QUATERNARY ENVIRONMENTAL CHANGES IN SOUTHERN AFRICA

E. M. VAN ZINDEREN BAKKER AND KARL W. BUTZER

University of the O.F.S., Bloemfontein, and University of Chicago

Received for publication May 4, 1973

ABSTRACT

Evidence for Quaternary environmental changes in southern Africa has been obtained from palynologic, geomorphologic, and pedologic studies. It appears that in the Vaal and Orange drainage of the interior (26–30°S) colder periods with heavier precipitation alternated with warmer and drier periods, broadly synchronous with glacial and non-glacial trends in high latitudes of the northern hemisphere. By contrast, pedogenesis in the coastal regions (32-34°30′S) corresponded with warmer climates, whereas colder phases were relatively dry.

The study of Quaternary environments in Africa has recieved much attention from many different disciplines in recent years. This research has been mainly focused on physical and biological aspects of northern Africa, the Sahara, the Nile region, and particularly on East Africa. Southern Africa, where most of the initial work had been done, has lagged behind the other regions in this new development. However, that part of the continent has great potential for Quaternary studies in its wealth of paleontological and archeological material and in its rich flora and fauna.

The climatological and biological setting of southern Africa is mostly of an arid and semiarid nature. Schematically the subcontinent can be divided into four quarters south of 22°S latitude (see Fig. 1). The east-west dividing line is roughly formed by the 26° longitude. The big northwestern quarter is occupied by the vast arid basin of Botswana and southwestern Africa, the southwest quarter is the domain of the semiarid Karroo. The eastern half of the subcontinent is more elevated and this plateau is delimited to the east by mountain ranges that stretch as far as Cape Town and which separate the humid and more equable climate of the coastal plain from the drier interior parts. The southwestern coastal periphery has an ecological character quite different from the rest of the subcontinent. It receives winter rain and is the home of the famous Flora Capensis. An interesting ecological gradient occurs between the very dry coastal Namib desert and the more humid biomes of the eastern plateau. North of about 27°S this plateau is covered with woodland, while south of that parallel, temperate grassveld is found.

The arid and semiarid central and western parts of Southern Africa are under the aridifying influence of the high pressure system of the southern Atlantic and the cold Benguela Current. From the climatological point of view the subcontinent is similar to other southern hemisphere land masses. In summer the Intertropical Convergence Zone moves far south, bringing summer rain, especially to the northeast sector. This rain diminishes towards the south. The southeast trades bring rain to the eastern coastal region.

The ecological patterning of southern Africa changed significantly during Quaternary times as has been shown by recent geomorphological and pedological studies, by fossil pollen evidence, and by archeological and paleontological research. Changes in temperature and humidity had important influences in this mainly semiarid region by affecting the geomorphic equilibrium and causing large-scale migrations of biota over wide areas where no orographic barriers occur. These changes left their marks in river valleys and lake basins, on exposed mountain ranges, and on the coastal platform. A growing number of C^{14} dates now makes it possible to establish a preliminary chronological framework for the late Quaternary of the interior and the southeastern coastal regions.

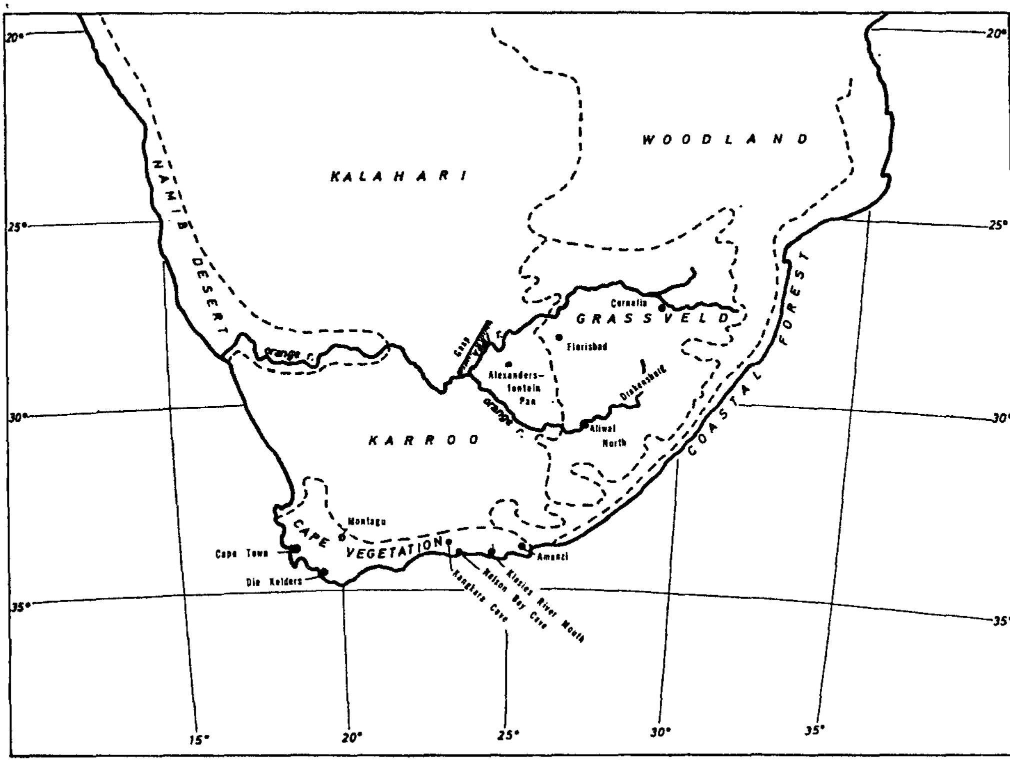

FIG. 1. Southern Africa.

FOSSIL POLLEN EVIDENCE

Quaternary deposits that contain fossil pollen are very rare in southern Africa. The main reason for this scarcity is that the aridity and the high temperatures that prevail in many parts are unfavorable for the preservation of these subfossils. The only sites where fossil pollen is found are the more humid coastal regions, the high mountains, and around old springs. The deposits of two such springs in particular have yielded results of great value for a climatic interpretation of the stratigraphy of the Upper Pleistocene and Holocene of the interior of southern Africa.

