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EVIDENCE FOR THE FORMER EXISTENCE OF A MAJOR, SOUTHERLY FLOWING RIVER IN GRIQUALAND WEST

by

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ABSTRACT

Banded iron-formation clasts appear in quantity in the alluvial gravels on high level Orange River terraces some 30 km downstream from the Vaal River confluence at a point a considerable distance upstream from where this rock type becomes a significant constituent of the catchment bedrock. The point of first appearance of banded iron-formation is associated with an impressive valley on the right bank of the Orange River, extending northwards towards Griekwastad and beyond, and containing vast gravel deposits in which banded iron-formation is the dominant clast type. Comparison of the clast assemblages of these gravels with those along the Orange River both upstream and downstream of the right bank valley indicates that the change in Orange River gravel composition was caused by entry into the Orange of a major river. Over the time period recorded in the high level terraces this river was at least equal to, or may have contained as much as four times the volume of water of the Orange. The river appears to have ceased flowing very suddenly, its demise apparently related to crustal tectonism along the Transvaal–Griqualand Axis. Cessation of discharge from this river may have coincided with the commencement of the infilling of the Kalahari Basin.

CONTENTS

I. INTRODUCTION

Speculation concerning the evolution of the drainage system in Griqualand West has a long history. Du Toit (1910) first drew attention to the importance of exhumation of old drainage systems in the area, and suggested that the present courses of the Harts, Orange and Vaal rivers between Vryburg and Prieska mirror a river system which existed in pre-Dwyka times. More recently, Helgren (1979a) provided additional detailed data to support this suggestion and went so far as to state that the extent of erosion of the pre-Dwyka Formation basement rocks (Ventersdorp and Transvaal sequences) by the Vaal and the Orange rivers was minimal. Helgren also drew attention to the influence which major crustal fractures have had on modern river systems.

Du Toit (1933) speculated on the influence exerted by crustal flexure on drainage patterns in Griqualand West, and further afield. A major warp axis was proposed, extending from Bushmanland in a north-easterly direction, crossing the Orange River between Upington and Prieska and terminating in the Witwatersrand. This he termed the Transvaal–Griqualand Axis. Northerly subsidence along this axis was considered to be the dominant tectonic factor, and was thought to have had considerable effect on the drainage basins of the Molopo, Limpopo and Vaal–Orange river systems. Mayer (1973) examined the drainage in the lower Vaal basin in this context, and suggested certain modifications to the location of the warp axis. His interpretation represented a deviation from Du Toit (1933), in that he considered actual upwarp to have occurred along the axis.

A remarkable feature of the Western Transvaal is the abundance of old river gravels many of which extend from the Vaal River in a north-westerly direction towards the Botswana border. These gravels are unrelated to present-day drainage patterns and have been described in detail by Williams (1932), Du Toit (1951) and, more recently, by Stratten (1979). Sedimentological studies indicated that certain of these gravels were derived from areas well into Eastern Botswana and the Western Transvaal (Stratten, 1979). Stratten, following Du Toit (1933), attributed the cessation of this drainage to Tertiary crustal warping.

In general, the clast assemblage within an alluvial gravel deposit reflects the geology of the catchment area associated with that deposit and, in principle, can be used to define the extent of catchment areas of ancient river systems. In the case of the Vaal and Orange river gravels, particularly in Griqualand West, this principle cannot easily be applied as the rivers traverse the Dwyka

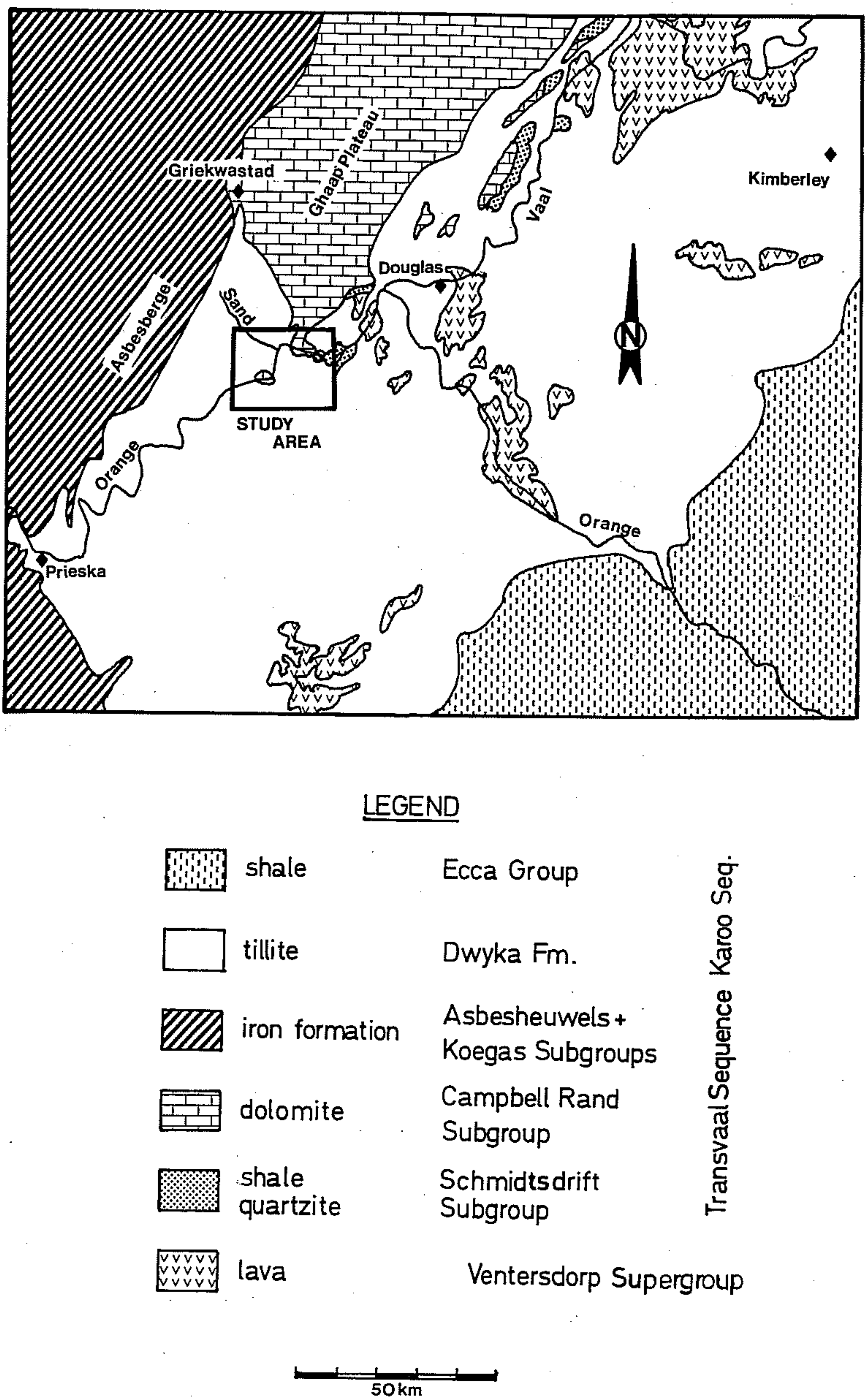

Figure 1
Regional geology and location of the study area.

Formation and it has been shown that the majority of gravel clasts have been derived from this rock type (Helgren, 1979a). Nevertheless, if distinctive lithologies occur, the concept of tracing gravel sources by their clasts can still be used.

