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Geomorphological evolution of the southern coastal zone of Kenya

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Abstract

Information on chronology, terrace levels and morphological elements associated with former sea level of southern Kenya are reviewed. Additional palaeontological and lithostratigraphical evidence is presented that may be relevant to the interpretation of major palaeogeographical events that occurred particularly during the Pleistocene. This paper therefore provides a description of the active processes in the genesis and development of coastal landscape in southern Kenya.

Up to eight former sea level still-stands can be recognized by comparing the heights of coastal terraces formed during the Pleistocene. These can be grouped into four main levels; namely the Ganda Terrace (over 20 m + MSL), the Kilifi Terrace (15–20 m + MSL), the Malindi Terrace (7–10 m + MSL) and the Shelly Beach Terrace (5 m + MSL). The lowest terrace is the modem reef platform, and which occupies a level of 2 m + MSL. Although it is difficult to ascertain whether this lowest marine terrace is of Pleistocene or Holocene age, it can be related to a sea level pause at its present position about 30000 years ago.

The present study demonstrates that the development of this coast is associated with eustatic, isostatic and tectonic movements. The raised coral reef formations and other marine features present between 10 and 20 m + MSL are the evidence for such processes. They have been dated at Middle Pleistocene to Early Holocene, a period which encompasses the Würmian Glaciation, when sea level regressed to 120 m—MSL lower than that of modem times. There is no evidence to support the contention that the postglaciation eustatic rise in sea level ever reached above 10 m + MSL levels over the last 20000 years. This is because during that time the Kenyan coast had been uplifted by above 60 m as a result of neotectonic movements associated with isostatic response to crustal loading. The presence of parallel step faults on all but the Shelly Beach terrace and the absence of a true fore-reef facies in the raised coral reefs are the main evidence of tectonic uplift of the coastal zone.

Keywords: Coastal geomorphology; Pleistocene; Terraces; Sea level indicators; Eustatic; Isostatic and tectonic movements; Raised coral reefs; Mechanisms; Glacial Control Theory

1. Introduction

The East African seaboard is an Afro-trailing edge type of coastal margin (Inman and Nordstrom, 1971). Its configuration has been moulded by the action of eustatic sea level oscillations in combination with isostatic and differential tectonic movements.

The number of creeks north of Mombasa (Fig. 1) suggests that there has been a recent rise in sea level. The submerged marine terraces are in keeping with this transgression; however several raised beach terraces, now occurring 3–5 km inland, indicate a substantial fall in sea level at some time in the past.

Fig. 1 shows the wide variety of deposits in the Malindi area; the estimated ages of stratigraphical succession are given in Table 1 and a geological cross-section for the Mombasa area is shown in Fig. 2. The Taru Grits and the Duruma Sandstone series, not represented on the map, are believed to form the immediate substrate to this area (Oosterom, 1988). Recent deposits comprise Recent Dune Sands, Recent Beach Sands and Tidal Flat Deposits, and Recent Alluvium.

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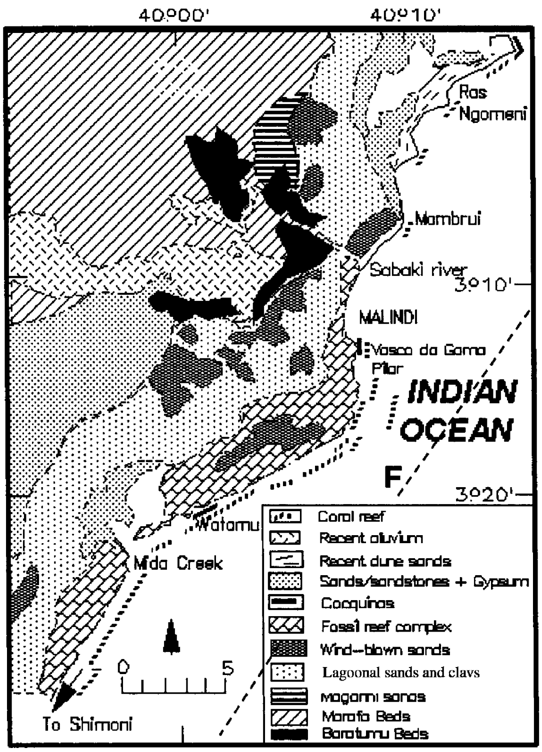

Fig. 1. Geological map of the coastal zone in the Malindi area north of Mombasa (redrawn from Thompson, 1956); the postulated Ruvu Mombasa Fault (F) is shown (scale is in km). The estimated ages of stratigraphical succession are given in Fig. 2.