Florisbad

Florisbad, the site of the oldest of the two profiles, is located at 28°42'S in the *Cymbopogon-Themeda* grassveld of the central Orange Free State at an altitude of 1275 m. The present rainfall of this area varies from 370–500 mm per annum. The peaty and sandy deposits that occur around this spring have revealed an extremely important picture of the past as they have provided a wealth of paleontological material, including fossil human remains, many Middle Stone Age artifacts, and fossil pollen. A number of important radiocarbon dates have also been procured for these deposits. A low hill of alternating layers of sand and organic clay has been deposited round the oligohaline spring. The hill occurs on the southern edge of a large deflation pan which at some stage must have contained a vast lake. At present it is characterized by some marshy vegetation and salt flats. The thermal spring originated along a dolerite intrusion. The pollen profile of the climatic indicators consists of the following strata (Zinderen Bakker 1957, 1967), which are summarized by Table 1, in terms of their key aspects and simplified interpretation.

Fossil pollen was thus found in the Beds 1, 2, 4, 5, and 6. Owing to the scarcity of pollen grains only 50 to 180 grains could be counted for each spectrum at 10-cm vertical intervals.

The pollen produced by the local halophytic vegetation consists of *Chenopodiaceae* and *Zygophyllum*, while abundant cyperaceous pollen and

TABLE 1

Pollen stratigraphy of Florisbad

10. "Recent Peat," removed by man.
9. Loamy surface soil. No pollen preserved.
8. "Peat IV." Contains diatoms but no pollen.
7. Sand with macrobotanical remains, similar to those of Beds 3 and 5. No pollen present.
6. "Peat III." Radiocarbon date: 19,350 ± 650 B.P. (L-271D). Grass pollen percentages from 75 to 94 percent, indicate wetter and cool conditions.
5. Sand with Middle Stone Age Artifacts. Contains similar plant macroremains to those of Bed 3. Pollen spectra indicate wetter and cooler conditions.
4. "Peat II." Radiocarbon date: 28,450 ± 2,200 B.P. (L-271C). The pollen curves of grasses and *Compositae* oscillate in opposite directions indicating changes from very arid to semiarid, warm conditions.
3. Bluish clay or loam containing vertical, parallel stems and roots of reeds and rushes; possibly deposited under gley conditions. Middle Stone Age artifacts. No pollen.
2. "Peat I." Radiocarbon dates: >48,900 B.P. (GrN-4208), >44,000 B.P. (Y-103), >41,000 B.P. (C-850), and >35,000 B.P. (L-271B). Most of the paleontological material occurs beneath this layer. The pollen spectra are the same as in Bed 1, except for one slightly wetter oscillation.
1. "Basal Peat." Equal proportions of *Compositae* and *Gramineae* pollen indicate very dry and warm conditions.

Tetraploa fungus spores are produced by the marshy areas. These sporomorphae give an indication of the spring cycle and cannot be used for climatic interpretations.

The pollen sum, representing the dominant pollen types of the wide surroundings, is of very simple composition. No tree pollen was found except for a few grains of *Rhamnus, Olea,* and *Podocarpus*. The central Orange Free State must therefore have been virtually treeless during the deposition of the Beds 1, 2, 3, 4, 5, and 6.

The remaining pollen types are predominantly grass and *Compositae* pollen. An extensive comparison of pollen spectra produced by the present-day grassveld vegetation, both from atmospheric pollen counts and from analyses of surface soil samples, showed that grass pollen is predominant at present, with percentages of 95–96 percent, while *Compositae* pollen account for almost 4 percent.

At the base of the diagram, very high percentages of *Compositae* pollen were found in all the spectra of Beds 1 and 2. These percentages equal or even surpass the grass pollen percentages from which a very dry and warm climate can be inferred. Conditions of this type occur at present in the warm Great Karroo and the Central Upper Karroo where *Compositae* undershrubs are dominant with a rainfall of only 125 to 250 mm per annum. Near the top of Bed 2, the curves oscillate and suggest slightly wetter conditions, followed by a period of great aridity. These oscillations are much more pronounced in Bed 4 and show an alternation of drier and wetter periods. At a calculated date of about 25,000 years B.P., a definite and important change takes place, and those pollen spectra between the ages of roughly 25,000 and 19,400 years B.P. greatly resemble the present-day pollen production of the *Cymbopogon-Themeda* grassveld. This significant ecological change can be interpreted by a lower temperature, or by a considerable increase in precipitation, or both. If increased rainfall only were responsible, then precipitation must have at least doubled.

Aliwal North

The second important site for fossil pollen studies is the area around the thermal springs of Aliwal North, situated at 1355 m elevation near the Orange River. Aliwal North was originally surrounded by *Cymbopogon-Themeda* grassveld with shrubs and small trees. In recent years this grassveld has been degraded to False Upper Karroo, which is an intrusion of the dry semidesert to the west of Aliwal North. This site lies near an important vegetational ecotone since it is situated in a critical climatic region: summer rainfall rapidly diminishes to the west, and increases in a northeast and southeast direction. It is also noteworthy that the winter precipitation which is characteristic of the Cape coast presently reaches inland to about 100 km Southeast of Aliwal North.

TABLE 2

Pollen stratigraphy from Aliwal North. (After Coetzee, 1967)

Pollen zone Z:	Age ca. 10,000 B.P. Initially a dry grassveld, then replaced by Karroid vegetation indicating a very dry and warm period that was optimal just before 9,650 ± 150 B.P. (GrN-4012).
Pollen zone Y:	Return of the grassveld with a colder climate, which was, however, not as cold as in zone W.
Pollen zone X:	Age approximately from 11,650 ± 170 B.P. (I-2109) to 11,250 ± 180 B.P. (I-2110). A rapid change to Karroo vegetation in which *Compositae* and *Chenopodiaceae* predominated points to a drier and warmer climate.
Pollen zone W:	Age from ca. 12,200 ± 180 B.P. (I-2107) to 11,650 ± 170 B.P. (I-2109). A decline in pollen of *Chenopodiaceae* and *Compositae* and an increase of grass pollen shows that a pure grassveld occurred under cooler and moister conditions, but not as cold as in zone U.
Pollen zone V:	Calculated age 12,600 to 12,200 B.P. Karroid vegetation with much pollen of *Chenopodiaceae* indicates a warmer and drier climate.
Pollen zone U:	Age between 12,600 ± 110 (GrN 4011) and 13,185 B.P. (calculated age). The site was much wetter, especially in the lower part of the zone, and was surrounded by pure grassveld in which *Stoebe plumosa* indicates cooler conditions similar to those now prevailing at higher altitudes.