During the course of reconnaissance surveys of high level gravel terraces along the Orange River in Griqualand West, an abrupt change in gravel composition was observed to occur some 30 km downstream from the Vaal River confluence (see Fig. 1). At this point, banded iron-formation suddenly becomes a major constituent of the high level gravel terraces. This communication reports results of a detailed examination of this phenomenon.

II. GEOLOGY

A. Basement Geology

The regional geology in the vicinity of the confluence of the Vaal and Orange rivers is shown in Fig. 1. Tillite and shales of the Dwyka Formation are regionally the most important rock types. Lavas of the Ventersdorp Supergroup and sediments of the Transvaal Sequence, both of which underlie the Dwyka Formation in the area, are exposed along the Vaal and Orange river valleys. To the north of the Vaal–Orange confluence lies the Ghaap Plateau, a flat, triangular tract of high ground formed by dolomitic rocks of the Transvaal Sequence. West of the Ghaap Plateau lie the Asbesberge, formed by resistant siliceous banded iron-formation which overlies the dolomitic rocks.

The Ventersdorp Supergroup is represented in the area mainly by green amygdaloidal lava, containing chlorite or agate amygdales. Within the study area (Fig. 2) the lower units of the Transvaal Sequence (Ghaap Group; Beukes, 1980) are represented by shales with interbedded fine-grained quartzites (Schmidtsdrift Subgroup). The quartzites decrease in frequency upwards, while dolomitic horizons make their appearance, becoming more common upwards, until dolomite is the only lithology present (Campbellrand Subgroup). Dolomitic rocks predominate in the triangular tract forming the Ghaap Plateau, but in the west, iron-formations (of the Asbesheuvels Subgroup) are the dominant lithology (Beukes, 1980; and Fig. 1).

The major Dwyka Formation lithology is green to grey diamictite. Shales are locally developed, but were not observed in the study area. The glaciated pre-Karoo surface was observed at a single locality in the study area. At this locality, the base of the Dwyka Formation consists of angular, poorly-sorted grits underlain by conglomerates which have been deeply grooved by ice. The grooves are orientated on a bearing of 230°. The basement Transvaal Sequence shales also reveal fine grooves having the same orientation. The thin basal deposits are overlain by diamictite.

Clast lithologies within the diamictite were recorded throughout the study area. Between one hundred and two hundred clasts, of diameter exceeding 3 cm, were identified at each sample site. In areas free of gravel terrace debris, sample sites were chosen on surfaces where clasts had been released by weathering. At sample sites where gravel terraces were developed, only clasts actually embedded in the diamictite were recorded. Three rock types dominate the clast assemblage; Ventersdorp Supergroup lava, and Transvaal Sequence dolomite and quartzite. Minor clast types observed included granites, gneisses, cherts and various silicic volcanics. Shale clasts greater than 3 cm are rare, presumably reflecting relative ease of comminution.

The results of the clast analysis are shown diagramatically in Fig. 3. In the southern portion of the study area, lavas of the Ventersdorp Supergroup dominate the clast assemblage but their abundance decreases to the north and especially the north-west,

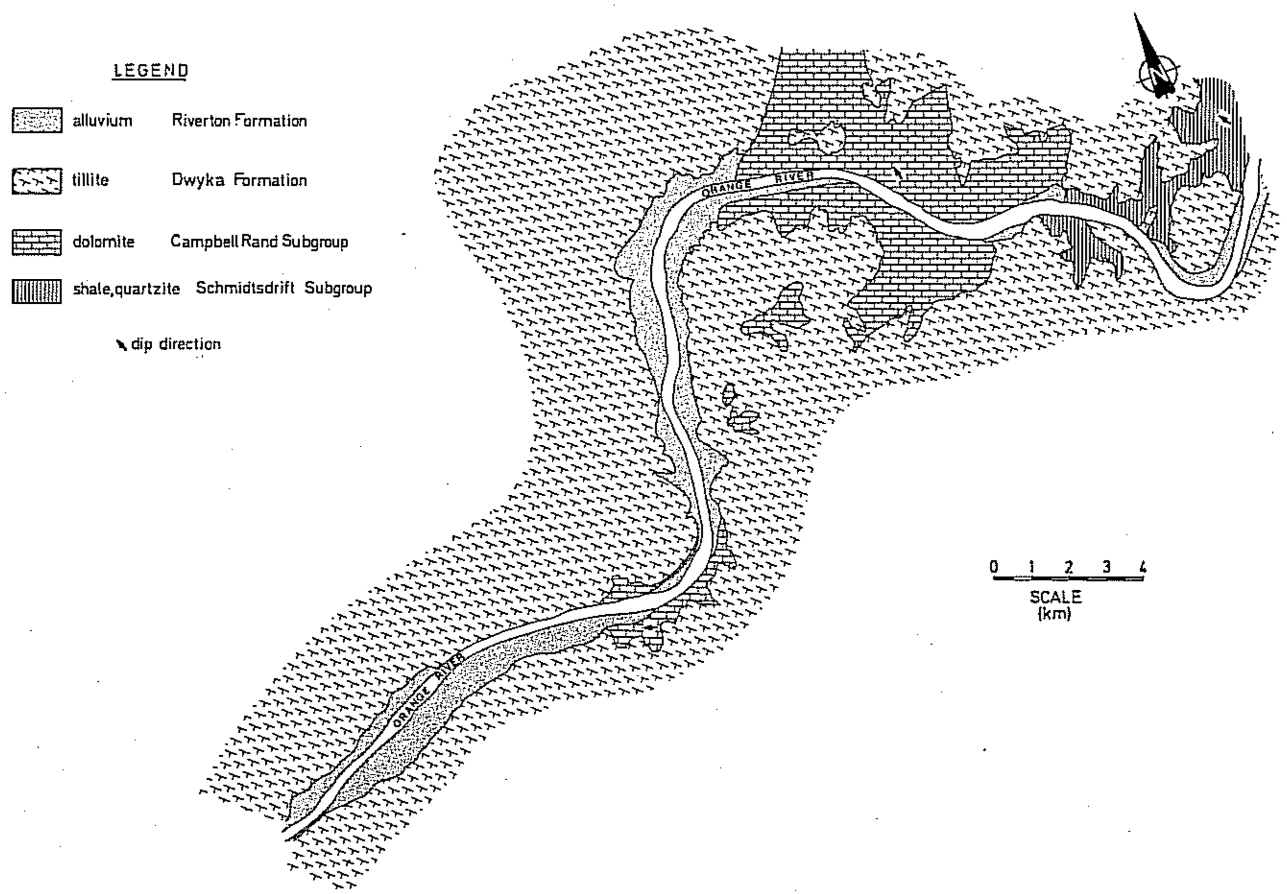

Figure 2
Geology of the study area.