2. Geological setting

2.1. Palaeozoic, Mesozoic and Tertiary rocks

The Palaeozoic-Mesozoic outcrops, stratigraphy and plate tectonic processes of coastal Kenya have been described by Cannon et al. (1981) and Rais-Assa (1988). Rifting occurred along the northeastern part of Gondwanaland at the inception of the proto Indian Ocean during Upper Carboniferous period. Vertical and horizontal tectonic movements, associated with the final phase of breakup of Gondwanaland, took place during Permo-Triassic times and final drowning of the continental margin is reflected by major marine incursion during the Jurassic (Kent, 1974).

Along certain straight segments of the coast, the shelf is markedly absent, suggesting a fault origin. This occurrence is supported by a sudden drop of the sea-floor topography off the Kenyan coast which is attributed to the postulated Ruvu-Mombasa fault (Abuodha, 1998). This fault apparently maintains a NNE-SSW orientation throughout (Fig. 1).

The underlying sedimentary sequence consists of the Upper Carboniferous Taru Grits overlain by Permo Triassic Karroo-strata represented by the Duruma Sandstone Series (Caswell, 1956; Thompson, 1956). The Tertiary sediments are represented by the Baratumu.

Table 1
Correlation of stratigraphies used by previous authors

		Gregory (1921)	Mombasa-Kwale Caswell (1953)	Malindi Thompson (1956)	Hadu-Fundisa Williams (1962)	Southeast Kenya Oosterom (1988)
Pleistocene	Upper		*Exact correlation uncertain* Red wind-blown sands and raised alluvial deposits	Gedi Beacon Sands	Plesitocene Dune Sands	Younger Pleistocene Sands (aeolian)
				Pleistocene Sands	Pleistocene Lagoonal Sands and Clays	Older Pleistocene Sands (fluviatile)
	Middle	Raised coral reefs	Raised coral reef North Mombasa Crag Kilindini Sands	Fossil coral reef coral reef breccia	Raised coral reef	Pleistocene reef complex
						Lagoonal Sands and Clays Changamwe/Kipevu Beds
	Lower		Magarini Sands (aeolian)	Magarini Sands • Upper Member (aeolian) • Lower Member (fluviatile)		
				Cocquinas	Marafa Beds (fluviatile)	Upper Magarini Formation Lower Magarini Formation
U. Pliocene		North Mombasa Crag Magarini Sands	Magarini Sands (Fluviatile) Changamwe Deposits (marine)	Marafa Beds (marine-deltaic deposits)	Midadoni Beds (fluviatile)	Marafa Formation

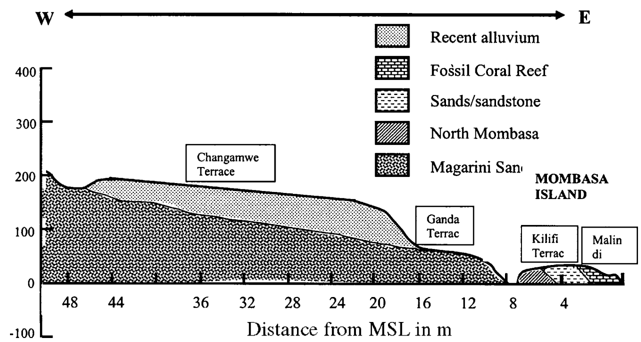

Fig. 2. Geological cross-section along Mombasa Island and Mainland showing coastal terraces and probable variations of sea level in relation to land level during Pleistocene time.

