928	1327
B 84	32° 06'	25° 34,9'	110	4	60	90	264	146	484	108	738	1002
B 85	32° 07,3'	25° 32,8'	60	2	50	100	212	112	394	81	587	799
B 86	32° 06,5'	25° 35'	38	3	33	60	134	36	345	27	408	542
B 87	32° 06'	25° 33,2'	52	1	5	27	85	60	92	25	177	262
B 88	32° 06,3'	25° 31'	86	1	39	35	161	90	261	71	422	583
B 89	32° 05,9'	25° 32'	90	3	40	43	176	112	246	86	444	620
B 90	32° 05,2'	25° 32,7'	100	2	25	39	166	105	272	27	404	570
B 91	32° 03,2'	25° 32,1'	180	4	54	46	284	145	439	140	724	1008
B 92	32° 01,1'	25° 32,5'	190	9	48	53	300	155	468	145	768	1068
B 93	32° 00,4'	25° 30,1'	320	3	96	150	569	475	355	445	1275	1844
B 94	32° 00,2'	25° 30,3'	125	3	42	77	247	150	283	170	603	850
K 13	31° 58,4'	25° 30,7'	200	5	80	91	376	290	298	305	893	1269
K 31	31° 58,7'	25° 32,7'	175	3	52	120	350	390	216	195	801	1151
K 34	31° 59,8'	25° 33,1'	180	3	77	140	400	390	228	350	968	1368
KA	31° 58,7'	25° 34,7'	47	5	38	36	126	50	283	36	369	495
KB	31° 58,9'	25° 32,8'	350	2	17	280	649	920	65	135	1120	1769
KC	31° 59,5'	25° 33,3'	280	8	76	110	474	400	229	430	1059	1533
KD	31° 59,7'	25° 32,1'	1200	34	530	490	2254	3100	334	1500	4934	7188
KE	31° 58,9'	25° 31,5'	300	2	21	18	341	375	440	320	1135	1676
KG	31° 58,9'	25° 30,4'	138	4	40	77	259	115	409	140	664	923
KH	31° 58,8'	25° 30,4'	110	4	38	48	200	90	323	110	523	723
KI	31° 58,7'	25° 30,6'	150	5	71	92	318	166	482	195	843	1161

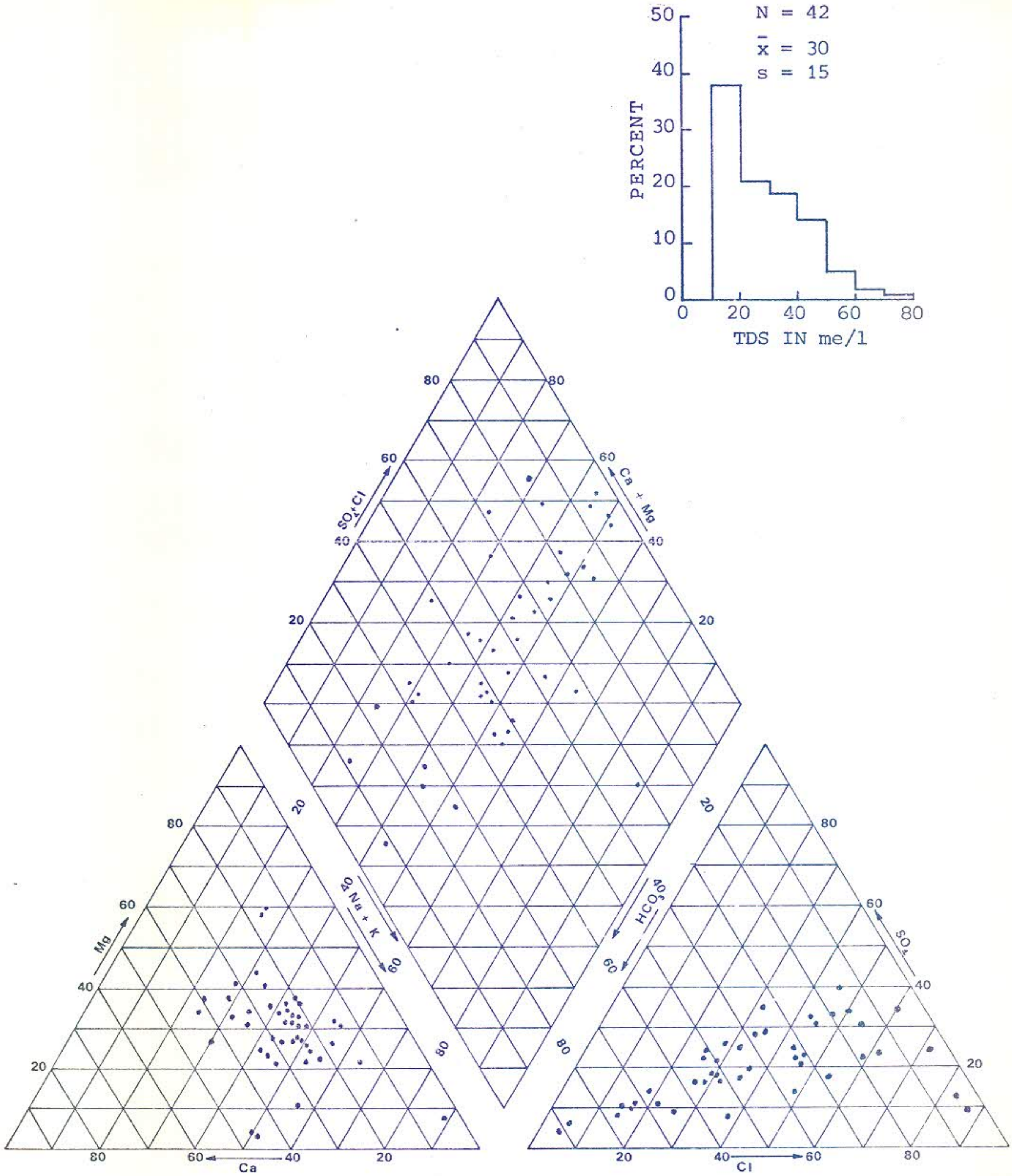


Fig. 6 - 3 Graphic presentation of the groundwater quality in the Baroda Compartment showing a histogram of the salinity distribution (epm) as well as the percentages (epm) of the major ions.

HCO_3^- or a Na^+ - Cl^- - character. The change from a bicarbonate to a chloride - character must explain the same trend observed in Fig. 6 - 1 and 6 - 2.

In Fig. 6 - 1 a gradual increase in the seepage-flow is recorded between the Baroda and the Scanlen weir, the former occuring upstream and the latter downstream of the compartment. The barrier dyke responsible for forming the compartment lies a few kilometres downstream of the Marlow weir and therefore the increased flow may be attributed to groundwater seepage from the compartment into the river.

6. 2. 2. 2 Golden Valley Compartment

This compartment is caused by the transgressive dolerite sill upstream of Middleton, which is the most southerly occurrence of dolerite in the area. According to Fig. 5 - 1 this compartment is much wider than the Baroda Compartment and extends as far north as Cookhouse. Far more groundwater is therefore expected in this compartment than in the previous one.

The chemical analyses of 48 groundwater samples of the Golden Valley Compartment are presented in Table 6 - 3, whilst Fig. 6 - 4 illustrates the variation in the salinity as well as the percentages of the various ions in the samples.

Fig. 6 - 4 illustrates that two different types of groundwater are encountered in the compartment. Exactly one third

TABLE 6 - 3 Inorganic Quality Of Groundwater Samples From The Golden Valley Compartment.

SAMPLE	CO-ORDINATES		CATIONS					ANIONS				TDS
No	Lat. S	Long. E	Na^{+} mg/l	K^{+} mg/l	Mg^{++} mg/l	Ca^{++} mg/l	Total mg/l	Cl^{-} mg/l	$SO_4^{=}$ mg/l	HCO_3^{-} mg/l	Total mg/l	
G - 8	32° 55'	25° 47,4'	232	9	83	92	416	390	61	550	1001	1417
-10	32° 54,2'	25° 44,2'	379	8	130	148	665	700	210	607	1517	2182
-11	32° 47.2	25° 47,8'	290	35	200	360	885	1250	390	194	1834	2719
-12	32° 45,6'	25° 46,9'	185	7	75	154	421	400	60	568	1028	1449
-13	32° 45,7'	25° 48,8'	785	5	205	242	1237	1340	670	560	2570	3807
-14	32° 46,9'	25° 50,2'	571	5	150	270	996	1180	360	506	2046	3042
-15	32° 47,7'	25° 53,4'	369	2	5	20	396	530	44	112	686	1082
-31	32° 47,6'	25° 47,8'	379	37	223	440	1079	1480	480	241	2201	3280
-32	32° 48,3'	25° 48,1'	1580	27	200	135	1942	2280	680	752	3712	5654
-33	32° 48,9'	25° 48,9'	561	32	223	238	1054	1300	490	434	2224	3278
-34	32° 49,4'	25° 48,8'	790	3	29	100	922	1130	420	37	1587	2509
-35	32° 48,8'	25° 47,9'	996	41	335	360	1732	2200	820	489	3509	5241
-36	32° 49,5'	25° 48,5'	686	3	6	25	720	880	300	76	1256	1976
-37	32° 49,2'	25° 48,2'	259	2	1	3	265	350	41	82	473	738
-38	32° 48,7'	25° 47,7'	849	16	205	295	1365	1700	760	256	2716	4081
-39	32° 47,9'	25° 47,0'	600	19	158	360	1137	1460	420	383	2263	3400
-40	32° 47,7'	25° 45,5'	179	6	48	108	341	250	51	554	855	1196
-42	32° 49,4'	25° 46,3'	523	39	87	161	810	920	820	355	1555	2365
-43	32° 50,6'	25° 46,9'	292	5	78	161	536	520	220	446	1186	1722
-45	32° 46,3'	25° 48,7'	921	6	172	189	1288	1380	660	644	2684	3972
-46	32° 46,4'	25° 48,7	880	6	221	293	1400	1600	620	749	2969	4369
-47	32° 47,9'	25° 46,0'	325	3	27	45	400	218	172	539	929	1329
-48	32° 47,7'	25° 46,4'	453	14	175	350	992	1400	280	419	2099	3091
-49	32° 49,1'	25° 46,0'	463	78	238	400	1179	1320	590	557	2467	3646
-50	32° 48,9'	25° 45,8'	343	64	150	305	862	980	380	412	1772	2634
-51	32° 50,1'	25° 45,7'	207	5	68	142	422	390	100	482	972	1394
-52	32° 50,4'	25° 46,2'	240	6	58	118	422	370	120	463	953	1375
-53	32° 47,5'	25° 48,0'	344	40	146	278	808	960	380	377	1717	2525
-54	32° 49,2'	25° 48,8'	532	7	133	315	987	1200	380	418	1998	2985
-55	32° 48,4'	25° 48,8'	590	13	167	335	1105	1200	570	539	2309	3414
-56	32° 48,6'	25° 48,9'	489	18	140	300	947	1290	230	351	1871	2818
-57	32° 48,8'	25° 48,9'	428	17	149	255	849	840	440	640	1920	2769
-58	32° 48,9'	25° 49,0'	539	16	83	146	784	450	250	291	991	1775
-60	32° 47,9'	25° 47,4'	515	31	177	330	1053	1640	260	228	2128	3181
-61	32° 48,7'	25° 48,8'	760	37	197	213	1207	1440	460	549	2449	3656
-63	32° 49,7'	25° 47,9'	561	24	290	485	1360	1960	460	662	3082	4442
-67	32° 52,3'	25° 49,3'	694	8	150	196	1048	1040	490	702	2232	3280
-68	32° 51,9'	25° 49,0'	541	4	12	41	598	860	25	67	952	1550
-70	32° 52,9'	25° 49,2'	828	13	128	219	1188	1280	500	575	2355	3543
-71	32° 52,4'	25° 49,8'	500	7	140	233	880	950	430	528	1908	2788
-73	32° 46,4'	25° 49,3	718	8	120	87	933	840	420	872	2132	3065
-74	32° 46,5'	25° 50,0'	569	3	85	193	850	1100	220	412	1732	2582
-76	32° 48,9'	25° 50,7'	373	11	80	135	601	580	150	607	1337	1938
-77	32° 50,2'	25° 50,3'	185	28	50	95	358	295	65	434	794	1152
-78	32° 51,0'	25° 47,9'	501	13	158	255	937	1000	530	461	1991	2728
-79	32° 50,2'	25° 48,8'	678	4	200	184	1066	1260	450	546	2256	3322
-81	32° 50,1'	25° 48,0'	710	25	265	480	1480	2000	610	460	3070	4550
-82	32° 48,3'	25° 45,4'	330	29	65	113	537	380	225	601	1206	1743

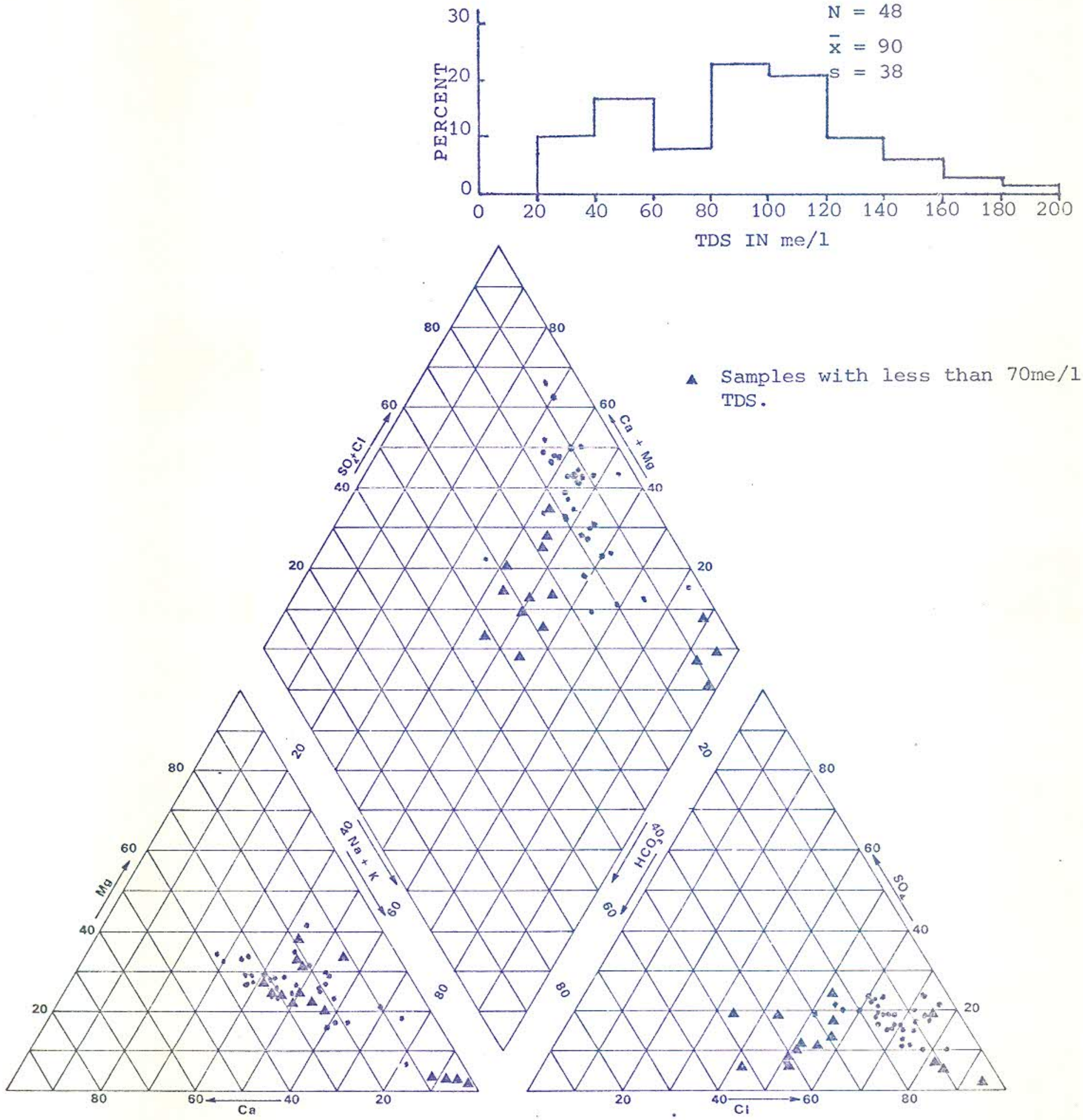


Fig. 6 - 4 Graphic presentation of the groundwater quality of the Golden Valley Compartment showing a histogram of the salinity distribution (epm) as well as the percentages (epm) of the major ions.

of the samples have a salinity of between 20 - 70 me/l, whilst the rest have salinities varying between 70 - 200 me/l. The groundwater with the lower salinity occurs along the outer perimeter of the compartment, whilst the more saline waters are encountered toward the centre. According to the histogram in Fig. 6 - 4 the mean salinity of the groundwater is 90/me/l, with a standard deviation of 38/me/l, which means that the groundwater in the Golden Valley compartment is distinctly more saline than that of the Baroda Compartment.

Whereas the cation-percentages of the groundwater in the Baroda Compartment show very little variation, those of the Golden Valley compartment have an almost linear trend, changing from sodium-rich waters to sodium-poor waters. In spite of the decreasing trend in the Na^{+}- percentage, neither Ca^{++} and Mg^{++} exceed 40 percent of the total cations (Fig. 6 - 1).

The anion-percentages, however, show less variation than in the Baroda Compartment. Most of the samples have chloride-percentages exceeding 50 percent of the total anions. A slight variation in the sulphate-percentages can also be discerned.

The samples with the lower salinity plot in two areas on the diamond (Fig. 6 - 4) which clearly distinguish them from those with a higher salinity. Four of these samples plot in the area dominated by $Na^{+} + Cl^{-}$ thus representing very old water, whilst the rest plot closer to the centre of the diagram, suggesting groundwater of various origin.

The more saline waters vary from $Na^+ + Cl^-$ to $Ca^{++} + Mg^{++} + Cl^-$ - rich, representing groundwater from a stagnant environment. It is, however, interesting to note that the groundwater with the low salinity generally has a higher HCO_3^- - percentage.

According to Fig. 1 - 1 and Fig. 5 - 1 the Middleton weir is located near the upstream side of the Golden Valley Compartment, whilst the Sheldon weir is located at the Sheldon bridge, downstream of the dolerite barrier. The increase in seepage-flow between the Middleton and Sheldon weirs must therefore be attributed to a high seepage into the river from the Golden Valley Compartment.

The groundwater compartments in the basin therefore appear to have a great influence on the chemical quality of the seepage water in the river, as well as on the amount of seepage-flow. The seepage water will therefore have a vital influence on the irrigation water flowing down the river; the higher the seepage-flow and the higher the salinity, the greater will the effect be on the irrigation water.

6. 3. INFLUENCE OF SEEPAGE WATER ON IRRIGATION WATER

In order to determine the influence of seepage water on irrigation water, Viljoen and Liebenberg (1974b), during an irrigation lead (where 1377 Ml of water were released from the Grass Ridge Dam), sampled the water and measured the flow at the outlet of the dam, as well as at various downstream-weirs at hourly intervals. As this lead was meant for the Grass Ridge area (Table 1 - 2), the survey was conducted only as far as the Marlow weir. The total load of soluble salts released from the dam, as well as the load reaching each weir and diverted on to each irrigation area, could hereby be determined. Another great advantage of the hourly sampling is that the variation in the salt load with time could be determined at each weir. The results of this investigation are summarized in Fig. 6 - 5 and 6 - 6.

According to Fig. 6 - 5 the salt load increases by 387,8 tons from Grass Ridge Dam to Marlow weir if no withdrawl of salts occur over this section of the river. Viljoen and Liebenberg (1974b, p. 12), however, give the amount of withdrawl of salts at each weir. By comparing the amount of salts withdrawn at each weir with the area irrigated from the weir, it was found that during the lead in question 193 kg/ha of salts were withdrawn at Katkop weir, whilst 854 kg/ha were withdrawn at Marlow weir. Although the area irrigated from the Katkop weir is 1,6 times larger than that irrigated from the Marlow weir, the salt load applied to these areas differs by a factor of 4, 4, which indicates an increase in the salt load applied to irrigation lands down the river.

A greater increase in the total salt load is recorded between Katkop and Marlow weir. The former is located upstream of the Baroda Groundwater Compartment, whilst the latter is located farther downstream, near the dolerite barrier which

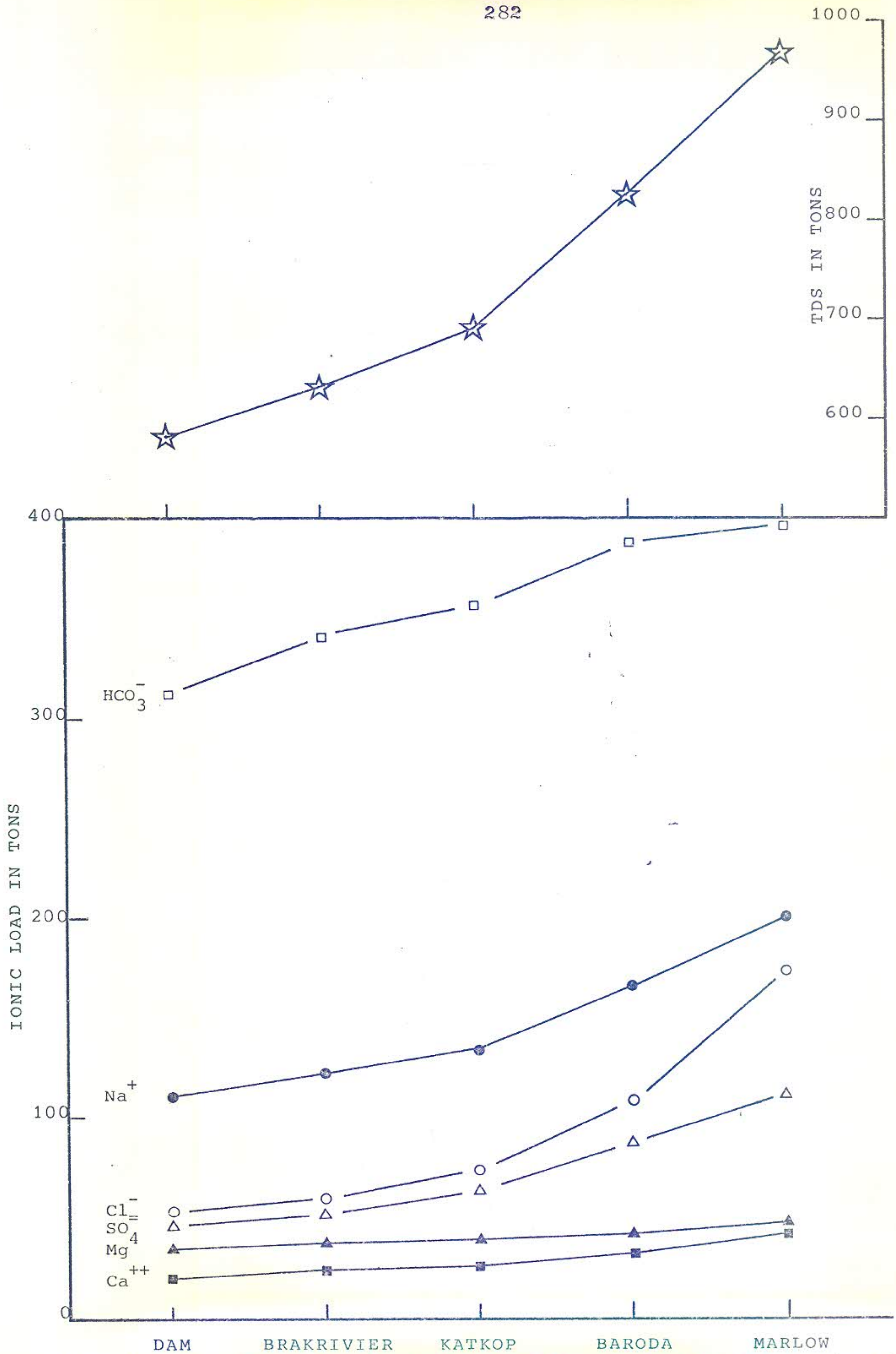


Fig. 6 - 5 Salt load contributed by the various ions to irrigation water at consecutive weirs down the Great Fish River after a release of 1377 Ml from Grassridge Dam.