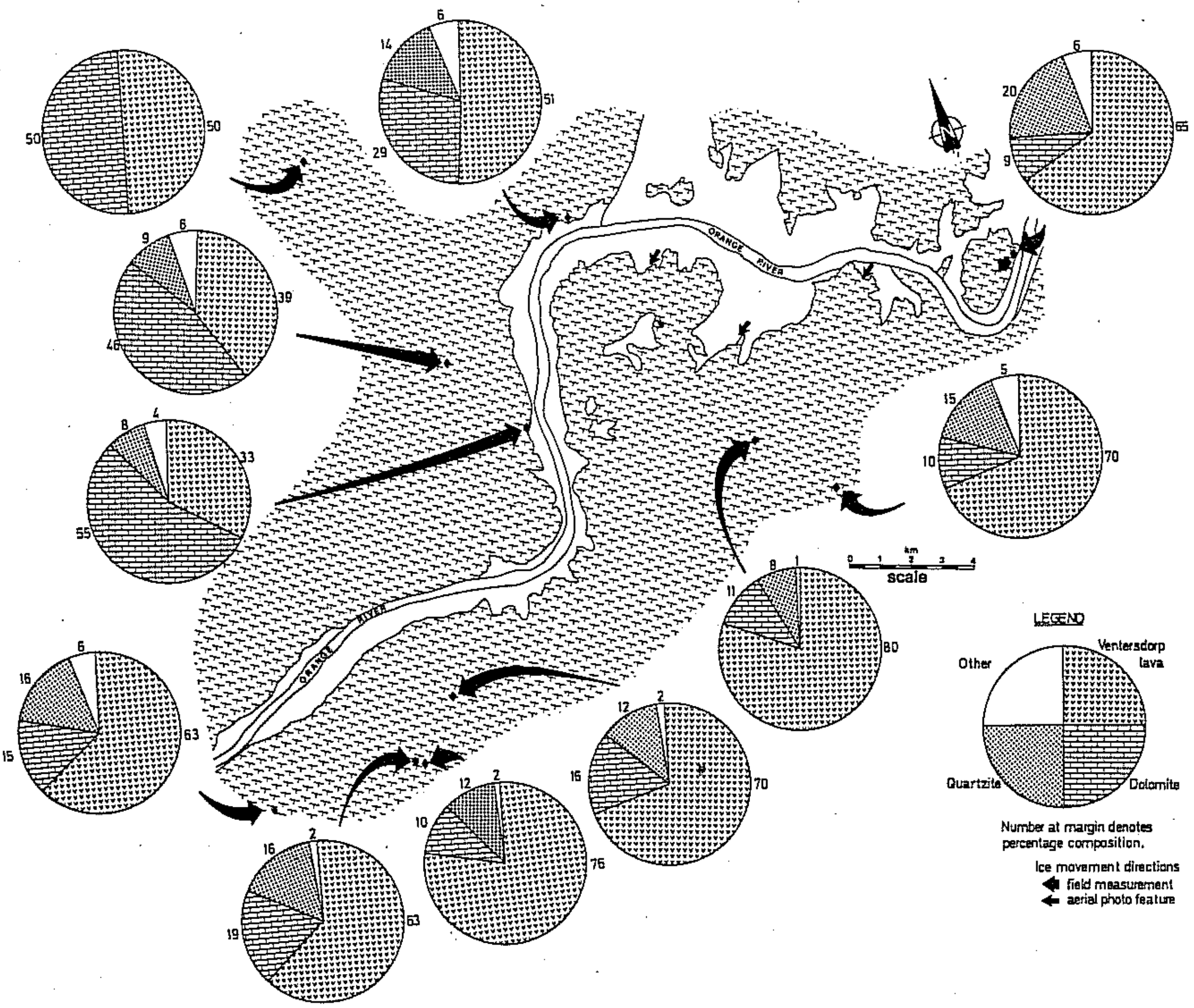

Figure 3
Distribution of the Dwyka tillite ⊤, and the composition of its clasts in the study area.

where dolomite becomes the major clast type. The clast distribution can be correlated with the ice movement direction (see Fig. 3). For example, ice traversing the northern portion of the area would move a greater distance across dolomitic rocks than ice which moved across the southern portion of the area. This is reflected in the clast compositions.

B. Alluvial Gravels

1. Distribution

Alluvial deposits are widely distributed in the study area (Fig. 4). Extensive gravel bars are developed within the present bed of the Orange River, but these are only exposed when the river is very low. There are several, distinct, high-level terraces with associated gravels. Some are laterally very extensive within the study area while others are restricted to single localities. The most extensive accumulation of alluvial material is situated along the immediate banks of the Orange River (see Fig. 2) and consists predominantly of sand, although the presence of underlying gravels has been reported by farmers in the area. This alluvial material correlates with the Riverton Formation of Helgren (1979a). Because of the scarcity of exposure of the associated gravels this terrace is not considered further in this paper.

Gravels are developed at successively higher elevations but generally the extent of the terraces decreases with increasing elevation. A laterally extensive terrace is associated with the 975 m to 991 m (3 200′ to 3 250′) topographic contours. Gravels lying on this terrace occur at Vaalkrans, Gewonne, Kaffersfontein, Riets Drift, Beatrys and Saxendrift. This terrace may be a correlative of the Rietputs terrace of Helgren (1979b). However, detailed examination indicates that this terrace does not represent a single chronological marker. At Vaalkrans, two discrete terraces are in fact developed, one slightly above and the other below the 975 m contour. A similar situation probably exists at Beatrys. Furthermore, the occurrences at Kaffersfontein and Gewonne lie at slightly higher elevations than the upper gravels at the other localities mentioned. Taking into account likely gradients in the Orange River during development of these deposits, it becomes evident that the 975 to 991 m contour deposits are only broadly coeval, but in detail must reflect gravel accumulation over a considerable period of time. The higher lying deposits on this terrace may actually be correlatives of Helgren's (1979a) Wedburg terrace.

Gravels associated with a terrace lying at an elevation of 1 006 m (3 300′) are developed at Saxendrift and Beatrys and may correlate with the Wedburg terrace of Helgren (1979b). Gravels also occur at an elevation of 1 021 m (3 350′) at Brakfontein, Beatrys and Brakkies, while at Kransfontein, gravel remnants occur above the 1 052 m (3 450′) contour. The highest lying gravels in the study area occur at Banghoek at an elevation of 1 067 m (3 500′).