Beds (sandstones with subordinate shales and limestones) of Miocene age and the Pliocene Marafa Beds (Fig. 1). The latter comprises unconsolidated sands, and sandstones with subordinate shales and marls. Thompson (1956) dated them as Upper Pliocene to Early Pleistocene based on the determination of fossil foraminifera. The Magarini Sands comprise unconsolidated quartzose sands, which locally include gravels or clays, and can broadly be subdivided into two different stratigraphic members. The Lower Member is formed by Plio-Pleistocene fluviatile sands (Caswell, 1953; Williams, 1962; Oosterom, 1988). The Upper Member was distinguished in the Malindi area by Thompson (1956) as sands of aeolian origin; both Caswell (1953) and Thompson (1956) considered the Upper Member to be of Pleistocene age.

2.2. Pleistocene and holocene deposits

During the Pleistocene period, the Malindi coast was affected by global eustatic sea level oscillations, which are reflected in geomorphology and stratigraphy of the area. Oosterom (1988) has correlated the sequence of events that operated along the Kenyan coast with events that occurred during the glacial history of Europe. The existing correlations of the Pleistocene sequence described above with particular sea level stands have neither been satisfactory nor have they been unanimously accepted, however. The following account is based on the work of Caswell (1956), Thompson (1956), Williams (1962), Braithwaite (1984) and Oosterom (1988).

Brief descriptions of the Pleistocene lithological units are given by Oosterom (1988) for the coastal rocks and unconsolidated deposits. Three main types of Pleistocene formations are identified along the coastal plain; from west to east these are: Lagoonal Sands and Clays, Fossil Reef Complex and Wind-blown Sands (Fig. 1). The basement to these units is probably a narrow wave cut platform that was beveled into the Cretaceous and Jurassic formations during the Middle Pleistocene. Table 1 gives a correlation of stratigraphies used by previous authors.

The Lagoonal Sands and Clays (Fig. 1; Table 1) were recognized to be equivalents of the Kilindini Sands by Caswell (1953), the Pleistocene Sands by Thompson (1956) and the red sandy laterite by Braithwaite (1984). They comprise mainly quartz sands with subordinate silts and clays. Williams (1962) proposed a lower Upper Pleistocene age based on the presumed age of a terrace he identified at 12 m + MSL elevation.

The quartzose Lagoonal Sands and Clays are considered to be contemporaneous with the Fossil Reef Complex, based on their fossil fauna (Caswell, 1953; Thompson, 1956). Due to the small extent of the local drainage basin, only covering the coastal zone, Caswell (1953) concluded that the clastic constituents were derived from the nearby Duruma Sandstones of Permo Triassic age, and from the Early Pleistocene Magarini Sands. The fact that the Lagoonal Sands and Clays and coral reef formations are more or less contemporaneous suggests that the rate of clastic sedimentation was rather low, allowing the corals to flourish despite the siltation.

The *Fossil Reef Complex* consists of an assemblage of coral limestone and calcarenites with intercalations of quartz sands, pebbly sandstone pebbles, siltstone and algal limestone. The main outcrops of the Fossil Reef Complex are found from the Sabaki river southward to Shimoni (Fig. 1), a coastline distance of approximately 200 km. This complex extends 3–5 km inland from the present shoreline, underlying the coastal plain, and attains elevations of up to 30 m + MSL level. However, Caswell (1956) determined from borehole records that the limestone might reach to a depth of 60 m − MSL level, so that a maximum thickness of about 90 m may be assumed for the Fossil Reef Complex.

Caswell (1953) proposed a Middle Pleistocene age for the Fossil Reef Complex, by connecting the reef buildup with the second interpluvial. In a later study of the Pleistocene limestones of the Kenyan coast, Braithwaite (1984) concluded that these deposits formed about 125 000 years ago, when sea level stood at between 15 and 20 m + MSL, based on radiocarbon dates of the raised coral reef component.

The North Mombasa Crag, comprising calcareous sand, shelly sands and clays (marls) are the sediments which form the foundation of the Fossil Reef Complex. This "basal" unit has been interpreted by Braithwaite (1984), (1) to be a product of mid-Pliocene crustal movements, and (2) to have formed in a subaerial environment during a period of low sea level stand.