The modern vegetation, the climatic setting, and the pollen evidence of four cores with a calculated age of 13,200 years have been studied in detail by Coetzee (1967). The pollen spectra show a number of extremely interesting alternations of northward (eastward) and southward (westward) movements of the vegetation boundary between 13,200 and 9,650 years B.P. Expansion of the grassveld area indicates wetter and cooler conditions, while invasions of the Karroo, similar to those at Florisbad, point to a drier and warmer climate. The pollen sequence studied by Coetzee (1967) is summarized in Table 2.

The close correspondence of the Aliwal North pollen zones with those of N.W. Europe is quite remarkable. In particular, zone U was coeval with the end of the Oldest Dryas, zone V with the Bölling oscillation, zone W with the Older Dryas, zone X with the Alleröd Interstadial, zone Y with the Upper Dryas, and zone Z with the initial phases of the Holocene (see Hamen et al. 1967).

GEOMORPHOLOGIC AND PEDOLOGIC EVIDENCE

Offsetting the paucity of good polliniferous deposits is a wide range of geomorphic and pedologic indicators of past environmental changes in most parts of southern Africa. However, few local sequences record any great time depth, and the general prevalence of erosion during the late Cenozoic necessitates a great deal of horizontal stratigraphy that includes quite diverse phenomena. Similarly, interpretation is rendered difficult by the inadequate understanding of contemporary processes, by the limited intensity of geomorphologic research in recent decades, and by the lack of appreciation by most pedologists that many soils on nonfunctional deposits are in fact relict. Despite these limitations and the pessimistic prognoses of several authors (e.g. Flint 1959; Partridge 1969), there is increasingly substantial evidence for complex environmental changes during the course of the Quaternary. A few of the more informative sequences studied since 1969 are summarized here.

Paleosol stratigraphy of the southeastern Cape coast

The southwestern coastal sector of southern Africa at 33–40°S latitude includes several paleosol horizons within a detailed record of distinctive sediments and erosional forms (Butzer and Helgren 1972).

An almost ubiquitous planation surface, the 200 m Coastal Platform is associated with the fanglomeratic and deltaic Keurbooms Formation (late Tertiary?). A major sea-level stage at +120 m truncated a laterite paleosol that is still preserved by extensive caps of massive, indurated plinthinte, consisting of geothite and quartz, and local exposures of truncated B2 and B3 horizons under reworked plinthite nodules. Developed on pure quartzites and quartzitic sandstones, these profile remnants consist of reddish to strong brown, sandy clay loams that have a clay fraction of finely comminuted quartz, some goethite (ca. 6 percent), and only a trace of kaolinitic clay. The +120 m sea level was followed by accumulation of (?frost-weathered) land rubble and littoral deposits to +101 m, constituting the Formosa Formation of apparent early Pleistocene age. Then follow sea-level stages at +60, +30, and +15–20 m, as well as several generations of weathered eolianite, including the mid-Pleistocene Brakkloof Formation and its superposed Brakkloof Soil, which rest directly on the 60 m platform. The latter paleosol preserves a 50–130 cm horizon of reworked plinthite nodules (up to 13 percent goethite) over thick polygenetic and mottled horizons (250–300 cm) of yellowish brown clay, in turn overlying a light gray or white, pallid horizon (100–150 cm) of sandy clay. Developed on quartzite-derived eolian sands that, in terms of modern analogs, have only a trace of glauconite and the heavy minerals, in addition to shell debris, the clay fraction again consists of finely comminuted quartz with goethite and mere traces of kaolinite. While also fitting the morphological description of a lateritic ferrisol, the Brakkloof Soil appears to have been a plinthudult.

Beaches and estuarine terraces at +5–12 m, with thermophile mollusca and C^{14} dates of >40,000 years, mark the Swartkops horizon of probable Eemian interglacial age. Next are cryoclastic screes and cave deposits (see below), and the weak Brenton Soil of the Würm Interpleniglacial. Where covered by younger sediments, this paleosol has a 100–150 cm, decalcified B2 horizon of yellowish brown to reddish yellow loamy sand with incipient ferruginization, over a deeper zone of oxidized sands. Massive eolianites, related to the Flandrian Transgression (Dingle and Rogers 1972), locally rest on a humic, partly decalcified soil dated 16,000 B.P. Temporary stabilization of these coastal dunes is indicated a little before 7,000 B.P. by formation of the weak, podsolic Beacon Island Soil and a temporary recolonization of the littoral arenaceous environment by woodland vegetation (see Martin 1968). This last paleosol has a slightly humic and acid sandy 50 cm A1-horizon over a 50-cm brown-to-pink B2 of loamy sand. Renewed eolian activity and forest recession (see Martin 1968) followed, with more permanent stabilization beginning only after 4,200 B.P. when sea level reached +2.5 m and humic soils formed on the coastal dunes, with peats accumulating in a very few valley bottoms. Geomorphic instability is again indicated in the stream valleys since 1000 B.P., followed by man-induced activation of the coastal dunes since the late eighteenth century.

The Beacon Island Soil and more recent humic soils represent different phases of the standard Holocene profile on nutrient-poor, highly permeable silica sands, although more effective podsolization is clearly taking place in mesic inland environments. The Brenton Soil differs only in degree from Holocene pedogenesis, and the apparent difference of weathering intensity may reflect more on time than on rates of weathering. However, the Brakkloof Soil and earlier weathering horizons fall well outside this norm. These older mid-Pleistocene-to-late-Tertiary paleosols infer a high degree of geomorphic stability and a completely closed vegetation, in contrast to those episodes with (i) active planation and fanglomerate production, (ii) interior dune formation, or (iii) intensive cold-climate denudation. Comminution of quartz sands to clay sizes requires very intensive chemical weathering, so that a warm, perhumid climate is indicated by the Brakkloof Soil. This paleosol appears to be correlated with a mid-Pleistocene interglacial, thus fitting the pattern of the late Quaternary paleosols, which appear to coincide with relatively warm, interstadial oscillations or with parts of the Holocene.

Cave sediments from the Cape Province

There are a good number of cave sedimentary sequences in the Cape Province that span much or most of the Upper Pleistocene; several of these have been studied analytically. The most complete and informative is from Nelson Bay Cave (Butzer 1973b), a poleward-facing cave just above sea level near Plettenberg Bay.