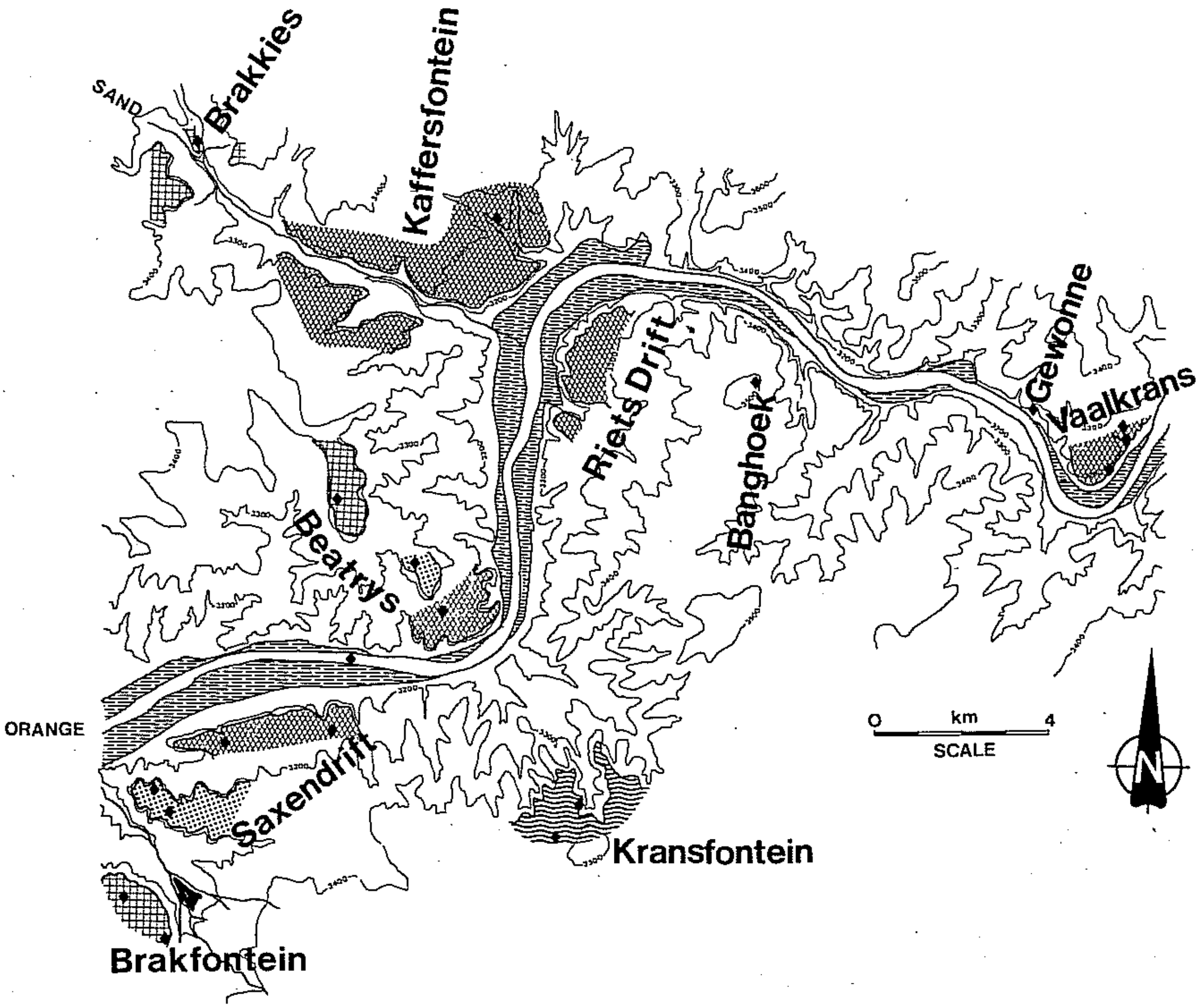

Figure 4
Alluvial deposits in the study area. Topographic contours are in English feet. Terraces which are correlated are shown in the same hatch pattern.

In addition to these terraces, which are associated with the Orange River, most of the larger tributaries also have associated gravel accumulations, some of which are quite large. Aggradation in these streams presumably coincided with aggradation in the main channel, although relative chronologies are not easily established. For the purposes of the present work, only one of these accumulations will be considered, namely that associated with two minor streams which converge to the east of the Brakfontein gravel deposit. The minor tributary deposit lies on the interfluve between the two streams at an elevation of 991 m. Both streams are flanked by at least two younger gravel accumulations but these are not considered here.

2. Nature of the Gravels

The gravel deposits differ widely in their morphology both within an individual deposit and across an entire terrace. Representative gravel profiles are shown in Fig. 5.

The most extensive gravel terraces generally occur on a planed Dwyka Formation surface. However, the small occurrence at Gewonne lies within a vertically-sided, 20 m-wide channel cut in Transvaal Sequence shales. In addition, the small derived gravel occurrence at Banghoek has, by virtue of erosion and weathering, been superimposed on karstic features in Transvaal dolomite (Fig. 6). The common tillite bedrock is typically overlain by at least 2 m, but seldom more than 5 m, of boulder gravel. The gravel is poorly sorted and boulders up to 50 cm in diameter are not uncommon, although the larger clasts typically measure about 25 cm along the major axis. Sandy partings containing small pebbles are often developed (Figs. 5 and 7). These partings are often cross bedded. In some exposures, cross bedded, gritty units form most of the profile (Fig. 8). In some cases no partings are present and the gravel shows a general upward decrease in boulder size. Elsewhere no obvious upward decrease in boulder size is evident. Boulders and cobbles are separated from one another by smaller pebbles but the coarser gravels may be regarded as being clast supported. The gravel matrix is calcretized throughout to varying degrees. The terraces at Brakkies and Kaffersfontein are strongly cemented, while calcretization of gravels more remote from the dolomitic rocks of the Ghaap plateau tends to be less intense.

Towards the top of the profile, matrix calcrete becomes nodular and passes upwards into a very hard, laminated calcrete in which non-siliceous clasts show evidence of being replaced by calcrete (Fig. 9). Furthermore, clasts are widely separated in this zone. Overlying this is a zone of rubefied gravel, the "derived gravel" of Partridge and Brink (1967), which also occupies solution pockets or makondos in the hard calcrete layer. This gravel consists almost entirely of siliceous clast types.