The *Wind blown Sands* were recognized to be equivalents of the Gedi Beacon Sands by Thompson (1956) and Pleistocene Dune Sands by Williams (1962). The latter differentiated three dune ridges on the basis of platforms on which they accumulated, at 60 m, 36 m and 9 m + MSL, respectively. The suggested age, of Upper Pleistocene, for the Gedi Beacon Sands (Thompson, 1956) was based on the assumption that the retreat of the sea from the 36 m + MSL terrace took place during that time, exposing offshore bar deposits to aeolian reworking to produce the dunes. Cocquinas are prominently exposed at Watamu Beach, Vasco da Gama Pillar at Malindi and Ras Ngomeni Peninsula (Fig. 1) and consist of wind blown, carbonate-rich deposits, derived from beach materials, mainly shells. These aeolian deposits are strongly cross-bedded which led Thompson (1956) to interpret them incorrectly as offshore bar deposits. Dunes, sporadically distributed and related to the cocquinas, are present along the southern coast of Kenya. Ase (1981) has shown that the dune system continues northward into southern Somalia, a coastline distance of approximately 2000 km.

3. Terrace correlations

3.1. Previous work on coastal terraces

The marine terraces of the coastal zone of Kenya (Figs. 2 and 3; Table 2) have been classified as the Marafa Terrace (80–120 m + MSL), Changamwe Terrace (45–70 m + MSL), Ganda Terrace (20 m + MSL), Kilifi Terrace (15–18 m + MSL), Malindi Terrace (7–10 m + MSL) and the Shelly Beach Terrace (4.5 m + MSL) in decreasing order of height and probably of age (Hori, 1970; Toya et al., 1973; Ase, 1978, 1981; Read, 1981; Braithwaite, 1984; Oosterom, 1988). In this review, however, only the four lower levels on which sufficient data are available are considered. Battistini (1969) and Braithwaite (1984) have also provided a correlation with terraces of coastal Madagascar and northern Tanzania, respectively.

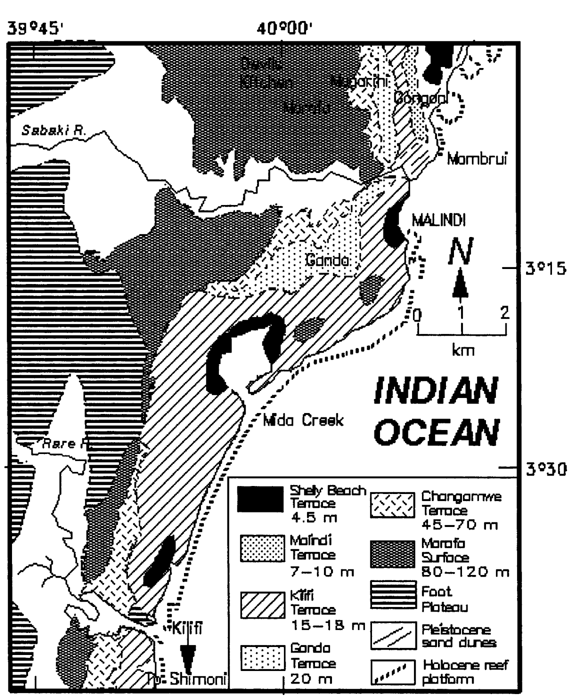

Fig. 3. Geomorphological map showing terraces of the coastal zone between Kilifi and Gongoni (redrawn from Hori, 1970).

Caswell (1953) identified four marine terraces at 61, 31, 9 and 4.5 m + MSL levels and designated them as post Pliocene. Thompson (1956) was the first to observe underwater terraces at 8 m and 35 m − MSL. Ase (1978, 1981) classified the terraces into eight groups, with the youngest terrace level I corresponding with the present reef platform, level II with the Shelly Beach Terrace, terrace levels III–IV with the Malindi Terrace and levels V–VIII with the Kilifi Terrace. Read (1981) recognized a buried terrace (?) at 3–7 m + MSL; in this study, evidence for lower sea levels is supported by submerged platforms at 5 m and 15 m − MSL. Perhaps the simplest description of fossil shorelines was the list of benches recognized by Braithwaite (1984) at 20, 10 tol 2, 8, 6, 4, 2 and 0 + MSL and, 8 and 35 m − MSL. They are considered to represent pauses and reversals in the local changes of sea levels during the postglacial period (see subsequent section). A later study by Oosterom (1988) showed the existence of the lowest exposed level at 4 m + MSL which he referred to as the Uhuru I level. The present reef platform is conspicuous at about 2 m + MSL; according to Braithwaite (1984), it was defined 30 000 years ago when sea level paused at its present position during a general decline to 120 m − MSL.