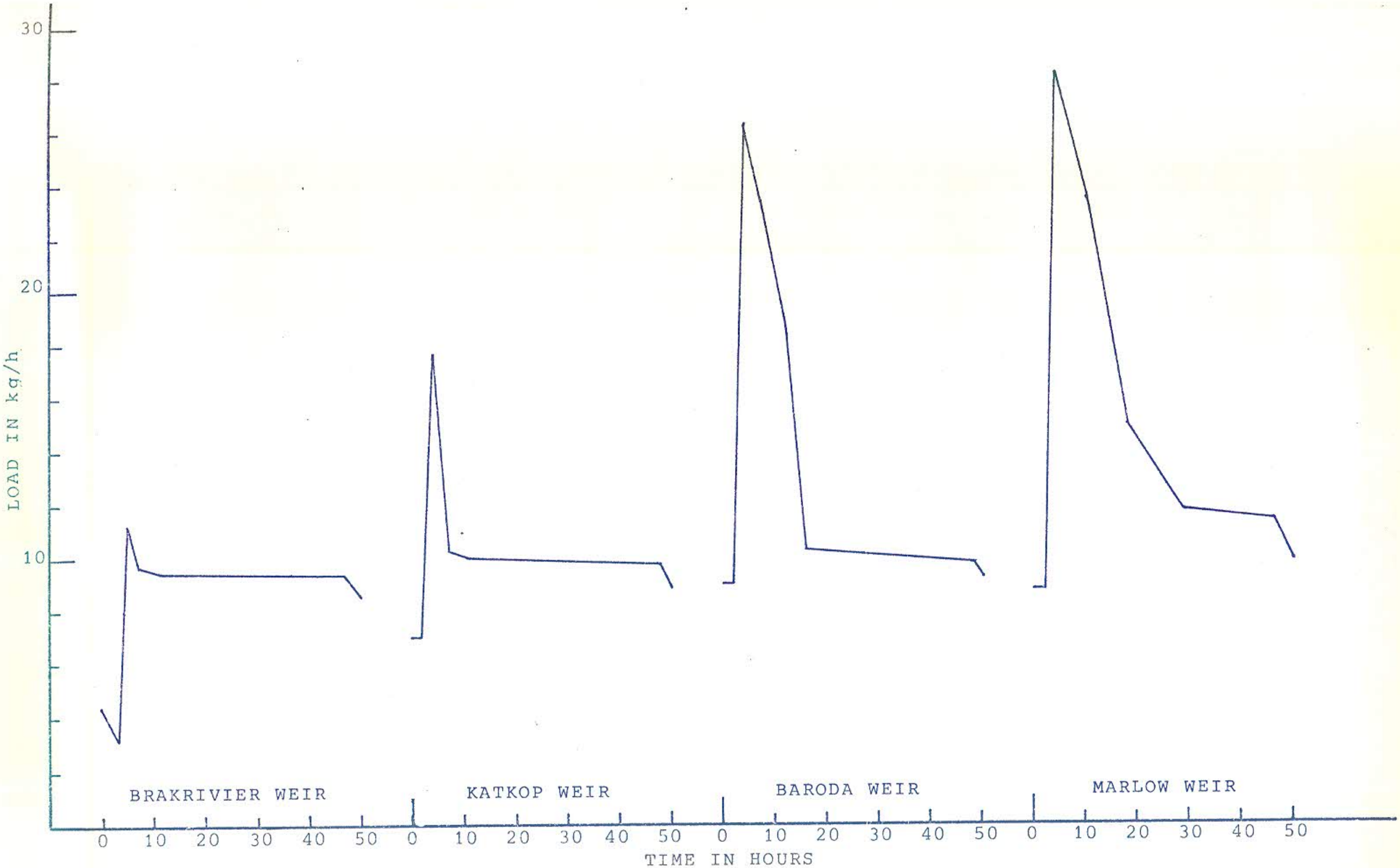


Fig. 6 - 6 Fluctuation in total salt loads at the various weirs from the time of arrival of the irrigation water until stability is reached. Experiment conducted after a release of 1377 M/l from Grassridge Dam.

is responsible for this compartment. It therefore appears that this compartment has a definite influence on the quality of the irrigation water in the area.

Bicarbonate constitutes the greatest part of the total salt load and increases by 83,5 tons from the dam to Marlow weir (Viljoen and Liebenberg, 1974b, p. 10). Chloride and sodium, however, respectively increase over the same area by 120,6 and 87 tons, while the sulphate increases by 63,2 tons.

The increase in magnesium (20,2 tons) and calcium (12,8 tons) has a very minor influence on the variation in the salt load of the water.

Fig. 6 - 6 shows the variation in the salt load with time at the different weirs during this particular irrigation lead. It is interesting to note that the first irrigation water to reach a particular weir has a high salinity and this prevails for some time before decreasing to a normal value. This phenomenon is attributed to the accumulation of saline water in pools along the river and to the precipitation of salts in the river bed (Fig. 6 - 7).

The maximum load of dissolved salts, as well as the time over which the head of saline water prevails at each weir, increases downstream (Fig. 6 - 6) and is in accordance with the increase in the total load of salts reaching the consecutive weirs. Once the head of saline water has passed the weir, only the seepage water can influence the

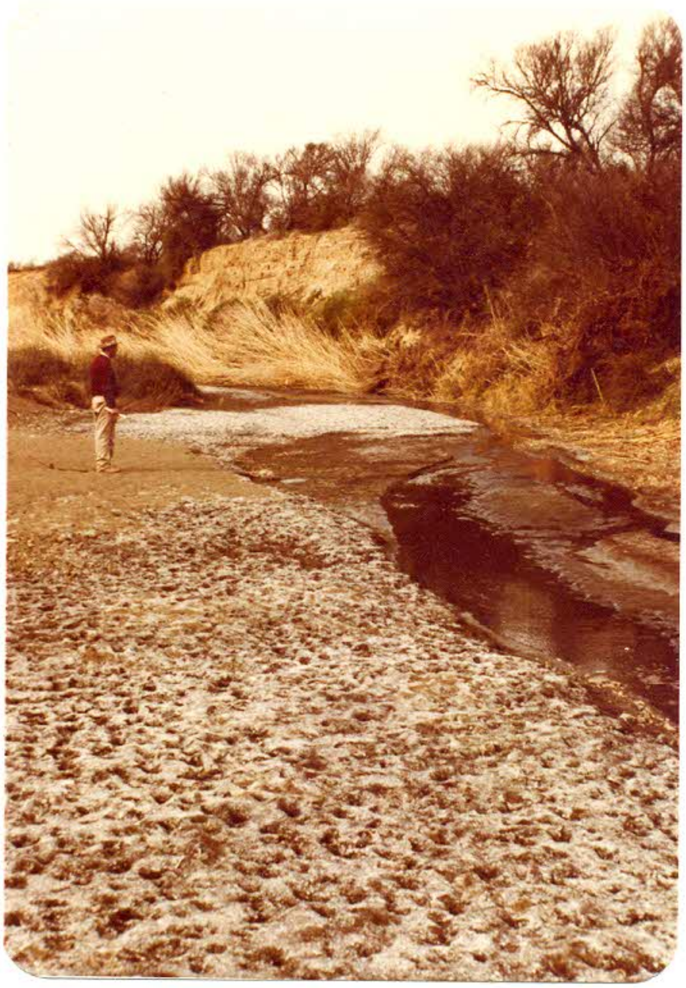


Fig. 6 - 7 The precipitation of salts in the Great Fish River downstream of Katkop weir.

quality of the irrigation water. This influence is reflected by the increase in the stable load at each consecutive weir.

Once the effect of seepage water on a particular volume of irrigation water down the river has been determined and the losses of water are known, projections on the quality of other water reaching the various weirs down the river can be made, providing the salinity of this water, as it leaves the storage dam, is known. Viljoen and Liebenberg (1974c) have attempted such a projection on water from the Hendrik Verwoerd Dam (having a salinity of 172 mg/l by the time it reaches the Grass Ridge Dam) which is released at a rate of 20 000 m^3/h from the Grass Ridge Dam into the Great Fish River. According to their calculations the salt load released at Grass Ridge Dam will be at a rate of 3,4 t/h, whilst at Katkop weir the load increases to 3,8 t/h and at Middleton weir to 8,2 t/h. These calculations are based on the assumption that no water is withdrawn at the various weirs and that the head of saline water has passed and therefore the salt load has already stabilized. Should the head of saline water be included into such calculations, the increase would be considerably higher.

It therefore appears that a progressive deterioration of any irrigation water down the Great Fish River is bound to occur, because of an increase in the salt load especially of sodium and chloride. This deterioration is also observed in the groundwater down the basin and therefore groundwater seepage in the river may be one of the main causes of the increasing salinity of the water down the river.

7. CONCLUSIONS

Since the turn of this century, irrigation along the banks of the Great Fish River commenced on a relatively organized scale. It, however, appears that until recently no concern existed as to the quality of the water used for irrigation or to its effect on the irrigable land in the area. Farmers in the area, because of the prevailing drought conditions and also because of insufficient water in the various storage dams, have often been only to eager to divert saline seepage water onto their lands. The result is that apricot orchards which once flourished in the Golden Valley area, have completely succumbed to salinisation. Even to this day, those living in the Great Fish River Valley do not want to believe that the deterioration of the crops is due largely to mineralisation. This deterioration is attributed by them to the lack of sufficient water and all have been looking foreward with great expectations to the arrival of water from the Orange River. It is questionable, however, whether this water will solve all problems, especially in the long term.

It is pointed out in this investigation that the arid climate of the area presents the first obstacle. Throughout the year the rate of evapotranspiration exceeds the rate of precipitation and therefore any water penetrating the soil is soon lost to the atmosphere, leaving a concentration of soluble salts behind. Only during exceptional periods of excessive rainfall, as in 1974, does the precipitation exceed the evapotranspiration for a long enough period to permit substantial amounts of water to percolate down to the groundwater level, taking soluble salts with it.

The soils of the area, due to their clay-content, prevent infiltration to any great extent. Most of these soils have a high adsorption capacity for cations and are therefore likely to retain most of the soluble substances present in the water. They also can retain chloride by acting as an impervious clay-membrane to this ion. Water is often encountered at the contact of the soil with the underlying rock strata. This zone is normally more pervious, allowing groundwater discharge to the river.

There is a definite change in the groundwater quality down the Great Fish River Basin and this change is reflected by the seepage water in the river. This phenomenon is attributed mainly to topography and partly to the character of the geological strata in the area. The main influence of topography lies in the fact that groundwater recharge occurs mainly in the areas of medium to high relief and discharge in the lower lying areas. Circulation of the groundwater is also more dynamic in the higher lying areas, while stagnant conditions, suitable for the accumulation of salts, prevail in the low lying areas. Because most of the groundwater occurs within the rock strata itself, it is inevitable that the geology must have some influence on the chemical quality. This influence is due mainly to (a) mixing of connate water in the sedimentary strata with the meteoric water, especially in the lower stratigraphic units,

(b) cation exchange, especially between Na^{+} and Ca^{++} in the argillaceous sediments and (c) the weathering of primary minerals especially in the igneous rocks. Only in the higher stratigraphic units in the areas of high relief does chemical weathering of primary minerals in the sedimentary rocks influence the groundwater quality slightly.

The mode of occurrence of the groundwater within the rock strata also has some influence on the quality. The primary hydrological properties of the sedimentary rocks do not permit much water to flow through them. Most of the water therefore occurs within secondary structures such as joints and fractures in or near dolerite intrusions. In addition, large groundwater compartments, which have a great influence on the seepage water in the river are formed by dolerites restricting the normal, regional, gravitational flow of groundwater.

The amount and quality of seepage water in the river is controlled mainly by groundwater discharge. This seepage water in turn has a considerable effect on the quality of irrigation water in the river. Irrespective of the quality of the initial irrigation water, the fact remains that because of the presence of saline seepage water in the river, a constant load of salts per hectare of irrigated land will be applied during each lead. This salt, because of the climatic conditions and because of the nature of the soils, is unlikely to be leached out of the soils and will tend to accumulate. Because of the lack of initial research in this field, no data regarding the accumulation of salts since the commencement of irrigation are available.

The standard practice thus far was to use all the water during a particular irrigation lead. Fortunately for the irrigators, farther upstream, the water was allowed to pass for some time before it was diverted onto their lands. The irrigators lower, downstream, however, had to be content with what water they received and always therefore received the bulk of the saline head of irrigation water. As no data are available on the amount of salts applied to the lands by this means, one can only imagine the enormity of the effect.

In order to prevent any further salinisation of the soils it therefore appears that some control must be exercised on the amount of seepage water applied to the lands. Due to the complexity and extent of the major groundwater compartments, it would be unrealistic and too costly to control the flow of seepage water into the river. The following alternate measures are, however, suggested:

(i) The construction of a concrete-lined canal to take the required water down the river. One of the main advantages of such an irrigation canal, is that contamination from human or other sources will almost be negligable and the loss of expensive water by means other than evaporation will also be eliminated. Better control on the amount of water supplied to each irrigation area can also be applied. The disadvantage of such a canal is the tremendous cost involved. As a long term investment, this, however, still seems to be the only real solution, because once the soils reach a certain high salinity, the crop production will decrease and to reclaim salinised soils will be even more costly.

(ii) Should the construction of a canal not be considered, at least the perpetual presence of a saline head during each lead should be avoided by maintaining a continual flow of water down the river. This way the river need only be flushed once and although the normal seepage water will still be applied to the lands, at least the load in the saline head will be avoided. This method will not permenantly solve the problem, but will at least prolong salinisation of the soil.

(iii) To maintain a low-flow in the river may be impractical as the demand for water along the river is periodic rather than continual. Should irrigation water therefore be released periodically the saline head at each weir should be allowed to pass before actual irrigation commences. The total volume of water lost by this method is determined by the duration of the head at the lower most weir in the system, i.e. that at Middleton. If the duration at Marlow weir already exceeds 30 hours it may well be expected that the time at Middleton will double. Taking into account the annual cost of water lost by these means and comparing it with the annual interest on the capital cost of a canal, one may be able to reach some conclusion as to the most economical way of preserving the soil.

There, however, appears to be no doubt that should the present practice of irrigation along the Great Fish River continue, the soils will become more saline in spite of an increased supply of water which merely introduces a greater load of salts. Further research is therefore essential to determine the rate of increase of salts in the soils as a result of irrigation and to what extent the increase occurs.

8. ACKNOWLEDGEMENTS

The author wishes to express his most sincere appreciation to Professor B.J.V. Botha, head of the Geology Department of the University of the Orange Free State, for his efforts in initiating this project through the NIWR of the CSIR. Without the substantial grant by the NIWR this study would, however, have been impossible and therefore thanks must go to the Hydro-Chemical Working Group for the ORP as well as to the Director of the NIWR.

Chemical analyses on rock samples were performed by the NPRL of the CSIR and by Mr. W.A. van der Westhuizen of the Geology Department of the UOVS. These persons are thanked for their efforts.

A special word of thanks is also expressed to Mrs. M. Tordiffe for the typing of the script, Mrs. A. Botha for preparing most of the Plates and figures of this study and again to Professor B.J.V. Botha for his advice and suggestions during the study and for critically reading through this script.

The farmers of the Great Fish River Basin with whom the author had contact during the field visits are thanked for their co-operation and hospitality. A special word of appreciation must, however, be extended to Mr. G.D. Saunders and his family of the farm "Kingsclere" near Baroda for their hospitality, friendship and extreme interest in this entire study.

9. REFERENCES

ACOCKS, J.H.P. ,(1975). Veld Types of South Africa. Bot. Surv. S. Afr., Mem. 40, 128pp.

ALLEN, J.R.L. ,(1964). Studies in fluviatile sedimentation: Six cyclothems from the lower Old Red Sandstone, Anglo-Welsh Basin. Sedimentology, 3, p. 163 - 198.

-------------- ,(1965a). Fining-upwards cycles in alluvial successions. Geol. J. , 4, p. 229 - 246.

------------- ,(1965b). A review of the origin and characteristics of recent alluvial sediments. Sedimentology, 5, p. 89 - 191.

------------- ,(1970). Studies in fluviatile sedimentation: A comparison of fining-upward cyclothems, with special reference to coarse-member composition and interpretation. J. Sed. Petr. ,40/1 , p. 298 - 323.

ATKINSON, D.J. ,(1962). Tectonic Control of Sedimentation and the Interpretation of Sediment Alteration in the Tertiary of Prince Charles Foreland, Spitsbergen. Geol. Soc. Am. Bull. ,73 , p. 343 - 364.

BAIN, A.G. ,(1845). On the discovery of fossil remains of bidental and other reptiles in South Africa. Trans. geol. Soc. London, V11, p. 53 - 55.

---------- ,(1856). On the geology of Southern Africa. Trans. geol. Soc. London, V11, p. 175 - 192.

BEUKES, N.J. ,(1969). Die Sedimentologie van die Etage Holkranssandsteen, Sisteem Karoo. Unpub. M.Sc.-thesis, Univ. O.F.S. , Bloemfontein, 138pp.

BLATT, H. , MIDDLETON, G. and MURRAY, R. ,(1972). Origin of Sedimentary Rocks. Prentice-Hall, Inc. , New Jersey, 634pp.

BOLTENHAGEN, C. ,(1969). Etude de la cristallinite' de L'illite de divers échantilons d'Afrique du Sud. Unpub. rep. , R/GS 65/69.

BOND, G.W. ,(1946). A Geochemical Survey of the Ground-water Supplies of the Union of South Africa. Geol. Surv. S. Afr. , Mem. 41 , 208pp.

BOTHA, B.J.V. and HODGSON, F.D.I. ,(1976). Karoo dolerites in northwestern Damaraland. Trans. geol. Soc. S. Afr. ,79 , p. 186 - 190.

BROOM, R. ,(1906). The classification of the Karroo beds of South Africa. Geol. Mag., 5 / 3, p.36.

CHEBOTAREV, I.I. ,(1955a). Metamorphism of natural waters in the crust of weathering- 1. Geochim. Cosmochim. Acta, 8, p. 22 -48.

---------------- ,(1955b). Metamorphism of natural waters in the crust of weathering - 3. Geochim. Cosmochim. Acta, 8 , p. 198 - 212.

DAPPLES, E.C. ,(1967). Diagenesis of Sandstones. In; Larsen, G. and Chilingar, G.V. ,(1967). Diagenesis in Sediments. Elsevier Publishing Coy. , Amsterdam, p. 91-126.

DEER, W.A. , HOWIE, R.A. and ZUSSMAN, J. ,(1963). Rock-Forming Minerals, Vol. 3. Sheet Silicates, second impression. Longmans, London, 270pp.

DEGENS, E.T. and CHILINGAR, G.V. ,(1967). Diagenesis of Subsurface Waters. In; Larsen, G. and Chilingar, G.V. ,(1967). Diagenesis in Sediments. Elsevier Publishing Coy., Amsterdam, p. 477 - 502.

DEPARTMENT OF MINES ,(1976). Geological Map 3226 King Williams Town. Geol. Surv. S. Afr. , Scale 1:250 000.

DEPARTMENT OF WATER AFFAIRS ,(1969). Aantekeninge oor etlike van die belangrikste besproeiingskemas en multi-doel-skemas wat deur die Departement van Waterwese gebou is en/of beheer word, 55pp.

DE VORE, G.W. ,(1959). The surface chemistry of feld-spars as an influence on their decomposition products. Proc. Nat. Conf. Clays and Clay Minerals, 6 , p. 26-41.

DIRECTOR OF IRRIGATION ,(1920). Report for the period 1st. April, 1918 to 31st. March, 1919. Cape Times Ltd. , Government Printers, UG 28.

--------------------- ,(1922). Report for the period 1st. April, 1920 to 31st. March, 1921. Cape Times Ltd. , Government Printers, UG 12.

--------------------- ,(1925). Report for the period 1st. April, 1923 to 31st. March, 1924. Cape Times Ltd. , Government Printers, UG 14.

--------------------- ,(1926). Report for the period 1st. April, 1924 to 31st. March, 1925. Cape Times Ltd. , Government Printers, UG 25.

DU PLESSIS, E.J. ,(1973). 'n Ondersoek na die oorsprong en betekenis van Suid-Afrikaanse berg- en riviername. Naamkundereeks, nr. 3. RGN , Tafelberg-uitgewers, Cape Town, 365pp.

DU PLESSIS, J.G. ,(1972). Beplanning van Oranjerivier-projek met spesiale verwysing na Hendrik Verwoerddam. The Civil Engineer in South Africa. Trans. S. Afr. Instn. Civil Engrs. , 14, 2,(Special Issue), p. 33 - 50.

DU TOIT, A.E. ,(1954). The Cape Frontier a study of native policy with special reference to the years 1847 - 1866. D. Phil. thesis, Argief-jaarboek vir Suid-Afrikaanse Geskiedenis, 7 , 1 , 299pp.

DU TOIT, A.L. ,(1904). Geological Survey of the Divisions of Aliwal North, Herchel, Barkly East and part of Wodehouse. Annu. Rep. geol. Commn. , Cape of Good Hope, p. 71 - 181.

-------------- ,(1915). On the Porosity of the Rocks in the Karroo System in South Africa. Trans. roy. Soc. S. Afr. , 4 , p. 169 - 180.

-------------- ,(1920). The Karroo dolerites of South Africa: a study in hypabyssal injection. Trans. geol. Soc. S. Afr. , 23, p. 1 - 42.

-------------- ,(1954). The Geology of South Africa, 3rd. edition. Oliver and Boyd, London, 611pp.

ELLIOT, D.H. and WATTS, D.R. ,(1974). The nature and origin of volcaniclastic material in some Karroo and Beacon rocks. Trans. geol. Soc. S. Afr. , 77, p.109 - 111.

ENSLIN, J.F. ,(1956). Pumping tests and safe yields of ground-water supplies. Trans. geol. Soc. S. Afr. , 59, p. xiii - xlvi.

FAIRBRIDGE, R.W. ,(1967). Phases of diagenesis and authigenesis. In; Larsen, G. and Chilingar, G.V. ,(1967). Diagenesis in Sediments. Elsevier Publishing Coy., Amsterdam, p. 19 - 90.

FROMMURZE, H.F. ,(1937). The water-bearing properties of the more important Geological Formations in the Union of South Africa. Geol. Surv. S. Afr., Mem. 34, 186pp.

------------------ ,(1953). Hydrological research in arid and semi-arid areas in the Union of South Africa. Arid Zone Hydrology, Unesco, p. 58 - 77.

GARRELS, R.M. and CHRIST, C.L. ,(1965). Solutions, Minerals, and Equilibria. Harper and Row, New York, 450pp.

GOLDICH, S.S. ,(1938). A study of rock weathering. J. Geol., 46, p. 17 - 58.

GOLUBEV, V.S. and GARIBYANTS, A.A. ,(1971). Heterogeneous Processes of Geochemical Migration. Consultants Bureau, New York, 150pp.

GREEF, G.J. ,(1968). Fracture systems and the kimberlite intrusions of Griqualand West. Unpub. M.Sc. thesis, Univ. Stellenbosch, 124pp.

GRIM, R.E. ,(1962). Applied Clay Mineralogy. McGraw-Hill Book Coy. INC., London, 422pp.

HAWKES, H.E. and WEBB, J.S. ,(1962). Geochemistry in mineral exploration. Harper and Row, New York, 415pp.

HEM, J.D. ,(1970). Study and interpretation of the chemical characteristics of natural water.. U.S. geol. Surv.

Water-Supply Paper, 1473, second edition, 363pp.

HODGSON, F.D.I. ,(1975). A preliminary investigation of the groundwater potential in the Vryburg Area. Bull. Inst. Groundwater Stud., Univ. O.F.S., Bloemfontein, 1,42pp.

HOUNAM, C.E. ,(1971). Problems of evaporation assessment in the water balance. W.M.O. tech. Note, 13 , p. 31 -51.

HUTCHISON, C.S. ,(1974). Laboritory handbook of petrographic techniques. John Wiley and Sons, New York, 527pp.

JENNY, H. ,(1950). Origin of soils. Applied Sedimentation, (P.D. Trask, ed.), John Wiley and Sons, New York, p. 41-61.

JOHNSON, J.H. ,(1974). Hydrochemistry in groundwater exploration. Groundwater Symp., Bulawayo, 12pp.

JOHNSON, M.R. ,(1966). The Stratigraphy of the Cape and Karroo Systems in the Eastern Cape Province. Unpub. M.Sc. thesis, Rhodes Univ., Grahamstown, 76pp.

-------------- ,(1976). Stratigraphy and Sedimentology of the Cape and Karoo Sequences in the Eastern Cape Province. Unpub. Ph.D. thesis, Rhodes Univ., Grahamstown, 336pp.

JOHNSON, M.R. ,BOTHA, B.J.V. ,ENGELBRECHT, L.N.J. ,HUGO, P.J. ,KEYSER, A.W. ,TURNER, B.R. and WINTER, H.DeLaR., (1975). Preliminary report on lithostratigraphic nomenclature in the Karoo Sequence. Unpub. rep. S. Afr. comm. Strat., (Karoo Working Group).