The Nelson Bay Cave strata begin with an Eemian beach sand and gravel to +10 m, following primary erosion of the cave in relation to a +15–20 m sea level. Coeval with and a little younger than this basal beach in emerged parts of the cave are fine residual sediments, mixed with organic products, extraneous silts, and beach sands of presumed eolian origin. Eventually frost-weathered roof spall became the most striking component, implying a minimal 10°C temperature drop on a now frost-free coast, during the Lower Würm Pleniglacial. This first complex of beds was ultimately impregnated by percolating waters or lateral seepage rich in ferruginous compounds, leading to ferricretion of the topmost 30–40 cm. An unusually wet external climate with increased chelation is inferred, contemporary with formation of the Brenton Soil. Thereafter, during the remainder of the Würm Interpleniglacial, sedimentation in Nelson Bay Cave was limited to very slow accumulation of edge-corroded roof rubble.

During the Upper Würm Pleniglacial, mixed organic-mineral sediments began to accumulate, with a protracted interruption marked by intensive waterlogging in relation to a perched water-table; this 'soil' has strong prismatic structure and suggests alternate wetting and drying. Eventually roof spall increased to a secondary maximum ca. 18,000 B.P., after which another superficial oxidation horizon was formed, with limonitic mottling, banding, or general discoloration of the topmost 10–20 cm, despite incomplete leaching of subjacent strata. The mammalian fauna from these last deposits include an unusually large proportion of open-country forms, in contrast to the closed forest and its restricted fauna of historical times (Klein 1972a, 1972b).

Later strata at Nelson Bay Cave record increasingly strong cultural components, with a rapid rise of sea level ca. 12,000–8,000 B.P. documented by the accelerating influx of marine elements in a faunal spectrum including fewer open-country grazers (Klein 1972b), concomitant with a 2–3°C rise of offshore water temperatures as inferred from oxygen-isotope analysis of marine shell (N. J. Shackleton, personal communication). Beach-derived eolian sands became very prominent 5500 B.P., with a rapid decrease of eolian sedimentation indicated after 4200 B.P., both in a nearby parallel cave sequence as well as by external dunes and paleosols (Butzer and Helgren 1972).

The Nelson Bay Cave data is complemented by a long succession from Klasie's River Mouth (Butzer 1974), west of Cape St. Francis. Here a marine cave is floored by beach gravel related to a +7.5 m sea level. The early substages of the Würm Glacial were thereafter recorded by abundant roof spall, suggestive of frost-weathering, while a relative increase in moisture is indicated by initial stalagmites and again by terminal leaching. Later parts of the Würm Glacial saw deposition of a primary eolianite lense in the cave—mixed with a littl roof spall—under dry conditions. Sedimentation thereafter was minimal except for shell middens that postdate 5000 B.P.

Equally close parallels can be cited from Die Kelders, a coastal cave near Hermanus, in the southwestern Cape Province. Here, cemented angular roof spall accumulated during the regression from a +7 m shoreline but well before accumulation of a major eolianite (Butzer, unpublished).

From interior locations of the Cape Folded Ranges, only two exemplary sequences can be cited here. One of these is Kangkara, an equatorward-facing, open cave at 450 m elevation north of Knysna (Butzer, unpublished). Here, frost-weathered, crude detritus accumulated prior to 44,000 B.P., and again after a long interval of leaching. Deposits apparently younger than 12,000 B.P. are notably fine-grained. The second case is a cave at 210 m elevation near Montagu, on the margin of the subarid Karroo. Here a long Acheulian sequence, of presumed late Middle Pleistocene age, is interbedded with lenticles that contain sesquioxide-stained silcrete grains in the 25–500 micron grade and that are probably of eolian origin (Butzer 1973c). The overyling Middle Stone Age deposits, which date from 20,000 to greater than 50,000 B.P., include several major rockfalls and equally abundant smaller roof detritus; however, the friable sandstone allows no further precisions on spall morphometry, while the stratigraphic detail is inadequate.

Altogether, these various cave sediments substantiate the external pedo- and morpho-stratigraphic evidence of the Cape coastal region, to the effect that repeated changes of the weathering–erosion balance have occurred, that significant oscillations of moisture and vegetation are indicated, and that the coldest episodes of the Upper Pleistocene were reflected by geomorphically-effective freeze-thaw alternations and substantially colder (winter?) climate.

Periglacial geomorphology

A critical survey of Pleistocene 'periglacial' phenomena in southern Africa has already been made elsewhere (Butzer 1973a). This has suggested that the available data are of uneven quality, concepts often vague or erroneous, and interpretations sometimes based on fallacious parallelisms reflecting experience in higher rather than lower latitudes. True 'periglacial' forms and deposits of apparent Upper (and Middle?) Pleistocene age have only been recognized in the Drakensberg and adjacent parts of the Cape Province in latitudes 28°30′–31°20′S. Lower limits of these phenomena lie near 1500–1800 m in the eastern Cape and Natal, but rise from southwest to northeast, to 2600 m in Lesotho. Significant nivation in the Drakensberg is also indicated, but at even higher elevations. Despite the evidence for several phases of accelerated Pleistocene frost-weathering at modern sea level, alleged 'periglacial' phenomena along the coastal margins of the Cape Folded Ranges are as unacceptable as other claims from the Transvaal and Rhodesia, since there is no evidence for cryonival or geliflual processes.

Alluvial terraces of the Vaal Basin

At somewhat lower latitudes (26–29°S) in the continental interior, the alluvial formations of the lower Vaal River are well known for their unique record of the Pleistocene. Recent field examination (Butzer, Helgren, et al. 1973) indicates that this terrace suite is both more complex and more systematic than previously understood, allowing a more detailed stratigraphy and firmer paleo-environmental inferences. The succession consists of three main categories: (i) a group of late Pliocene to early Pleistocene 'Older Gravels,' (ii) a complex of mid-Pleistocene 'Younger Gravels,' and (iii) the fine-grained silt or sand facies of the late Pleistocene to Holocene Riverton Formation, with five members.

The 'Older Gravels' include four discrete fluvial platforms with thin, discontinuous mantles of related cobble grade gravel. These range in elevation from +18 to +90 m and vary considerably along complex longitudinal gradients. In primary contexts the gravels are normally indurated by several meters of calcrete; otherwise, they are intensively weathered to a residual of resistant rocks with a reddish, sandy matrix. There appear to be no in situ faunas or artifacts, and dating is primarily based on extrapolation from the 'Younger Gravels' and consideration of the Tertiary planation surfaces that cut across the basin watersheds. As such, the 'Older Gravels' record four long periods of fluvial planation in relation to a high competence stream, alternating with episodes of effecting bedrock downcutting.