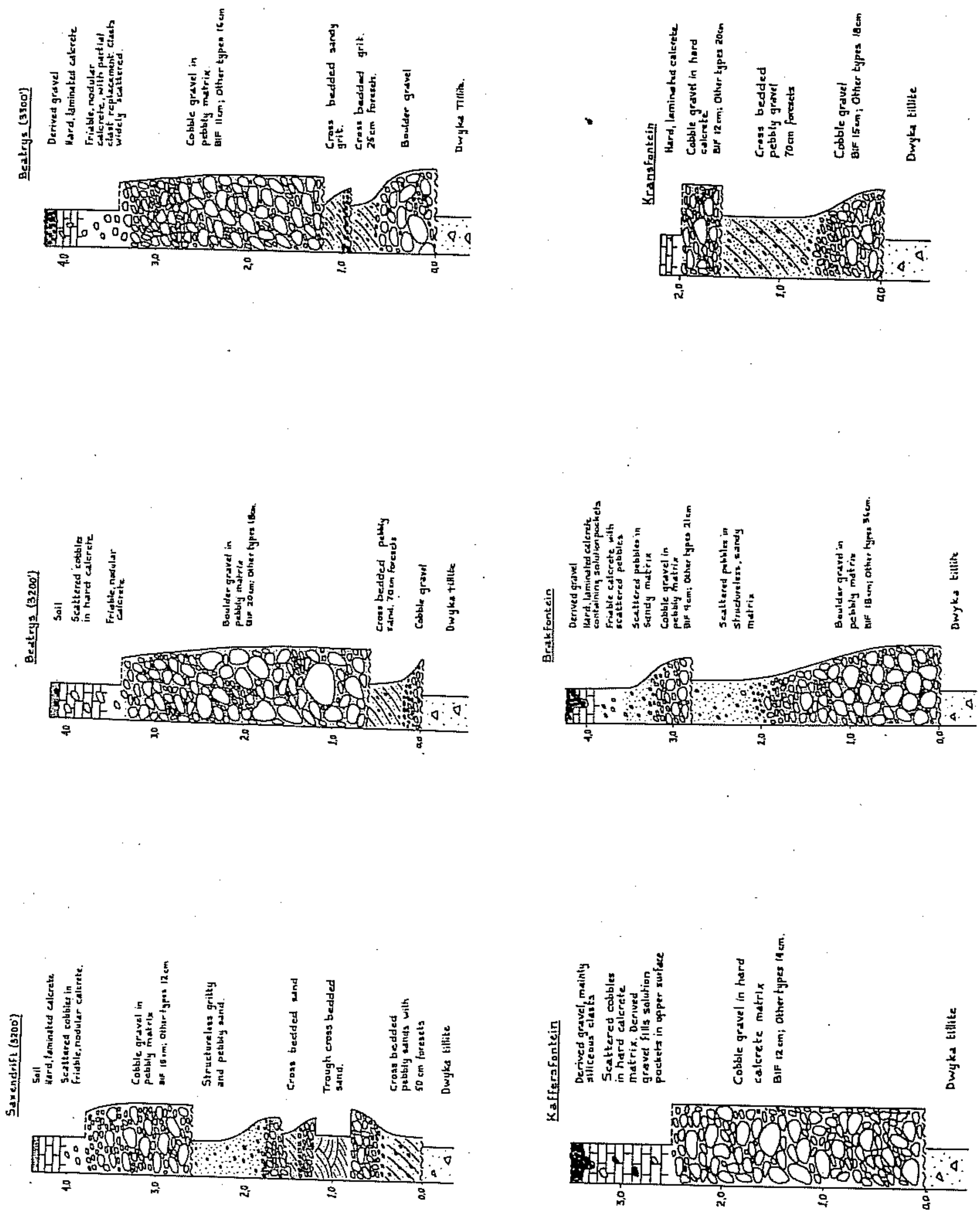

Figure 5
Representative gravel profiles for the area. Gravel thicknesses are in metres. Clast size measurements refer to the average of the ten largest clasts in unit area of exposed gravel.

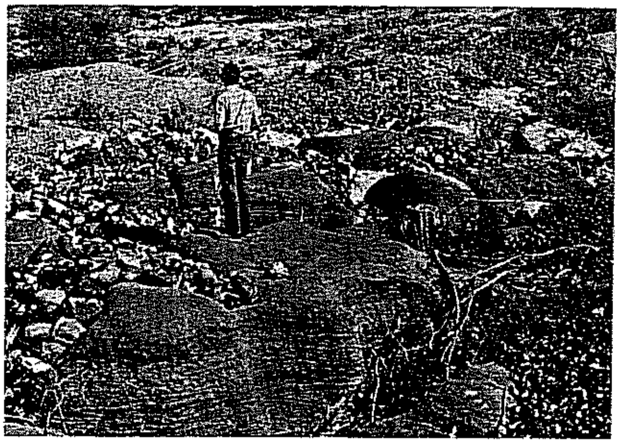

Figure 6
Karst features containing calcified derived gravel in Transvaal dolomite, Banghoek.

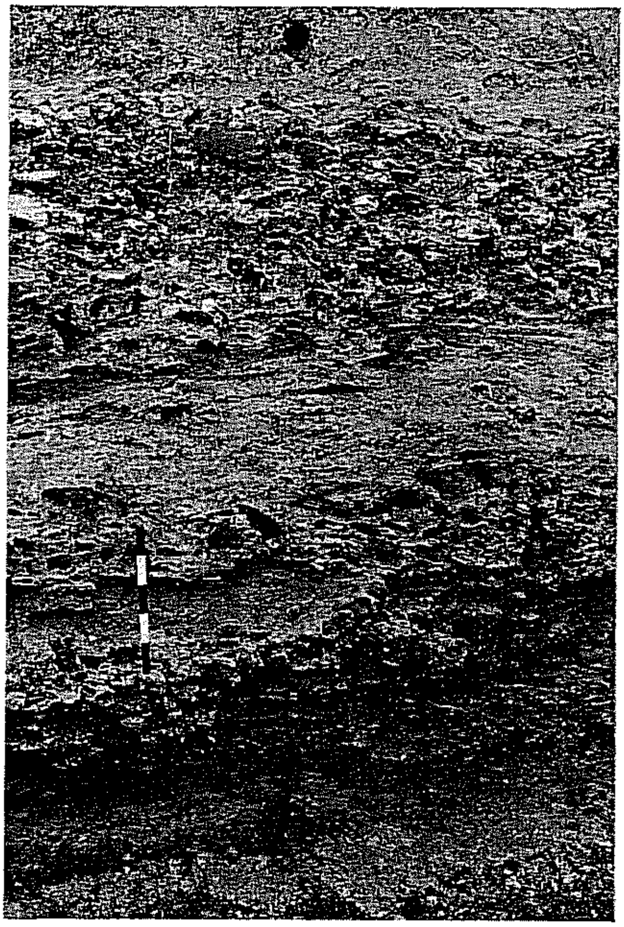

Figure 7
Cross-bedded sandy partings in gravel (Saxendrift, 975 m (3 200′) terrace). Scale intervals are 10 cm.

The boundary between hard calcrete and derived gravel is often extremely sharp (Fig.10). The hard calcrete frequently contains clusters of rubefied gravel suggesting recementation of old, gravel-filled makondos (Fig. 11).

III. CLAST ASSEMBLAGES IN THE GRAVELS

A. Methods

Where possible, diggers, excavations or natural exposures which had reached bedrock were used as sample sites for determining proportions of rock types in clast assemblages in the lower portions of the profiles. All clasts exceeding 3 cm in size were identified down vertical profiles at the sample sites until the number of clasts

Figure 8
Cross-bedded gritty gravels (Kransfontein). Scale intervals are 10 cm.

Figure 9
Replacement of lava cobbles by calcrete. Note the separation of fragments.

recorded exceeded 150. The clast assemblage thus obtained represents the entire profile. Small differences in clast composition were observed within individual zones in certain profiles, and between profiles measured at different sites on the same gravel body. These invariably relate to average size such that the coarser and

Figure 10
Derived gravel lying on hard calcrete surface showing solution pocket development. Scale intervals are 10 cm.

Figure 11
Recemented derived gravel in hard calcrete, probably representing a former solution pocket fill. Scale intervals are 10 cm.

more compact the gravel, the greater the proportion of lava boulders.

The precision of the data obtained is difficult to estimate because of the inherent variability of clast composition, and undoubtedly varies considerably with absolute abundance. For the more abundant clast types (tens of per cent abundance), errors are probably of the order of 20 per cent relative, while for the very minor clast types (around one per cent), relative errors of several hundred per cent are probable. For this reason, minor clast types are of little quantitative value.

Where possible, clast assemblages in the derived gravels were determined in the same manner. In cases where no gravel faces were exposed, data were recorded on gravel littering the surface, or on diggers' heaps.

Current directions were also recorded from sedimentary structures wherever possible.

B. Results

A great variety of clast lithologies was identified in the gravels. By far the most common types were green Ventersdorp lavas, quartzite, banded iron-formation, including riebeckite ("blou bliksem") and jaspilite, and chert with less common varieties including lydianite (pyroxene hornfels), shale, agate, silicified wood and felsic lavas of various types. A particularly important minor component, normally of too low an abundance to quantify reliably, was red-weathering lava (typically containing zeolite amygdales) of the Drakensberg Basalt Formation of the Lebombo Group (formerly Stormberg Series). This rock type is characteristic of present-day and older Orange River gravels and constitutes an important marker of an Orange River source. The largest clast of this type observed measured 7 cm along the major axis. Of paramount importance to this study were the banded iron-formation clasts as these were derived from outside the study area and, as such, also constituted an important marker. Dolomite is rare on the higher level terraces but is a major constituent in the modern river gravels, perhaps reflecting greater susceptibility to *in situ* weathering.

Results of the gravel clast analysis are shown in Fig. 12a, b. In these data, Lebombo Group ("Stormberg") lavas are expressed on a presence or absence basis, while data for the more abundant clast types are quantified. Where possible, data for both the derived gravel and the deeper, primary gravel are given. It can be seen that there is a general correlation between the data from the two types; where banded iron-formation is a major constituent of the deeper gravel it is the dominant type in the derived gravel, and vice versa.