JORDAAN, J.M. ,(1968). Monthly flow records of gauging stations up to September,1960, Vol.II. Hydrgraphic Surv. Pub., No.8, Division of Hydrological Research, Dept. Water Affairs.

KENT, L.E. ,(1949). The thermal waters of the Union of South Africa and South West Africa. Trans. geol. Soc. S. Afr., 52 , p. 231 - 264.

KEYSER, A.W. ,(1973). A preliminary study of the type area of the Cistecephalus Zone of the Beaufort Series, and a revision of the Anomodont family Cistecephalidae. Geol. Surv. S. Afr,,Mem. 62, p. 1 - 71.

KING, L.C. ,(1963). South African Scenery, third edition. Oliver and Boyd, London, 308pp.

KINGSLEY, C.S. ,(1977). Stratigraphy and Sedimentology of the Ecca Group in the Eastern Cape Province, South Africa. Unpub. Ph.D. thesis, Univ. Port Elizabeth, 290pp.

KITCHING, J.W. ,(1977). The distribution of the Karroo vertebrate fauna. Bernard Price Inst. for Palaeontological Res., Mem.1, Univ. Witwatersrand, Johannesburg, 131pp.

KRAUSKOPF, K.B. ,(1967). Introduction to geochemistry. McGraw-Hill Book Coy., New York, 721pp.

KRAUSS, F. ,(1843). Über die Quellen des südlichen Afrikas. Neues Jb. Miner. Geol. Paläont. , p. 150 - 164.

KRIEL, J.P. ,(1970). Monthly rainfall and evaporation records of evaporation stations up to September,1967. Hydrographic Surv. Pub., No.9, Division of Hydrological Research, Dept. Water Affairs.

KÜBLER, B. ,(1966). La cristallinité de L'illite et les zones tout á fait superieures du metamorphisme. Inst. Geol. Univ. Neuchâtel, Colloque sur les étages tectoniques Neuchâtel.

LARSEN, G. and CHILINGAR, G.V. ,(1967). Concluding remarks on diagenesis. In; Larsen, G. and Chilingar, G.V., (1967). Diagenesis in Sediments. Elsevier Publishing Coy., Amsterdam, p. 523 - 524.

LAWRENCE, C.R. ,(1975). Geology, Hydrodynamics and Hydrochemistry of the Southern Murry Basin. Geol. Surv. Victoria, Mem. 30, 359pp.

LEVINSON, A.A. ,(1974). Introduction to Exploration Geochemistry. Applied Publishing Ltd., Calgary, 612pp.

LOCK, B.E. and WILSON, J.D. ,(1975). Discussion: "The nature and origin of volcaniclastic material in some Karoo and Beacon Rocks", by D.H. Elliot and D.R. Watts. Trans. geol. Soc. S. Afr., 78 , p.171.

LOMBAARD, B.V. ,(1952). Karroo dolerites and lavas. Trans. geol. Soc. S. Afr., 55 , p. 175 - 198.

LOUGHNAN, F.C. ,(1969). Chemical Weathering of silicate minerals. Elsevier Publishing Coy., Amsterdam, 154pp.

LOUW, W.J. and KRUGER, J.P. ,(1968). Potential evapotranspiration in South Africa. Notos, 17 , p. 3 - 14.

LUKASHEV, K.I. ,(1970). Lithology and geochemistry of the weathering crust. Izdatel′stvo Akademii Belorusskoi SSR Minsk 1958. Translated from Russian , Israel Program for Scientific Tranlations, Jerusalem, 368pp.

MARTINI, J.E.J. ,(1974). On the presence of ash beds and volcanic fragments in the graywackes of the Karroo System in the Southern Cape Province. Trans. geol. Soc. S. Afr., 77 , p. 113 - 116.

MASKE, S. ,(1966). The petrography of the Ingeli mountain range. Ann. Univ. Stellenbosch, 41 , A , 1 , 111pp.

McCLAREN, C.H. ,(1974). Die geologie van 'n gebied suid van Plooysburg, Noord-Kaapland. Unpub. M.Sc. thesis, Univ. O.F.S., Bloemfontein, 257pp.

MIALL, A.D. ,(1977). A review of the braided-river depositional environment. Earth-science Reviews, 13 , p. 1-62.

MILLOT, G. ,(1970). Geology of Clays. Springer-Verlag, New York, 429pp.

MOUNTAIN, E.D. ,(1946). The geology of an area east of Grahamstown. An explanation of sheet No. 136 (Grahamstown), Geol. Surv. S. Afr., 56pp.

---------------- ,(1960). Felsic material in Karroo dolerite. Trans. geol. Soc. S. Afr., 63 , p. 137 - 151.

MÜLLER, G. ,(1967). Diagenesis in argillaceous sediments. In; Larsen, G. and Chilingar, G.V. ,(1967). Diagenesis in Sediments. Elsevier Publishing Coy., Amsterdam, p. 127-178.

OLIVIER, H.J. ,(1972). Geohydrological investigation of the flooding at Shaft 2, Orange-Fish Tunnel, North- Eastern Cape Province. Trans. geol. Soc. S. Afr., 75, p. 197-224.

PARKER, A. ,(1970). An index of weathering for silicate rocks. Geol. Mag.,107, p. 501 - 504.

PETTJOHN, F.J. ,(1975). Sedimentary Rocks, third edition. Harper and Row, New York, 628pp.

PETTIJOHN, F.J., POTTER, P.E. and SIEVER, R. ,(1973). Sand and Sandstone. Springer-Verlag, New York, 618pp.

PINCHIN, R. ,(1875). A short description of the geology of part of the Eastern Province of the Colony of the Cape of Good Hope. Quart. J. geol. Soc. London, 31 , p. 106 - 107.

READ, H.H. and WATSON, J. ,(1962). Introduction to Geology, MacMillan & Co. Ltd. , London, 693pp.

REINECK, H.E. and SINGH, I.B. ,(1973). Depositional Sedimentary Environments. Springer-Verlag, Berlin, 439pp.

RHODES, R.C. and KROHN, D.H. ,(1972). Tectonic control over regional geochemical variation in the Karroo basaltic province of Southern Africa. Trans. geol. Soc. S. Afr. , 75 , p. 11 - 21.

ROMER, A.S. ,(1970). Tetrapod vertebrates and Gondwanaland. Proc. and Pap., Second Gondwana Symp., S. Afr. Coun. Sci. Ind. Res., Pretoria, p. 111 - 124.

ROSSOUW, P.J. ,(1953). General geology: the southern Karoo. In; Haughton, S.H. , etal., Results of an investigation into the possible presence of oil in Karoo rocks in parts of the Union of South Africa. Geol. Surv. S. Afr., Mem. 45, p. 14 - 36.

ROWSELL, D.M. and De SWARDT, A.M.J. ,(1976). Diagenesis in Cape and Karroo sediments, South Africa, and its bearing on their hydrocarbon potential. Trans. geol. Soc. S. Afr.,79, p.81 - 145.

RUBIDGE, R.N. ,(1857). Section of the Zuurberg. The Eastern Province Monthly Magazine, 11 , 17 , p.187 - 196.

RYAN, P.J. ,(1967). Stratgraphic and palaeocurrent analysis of the Ecca Series and lowermost Beaufort beds in the Karroo basin of South Africa. Unpub. Ph.D. thesis, Univ. Witwatersrand, Johannesburg, 210pp.

SAYLES, F.L. and MANGELSDORF, P.C.,Jr. ,(1977). The equilibration of clay minerals with seawater: exchange reactions. Geochim. Cosmochim. Acta, 41, p. 951 - 960.

SCHOLTZ, D.L. ,(1936). The magmatic nickeliferous ore deposits of East Griqualand and Pondoland. Trans. geol. Soc. S. Afr. , 39 , p. 81 - 210.

SOEKOR, (1966). Photogeological map of area 2-Southern Karoo. Compiled by Geomap, Scale 1 : 250 000.

SOUTH AFRICAN COMMITTEE FOR STRATIGRAPHY, (1971). South African code of stratigraphic terminology and nomenclature. Trans. geol. Soc. S. Afr., 74 , p. 111 - 131.

STRATTEN, T. ,(1968). The Dwyka glaciation and its relationship to the pre-Karoo surface. Unpub. Ph.D. thesis, Univ. Witwatersrand, Johannesburg, 196pp.

THERON, J.N. and BLIGNAULT, H.J. ,(1973). A model for the sedimentation of the Dwyka Glacials in the Southwestern Cape. In; Campbell, K.S.W.(ed.), Gondwana Geology. Papers presented at the third Gondwana Symposium, Canberra, Australia, Aust. Nat. Univ. Press, p. 347 - 356.

TORDIFFE, E.A.W. ,(1974). Unpublished annual report on the geohydrological research in the Great Fish River Basin. Ninth meeting of the ORP Hydrochemical Working Group, C.S.I.R., 20th August, 1974, Doc.G, 51pp.

TWENHOFEL, W.H. ,(1950). Principles of Sedimentation. McGraw-Hill Book Coy., London, 673pp.

TYSON, P.D., DYER, T.G.J. and MAMETSE, M.N., (1975). Secular changes in South African Rainfall: 1880 - 1972. Quart. J. roy. Met. Soc., 101, 430, p. 817 - 833.

VAN DER WALT, A.J.H., WIID, J.A., and GEYER, A.L., (1966). Geskiedenis van Suid-Afrika, tweede uitgawe. Nasionale Handelsdrukkery Bpk., Elsiesrivier, 620pp.

VAN NIEKERK, C.D., (1977). Die geologie noord-oos van Graaff-Reinet. Unpub. M.Sc. thesis, Univ. O.F.S., Bloemfontein, 145pp.

VAN ROOYEN, L., (1977). Die geologie en geohidrologie van 'n gebied noord van Reivilo Kaapprovinsie. Unpub. M.Sc. thesis, Univ. O.F.S., Bloemfontein, 155pp.

VAN WYK, W.L., (1960). Groundwater studies in Northern Natal, Zululand and surrounding areas. Unpub. Ph.D. thesis, Univ. Cape Town, 160pp.

VILJOEN, P.T., (1972). Opname van sugwaters in die Visrivier. Unpub. report for the fifth meeting of the Hydrochemical Working Group for the ORP, NIWR, C.S.I.R., Doc. H, S14/106/8A, 9pp.

VILJOEN, P.T. and LIEBENBERG, J.J.E., (1974a). Kwantitatiewe en kwalitatiewe opname van sugwater in die Visrivier stroomaf van die Grassridgedam tot by Sheldon. Unpub. report for the ninth meeting of the Hydrochemical Working Group for the ORP, NIWR, C.S.I.R., Doc. C, W6/553/3, 15pp.

-------------------------------------, (1974b). Mineralisasie van besproeiingswater in die Visrivier stroomaf van Grassridgedam. Unpub. report for the ninth meeting of the Hydrochemical Working Group for the ORP, NIWR, C.S.I.R., Doc.D, W6/533/3, 15pp.

VILJOEN, P.T. and LIEBENBERG, J.J.E. ,(1974c). Moontlike effek van bestaande mineralisasie in die Visrivier op die kwaliteit van Oranjerivierwater indien dit in die Visrivier sou afbeweeg. Unpub. report for the ninth meeting of the Hydrochemical Working Group for the ORP, NIWR, C.S.I.R., Doc. F, W6/533/3 , 8pp.

VISSER, H.N., et al. ,(1963). Gips in die Republiek van Suid-Afrika. Geol. Surv. S. Afr., 152pp.

VON ENGELHARDT, W. ,(1967). Interstitial solutions and diagenesis in sediments. In; Larsen, G. and Chilingar, G.V.,(1967). Diagenesis in Sediments. Elsevier Publishing Coy., Amsterdam, p. 503 - 522.

WALKER, F. and POLDERVAART, A. ,(1942). The petrology of the Karroo dolerites between Sutherland and Middelburg, Cape Province. Trans. geol. Soc. S. Afr., XLV , p. 55-64.

---------------------------------- ,(1949). Karroo dolerites of the Union of South Africa. Bull. geol. Soc. Amer.,60, p.591 - 706.

WILKE, P.P. ,(1962). Groundwater geology of the Fraserburg area. Ann. Univ. Stellenbosch, 37/A/9, p. 593 - 655.

WHITMORE, J.S. ,(1961). Agrohydrology. Unpub. report, 22, Hydrological Research Division, Dept. Water Affairs.

---------------- ,(1963). Some aspects of the hydrology of the semi-arid regions. Unpub. report, 27, Hydrological Research Division, Dept. Water Affairs.

WYBERG, W. ,(1932). The building stones of the Union of South Africa. Geol. Surv. S. Afr., Mem. 29, 238pp.

YOUNG, A. ,(1913). Tidal phenomena at inland boreholes near Crodock. Trans. roy. Soc. S. Afr., 3, p. 61.

--oOo--

Titel: ASPECTS OF THE HYDROGEOCHEMISTRY OF THE KAROO SEQUENCE IN THE GREAT FISH RIVER BASIN, EASTERN CAPE PROVINCE, WITH SPECIAL REFERENCE TO THE GROUNDWATER QUALITY.

Outeur: Eric Arthur Wolferstan Tordiffe, M.Sc.

Promotor: Prof. B.J.V. Botha

Graad: Ph.D.

Departement: Geologie

Medium: Engels

The aim of this study was to examine some of the major aspects responsible for the chemical quality of the groundwater in the Great Fish River Basin and its influence on the irrigation water.

The basin comprises an area of 25 000 km^2 and divided into four main geomorphologic provinces; the Marginal Region, the Great Escarpment, the Headbasin and the Interior Plateau, which play an important part in controling the movement and the chemical quality of the groundwater in the area.

Infiltration of meteoric water down to the groundwater level, is rather limited, because of the semi-arid climatic conditions as well as the nature of the soils in the area. Although the groundwater is recharged in the higher lying areas, by circulating meteoric water, there appears to be no direct relationship between the seasonal precipitation and the fluctuation in groundwater levels.

The area is underlain by sedimentary rocks of the Dwyka Tillite Formation, the Ecca Group, the Koonap Formation and the Beaufort Group. Purely on lithological grounds the Beaufort Group is sub-divided into the Adelaide Sub-group (Middleton and Balfour Formations), which was deposited in a reducing palaeoenvironment and the Tarkastad Sub-group (Katberg and Burgersdorp Formations), which was deposited in an oxidizing palaeoenvironment. The Balfour Formation is furthermore sub-divided into the Oudeberg Sandstone, Daggaboersnek, Barberskrans and Elandsberg Members which are easily recognizable lithostratigraphic units in the area.

Intrusive in the above sedimentary strata are numerous dykes and sills of Karoo dolerite. These bodies are of particular importance in that, when crossing the regional flow of the groundwater, they may act as impervious barriers, thus developing groundwater compartments.

Geochemical factors which play an important part in controling the groundwater quality of the area are the weathering of the various rocks, the diagenesis of the sediments and the ion exchange between the mudrock and the interstitial water. Dolerite is more prone to weathering than the sedimentary rocks which have already survived at least one cycle of weathering. In the Katberg and Burgersdorp Formations weathering, however, appears to be more prevalent whilst the influence of diagenesis is more prominent in the lower strata.

Because of the low porosity and permeability of the sedimentary rocks, the groundwater in the area is restricted mainly to joints and fracture zones near dolerite intrusions.

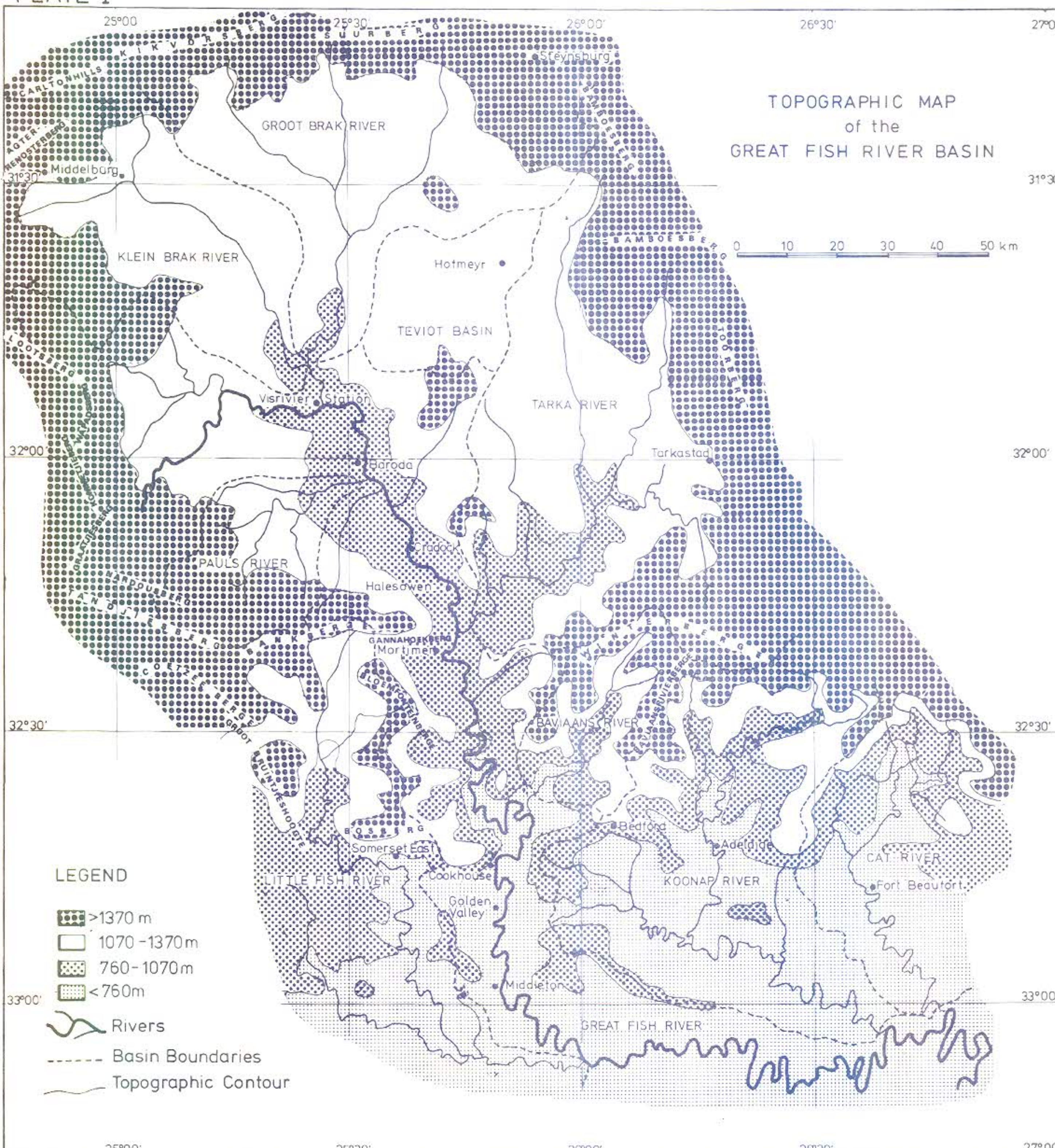
The cations in the groundwater appear to have originated mainly from the weathering of the rocks, whilst the anions accumulate from non-lithologic sources. Generally the groundwater in the areas of recharge has a pronounced Ca^{++} and HCO_3^- - character, whilst in the stagnant low-lying areas Na^+ and Cl^- are the predominant ions.

In spite of the overwhelming influence of topography on the groundwater quality, some influence of the geology can still be observed. High salinities, as a result of high Na^+ and Cl^- - concentrations prevail in the groundwater of the lower strata up to the Daggaboersnek Member. The high Cl^- - concentrations in this water are attributed mainly to ultrafiltration during the upward migration of connate water as a result of compaction. In the strata above the Daggaboersnek Member the concentration of the above ions decreases sharply. It is also interesting to note that higher $SO_4^=$ - concentrations are encountered in groundwater associated with strata deposited under reducing conditions. Under such conditions pyrite is formed which may later oxidize to release $SO_4^=$.

There is no doubt that, because of its effluent nature, the groundwater has a marked effect on the seepage water in the Great Fish River, which in turn has an adverse effect on the irrigation water.

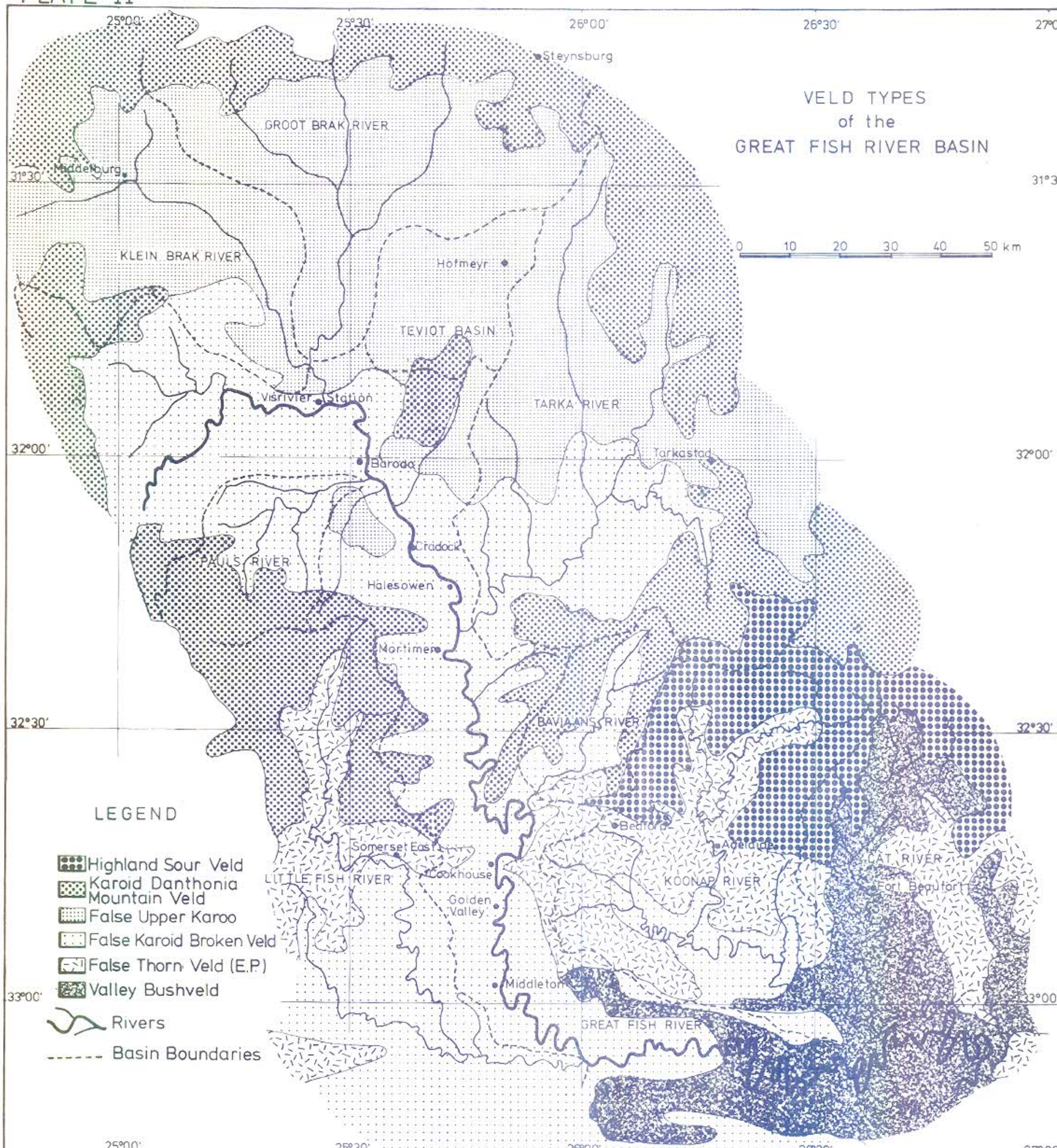
It is therefore suggested that measures be taken to prevent the contamination of irrigation water by the effluent groundwater in the river.