The 'Younger Gravels' include two distinct gravel bodies that fill a rock-cut valley bottom to +10–25 m, each gravel maintaining different longitudinal gradients and, where superposed, separated by a paleosol with a reddish cambic horizons. The 'Younger Gravels' have in the past yielded several facies of Acheulian artifacts and a diverse, Middle Pleistocene mammalian fauna —both unfortunately derived from uncontrolled collecting and uncertain stratigraphic provenance. These cobble gravels also presuppose a much greater stream competence and, significantly, there has been no significant bedrock incision of the Vaal since their aggradation. Surface 'Younger Gravels' are pedogenetically altered with reddish cambic horizons, frequently found under younger calcrete or silcrete duricrusts.

The Riverton Formation begins, as Member I, with over 6 m of massive-bedded, gleyed flood silts, interdigitated with tributary sands or gravels. Following post-depositional calcification and erosional truncation, Member II was deposited to a thickness of 7 m, ranging from flood silts to channel sands that are capped by a 60 cm, light gray, loamy Vertisol. After an interval of downcutting, Member III comprises 9 m of interdigitated lateral and channel deposits, primarily sands and gravels; shell horizons in the terminal bed are dated to a little before 17,000 B.P., and a calcareous paleosol formed on their surface prior to major downcutting of the Vaal through 15 m of older alluvium. Member IV includes 9 m of Vaal flood silts without tributary equivaments, and carbonate resolution within older deposits ca. 8000–6000 B.P. provides a possible clue to its age. Finally, Member V follows another 11 m of fill dissection, and consists of flood silts interdigitated with up to 6 m of tributary sands; the capping, dark gray, relict vertisol has a loamy texture and related subsurface carbonates at −1 to −1.5 m gave an apparent age of ca. 2700 B.P. Younger events include over 6 m downcutting of the Vaal channel, and a variety of colluvial deposits and channel wash, derived from upland dunes and other surficial materials.

The Riverton Formation, the oldest units of which may extend back to the early Upper or even late Middle Pleistocene, records a long history of repeated changes in climate, vegetation, hydrology, geomorphologic steady state, and pedogenesis. Although the specific facies and subdivisions described here are unique to the semiarid lower Vaal River, close parallels are apparent with the Cornelia Formation established in a subhumid part of the basin, in the northeastern Orange Free State (Butzer 1973d). Elsewhere in the Vaal and upper Orange drainage, the undifferentiated Riverton Formation corresponds collectively to the 'Older Fill' that is common to countless major and minor tributaries and which has been outlined by Butzer (1971a). This Riverton fill, as it may be informally called, is typically 6–15 m thick. Surface soils range from humic floodplain soils to dark, clayey vertisols or cambic horizons with subsoil carbonate enrichment, depending on parent material, drainage conditions, and the east-west gradient of precipitation.

From the air the dissected surface of the Riverton fill shows impressive traces of former flood plain features: cutoff meanders, channel bifurcations, basins or backswamps, and even some low dune forms with blowout scars. These once poorly-drained surfaces extend laterally into every tributary valley and peripheral basin and, depending on the restriction of local gully-cutting, there may still be flats of poorly drained ground. The gradient of the Riverton alluvial terraces is also irregular, steepening rapidly below valley constrictions and flattening out above major confluences. An additional element of interest is the channel trace, since rivers had greater sinuosity, with intricate meander loopings, whereas the geometry of the abbreviated modern channels is characterized by markedly reduced wave amplitude, together with increased meander wavelength and radius of curvature. Consequently the streams responsible for alluviation of the Riverton fill had a rather different dynamism, with longer channels, broader floodplains, reduced longitudinal gradients, and a higher proportion of suspended sediment.

Interpretation of the Riverton Formation

Basic to an understanding of alluviation and dissection is the close dependence of geomorphologic equilibrium upon vegetation cover. The "natural" vegetation of most of the Vaal and upper Orange basins is grassveld or grass savanna which, in an undisturbed situation, provides excellent soil protection and slows down surface runoff, indirectly smoothening discharge maxima. In the stream channels of even the largest tributaries such conditions favor finer-grained sedimentation across channel beds choked in grass, reeds, and sedges, and dotted by disconnected pools of water even at the end of the dry season. Situations like this, which contrast strongly with bare sand and gravel floors elsewhere, have been reconstituted below a series of water-control dams. Consequently, geomorphic equilibrium, with slow accumulation of suspended sediments, is best associated with shallow, vegetated channels and a high water-table even during the dry season; on the other hand, deep channels with exposed bed loads and low water-tables relate closely to stream incision and headward gully erosion.

The modern erosional trend in the Orange-Vaal drainage was established during the period 1880–1930 as the grassveld was overstocked with cattle and sheep, and repeated burning accelerated the process of range deterioration. Over much of the area former grassland has been invaded or replaced by communities of low, succulent shrubs where the soil is generally bare at the end of the dry season. Runoff is rapid and unimpeded, stripping topsoil by sheet erosion and cutting gullies wherever flow concentrates on steepening gradients. Stream discharge is marked by aperiodic flood surges that carry large quantities of sediment and mainly deposit bed-load along the major rivers; the intervening base-flow is minimal. As a result there is erosion and dissection, with rapid headward incision, everywhere except along the largest through-rivers.

Flowing across a low-gradient surface and separated from the ocean by repeated stretches of cataracts and waterfalls, the Vaal and upper Orange have not been affected by changes of base level nor, for that matter, is there evidence for tectonic interference during the Pleistocene. For these reasons the cut-and-fill cycles that predate human disturbance must be attributed to changes of effective groundcover and, indirectly, climate (Butzer 1971a).

In general, the Riverton fill suggests long periods with slow geomorphologic change, and a highly effective vegetation. Improved moisture conditions are implied, with a more complete mat of vegetation and increased percolation, so reducing the rate of runoff and discharge, with gradual aggradation of both the footslopes and valley floors. At such times a lush grassveld, neither disturbed by man nor overgrazed by native game, would have prevailed. Periods of dissection or accretion of crude slope detritus probably marked periods of drier climate, with a correspondingly more open vegetation. Such an interpretation closely matches that of the pollen records from Florisbad and Aliwal North, which indicate repeated and significant shifts of vegetation communities between grassveld and Karroo shrub during the late Pleistocene. Interestingly, the stratigraphic relationships and radiocarbon dates suggest that the closed vegetation was present during the cooler glacial stades, open shrub vegetation during the warmer interstadials. This picture is greatly strengthened by a consideration of other aspects of the geomorphologic record.

Spring, tufa, and lacustrine deposits

Significant, environmental changes during the South African Quaternary are equally recorded by such variable phenomena as spring activity, tufa-accretion cycles on the margins of the Kalahari Desert, and a variety of lake or pan deposits in the semiarid sector of the Orange-Vaal drainage. Unfortunately, only a limited number of sequences has been studied in full, although ongoing work promises to provide further information.