It will be noted from Fig. 12a that to the east of Banghoek, banded iron-formation is an insignificant constituent of the gravels. In these terraces, banded iron-formation clasts rarely exceed 10 cm along the major axis. However, west of Banghoek, and especially west of the Sand River, banded iron-formation becomes a major clast type in all terraces, and boulders up to 30 cm diameter are not uncommon. An interesting exception is the minor terrace east of Brakfontein. Lebombo Group lava characterizes all the terraces along the Orange River but is absent from terraces in the Sand River area (Kaffersfontein and Brakkies).

IV. DISCUSSION

A. Depositional Environments of the Gravels

The extreme coarseness of the bulk of the gravels suggests formation in abnormally rapid flowing water, particularly in view of the maturity of the river systems in which they occur. It is therefore inferred that the bulk of the coarse gravels represent gravel bar accumulations formed during periodic flood events.

The occasional occurrence of a single upward-fining cycle in an entire gravel profile even suggests formation of the entire profile in a single event although, of course, such profiles may represent the last event in a long history of reworking. Other profiles, particularly those containing sand partings, reflect multiple events, although the time separating individual events cannot be deduced. Homogeneous gravel profiles which exhibit no upward-fining could represent single or multiple events. Gravel bar surfaces may remain bare of sand, as in the present day Orange River, and simply accrete incrementally at each new flood event. Within sandy partings sedimentary structures indicative of bar accretion, and of migrating transverse bars and dunes in excess of 50 cm in height, were recorded (Fig. 5). The latter again points to rapidly flowing, deep water, and, combined with underlying boulder gravel, suggests formation during waning floods. In general, these features tend to have heavily eroded tops.

Because of the predominance of coarse gravel it appears that these terraces largely record catastrophic events in the history of the river. Periods of normal flow leave little or no trace as their associated sedimentary features are destroyed by erosion. Catastrophic events will tend to deposit bed load at localities remote from the normal thalweg which can thus be preserved from erosion during periods of normal flow.

B. Derived Gravel Development

Cooke (1946) recognized the fact that the rubefied gravels represent the residual weathering product of the deeper, calcretized, primary gravels. Partridge and Brink (1967) noted possible lateral movement and coined the term "derived gravels" for these gravels.

The derived gravels are composed of the siliceous rock types present in the underlying gravels. Destruction, largely by replacement, of the non-siliceous types can be observed taking place in the hard, laminated calcrete layer (Fig. 9) and the underlying nodular calcrete. However, the widely scattered nature of pebbles and boulders in typical hard calcrete is problematic. From the writer's observation in the area, it appears that the development of the hard calcrete is associated with considerable volume increase. Thus, for example, partly replaced fragments of a single lava boulder become separated (Fig. 9). Similar expansion features, notably anticlinal structures, which become steeper and tighter as the overlying solid calcrete layer is approached, are commonly observed in tillites and shales which have been calcretized. The resulting steeply-dipping shale remnants in the surface calcrete are often mistakenly regarded as indicative of the presence of kimberlite fissures by prospectors.

It appears, therefore, that the hard calcrete layer, which is continually being leached downwards by meteoric water, not only induces destructive replacement of non-siliceous clasts, but also, because of the associated volume increase, causes dispersal of the unreplaced clasts. As the layer migrates downwards these clasts are released at surface contributing to the derived gravels. Surface weathering finally eliminates remaining traces of less resistant rock types. Thus, as the age (maturity) of the calcrete profile and of the derived gravel increases, so does the proportion of siliceous clasts in that gravel. Occasionally, the hard layer may migrate upwards, perhaps due to climatic changes, during which time derived gravel occupying makondos may become cemented (Fig. 11).

C. Palaeodrainage

Although gravels occur at several different elevations to the west of the Sand–Orange River confluence (Fig. 12) the majority have no counterparts to the east, making comparison difficult. However, one terrace is widely developed in both areas, namely that at an elevation of 975 m.

East of the Sand River, at Vaalkrans and Gewonne (Fig. 12a) the gravels at this elevation consist predominantly of lava, quartzite, lydianite and shale. The lava and much of the quartzite is undoubtedly derived from the tillite (see Fig. 3) the latter having been augmented by contributions from the Transvaal Sequence outcrops in the area. The shales are probably also locally derived. Lydianite is derived from the Karoo Sequence (Ecca Group). Banded iron-formation is a negligible constituent in the gravels, its presence being due to old drainage systems through the Western Transvaal (Cooke, 1946). Lebombo Group ("Stormberg") lava, indicative of Orange River material, is also present.

In the gravels at an equivalent elevation to the west of the Sand River confluence (Saxendrift and Beatrys; Figs. 4, 12b), banded iron-formation is a major component, accompanied by Ventersdorp lava. Quartzite is a minor constituent of these gravels. Banded iron-formation does not occur within the Transvaal Sequence rocks exposed in the area studied, nor was this rock type recorded in the tillite of the Dwyka Formation in the study area. The banded iron-formation in the gravels must thus have been fluvially introduced from outside the area investigated. In spite of the high banded iron-formation content in the Saxendrift and Beatrys 975 m gravels, the presence of Lebombo Group ("Stormberg") lava indicates that these deposits have nevertheless experienced input of Orange River material (Fig. 12a).

An immediate source for the banded iron-formation is recorded in the area in the form of an extensive terrace in the Sand River area, lying at an elevation of about 991 m, which extends up the broad valley between the Ghaap Plateau in the east and the Asbesberge in the west (Kaffersfontein and Brakkies in Figs. 4, 12a, b). Gravels on the terraces within this valley are dominated by banded iron-formation and, furthermore, they contain no Lebombo Group lava. The ultimate source for the banded iron-formation is within the catchment associated with these terraces, presumably the iron-formation forming the Asbesberge to the north-west (Fig.1).

It must be inferred that, at the time the 975 m gravels were accumulating, sediment was entering the Orange River from the north via this valley.

In contrast to the above, the clast assemblage in the modern river bed adjacent to the 975 m terrace on Saxendrift (Fig. 12a) contains very little banded iron-formation, and, except for the presence of dolomite, resembles Orange River gravels on the 975 m terrace at Vaalkrans. Clearly, the major input of banded iron-formation, which is reflected in the 975 m terrace, no longer occurs.

It is possible to trace the input of banded iron-formation back in time. In the 1 006 m gravels on Beatrys and on Saxendrift (Fig. 12b), banded iron-formation continues to be a major clast type, in spite of the presence of lavas of the Lebombo Group ("Stormberg") in both occurrences. A similar clast assemblage is present at Brakfontein (Fig. 12b) (1 201 m). A possible contemporaneous terrace is developed across the river at Beatrys (Fig. 12b), but at this locality no deeper primary gravels were exposed. However, the derived gravels here indicate that iron-formation is a major constituent. Similarly, gravels containing lava of the Lebombo Group, lying at an elevation of 1 052 m at Kransfontein also contain iron-formation as a major component. Finally, the oldest gravels in the area (Banghoek) are also dominated by iron-formation, although a contribution from the Orange River at this locality cannot be verified.

Within this study area, there has clearly been a consistent input of large amounts of banded iron-formation from the earliest times up to, and including, the time of formation of the 975 m deposits. As mentioned earlier, the only possible source for this material is the iron-formation of the Asbesberge north-west of the study area. Clearly, these fell within the catchment of the northerly drainage.