PLATE I



TOPOGRAPHIC MAP
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
Middelburg
GROOT BRAK RIVER
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
TARKA RIVER
Visrivier Station
Baroda
Tarkastad
Cradock
PAULS RIVER
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
Somerset East
Cookhouse
LITTLE FISH RIVER
KOONAP RIVER
CAT RIVER
Fort Beaufort
Golden Valley
Middleton
GREAT FISH RIVER
LEGEND
>1370 m
1070 -1370 m
760-1070 m
<760m
Rivers
Basin Boundaries
Topographic Contour

PLATE II



VELD TYPES
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
GROOT BRAK RIVER
Middelburg
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
Visrivier Station
TARKA RIVER
Baroda
Tarkastad
Cradock
PAULS RIVER
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
Somerset East
Cookhouse
LITTLE FISH RIVER
KOONAP RIVER
CAT RIVER
Fort Beaufort
Golden Valley
Middleton
GREAT FISH RIVER
LEGEND
Highland Sour Veld
Karoid Danthonia Mountain Veld
False Upper Karoo
False Karoid Broken Veld
False Thorn Veld (E.P.)
Valley Bushveld
Rivers
Basin Boundaries

PLATE V

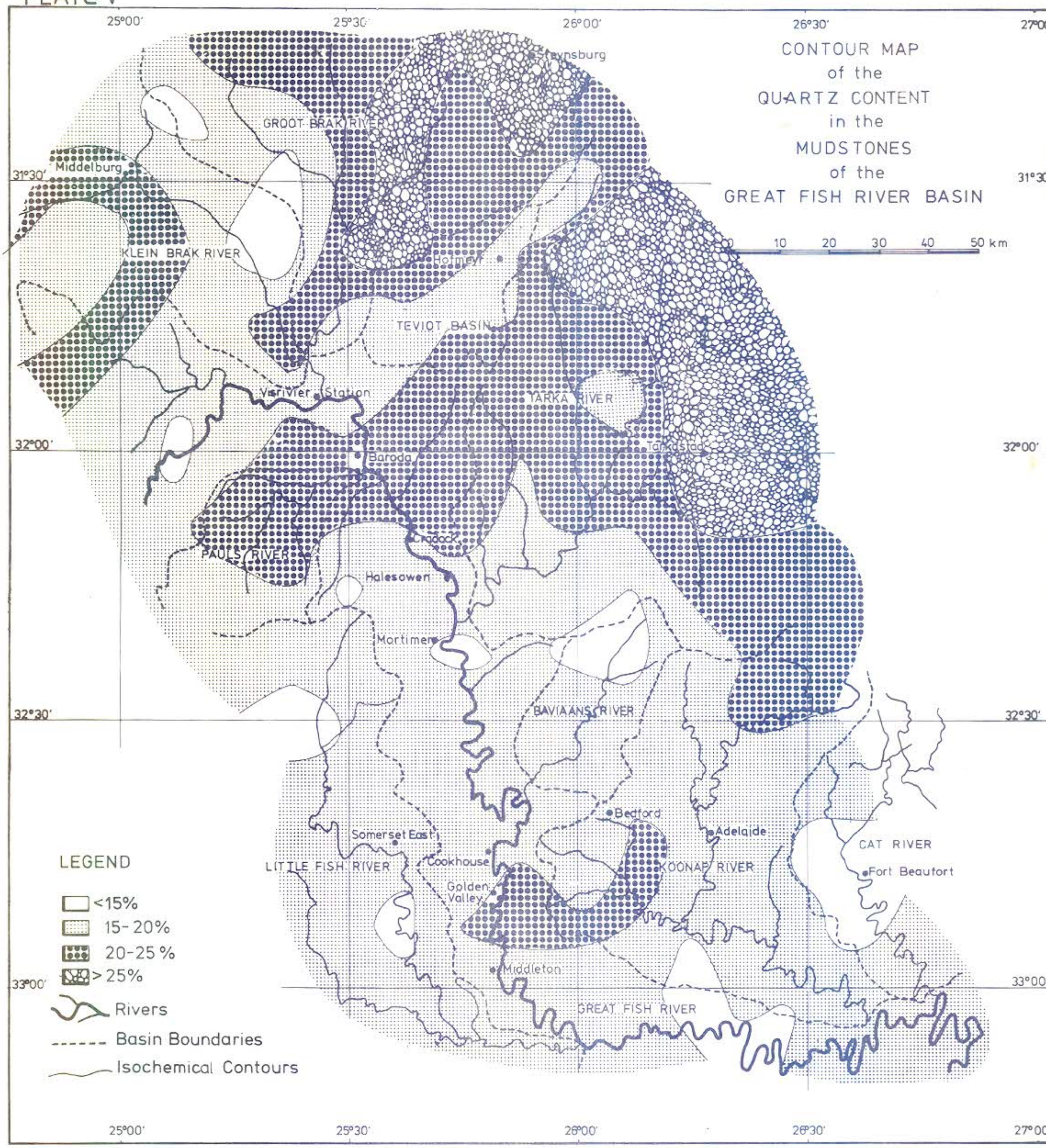

CONTOUR MAP
of the
QUARTZ CONTENT
in the
MUDSTONES
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Middelburg
GROOT BRAK RIVER
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
TARKA RIVER
Tarkastad
Baroda
PAULS RIVER
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
CAT RIVER
Fort Beaufort
Somerset East
LITTLE FISH RIVER
Cookhouse
Golden Valley
KOONAP RIVER
Middleton
GREAT FISH RIVER
LEGEND
<15%
15-20%
20-25 %
>25%
Rivers
Basin Boundaries
Isochemical Contours

PLATE VI

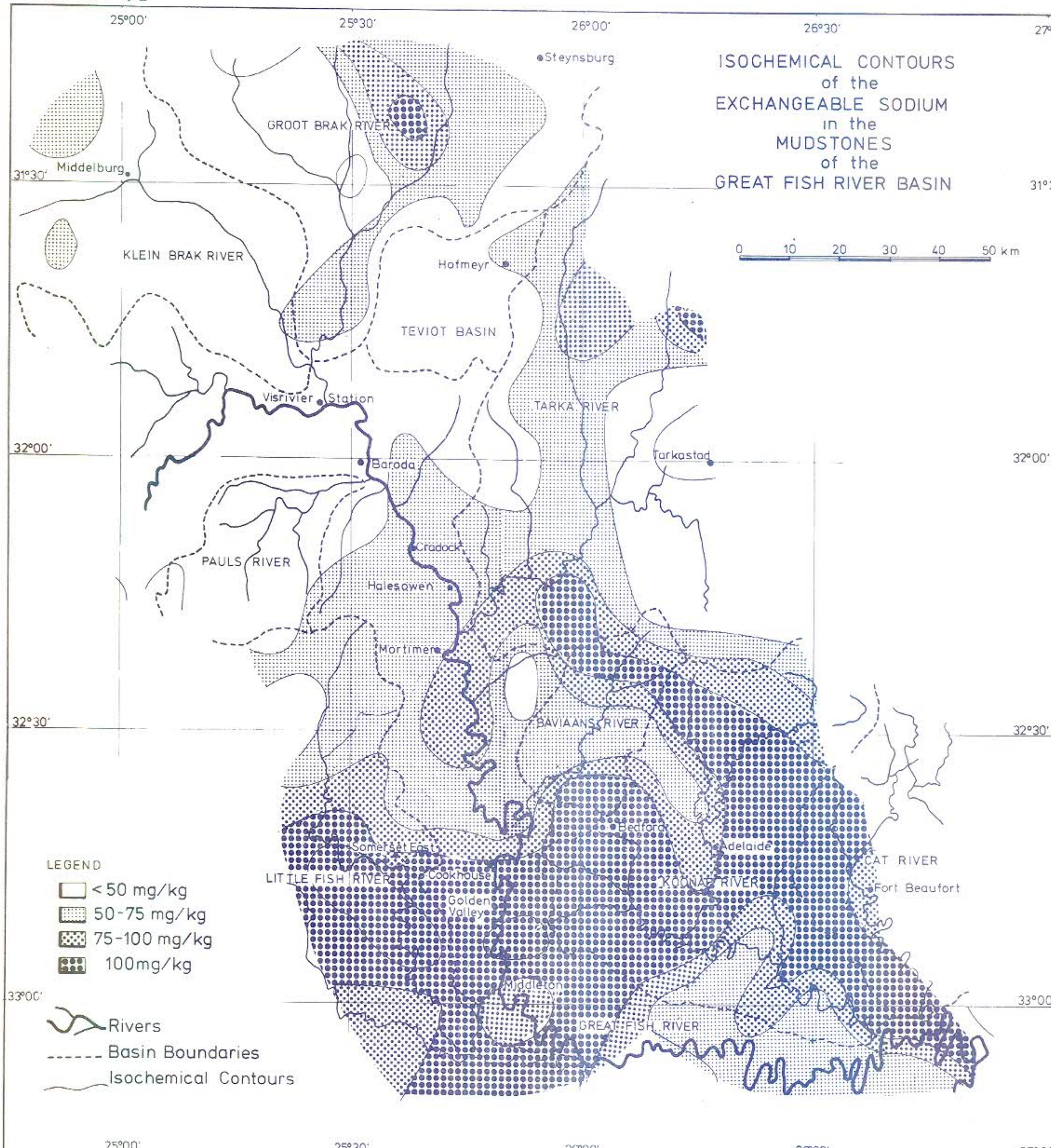

ISOCHEMICAL CONTOURS of the EXCHANGEABLE SODIUM in the MUDSTONES of the GREAT FISH RIVER BASIN
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
0 10 20 30 40 50 km
Steynsburg
GROOT BRAK RIVER
Middelburg
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
Visrivier Station
TARKA RIVER
Tarkastad
Baroda
PAULS RIVER
Cradock
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
Somerset East
Cookhouse
LITTLE FISH RIVER
KOONAP RIVER
CAT RIVER
Fort Beaufort
Golden Valley
Middleton
GREAT FISH RIVER
LEGEND
< 50 mg/kg
50-75 mg/kg
75-100 mg/kg
100mg/kg
Rivers
Basin Boundaries
Isochemical Contours

PLATE VII

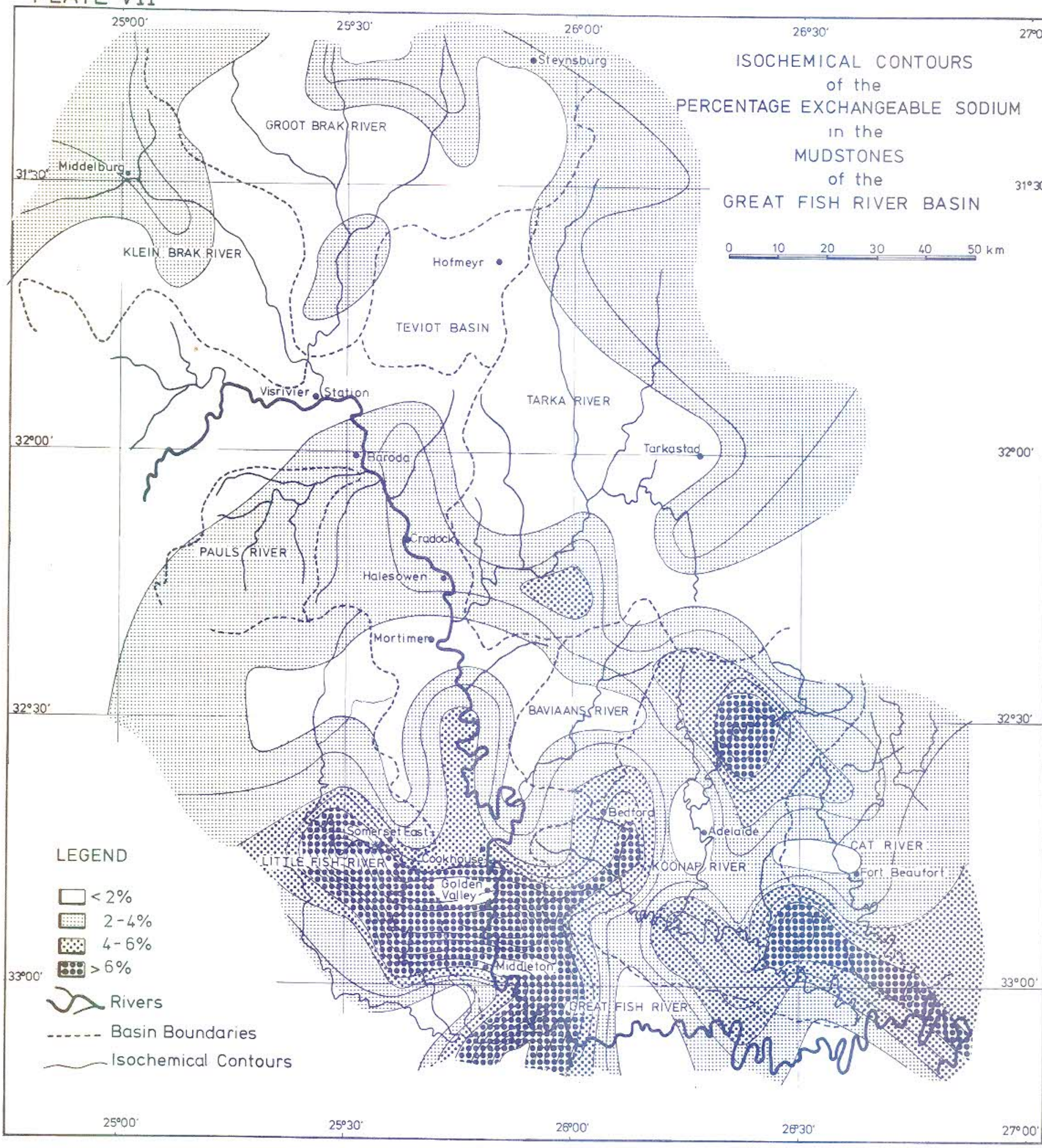

ISOCHEMICAL CONTOURS
of the
PERCENTAGE EXCHANGEABLE SODIUM
in the
MUDSTONES
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
Middelburg
GROOT BRAK RIVER
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
Visrivier Station
TARKA RIVER
Tarkastad
Baroda
Cradock
PAULS RIVER
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
Somerset East
LITTLE FISH RIVER
Cookhouse
Golden Valley
KOONAP RIVER
CAT RIVER
Fort Beaufort
Middleton
GREAT FISH RIVER
LEGEND
< 2%
2 - 4%
4 - 6%
> 6%
Rivers
Basin Boundaries
Isochemical Contours

PLATE VIII

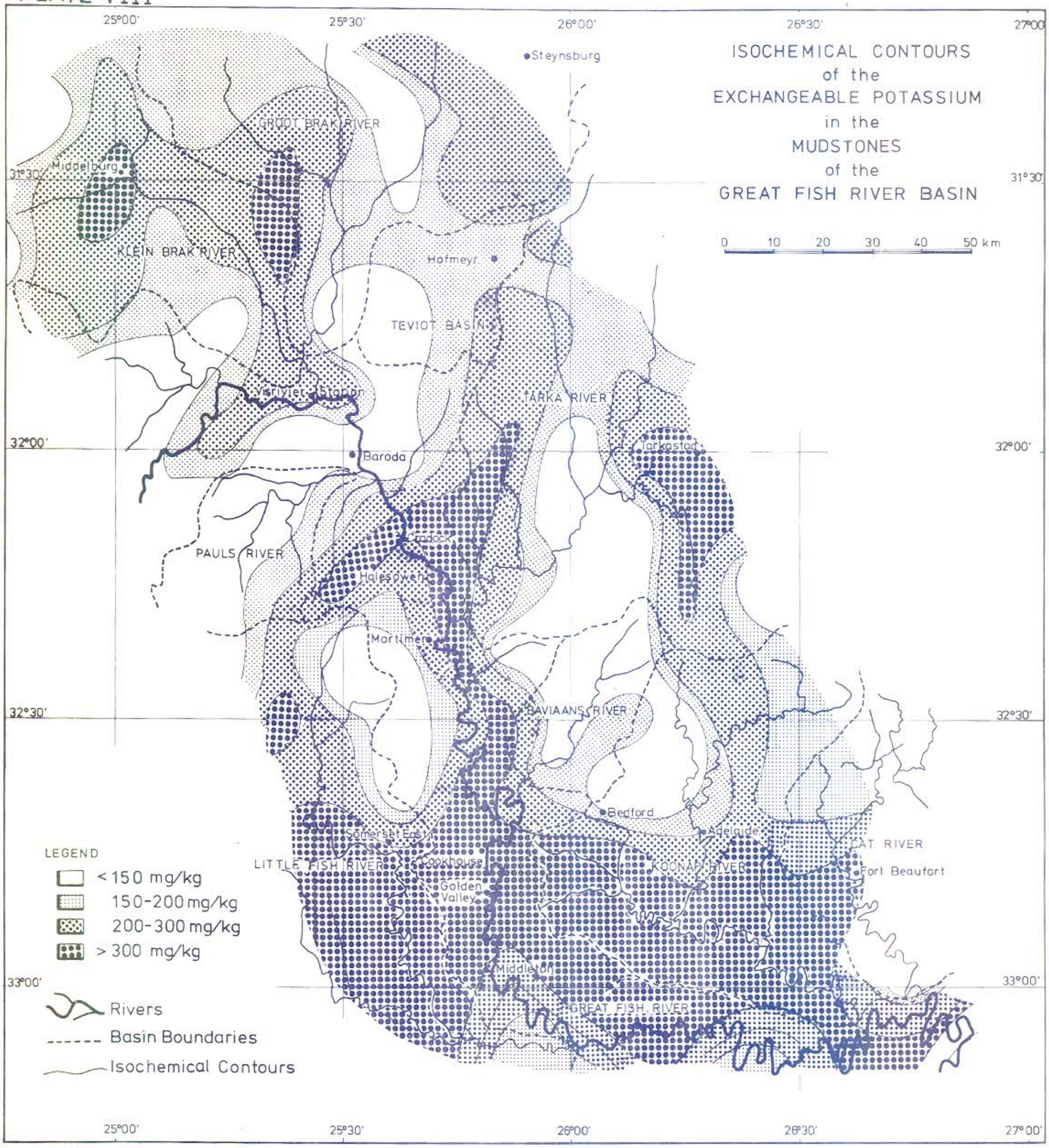
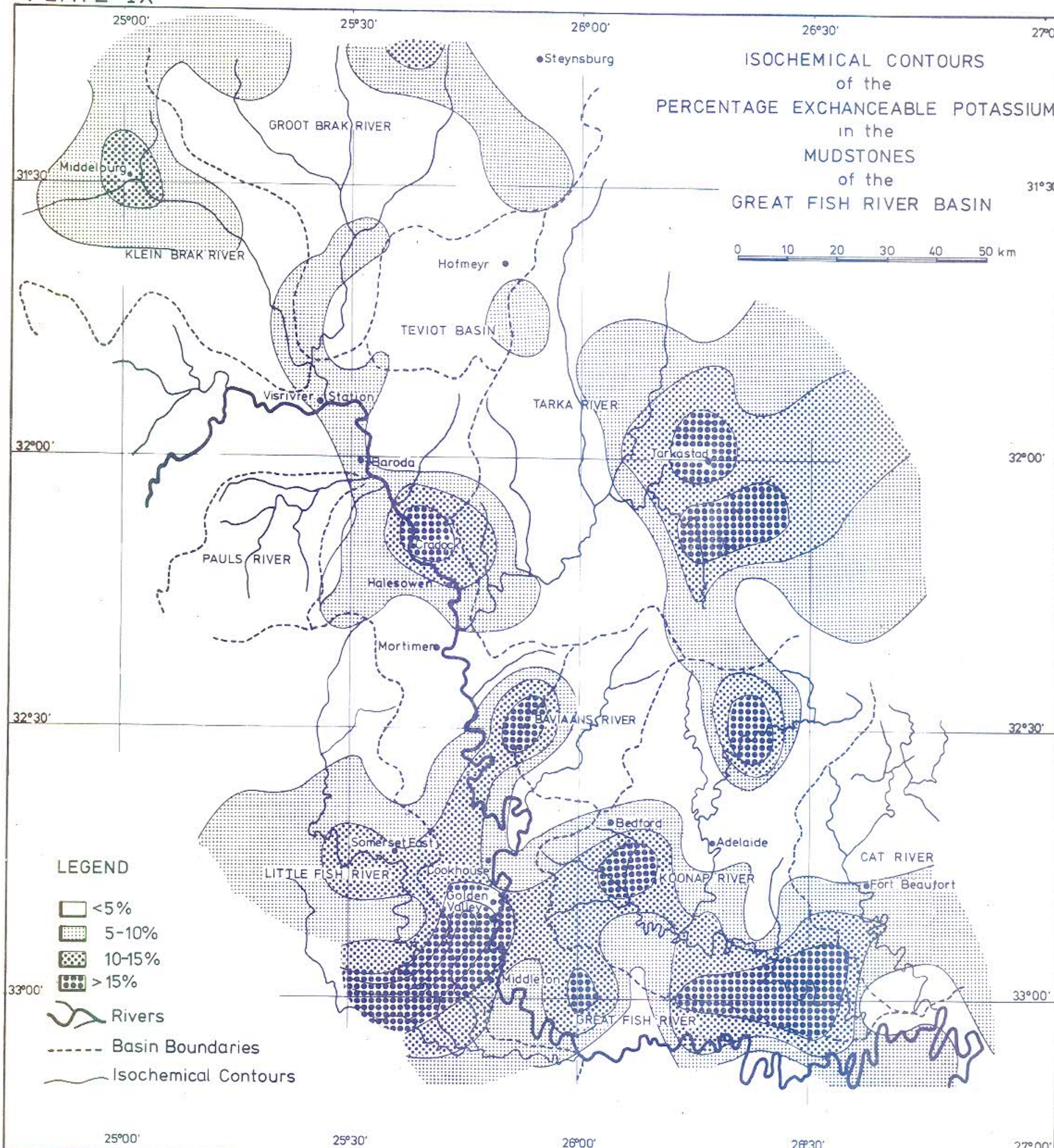

ISOCHEMICAL CONTOURS
of the
EXCHANGEABLE POTASSIUM
in the
MUDSTONES
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
GROOT BRAK RIVER
Middelburg
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
TARKA RIVER
Baroda
Tarkastad
PAULS RIVER
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
CAT RIVER
Fort Beaufort
Somerset East
LITTLE FISH RIVER
Cookhouse
Golden Valley
KOONAP RIVER
Middleton
GREAT FISH RIVER
LEGEND
< 150 mg/kg
150-200 mg/kg
200-300 mg/kg
> 300 mg/kg
Rivers
Basin Boundaries
Isochemical Contours

PLATE IX



ISOCHEMICAL CONTOURS
of the
PERCENTAGE EXCHANCEABLE POTASSIUM
in the
MUDSTONES
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
GROOT BRAK RIVER
Middelburg
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
Visrivier Station
TARKA RIVER
Tarkastad
Baroda
Cradock
PAULS RIVER
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
CAT RIVER
Somerset East
LITTLE FISH RIVER
Cookhouse
KOONAP RIVER
Fort Beaufort
Golden Valley
Middleton
GREAT FISH RIVER
LEGEND
<5%
5-10%
10-15%
> 15%
Rivers
Basin Boundaries
Isochemical Contours

PLATE X

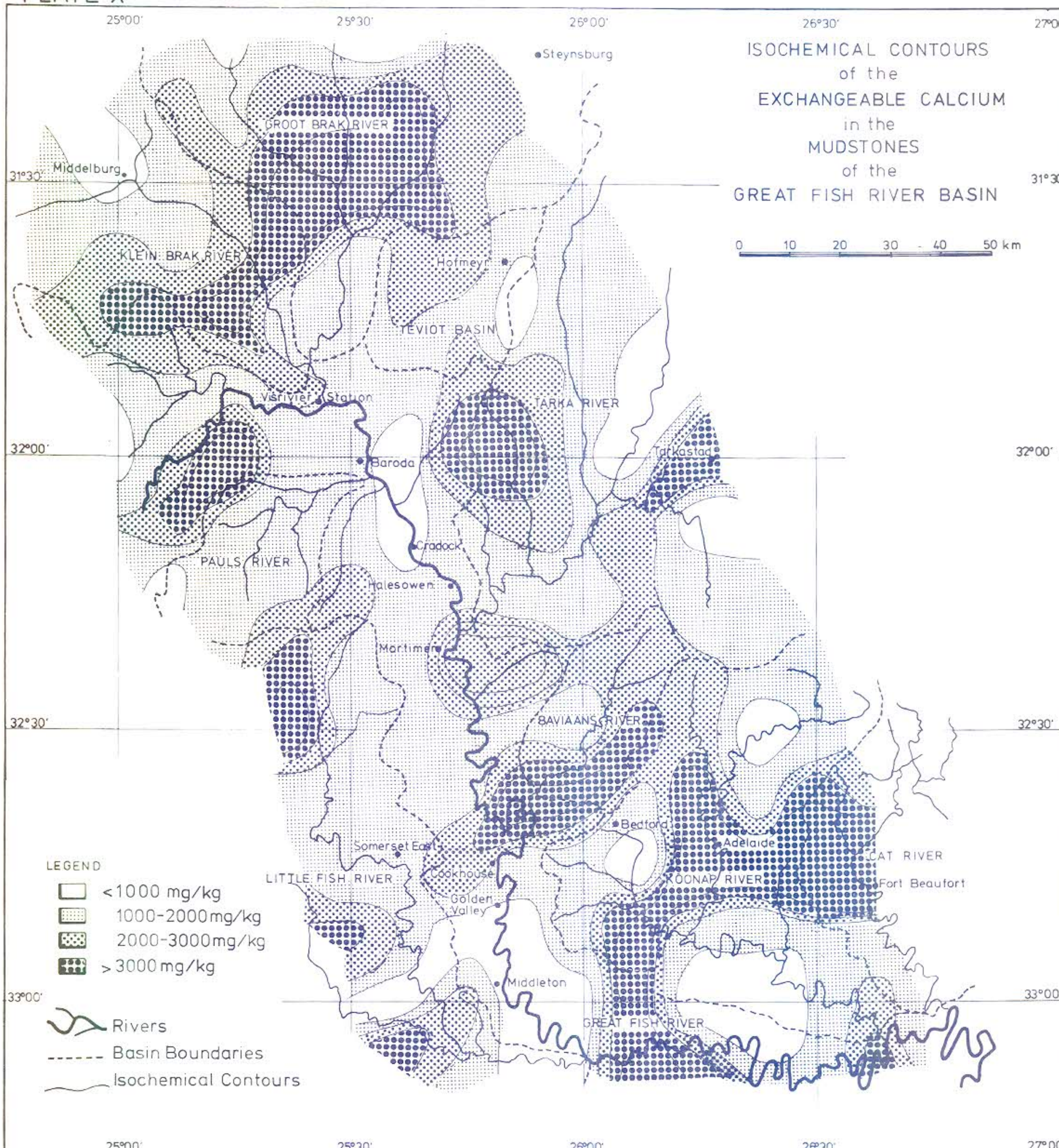

ISOCHEMICAL CONTOURS
of the
EXCHANGEABLE CALCIUM
in the
MUDSTONES
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
Middelburg
GROOT BRAK RIVER
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
Visrivier Station
TARKA RIVER
Tarkastad
Baroda
Cradock
PAULS RIVER
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
CAT RIVER
Fort Beaufort
Somerset East
Cookhouse
LITTLE FISH RIVER
KOONAP RIVER
Golden Valley
Middleton
GREAT FISH RIVER
LEGEND
<1000 mg/kg
1000-2000mg/kg
2000-3000mg/kg
>3000 mg/kg
Rivers
Basin Boundaries
Isochemical Contours