The Florisbad pollen sequence as already discussed, reflects both on regional changes of vegetation and on cycles of a deepseated thermal spring. The geology of the site confirms a multiplicity of spring vents within the complex accretional mound of quartz sands and organic lenses, indicating repeated episodes of flush flow and subsequent, gradual quiescence (K. W. Butzer, unpublished). These immediate spring deposits interfinger with a local alluvial sequence, terminating in the nearby pan. A similar sequence occurs in relation to the nearby spring site of Vlakkraal, also with an Upper Pleistocene faunal assemblage (see Cooke 1963), but where the sediments intergrade with the Riverton fill of the Modder River. On geological, faunal, and radiometric reasoning both the Vlakkraal and the younger part of the Florisbad strata can be correlated with Riverton Member III.

Another key spring site is Amanzi, near Port Elizabeth, in the southeastern Cape (Butzer, 1973e). Here two major episodes of flush flow, related to a small, artesian basin, occurred during late Middle or early Upper Pleistocene times, possibly contemporary with a high interglacial sea level. Markedly out of phase with Florisbad is the third and final phase of accelerated spring flow at Amanzi a little before 30,000 B.P., coeval with the Würm Interpleniglacial.

A longer stratigraphic succession is provided by the tufa fans and plugs associated with drainage lines coming off the Gaap Escarpment, immediately west of the lower Vaal River. Here at least four major cycles can be recognized, the basic patterns of which are outlined in Table 3 (see Butzer 1973f). The youngest of these cycles is correlated with Riverton Member III, both on the basis of the regional stratigraphy and a C^{14} date of 26,100 B.P. This places the last interval of accelerated frost-weathering just prior of a significant moist phase, in either the Lower or Upper Würm Pleniglacial. The Gaap tufa cycles bear some relation to the australopithecine-bearing, early Pleistocene cave breccias of the Transvaal. Despite earlier claims that these breccias reflect on complex climatic oscillations (Brain 1958), they can better be attributed to ruptures of weather/erosion balances (Butzer 1971b). Ongoing studies show that these cave fills were derived from deep soil mantles or hillwash accumulations, not from hillside lithosols (Butzer, unpublished).

The oldest lacustrine deposits studied in South Africa to date are calcrete lake chalks that alter-

TABLE 3

An explanatory model for geomorphologic cycles on the Gaap escarpment

Phase	Dominant geomorphic process	Vegetation and climate
4.	Declining spring and fluvial activity leading to erosion; karstic solution with fissure and cave formation.	Grassveld but groundcover deteriorating; warm and increasingly dry.
3.	Maximum spring discharge and protracted fluvial activity, leading to large-scale tufa accumulation—initially of subhorizontal aprons, later of steep carapaces; cementation of older breccias and cave fills.	Grassveld providing optimal groundcover and infiltration; cool and humid.
2.	Accelerated fluvial and colluvial activity wash together surficial sediments into valleys and cave fissures; accelerated frost-weathering leads to talus accumulation.	Incomplete groundcover with poor infiltration; cool and inincreasingly wet.
1.	Accelerated eolian activity, with limited colluvial and fluvial processes, leading to thin but extensive mantles of eolian sediments and wash.	Thorn savanna with limited groundcover and infiltration; warm and dry, in part drier than today.

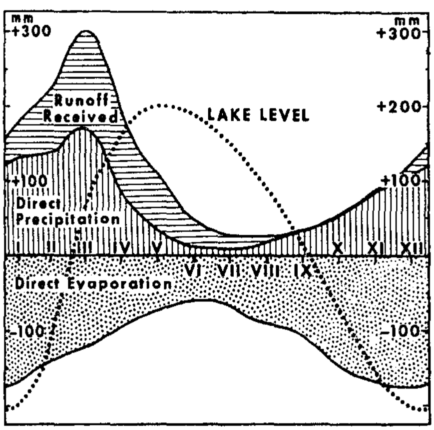

FIG. 2. Hypothetical hydrological budget for Pleniglacial Lake Alexandersfontein (non-outlet) (after Butzer et al. 1973).

nate with reworked eolian sands in a number of deflation pans of the Kimberley region (Butzer 1973g). Two such sequences contain in situ Acheulian assemblages, the younger of which has a tentative Th/U date of 115,000 B.P. Presumably of mid-Pleistocene age, such strata again record a long succession of hydrological oscillations.

More immediate are the younger lacustrine beds and beach forms of the Alexandersfontein Pan. These record several nonoutlet lakes, the largest of which was up to 19 m deep and had a surface area of 44 km². This dates from the Upper Würm Pleniglacial, a little before 16,000 B.P. (Butzer, Fock, et al. 1973). An obvious correlative of Riverton Member III, this lake was followed by another, with a maximum depth of 12 m, an area of 24 km², but as yet undated.

A possible hydrological budget for the 19 m Pleniglacial lake is shown by Fig. 2, assuming a 6°C drop in mean temperature, evaporation of water correspondingly reduced from 2120 to 1400 mm, and a runoff quotient of 9.3 percent. These assumptions are documented or discussed by Butzer, Fock, et al. (1973); they specifically presume a seasonal precipitation and evaporation regime parallel to that of today. Although the modern precipitation mean for the Alexandersfontein basin is only 397 mm, the hydrological computations would require 878 mm rainfall to balance with a 19 m lake, 671 mm with a 12 m lake. In other words, even with an appreciable cooling, a rainfall increase of at least 50 percent is implied by the 12 m lake, an increase of at least 100 percent by the 19 m lake. This shows conclusively that the Upper Würm Pleniglacial climate of the Vaal-upper Orange drainage was both substantially cooler and wetter than today, in a relative as well as in an absolute sense. This bears out the more equivocal interpretation of the Florisbad and Aliwal North pollen sequences, of the Riverton Formation, and of the Gaap tufas.