It is obviously of some interest to establish relative sizes of the rivers in this catchment area. It can probably be reasonably assumed that relative sizes of rivers are more or less related to their respective sediment contributions at a confluence, and thus, by examining the clast types within individual and mixed populations, relative sizes or volumes of rivers can be deduced. Applying this concept to a single terrace could lead to uncertain results as only one of the rivers may have been in flood at the time of formation of that terrace. This problem is unlikely to arise if many terraces are considered.

Downstream of the Sand River confluence, all Orange River-associated terraces are dominated by banded iron-formation (Fig. 12a, b) indicating consistent input of sediment from the north throughout the time period recorded in the terraces and probably spanning many millions of years. At no stage was banded iron-formation a major component of either the Vaal or the Orange River sediment input as this rock type is not recorded in quantity on the high level terraces associated with these rivers upstream of the study area. Thus the absence of comparative terrace material east of the Sand River

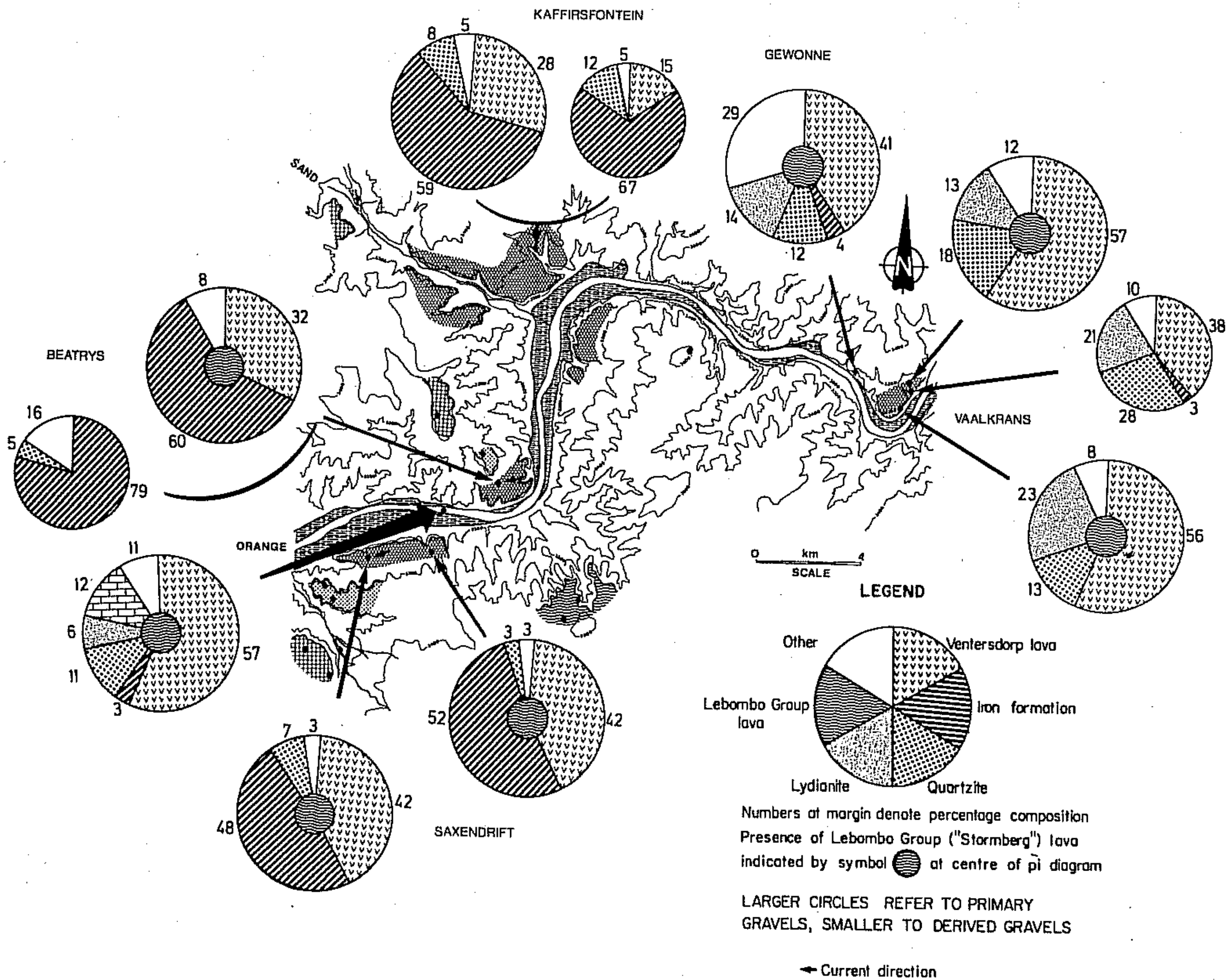

Figure 12

a. Composition of gravels on the younger terraces, in the study area. The bold arrow refers to gravels in the present river bed.