PLATE XI

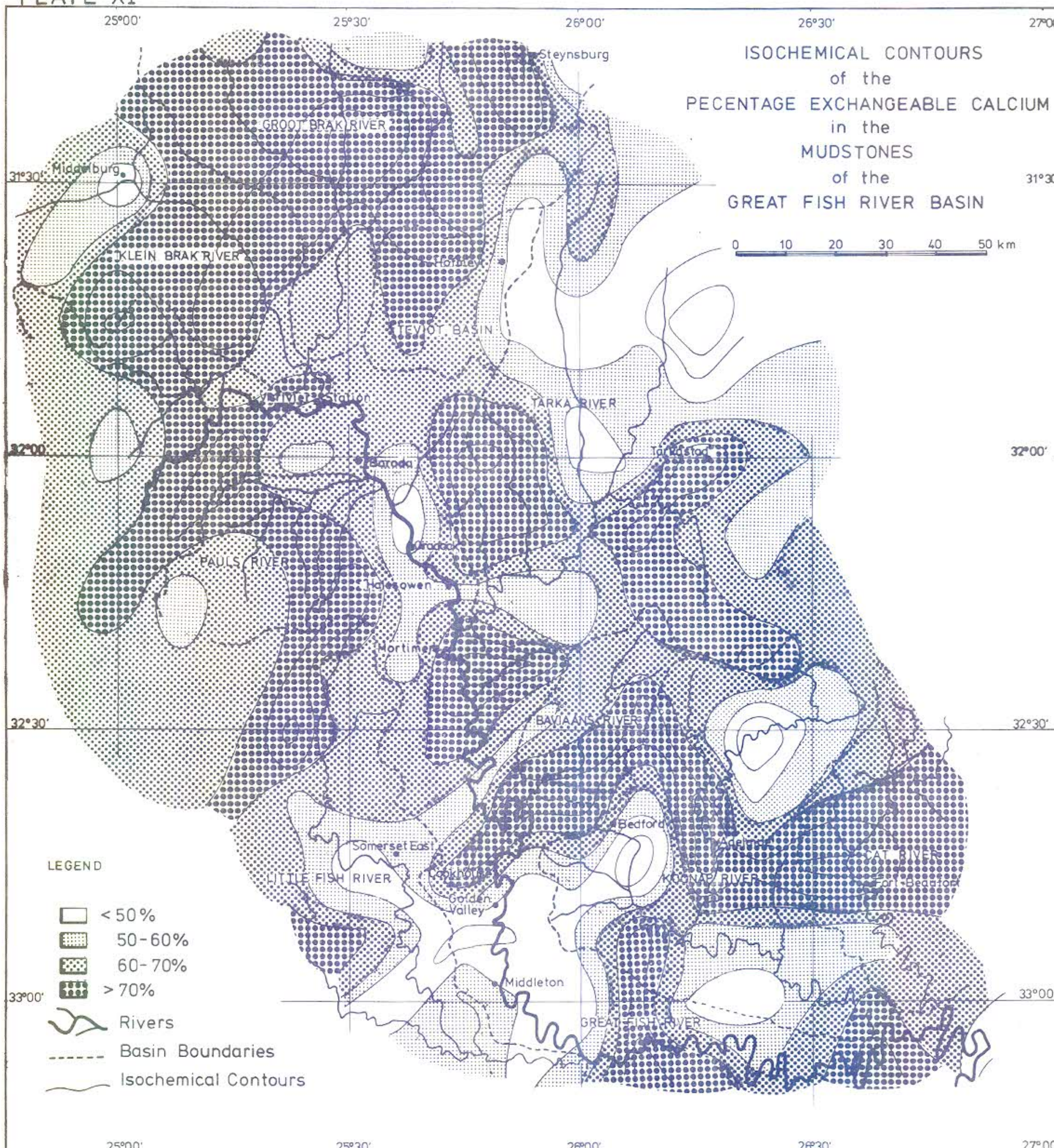
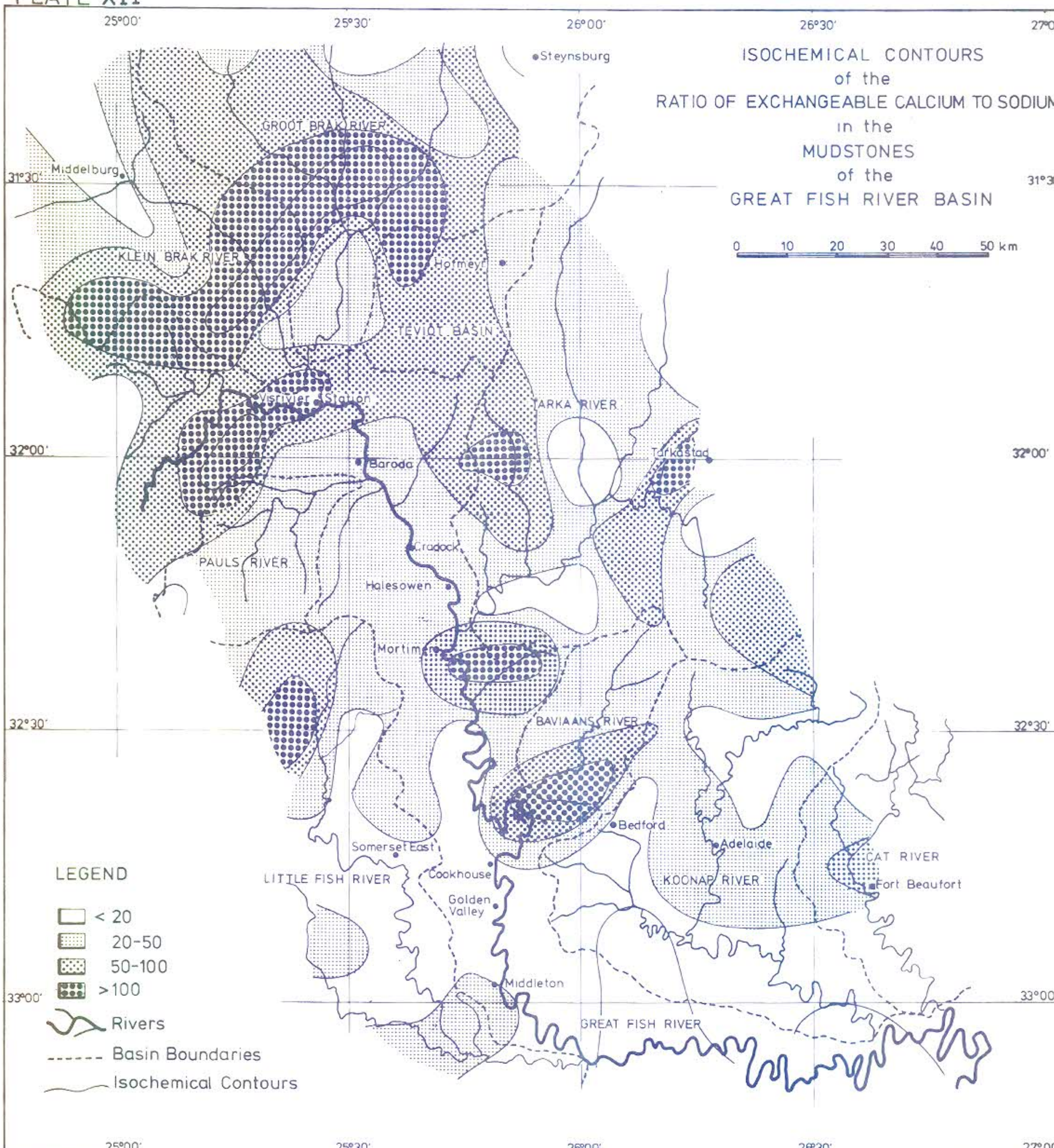

ISOCHEMICAL CONTOURS
of the
PECENTAGE EXCHANGEABLE CALCIUM
in the
MUDSTONES
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
GROOT BRAK RIVER
KLEIN BRAK RIVER
TEVIOT BASIN
TARKA RIVER
Baroda
Cradock
PAULS RIVER
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
Somerset East
LITTLE FISH RIVER
KOONAP RIVER
Golden Valley
Middleton
GREAT FISH RIVER
LEGEND
<50%
50-60%
60-70%
>70%
Rivers
Basin Boundaries
Isochemical Contours

PLATE XII



ISOCHEMICAL CONTOURS
of the
RATIO OF EXCHANGEABLE CALCIUM TO SODIUM
in the
MUDSTONES
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
GROOT BRAK RIVER
Middelburg
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
Visrivier Station
TARKA RIVER
Tarkastad
Baroda
Cradock
PAULS RIVER
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
Somerset East
Cookhouse
CAT RIVER
LITTLE FISH RIVER
KOONAP RIVER
Fort Beaufort
Golden Valley
Middleton
GREAT FISH RIVER
LEGEND
< 20
20-50
50-100
>100
Rivers
Basin Boundaries
Isochemical Contours

PLATE XIII

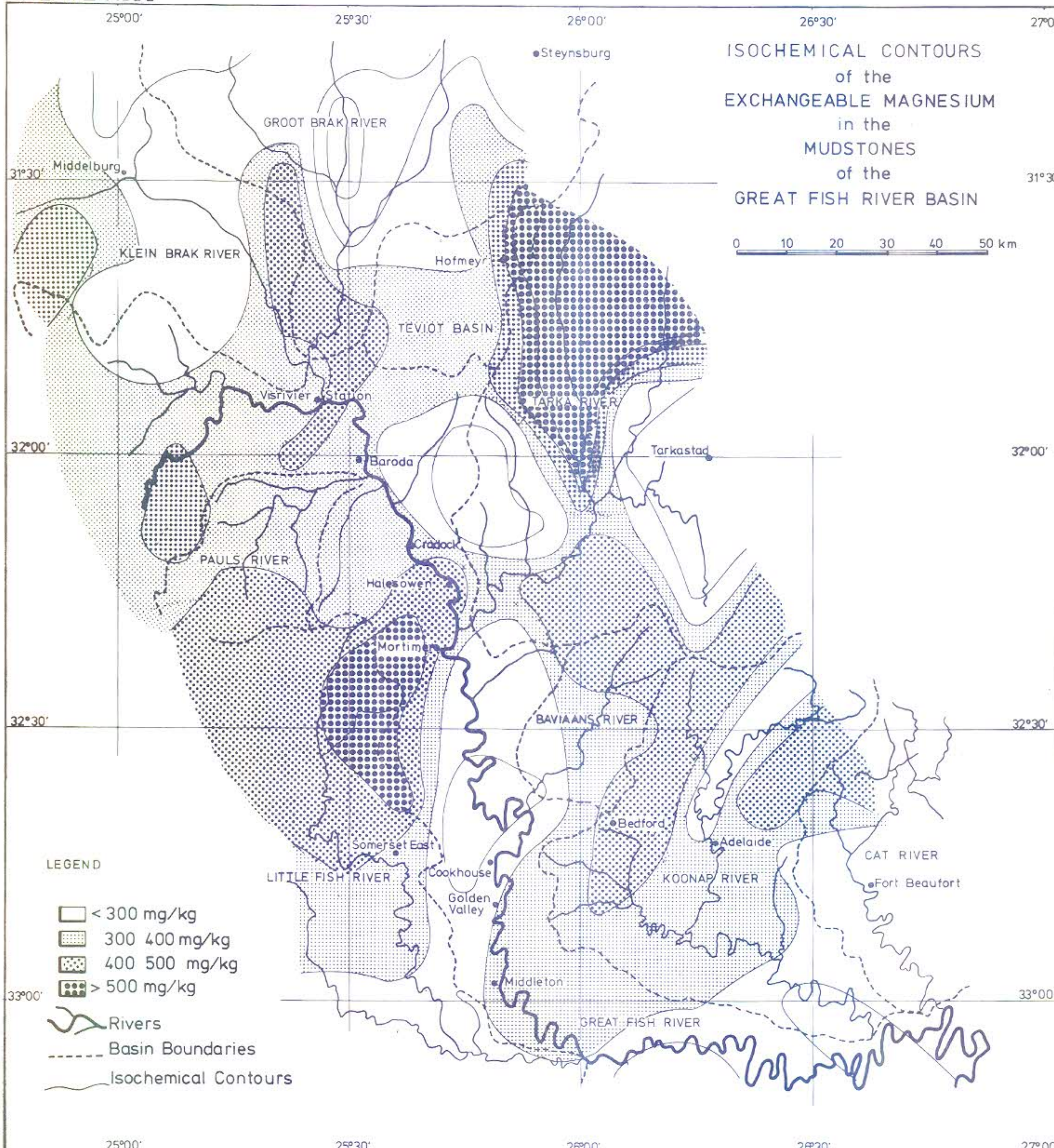
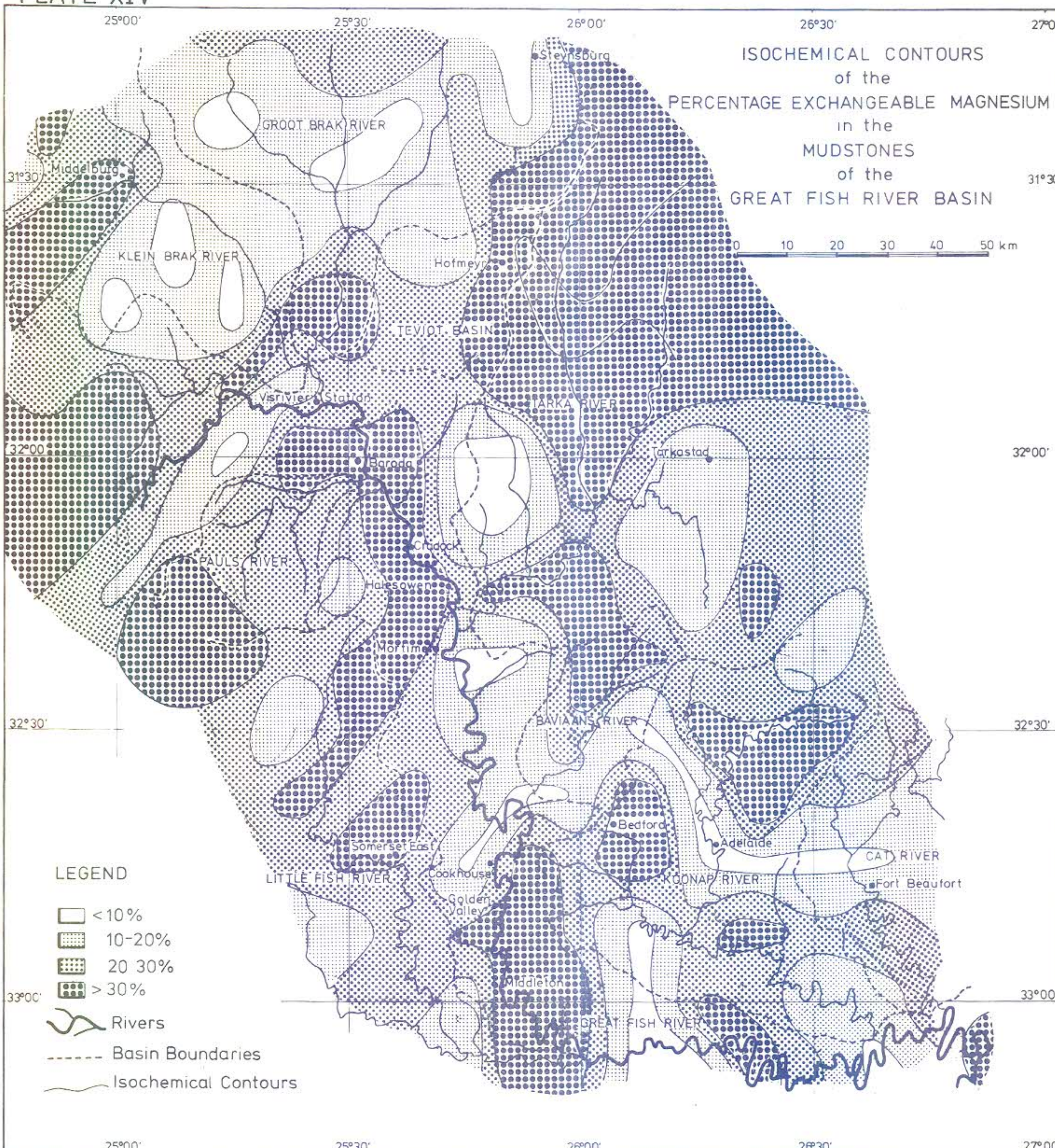
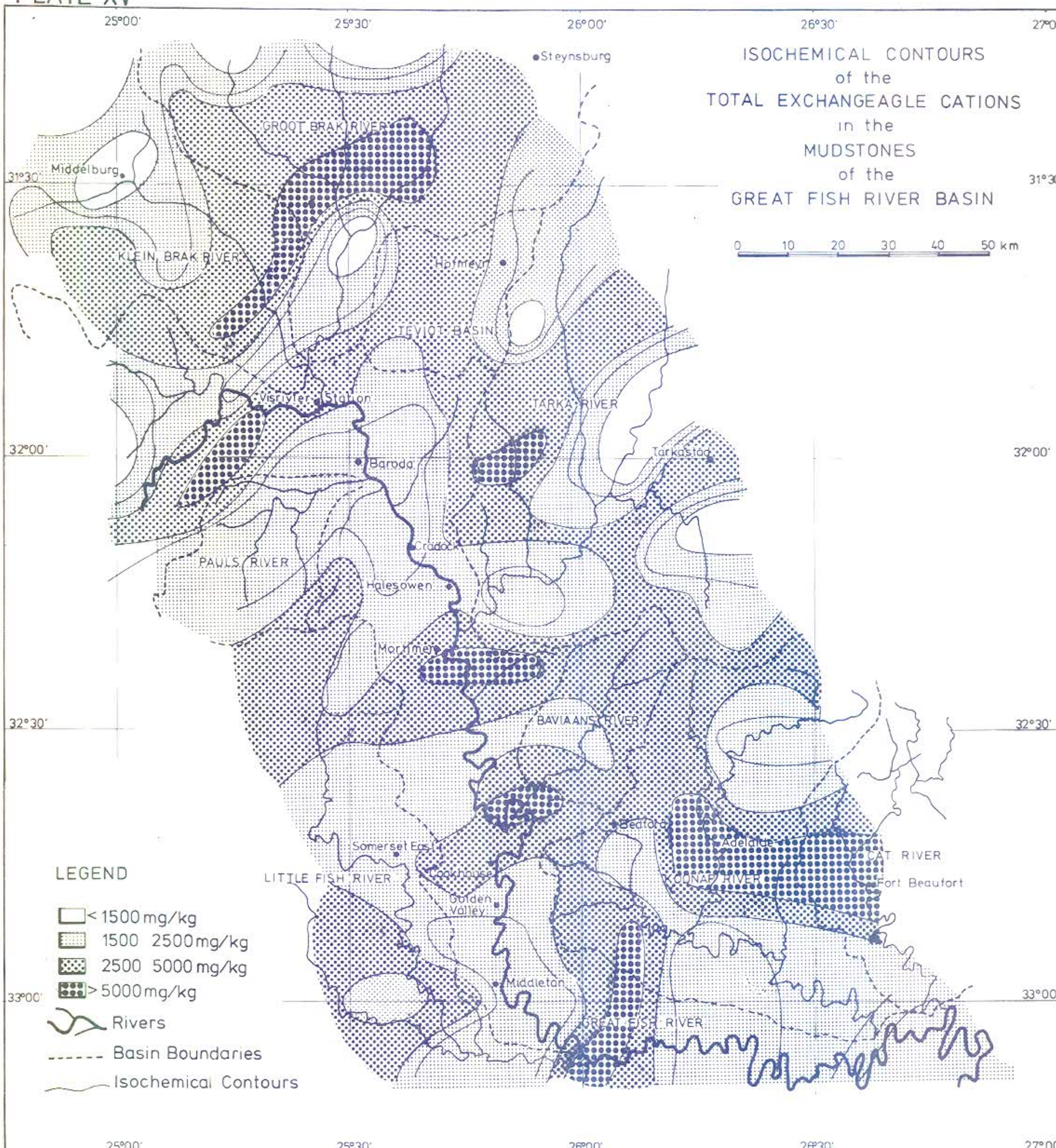

ISOCHEMICAL CONTOURS
of the
EXCHANGEABLE MAGNESIUM
in the
MUDSTONES
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
Middelburg
GROOT BRAK RIVER
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
Visrivier Station
TARKA RIVER
Tarkastad
Baroda
PAULS RIVER
Cradock
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
Somerset East
LITTLE FISH RIVER
Cookhouse
Golden Valley
KOONAP RIVER
CAT RIVER
Fort Beaufort
Middleton
GREAT FISH RIVER
LEGEND
< 300 mg/kg
300 400 mg/kg
400 500 mg/kg
> 500 mg/kg
Rivers
Basin Boundaries
Isochemical Contours