DISCUSSION

The available evidence indicates that repeated and significant environmental changes occurred in southern Africa during the Quaternary. There

TABLE 4

Tentative Late Quaternary sequences from interior (26–30°S) and coastal (32–34°30′S) environments of southern Africa

Geological age	Vaal and Upper Orange drainage	Southeastern Cape coast
Late Holocene	Overgrazing and range deterioration; active gullying with alluviation of major channels; local eolian activity. Since 1880's.	Deforestation and reactivation of coastal dunes; gullying. Since early 19th century.
	Brief cut-and-fill cycle ('Younger Fill'); initially, activation of local dune fields. Stabilization before 1600 A.D.	Brief cut-and-fill cycle, initiated ca. 900 A.D.
	Aggradation of Riverton Member V, with local vertisol development after stabilization (? by ca. 3000 B.P.).	Development of humic soils; local valley peat accumulation; stabilization of coastal dunes with forest readvance. Beginning 4200 B.P.
Middle Holocene	Stream dissection, local eolian activity.	Accumulation of littoral dunes, with forest recession; soil stripping on slopes leading to valley aggradation. Ca. 7100–4200 B.P.
	Aggradation of Riverton Member IV.	Development of weak podsolic or humic soils; stabilization of coastal dunes with temporary forest advance. 8th millenium B.P.
Early Holocene	Increasingly xeric vegetation; stream dissection. Beginning 10,000 B.P.	Accumulation of littoral dunes; faunal shift from mixed open-country and woodland forms to forest assemblage; offshore waters warming 2–3°C. Ca. 16,000–8,000 B.P.
Late Würm Glacial	Repeated oscillations of Karroo grassveld boundary (Aliwal North profile), reflecting changes from cooler and wetter to drier and warmer conditions; fluctuating water levels in Paleo-lake Alexandersfontein; pedocal and duricrust formation. Beginning after 16–17,000 B.P., terminating by 10,000 B.P.	
Upper Würm Pleniglacial	Maximum extent of mesic grassveld; high lake levels (+19 m Paleo-lake Alexandersfontein); aggradation of Riverton Member III; periglacial phenomena in mountains (Lesotho). Temperatures at least 6°C lower, precipitation double that of today. Beginning 29–25,000 B.P., terminating 17–16,000 B.P.	Slow development of humic soils on littoral sands, with two phases of accelerated pedogenesis (cave oxidation horizons); frost-shattering in upland caves; prominence of open-country mammals. Beginning after 30,000 B.P., terminating ca. 16,000 B.P.
Würm Interpleniglacial	Stream dissection, local eolian activity.	Development of weak podsolic soil, with increased chelation; limited cave sedimentation with leaching or ferruginization; accelerated spring discharge (Amanzi). Terminating by 30,000 B.P.
Lower Würm Pleniglacial	Shift from dry, warm conditions to increasingly mesic vegetation (partial Florisbad profile); aggradation of Riverton Member II, followed by local vertisol formation; frost-shattered talus and tufa accumulation (Gaap Escarpment);? periglacial phenomena in mountains.	Downcutting of valleys in response to low sea level; active frost-shattering in coastal and upland caves indicate 10°C temperature depression. Terminating by 40,00 B.P.?
Early Würm Glacial	(Stratigraphic positions uncertain)	Accelerated denudation, with deposition of slope breccias and screes with frost-shattered detritus; increasing frost-weathering in caves. Beginning after 75,000 B.P.?
Eem Interglacial		Aggradation or cutting of beach forms in relation to +5–12 m sea level with thermophile mollusca; related estuarine or fill terraces in coastal valleys; accelerated spring discharge (Amanzi). Before 75,000 B.P.?

are strong implications for drier climates and also for periods of greatly augmented precipitation. Much colder conditions can be inferred from geological and paleobotanical evidence. It is, however, still impossible to set up a climatic chronology for the subcontinent. In fact, it is premature to attempt detailed regional stratigraphic columns outside the two areas under discussion, and even here basic correlations are uncertain prior to the mid-Upper Pleistocene. With these reservations, a tentative and partly hypothetical chart of the two parallel sequences is offered in Table 4. The terminology of the chronostratigraphic units follows that of van der Hammen and others (1967). Approximate dating to the mid-Upper Pleistocene is based on C^{14} determinations.

Comparison of the two columns shows that ecological shifts were similar in the Vaal-Orange drainage and in the southeastern Cape coastal region throughout the Holocene time range. Morphostatic and morphodynamic trends in the sense of Erhart (1967) can be recognized, the former leaving a tangible record of soil development, the latter favoring a negative soils balance but creating a sedimentary or erosional record. Relatively moist conditions are indicated in the 8th millenium B.P. and again ca. 4200–1000 B.P., with drier conditions prevalent at other times. These changes find no parallels in the lacustrine record of East Africa, the Sahara, or northern Africa.

During the Upper Pleistocene, environmental anomalies in the interior and at the coast diverge and, except for the records of greater cold, appear to be out of phase if not inverse. Thus the pleniglacials were wet in the interior, but relatively dry on the coast, whereas all warm oscillations were dry in the interior, yet favorable to pedogenesis and relatively moist on the coast. Neither area shows any parallels with the climatic shifts evident in the contemporaneous record of innertropical Africa (Butzer 1971c; Butzer et al. 1972; Isaac et al. 1972; Zinderen Bakker 1972), and the Vaal-Orange patterns find only limited similarities in the Sahara at 26-30°N latitude. By contrast, the coastal region suggests surprising analogies with the eco-history of the Mediterranean Basin (see Butzer 1971c).

Two explanations can be offered for the observed paleoclimatic trends in southern Africa, assuming that the major changes of Quaternary temperature climate were of a planetary nature. It can, firstly, be posited that the glacial-age climatic belts were contracted, symmetrically about the equator, with the subtropical anticyclones of the South Atlantic and Indian oceans shifted equatorward (Hastenrath 1972; Zinderen Bakker 1967). This would reduce the impact of the South Atlantic high and the Benguela Current on the interior of southern Africa, while displacing the zonal boundary of the southern circumpolar westerlies and the tropical easterlies northward. Large parts of the subcontinent were therefore more influenced by the mid-latitude westerlies while cold fronts could regularly produce cyclonic rains here throughout the year. Such rains would have had an important influence on lake levels, on the vegetation, and on geomorphologic equilibrium, since evapotranspiration was lower because of the decrease in temperature. Under these conditions a dense sward of mountain grassveld could have covered vast areas of the Orange Free State, spreading southward at the expense of the Karroo semidesert vegetation. At the same time cyclonic activity at higher, coastal latitudes did not increase, although the harsh climate may have promoted denudation rather than pedogenesis. Changes in the opposite direction occurred with the warmer and drier, nonglacial pattern, when the Karroo occupied much of the interior and a positive soil balance was reestablished in the coastal region.

An alternative hypothesis, suggested to us by J. Bjerknes, posits that the northern hemisphere glacial-age circulation may have been greatly strengthened as a result of the new continental glaciers. Accordingly, the tropical circulations were displaced southwards, bringing West and East Africa into the dry Saharan periphery while placing the Vaal-Orange drainage completely within the summer rainfall belt.