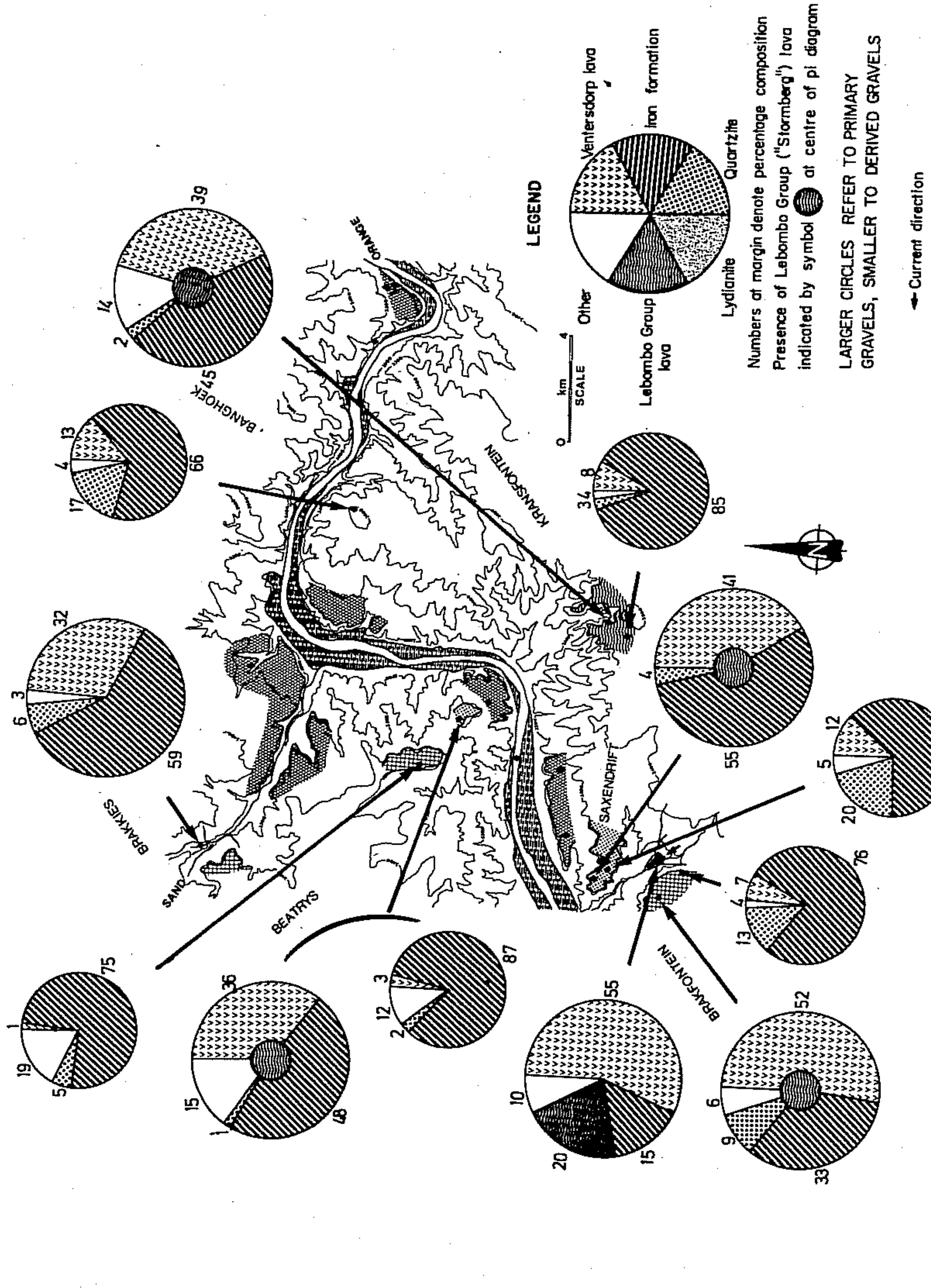

Figure 12
b. Composition of gravels on the older terraces in the study area.

confluence does not detract from the arguments presented here.

Typical gravel associated with the banded iron-formation source (Kaffersfontein and Brakkies, Fig. 12a, b) contains 59 per cent banded iron-formation. The primary gravels west of the Sand River (all terraces combined) contain on average 47 per cent banded iron-formation. These represent a mixture of Orange River gravel and material derived from the north. Since the Orange River east of this region has contributed negligible banded iron-formation to the mixed gravels, it must be concluded that around 80 per cent of the mixed gravels, averaged over all the terraces down to the 975 m level, were derived from the northerly source. Calculated on an individual terrace basis, the proportion of material derived from the northerly source is: 975 m terrace — 90 per cent; 1 006 m terrace — 88 per cent; 1 021 m terrace — 56 per cent; and 1 052 m terrace — 88 per cent. The implication is that the river entering the Orange River from the north consistently carried about four times as much sediment as the Orange River in equivalent size range (i.e. large clasts).

In principle one could cross check this result using other clast types. Quartzites are unsuitable for this purpose as abundances are low and the precision of the data is therefore poor. Lavas could be used although the possibility exists that additional lava may have been added to the sediment load below the confluence of the northerly river and the Orange. Indeed, the proportion of lava in the gravels shows a steady increase westwards of the inferrred confluence point (Fig. 12a, b), suggesting that such addition took place. Using only the data from the terraces on Beatrys, which are closest to the inferred confluence and which contain 34 per cent lava on average, the calculated proportion of sediment derived from the northerly source is 81 per cent, and is in agreement with the results obtained using iron-formation.

It is evident from these calculations that the vast bulk of the sediment deposited east of the inferred confluence was derived from the northerly source. This must relate to the relative sizes of the two rivers. If the size distribution of sediment in the two rivers was similar, then it is evident that the northerly river must have carried about four times as much water as the Orange. However, if the northerly river carried a greater proportion of large boulders, then direct equivalence of sediment contribution to the gravels and relative volumes of water does not hold.

Gravel terraces east of the inferred confluence are identical in appearance to those to the west and north, and clast sizes in all are similar. There is no indication that clast size distribution was substantially different in the two rivers.

However, because of the uncertainties involved, all that can be concluded is that the northerly river was of a substantial size, certainly equivalent in volume to the Orange River, but probably much larger. Helgren (1979b) has argued that the Orange River in Rietput's times was already a great river, certainly equal to the present Orange River. The river which entered the Orange from the north must likewise have been a great river; and it existed as such for a considerable period of time, as reflected in the gravels from 1 067 m down to 975 m.

The catchment area of this great river must have been very large, perhaps encompassing much of south-central Africa. Indeed, assuming that semi-arid conditions had already commenced in the Kalahari region, a catchment extending into tropical central Africa seems to be indicated. For this reason, the name Trans-Tswana is proposed for this river.

This river no longer exists today, and its remaining alluvial deposits are in the process of dissection by the Sand River and other minor seasonal streams. Information pertaining to the demise of the Trans-Tswana River system is sparse. During deposition of the 975 m terrace gravels this river was still very active. Between the time of formation of these deposits and the present, this great river had ceased to flow into the area. Examination of the distribution of the Riverton Formation (Figs. 2, 4) suggests that the Trans-Tswana River had already ceased by this time. In fact, the only visible incision of its 975 m terraces can be directly related to present minor stream activity, suggesting that its demise was extremely rapid following the deposition of the 975 m terrace gravels (Kaffersfontein).

The confluence of this great river with the Orange appears to have migrated from a point east of Banghoek to a final position south of the Sand River confluence, but at all recorded times remained within the study area.

The position of the Orange River may, however, have undergone drastic changes in the period under consideration. Helgren (1979b) suggested that the Orange River once flowed along a more southerly course entering the Vaal River near Brakfontein. A possible remnant of this early period is recorded in the minor terrace to the east of Brakfontein, in which banded iron-formation is a minor constituent in gravels containing Lebombo Group lava. It is possible that this gravel represents re-eroded material from older Orange River terraces (plus more recent terraces) which lay to the south of Brakfontein. However, the presence of Lebombo Group lava in the Kransfontein gravels indicates that by the time the 1 052 m gravels were being deposited, the Orange River had already migrated northwards, or experienced upstream capture, and flowed more or less along its present course.

D. Drainage Evolution

The axis of crustal warping (Transvaal–Griqualand Axis) proposed by Du Toit (1933) extends across the inferred course of the Trans-Tswana River. Although Mayer (1973) considered this to be an axis of uparching, Du Toit (1933) in fact suggested that northerly sinking was the main process along the axis. It is interesting to speculate on the effect such warping might have on a mature drainage system (low gradients and without deep incision) draining virtually at right angles across the axis of the warp, especially if the region of the warp was semi-arid. If the warping was sufficiently rapid (perhaps accompanied by faulting) in relation to rates of river incision, the river would dam up and form an internal drainage basin. A modern analogue may be the Okavango River and its inland delta (Wellington, 1955). If the dammed area was semi-arid, the increased evaporation associated with the increased surface area could lead to permanent closure and a gradual infilling of the basin, accompanied by retreat of the river system. In this regard it is perhaps of interest to note that the underlying formations of the Kalahari Basin include shales and marls, compatible with deposition in a lacustrine environment (Haughton, 1969). It is not inconceivable, therefore, that the demise of the Trans-Tswana River was associated with the commencement of infilling of the Kalahari Basin. Perhaps the Okavango delta, Lake Ngami and the Makgadikgadi pans are the last vestiges of this once vast drainage system.

CONCLUSIONS

This work illustrates that clast assemblages can be used in tracing old drainage systems even in areas underlain by the Dwyka tillite. In this case, banded iron-formation proved useful not only in locating provenance, but also enabled estimation of relative sizes of rivers.

It is evident from this study that a river of vast proportions at one time entered the Orange River in Griqualand West, and must have represented a catchment area extending well into south-central Africa. The rapid, perhaps even catastrophic, demise of this river system may be related to the commencement of infilling of the Kalahari Basin.

Speculations concerning the evolution of the river system in Griqualand West are far from over, and the findings of this work add a new dimension to drainage studies in this remarkable area.

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