PLATE XIV



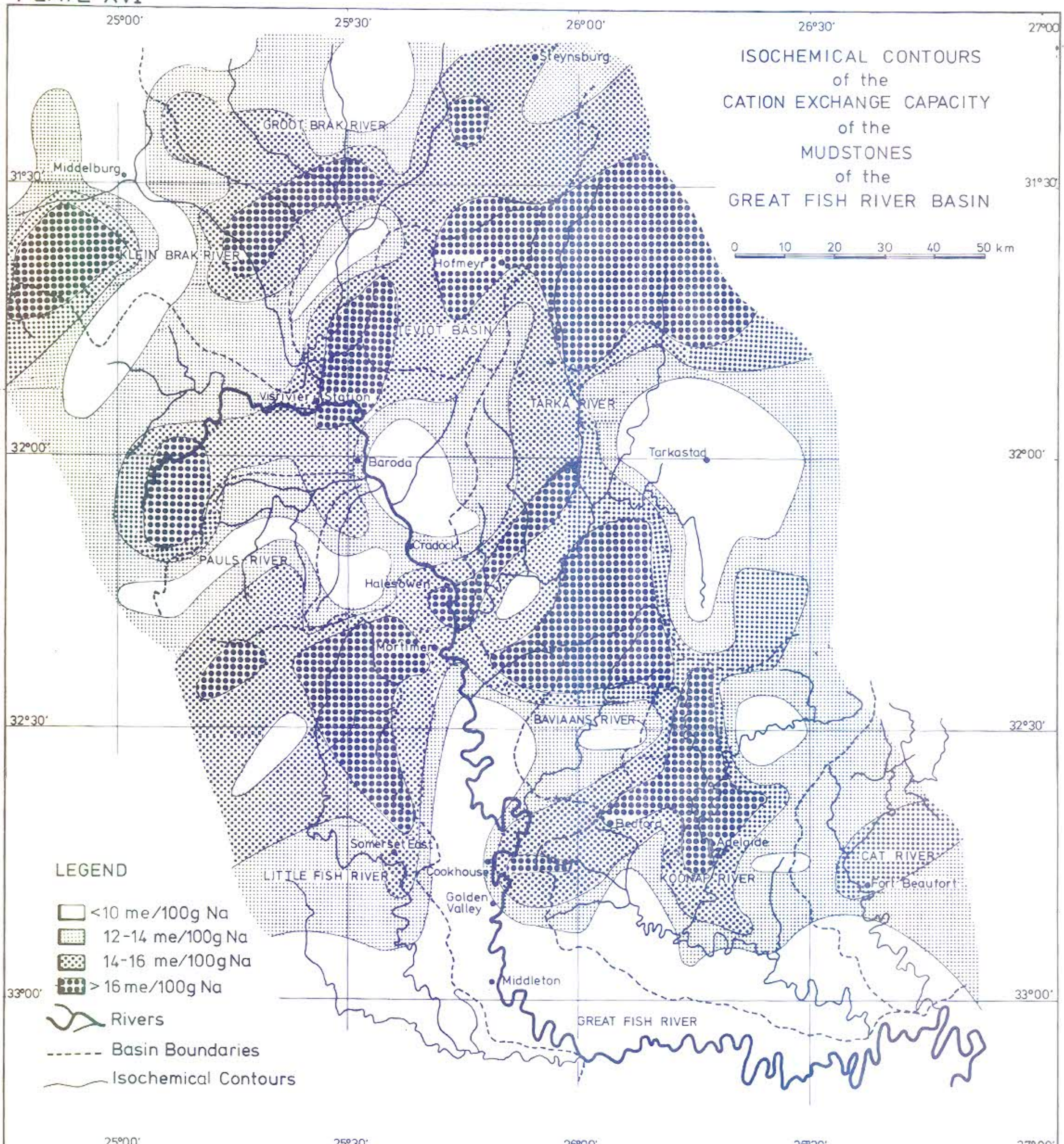
ISOCHEMICAL CONTOURS of the PERCENTAGE EXCHANGEABLE MAGNESIUM in the MUDSTONES of the GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
GROOT BRAK RIVER
Middelburg
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
Visrivier Station
TARKA RIVER
Tarkastad
Baroda
Cradock
PAULS RIVER
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
CAT RIVER
Fort Beaufort
Somerset East
LITTLE FISH RIVER
Cookhouse
KOONAP RIVER
Golden Valley
Middleton
GREAT FISH RIVER
LEGEND
< 10%
10-20%
20 30%
> 30%
Rivers
Basin Boundaries
Isochemical Contours

PLATE XV



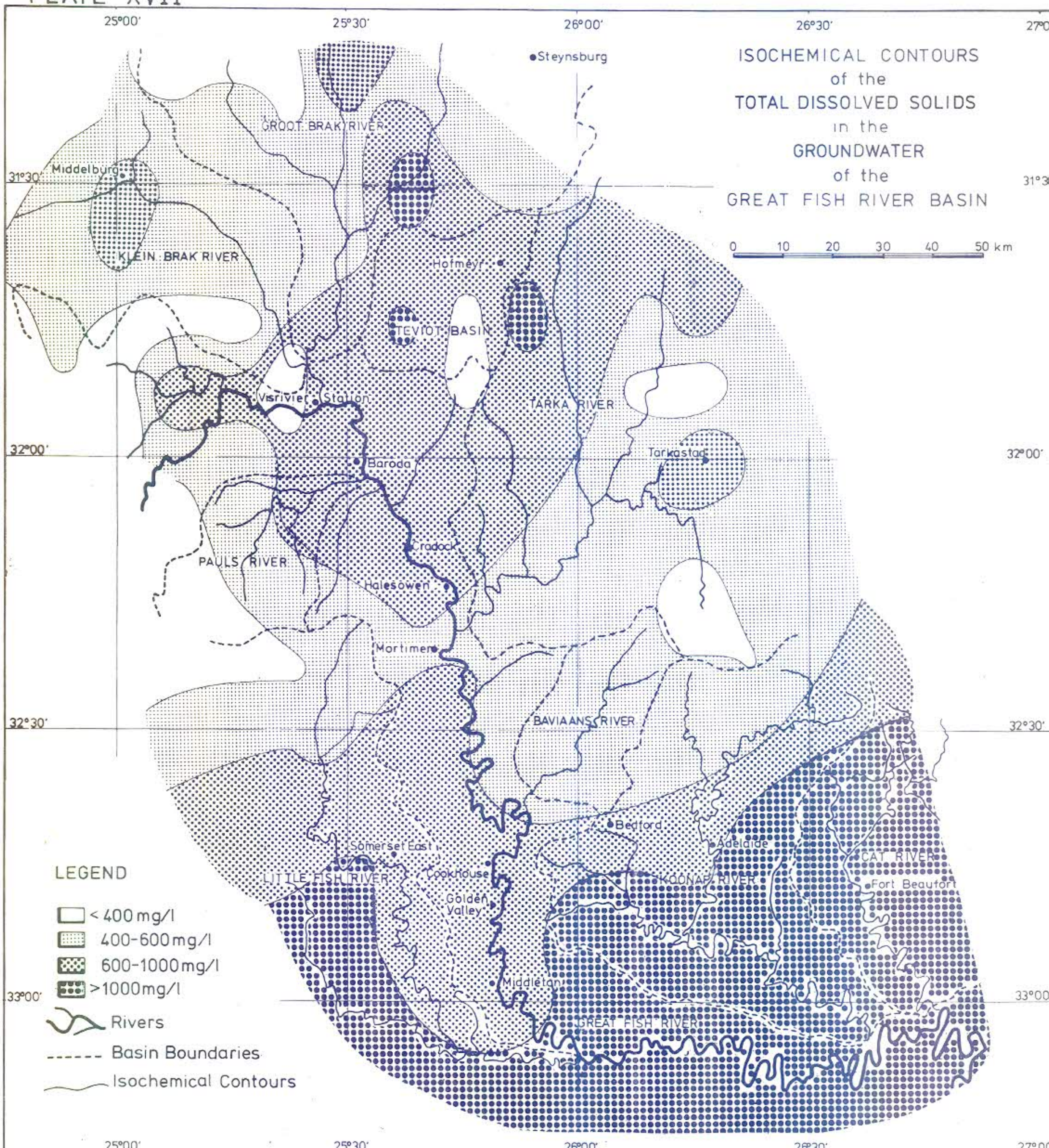
ISOCHEMICAL CONTOURS
of the
TOTAL EXCHANGEAGLE CATIONS
in the
MUDSTONES
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
GROOT BRAK RIVER
Middelburg
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
TARKA RIVER
Tarkastad
Baroda
Cradock
PAULS RIVER
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
CAT RIVER
Fort Beaufort
KOONAP RIVER
Somerset East
LITTLE FISH RIVER
Cookhouse
Golden Valley
Middleton
GREAT FISH RIVER
LEGEND
< 1500 mg/kg
1500 2500 mg/kg
2500 5000 mg/kg
> 5000 mg/kg
Rivers
Basin Boundaries
Isochemical Contours

PLATE XVI



ISOCHEMICAL CONTOURS
of the
CATION EXCHANGE CAPACITY
of the
MUDSTONES
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
GROOT BRAK RIVER
Middelburg
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
TARKA RIVER
Tarkastad
Baroda
Cradock
PAULS RIVER
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
Somerset East
LITTLE FISH RIVER
Cookhouse
Golden Valley
KOONAP RIVER
CAT RIVER
Fort Beaufort
Middleton
GREAT FISH RIVER
LEGEND
<10 me/100g Na
12-14 me/100g Na
14-16 me/100g Na
> 16 me/100g Na
Rivers
Basin Boundaries
Isochemical Contours

PLATE XVII



ISOCHEMICAL CONTOURS
of the
TOTAL DISSOLVED SOLIDS
in the
GROUNDWATER
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
GROOT BRAK RIVER
Middelburg
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
Visrivier Station
TARKA RIVER
Tarkastad
Baroda
Cradock
PAULS RIVER
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
Somerset East
LITTLE FISH RIVER
Cookhouse
KOONAP RIVER
CAT RIVER
Fort Beaufort
Golden Valley
Middleton
GREAT FISH RIVER
LEGEND
< 400 mg/l
400-600 mg/l
600-1000 mg/l
>1000 mg/l
Rivers
Basin Boundaries
Isochemical Contours

PLATE XVIII

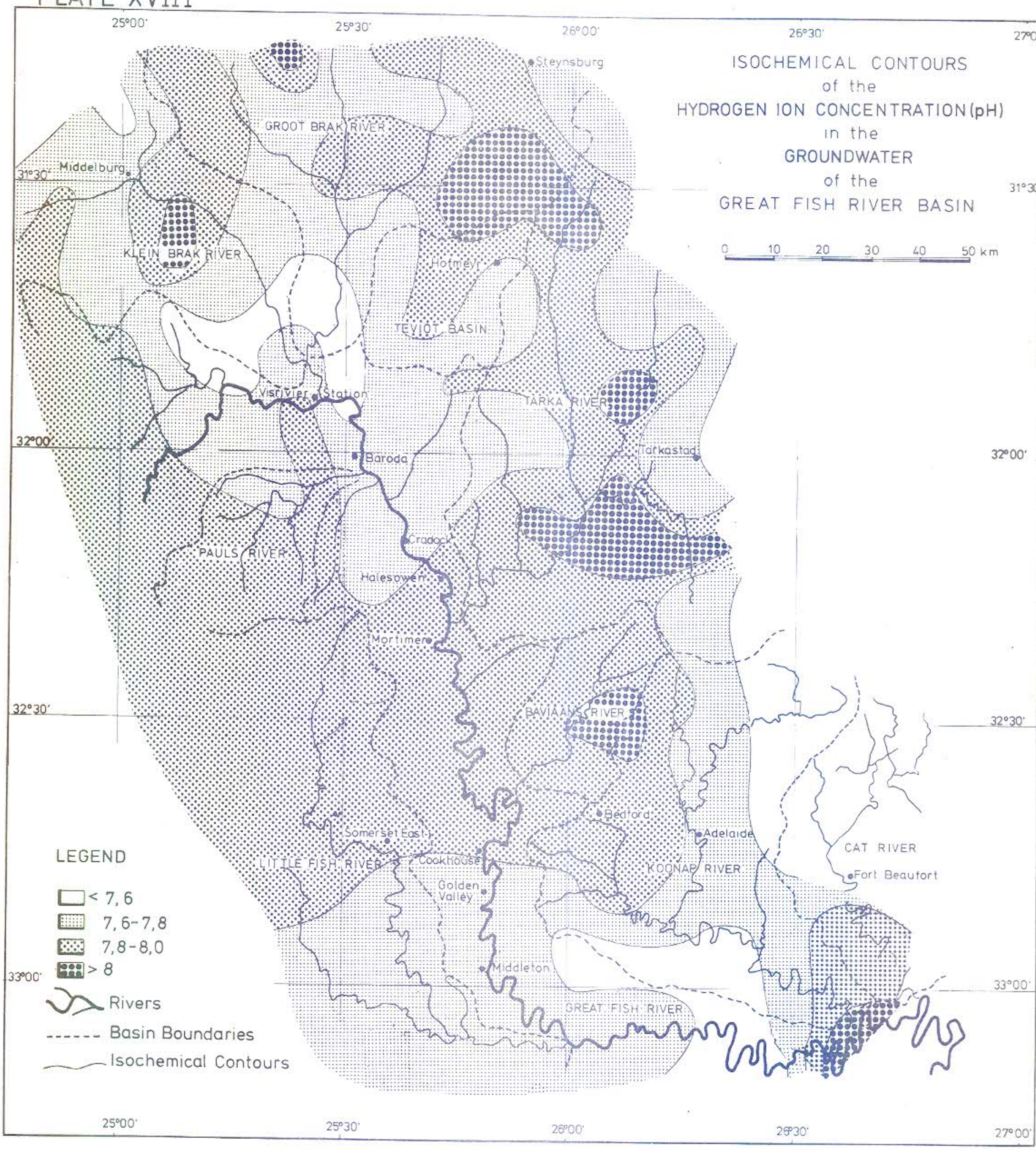
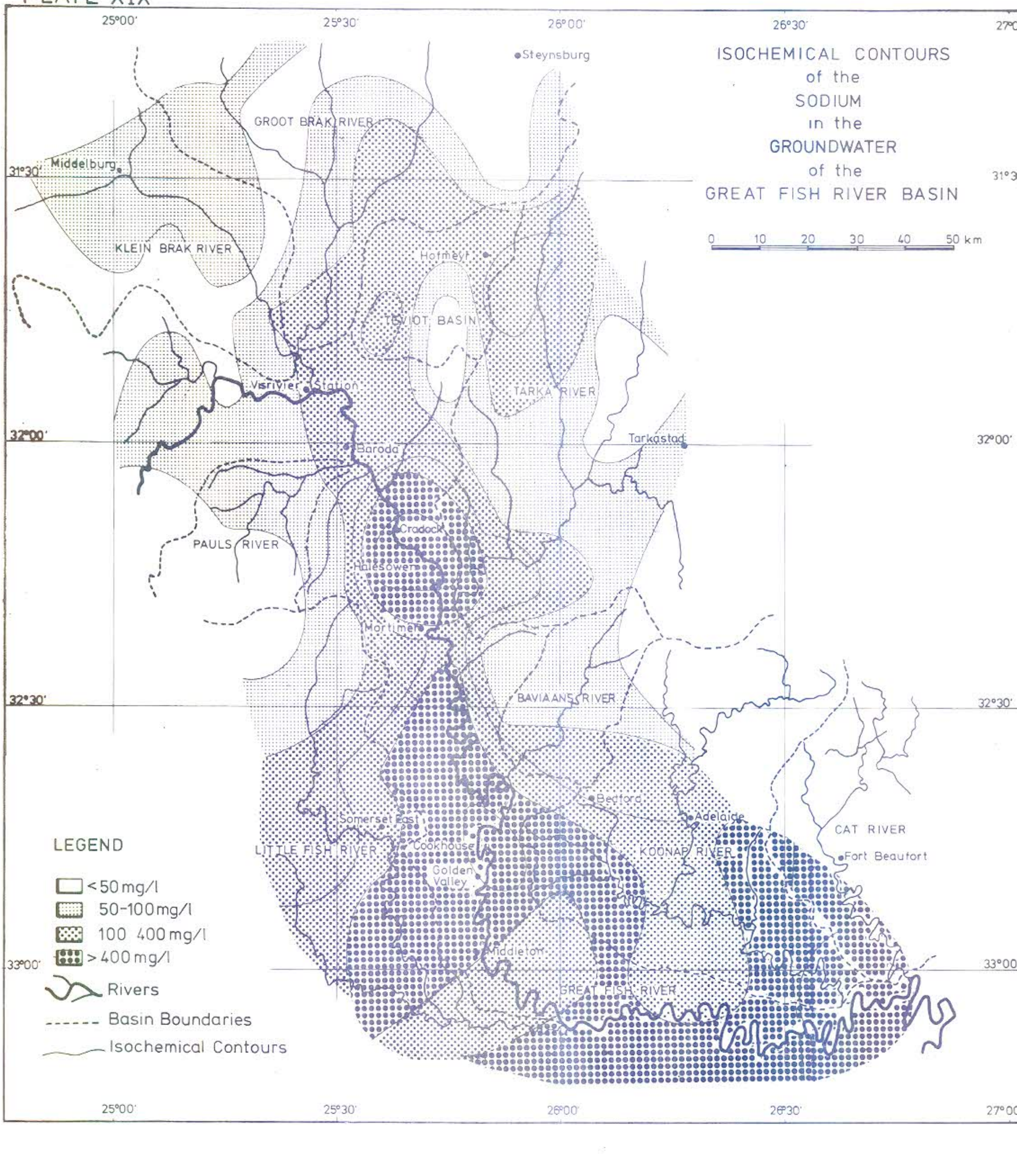

ISOCHEMICAL CONTOURS
of the
HYDROGEN ION CONCENTRATION (pH)
in the
GROUNDWATER
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
GROOT BRAK RIVER
Middelburg
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
Visrivier Station
TARKA RIVER
Baroda
Tarkastad
PAULS RIVER
Cradock
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Somerset East
Adelaide
LITTLE FISH RIVER
Cookhouse
KOONAP RIVER
CAT RIVER
Fort Beaufort
Golden Valley
Middleton
GREAT FISH RIVER
LEGEND
< 7,6
7,6-7,8
7,8-8,0
> 8
Rivers
Basin Boundaries
Isochemical Contours

PLATE XIX



ISOCHEMICAL CONTOURS of the SODIUM in the GROUNDWATER of the GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
Middelburg
GROOT BRAK RIVER
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
TARKA RIVER
Tarkastad
Visrivier Station
Baroda
PAULS RIVER
Cradock
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
Somerset East
LITTLE FISH RIVER
Cookhouse
Golden Valley
KOONAP RIVER
CAT RIVER
Fort Beaufort
Middleton
GREAT FISH RIVER
LEGEND
<50 mg/l
50-100 mg/l
100 400 mg/l
> 400 mg/l
Rivers
Basin Boundaries
Isochemical Contours

PLATE XX

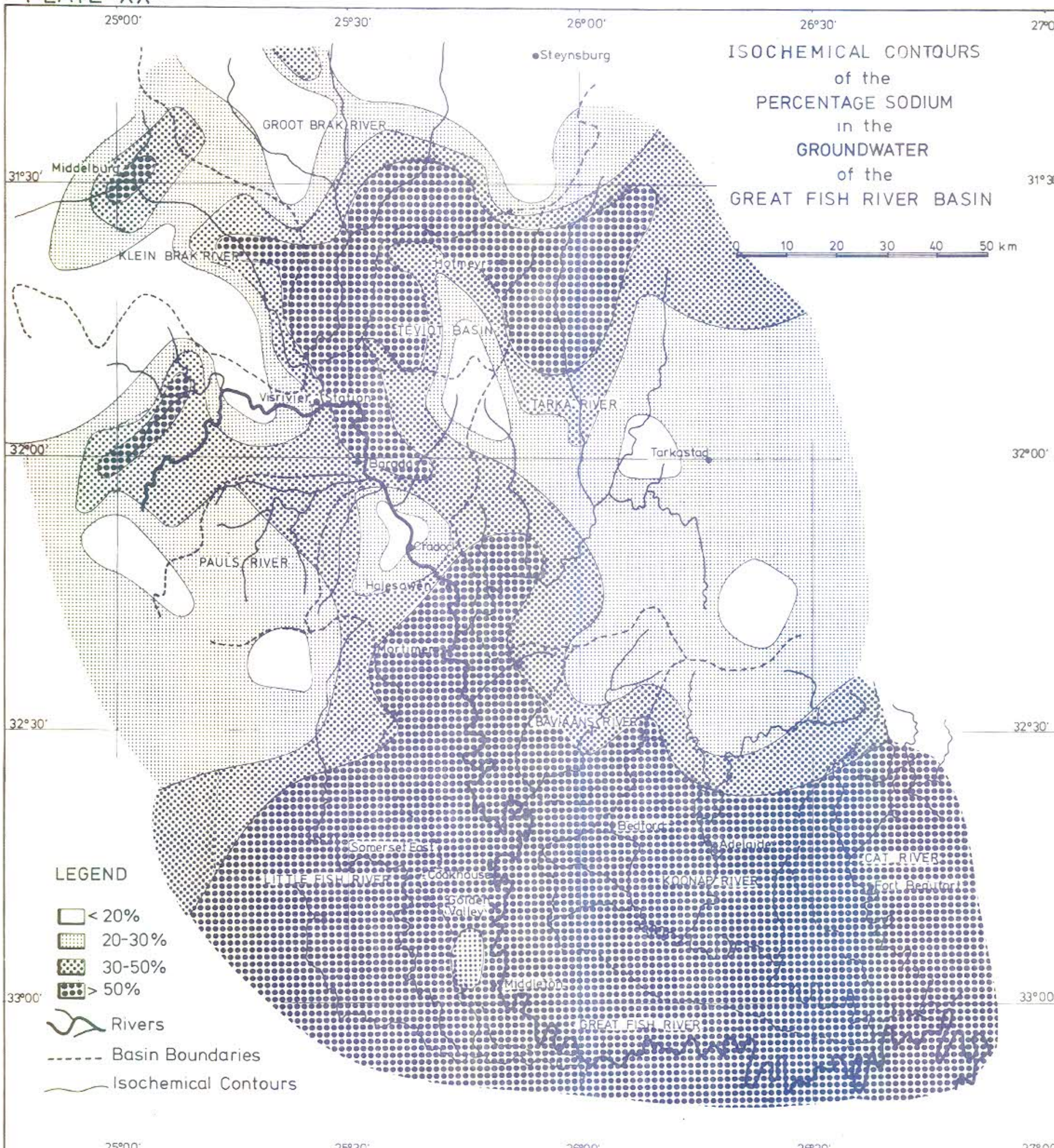
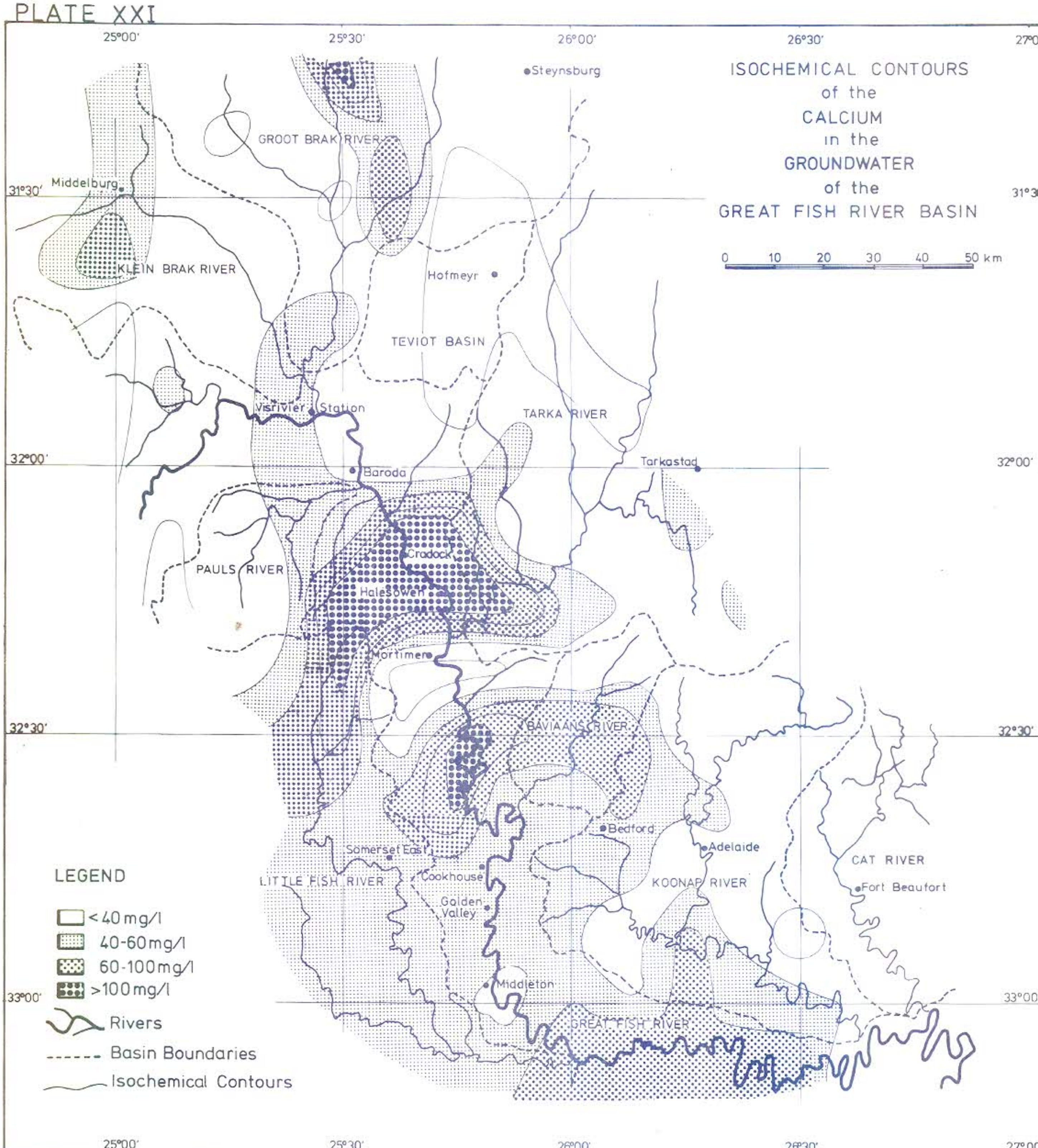
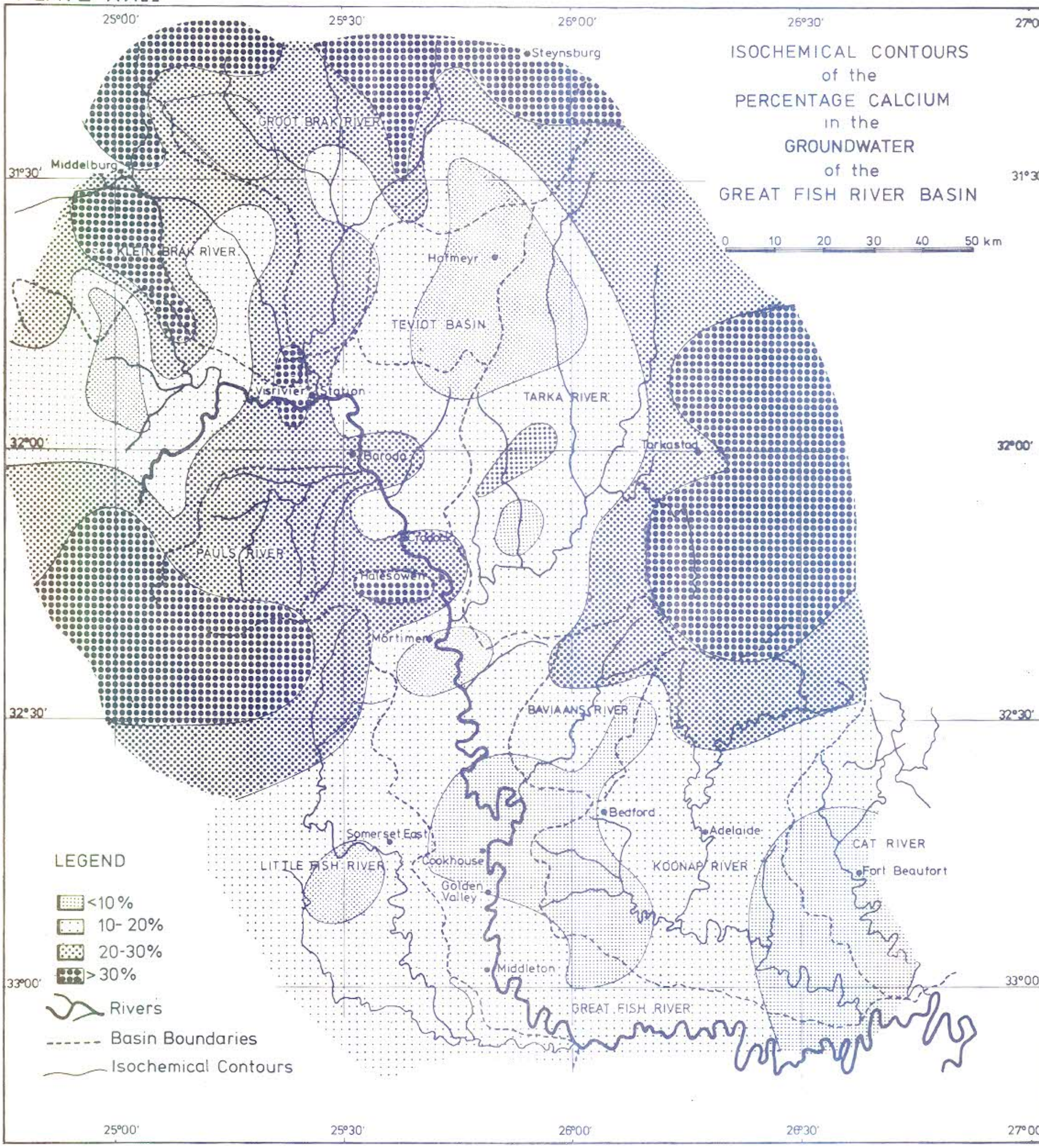