Whatever the best explanations, it should be realized that a great number of diverse ecological changes or fluctuations have taken place during the course of the South African Quaternary, and that only those which were of sufficient amplitude and duration are registered in the fossil record. Far more paleoclimatic evidence is needed as well as greater stratigraphic detail and, above all, improved radiometric precision. Nonetheless, the data available at present provide a provocative yet basically coherent picture that fits well within the emerging scheme of paleoclimatic evolution.

REFERENCES

Brain, C. K. 1958. The Transvaal ape-man bearing cave deposits. Mem. Transvaal Mus. 1: 1–121.

Butzer, K. W. 1971a. Fine alluvial fills in the

Orange and Vaal Vasins of South Africa. Proc. Assoc. Amer. Geog. 3: 41–48.

Butzer, K. W. 1971b. Another look at the australopithecine cave breccias of the Transvaal. Amer. Anthropol. 73: 1197—1201.

Butzer, K. W. 1971c. Environment and archeology: an ecological approach to prehistory. Chicago, Aldine-Atherton.

Butzer, K. W. 1973a. Pleistocene "periglacial" phenomena in Southern Africa. Boreas 2: 1–12.

Butzer, K. W. 1973b. Geology of Nelson Bay Cave, Robberg, South Africa. S. Afr. Arch. Bull. 28: in press.

Butzer, K. W. 1973c. A provisional interpretation of the sedimentary sequence from Montagu Cave (Cape Province), South Africa. (*In* Montagu Cave, in press, C. M. Keller (ed.), University of California Press, Berkeley and Los Angeles.)

Butzer, K. W. 1973d. Geology of the Cornelia Beds, northeastern Orange Free State. Mem. Nat. Mus. Bloemfontein: in press.

Butzer, K. W. 1973e. Spring sediments from the Acheulian site of Amanzi, Uitenhage District, South Africa. Quaternaria 16: in press.

Butzer, K. W. 1973f. Paleo-ecology of South African australopithecines: Tau'ng revisited. Current Anthropology 14: in press.

Butzer, K. W. 1973g. Geomorphology and geo-archeology of two Acheulian 'Pan' sites in the Kimberley District: Doornlaagte and Rooidam. Ann. Cape Prov. Mus. (Nat. Hist.) (Grahamstown). J. Arch. Sci. 1: in press.

Butzer, K. W. 1974. Geomorphology and sediment stratigraphy of archeological sites at Klasies River Mouth and Peelhoutboom. (*In* Excavations at Klasies River Mouth, in press, R. Singer and J. Wymer (eds.), Univ. Chicago Press.)

Butzer, K. W., G. J. Fock, A. Stuckenrath, and A. Zilch. 1973. Paleo-hydrology of late Pleistocene Lake Alexandersfontein, Kimberley, South Africa. Nature: in press.

Butzer, K. W. and D. M. Helgren. 1972. Late Cenozoic evolution of the Cape Coast from Knysna to Cape St. Francis, South Africa. Quart. Res. 2: 143–169.

Butzer, K. W., D. M. Helgren, G. J. Fock, and R. Stuckenrath. 1973. Alluvial terraces of the lower Vaal River, South Africa: a re-appraisal and re-investigation. J. Geol. 81: in press.

Butzer, K. W., G. Ll. Isaac., J. L. Richardson, and C. Washbourn-Kamau. 1972. Radiocarbon dating of East African Lake Levels. Science 175: 1069–1076.

Coetzee, J. A. 1967. Pollen analytical studies in East and Southern Africa. Palaeoecology of Africa 3: 1–146.

Cooks, H. B. S. 1963. Pleistocene mammal faunas of Africa, with particular reference to southern Africa. (*In* African ecology and human evolution, pp. 65–116, F. C. Howell and F. Bourlière (eds.), Aldine-Atherton, Chicago.)

Dingle, R. V. and J. Rogers. 1972. Effects of sea-level changes on the Pleistocene ecology of the Agulhas Bank. Palaeoecology of Africa 6: 55–58.

Erhart, M. 1967. La génèse des sols en tant que phénomene géologique, 2nd ed. Masson, Paris.

Flint, R. F. 1959. Pleistocene climates in eastern and southern Africa. Bull. Geol. Soc. Am. 70: 343–374.

Hammen, T. van der, G. C. Maarleveld, J. C. Vogel, and W. H. Zagwijn. 1967. Stratigraphy, climatic succession and radiocarbon dating of the Last Glacial in the Netherlands. Geol. en Mijnbouw. 46: 79–95.

Hastenrath, S. 1972. A note on Recent and Pleistocene altitudinal zonation in southern Africa. S. Afr. J. Sci. 68: 96–102.

Isaac, G. Ll., H. V. Merrick, and C. M. Nelson. 1972. Stratigraphic and archaeological studies in the Lake Nakuru Basin, Kenya. Palaeoecology of Africa 6: 225–232.

Klein, R. G. 1972a. Preliminary report on the July through September 1970 excavations at Nelson Bay Cave, Plettenberg Bay (Cape Province, South Africa). Palaeoecology of Africa 6: 177–208.

Klein, R. G. 1972b. The late Quaternary mammalian fauna of Nelson Bay Cave (Cape Province, South Africa): its implications for megafaunal extinctions and environmental and cultural change. Quat. Res. 2: 135–142.

Martin, A. R. H. 1968. Pollen analysis of Groenvlei lake sediments, Knysna. Rev. Palaeobot. Palyn. 7: 107–144.

Partridge, T. C. 1969. Fluvial features and climatic change during the Quaternary in South Africa. S. Afr. Arch. Bull. 24: 106–116.

Zinderen Bakker, E. M. van. 1957. A pollen analytical investigation of the Florisbad deposits (South Africa). (*In* Third Pan-African Congress on Prehistory, Livingstone 1955, pp. 56–67, J. D. Clark (ed.), Chatto & Windus, London.)

Zinderen Bakker, E. M. van. 1967. Upper Pleistocene and Holocene stratigraphy and ecology on the basis of vegetation changes in sub-Saharan Africa. (*In* Background to evolution in Africa, pp. 125–147, W. W. Bishop and J. D. Clark (eds.), Univ. Chicago Press.)

Zinderen Bakker, E. M. van. 1972. Late Quaternary lacustrine phases in the southern Sahara and East Africa. Palaeoecology of Africa 6: 15–27.