ISOCHEMICAL CONTOURS
of the
PERCENTAGE SODIUM
in the
GROUNDWATER
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
GROOT BRAK RIVER
Middelburg
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
Visrivier Station
TARKA RIVER
Tarkastad
Baroda
Cradock
PAULS RIVER
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Somerset East
Adelaide
CAT RIVER
LITTLE FISH RIVER
Cookhouse
KOONAP RIVER
Fort Beaufort
Golden Valley
Middleton
GREAT FISH RIVER
LEGEND
< 20%
20-30%
30-50%
> 50%
Rivers
Basin Boundaries
Isochemical Contours

PLATE XXI



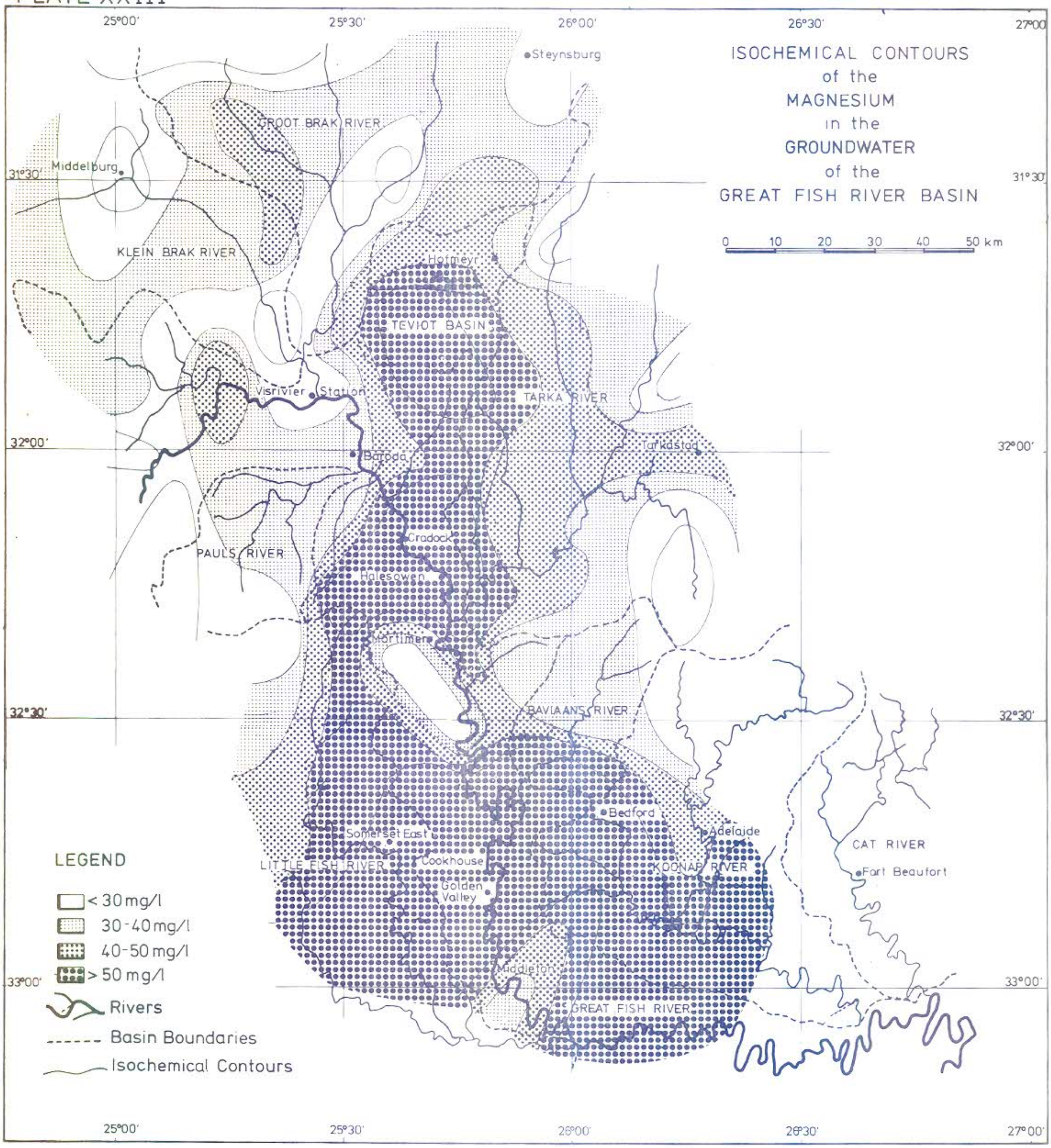
ISOCHEMICAL CONTOURS
of the
CALCIUM
in the
GROUNDWATER
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
GROOT BRAK RIVER
Middelburg
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
Visrivier Station
TARKA RIVER
Tarkastad
Baroda
PAULS RIVER
Cradock
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
Somerset East
Cookhouse
LITTLE FISH RIVER
KOONAP RIVER
CAT RIVER
Fort Beaufort
Golden Valley
Middleton
GREAT FISH RIVER
LEGEND
< 40 mg/l
40-60 mg/l
60-100 mg/l
>100 mg/l
Rivers
Basin Boundaries
Isochemical Contours

PLATE XXII



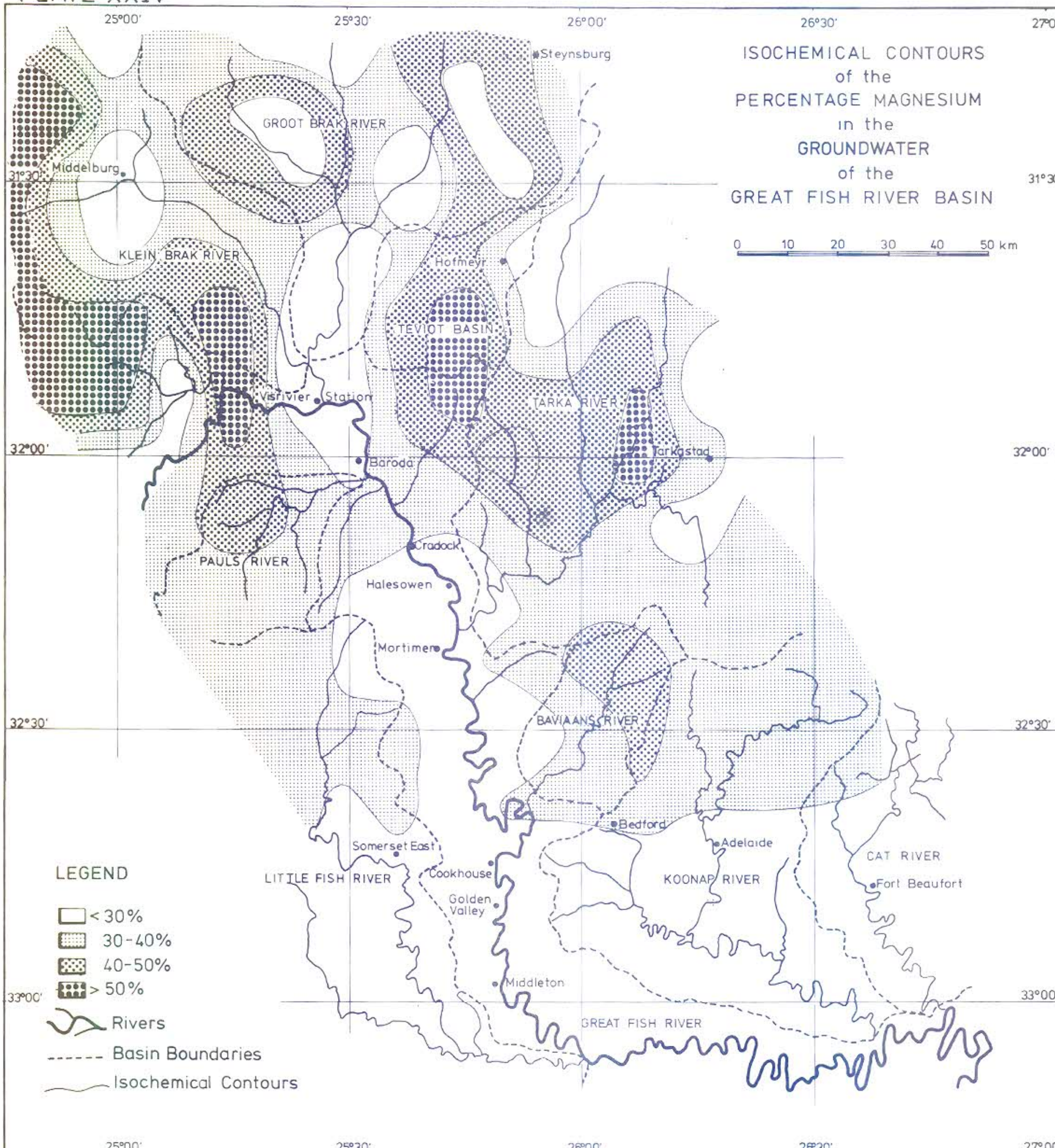
ISOCHEMICAL CONTOURS
of the
PERCENTAGE CALCIUM
in the
GROUNDWATER
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
Middelburg
GROOT BRAK RIVER
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
Visrivier Station
TARKA RIVER
Tarkastad
Baroda
PAULS RIVER
Cradock
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
Somerset East
CAT RIVER
LITTLE FISH RIVER
Cookhouse
KOONAP RIVER
Fort Beaufort
Golden Valley
Middleton
GREAT FISH RIVER
LEGEND
<10%
10- 20%
20-30%
> 30%
Rivers
Basin Boundaries
Isochemical Contours

PLATE XXIII



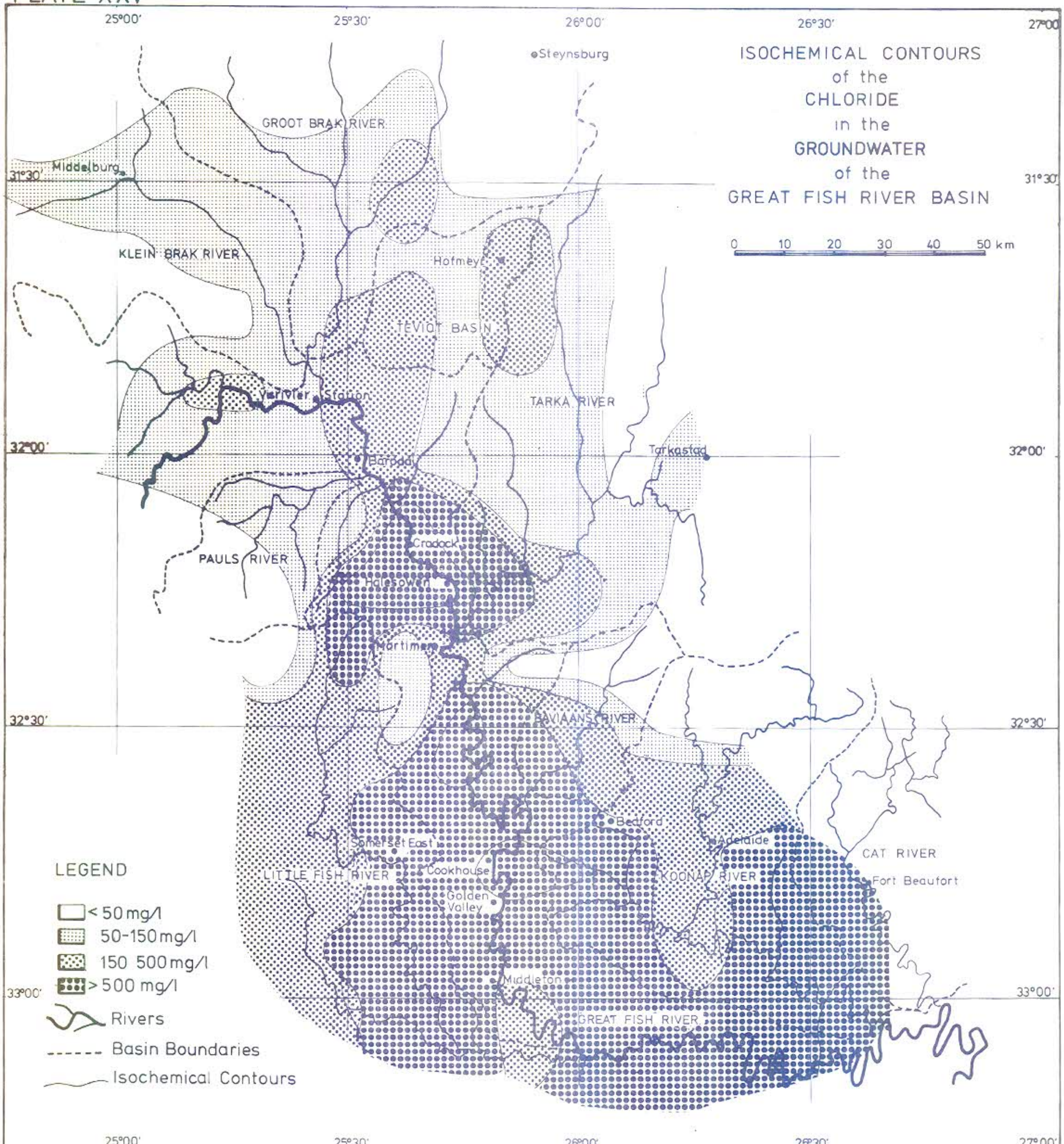
ISOCHEMICAL CONTOURS
of the
MAGNESIUM
in the
GROUNDWATER
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
GROOT BRAK RIVER
Middelburg
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
Visrivier
Station
TARKA RIVER
Tarkastad
Baroda
Cradock
PAULS RIVER
Halesoven
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
Somerset East
Cookhouse
LITTLE FISH RIVER
KOONAP RIVER
CAT RIVER
Fort Beaufort
Golden Valley
Middleton
GREAT FISH RIVER
LEGEND
< 30 mg/l
30-40 mg/l
40-50 mg/l
> 50 mg/l
Rivers
Basin Boundaries
Isochemical Contours

PLATE XXIV



ISOCHEMICAL CONTOURS of the PERCENTAGE MAGNESIUM in the GROUNDWATER of the GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
GROOT BRAK RIVER
Middelburg
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
Visrivier
Station
TARKA RIVER
Tarkastad
Baroda
PAULS RIVER
Cradock
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
Somerset East
CAT RIVER
LITTLE FISH RIVER
Cookhouse
KOONAP RIVER
Fort Beaufort
Golden Valley
Middleton
GREAT FISH RIVER
LEGEND
< 30%
30-40%
40-50%
> 50%
Rivers
Basin Boundaries
Isochemical Contours

PLATE XXV



ISOCHEMICAL CONTOURS
of the
CHLORIDE
in the
GROUNDWATER
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
GROOT BRAK RIVER
Middelburg
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
TARKA RIVER
Tarkastad
Baroda
Cradock
PAULS RIVER
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
CAT RIVER
Fort Beaufort
Somerset East
Cookhouse
LITTLE FISH RIVER
KOONAP RIVER
Golden Valley
Middleton
GREAT FISH RIVER
LEGEND
< 50 mg/l
50-150 mg/l
150 500 mg/l
> 500 mg/l
Rivers
Basin Boundaries
Isochemical Contours

PLATE XXVI

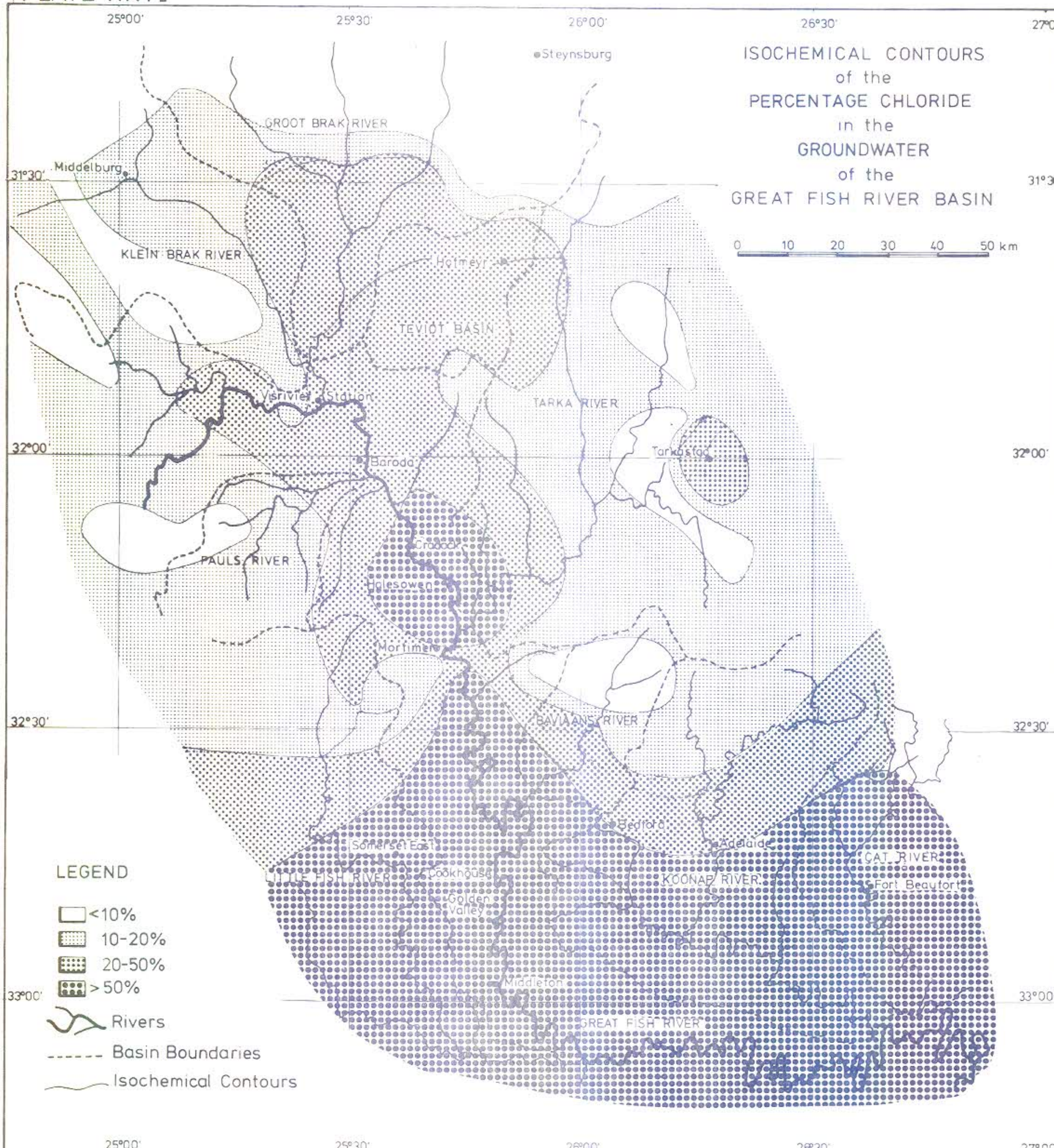

ISOCHEMICAL CONTOURS
of the
PERCENTAGE CHLORIDE
in the
GROUNDWATER
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
Middelburg
GROOT BRAK RIVER
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
TARKA RIVER
Tarkastad
Visrivier Station
Baroda
PAULS RIVER
Cradock
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
Somerset East
LITTLE FISH RIVER
Cookhouse
Golden Valley
KOONAP RIVER
CAT RIVER
Fort Beaufort
Middleton
GREAT FISH RIVER
LEGEND
<10%
10-20%
20-50%
>50%
Rivers
Basin Boundaries
Isochemical Contours

PLATE XXVII

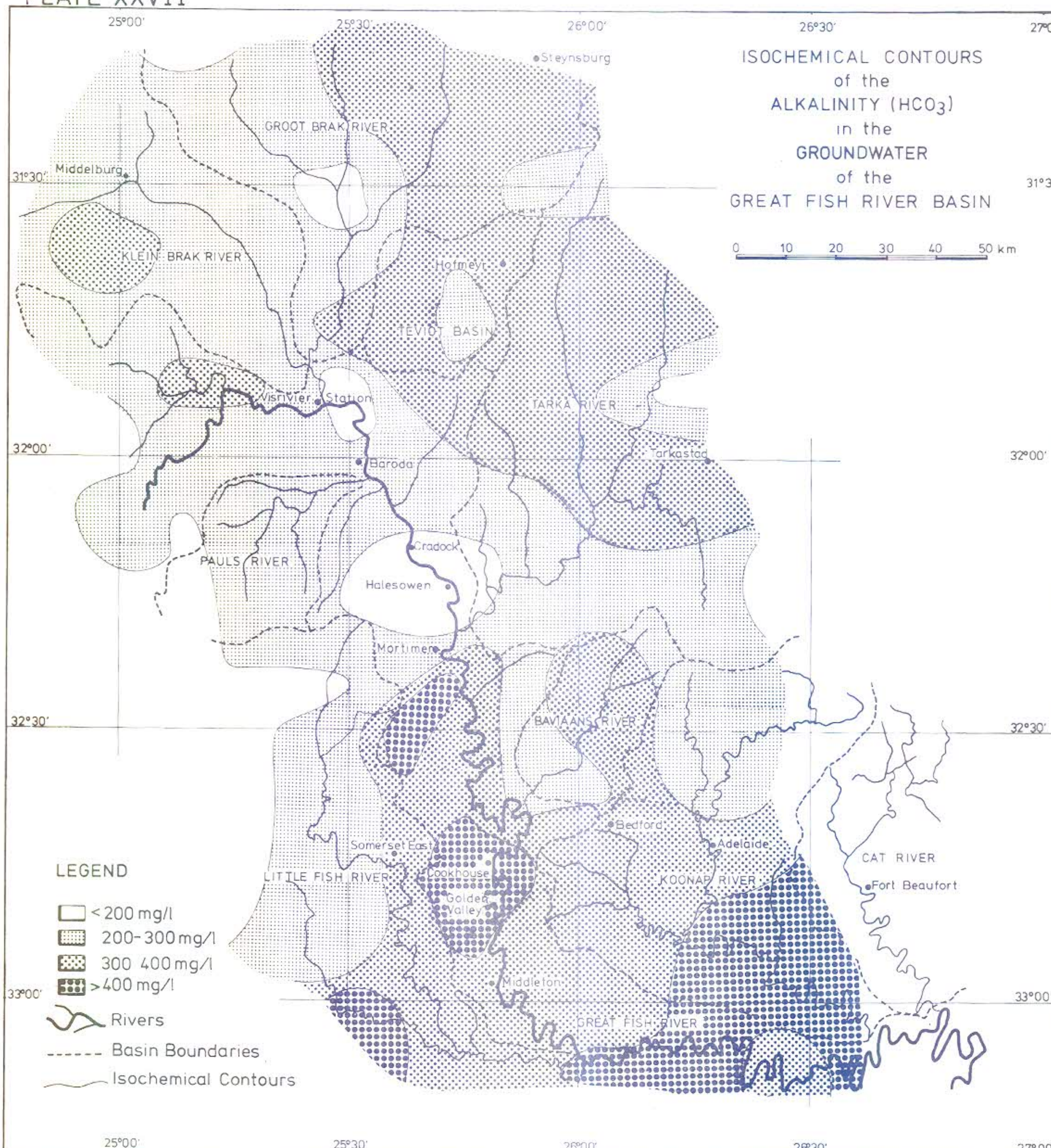

ISOCHEMICAL CONTOURS
of the
ALKALINITY (HCO_3)
in the
GROUNDWATER
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
GROOT BRAK RIVER
Middelburg
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
Visrivier Station
TARKA RIVER
Tarkastad
Baroda
PAULS RIVER
Cradock
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Somerset East
Adelaide
CAT RIVER
LITTLE FISH RIVER
Cookhouse
KOONAP RIVER
Fort Beaufort
Golden Valley
Middleton
GREAT FISH RIVER
LEGEND
< 200 mg/l
200-300 mg/l
300 400 mg/l
> 400 mg/l
Rivers
Basin Boundaries
Isochemical Contours

PLATE XXVIII

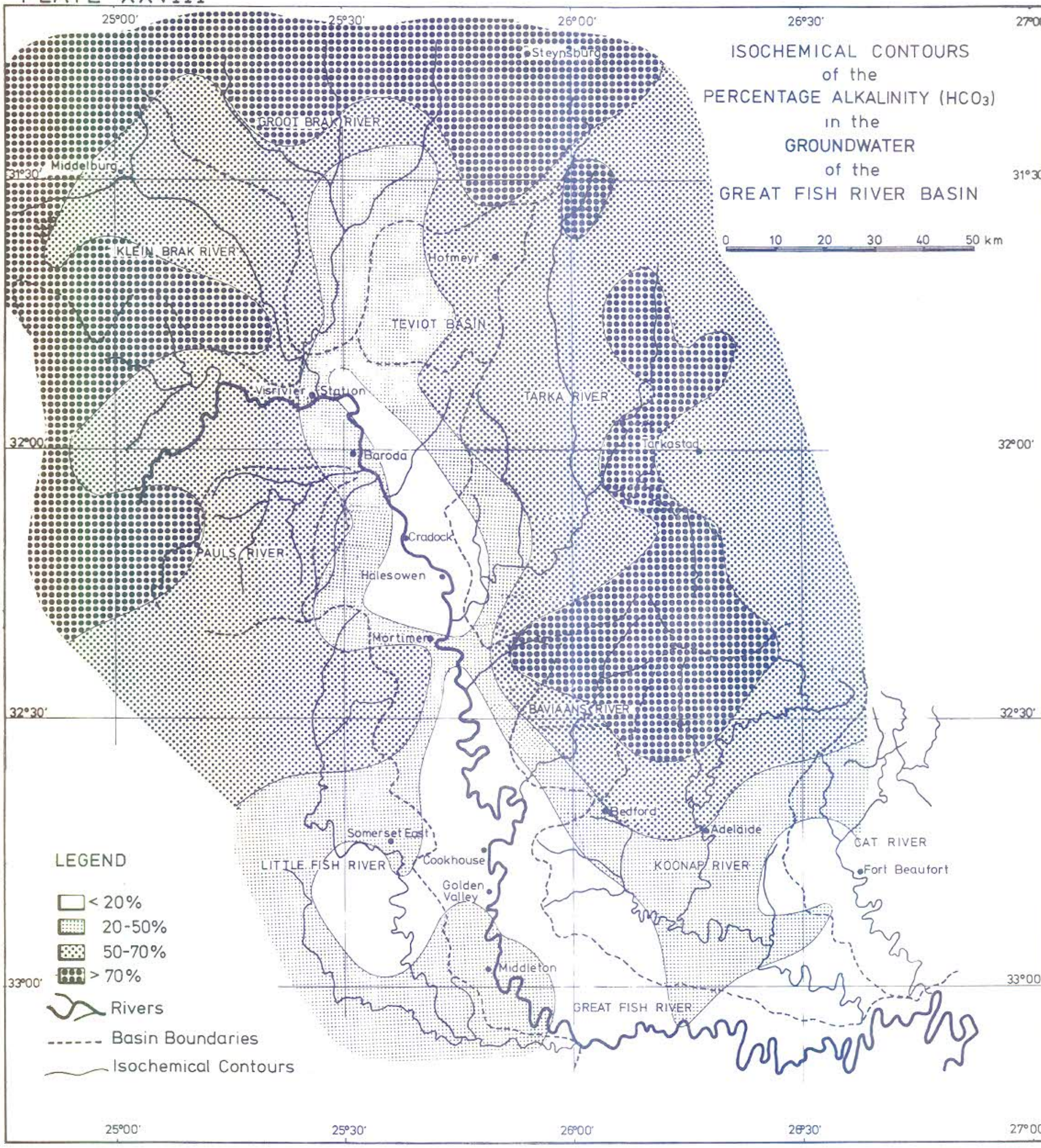


PLATE XXIX

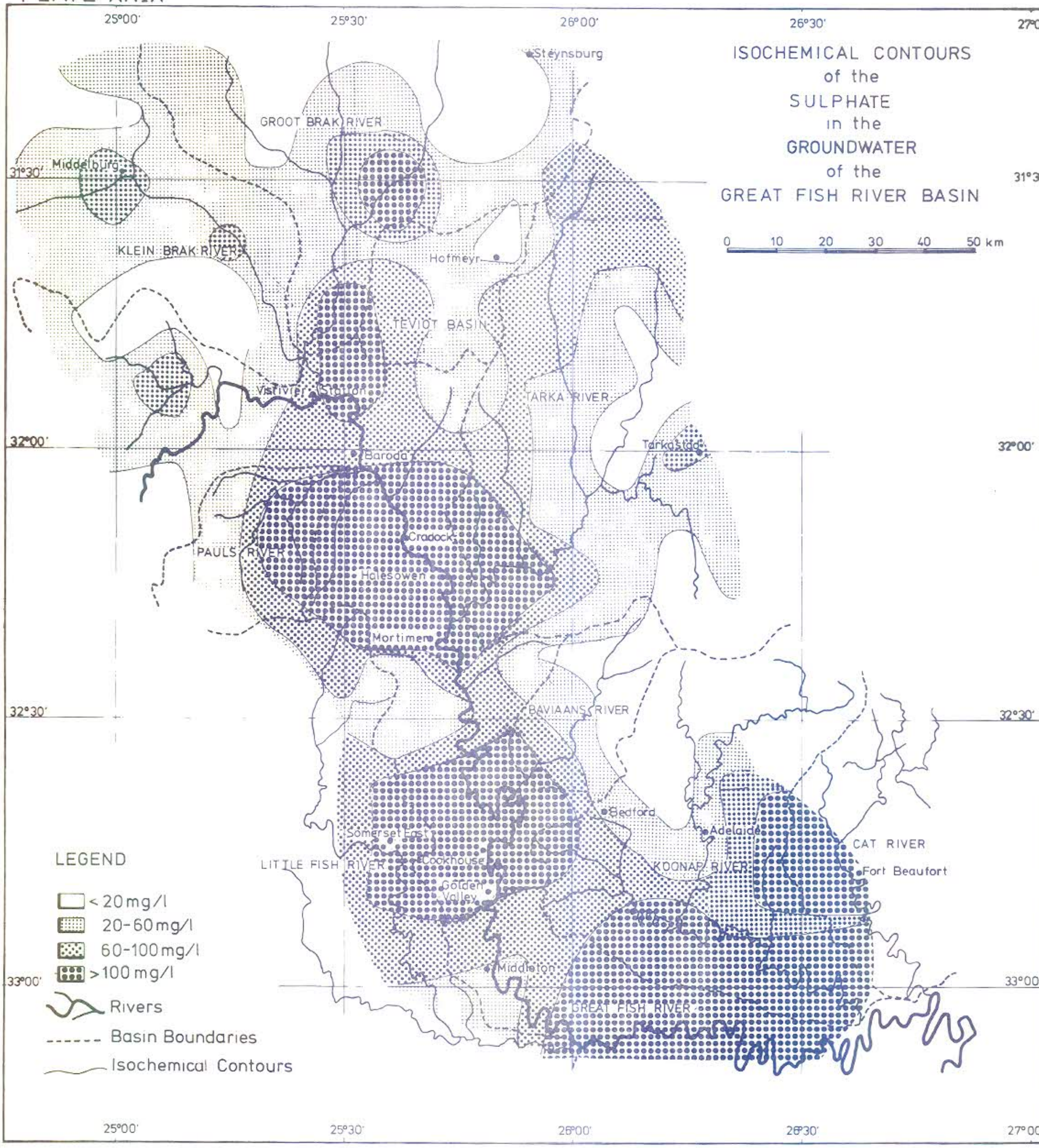

ISOCHEMICAL CONTOURS
of the
SULPHATE
in the
GROUNDWATER
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
GROOT BRAK RIVER
Middelburg
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
Visrivier Station
TARKA RIVER
Tarkastad
Baroda
Cradock
PAULS RIVER
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
CAT RIVER
Fort Beaufort
Somerset East
LITTLE FISH RIVER
Cookhouse
Golden Valley
KOONAP RIVER
Middleton
GREAT FISH RIVER
LEGEND
< 20 mg/l
20-60 mg/l
60-100 mg/l
> 100 mg/l
Rivers
Basin Boundaries
Isochemical Contours

PLATE XXX

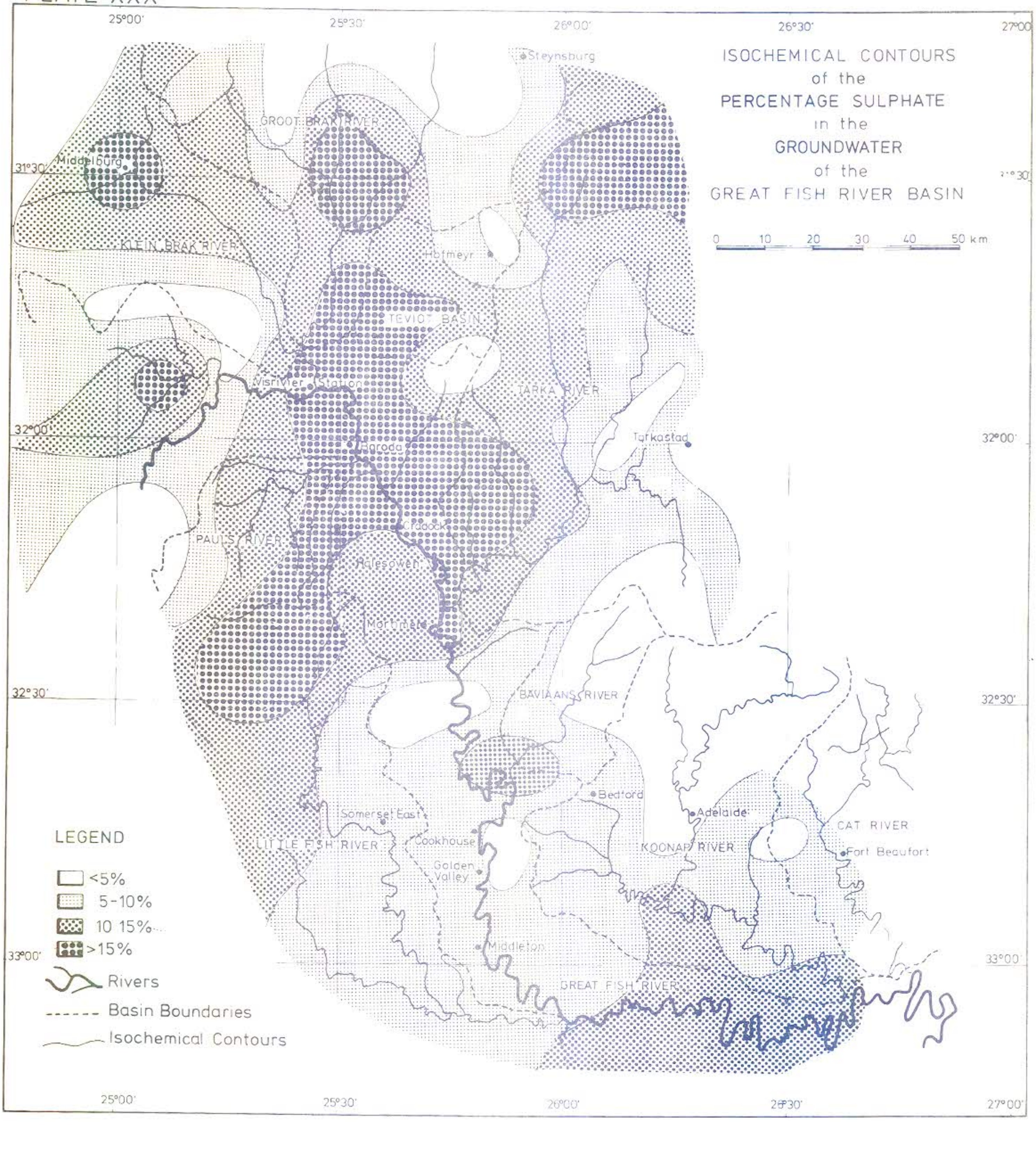
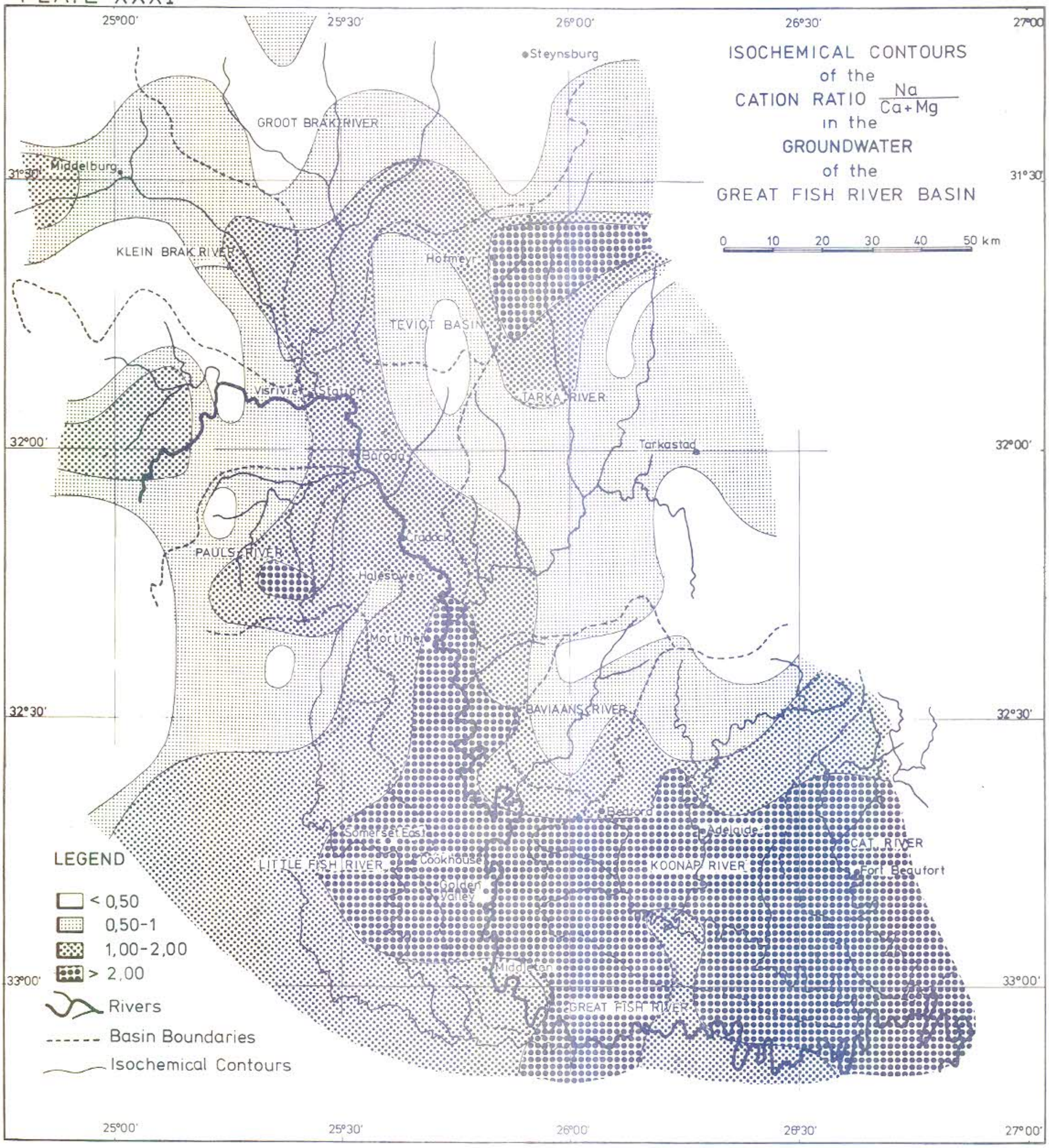
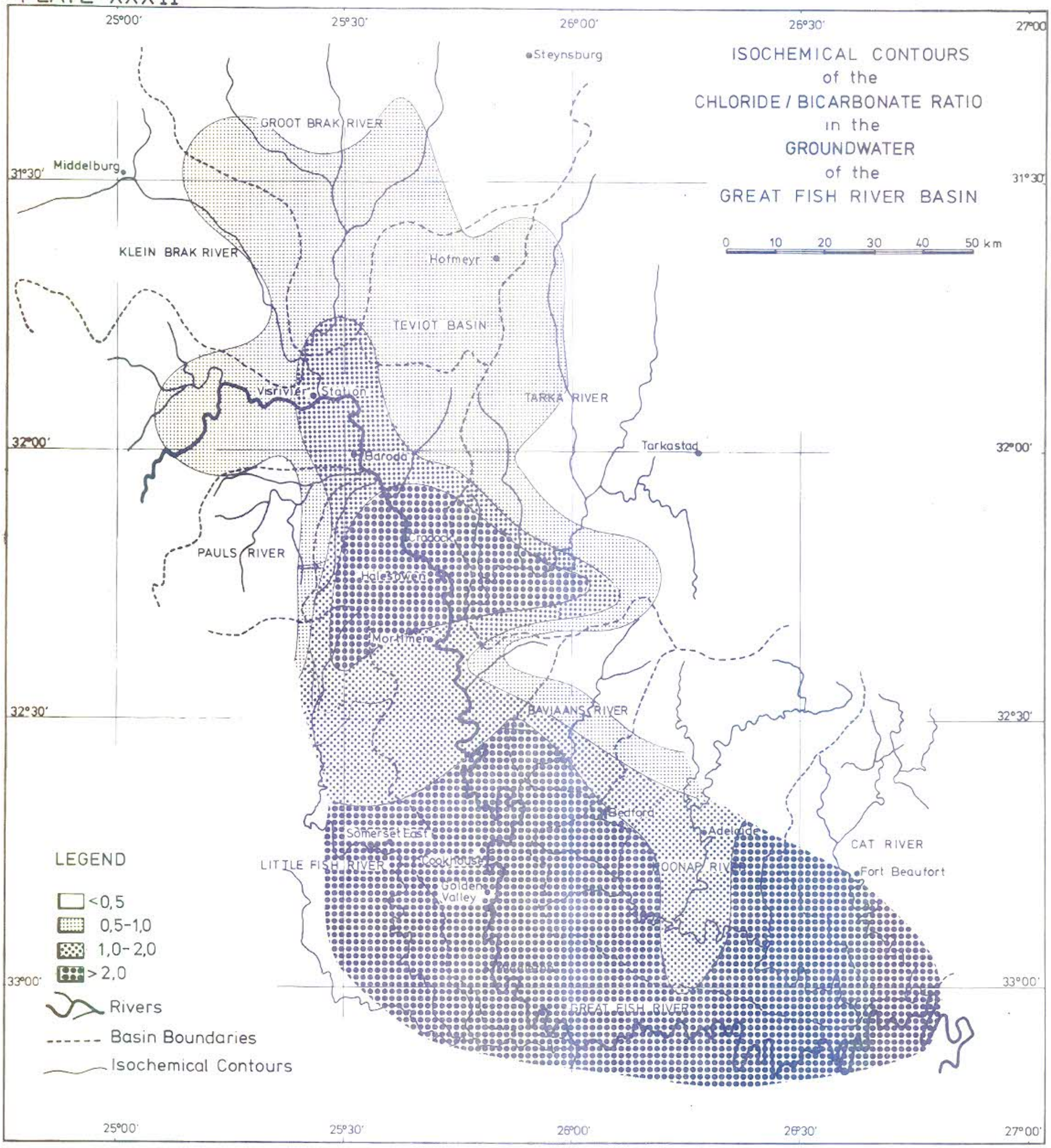

ISOCHEMICAL CONTOURS
of the
PERCENTAGE SULPHATE
in the
GROUNDWATER
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
Middelburg
GROOT BRAK RIVER
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
Visrivier Station
TARKA RIVER
Tarkastad
Baroda
Cradock
PAULS RIVER
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
Somerset East
LITTLE FISH RIVER
Cookhouse
Golden Valley
KOONAP RIVER
CAT RIVER
Fort Beaufort
Middleton
GREAT FISH RIVER
LEGEND
<5%
5-10%
10 15%
>15%
Rivers
Basin Boundaries
Isochemical Contours

PLATE XXXI



ISOCHEMICAL CONTOURS of the CATION RATIO $\frac{Na}{Ca+Mg}$ in the GROUNDWATER of the GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
GROOT BRAK RIVER
Middelburg
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
TARKA RIVER
Visrivier Station
Baroda
Tarkastad
PAULS RIVER
Cradock
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Somerset East
Adelaide
CAT RIVER
LITTLE FISH RIVER
Cookhouse
KOONAP RIVER
Fort Beaufort
Golden Valley
Middleton
GREAT FISH RIVER
LEGEND
< 0,50
0,50-1
1,00-2,00
> 2,00
Rivers
Basin Boundaries
Isochemical Contours

PLATE XXXII



ISOCHEMICAL CONTOURS
of the
CHLORIDE / BICARBONATE RATIO
in the
GROUNDWATER
of the
GREAT FISH RIVER BASIN
0 10 20 30 40 50 km
25°00'
25°30'
26°00'
26°30'
27°00'
31°30'
32°00'
32°30'
33°00'
Steynsburg
Middelburg
GROOT BRAK RIVER
KLEIN BRAK RIVER
Hofmeyr
TEVIOT BASIN
Visrivier
Station
TARKA RIVER
Tarkastad
Baroda
PAULS RIVER
Cradock
Halesowen
Mortimer
BAVIAANS RIVER
Bedford
Adelaide
Somerset East
Cookhouse
Golden Valley
LITTLE FISH RIVER
KOONAP RIVER
CAT RIVER
Fort Beaufort
GREAT FISH RIVER
LEGEND
<0,5
0,5-1,0
1,0-2,0
>2,0
Rivers
Basin Boundaries
Isochemical Contours

PLATE XXXIII