Previous work on terraces in other sections of the East African coast such as Alexander's (1968) Sukura, Tanga and Mtoni terraces in northern Tanzania and Battistini's (1969) Tatsimian (Reef I), Karimbolian (Reef II) and Flandrian Level in Madagascar, correlate very closely with the Kenyan Kilifi, Malindi and the Shelly Beach terraces, respectively.

3.2. An overview of marine terrace formation

There is a general consensus that marine terraces are associated with sea level fluctuations but contradictions arise on the exact mechanism(s) that contributed to their development. The answer to the active processes of terrace formation has direct implications on its age and must be clarified to avoid misleading interpretation on the nature, magnitude and timing of the Pleistocene events.

The first mechanism attributes the genesis of marine terraces to stability pauses in the process of abrupt falling of sea levels (Braithwaite, 1984). Braithwaite (1984) also concluded that the thin cover of marine deposits placed on top of terraces show episodes of brief reversals. However, this authigenic deposition is most likely a syngenetic feature. The second mechanism considers the

Table 2
Correlation of the coastal terraces in southern Kenya according to earlier studies

	Sikes (1930)	Caswell (1953)	Thompson (1956)	Battistini (1969)	Hori (1970)	Toya et al. (1973)	Ase (1978)	Read (1981)	Braithwaite (1984)	Oosterom (1988)	Age (Oosterom)
Study area	Southern Kenya	Mombasa area	Malindi area	Mombasa-Malindi area	Southern Kenya	Southern Kenya	Mombasa area	Mombasa-Mtwapa area	Southern Kenya	Mombasa area	
Terrace name and elevation above datum	75 m	Foot Plateau	Foot Plateau		Matuga Surface 80–120 m	Marafa Terrace 80–120 m				Cambini-I 85–130 m Cambini-II 70–90 m	950000 BP 750000 BP
	50 m	Foot Plateau	Foot Plateau		Changamwe Surface 45–70 m	Changamwe Terrace 45–70 m				Tezo (I + II) 40–85 m	750000 BP 650000 BP
	12 m level	37 m Terrace	36 m Terrace	18–25 m platform	Upper Mombasa Terrace 15–37 m	Ganda Terrace 20–37 m	VIII: 20 m	Ganda Terrace 20 m + ?	+ 20 m	Majaoni-I 35–50 m	650000 BP
	24 m level									Majaoni-II 25–25 m	475000 BP
						Kilifi Terrace 15–18 m	VII: 17 m VI: 15 m V: 13.5 m	Kilifi Terrace 15–18 m		Mtondia-I 15–25 m Mtondia-II 10–15 m	300000 BP 180000 BP
	9 m old sea beach	7.6 m platform	7.5 m beach Terrace	7–10 m platform	Lower Mombasa Terrace 10 m	Malindi Terrace 7–10 m	IV: 12 m III: 9 m II: 7 m	Malindi Terrace 7–10 m	+ 10 to + 12 m + 8 m	Mackenzie-I 8–10 m Mackenzie-II 6–8 m	130000 BP
	4.5 m beach	–	–		Shelly Beach Terrace 5 m	Shelly Beach Terrace 4.5 m	I: 5 m	Shelly Beach Terrace 4.5 m	+ 6 m + 4 m	Uhuru-I 4–6 m	10000 BP
			2 m level						+ 2 m Present-day reef platform	Uhuru-H 2–4 m	0

Submarine terraces are present at −8 m and −35 m (Thompson, 1956) and at −5 m and −15 m (Read, 1981).

formation of marine terraces to mark the initiation of transgressive phases (Oosterom, 1988).

If the falling sea level hypothesis is considered then the age of shore elements should decrease with height. To some extent, evidence of such a relationship can be seen by studying more detailed mapping and dating of shore levels by Hori (1970), Ase (1981) and Oosterom (1988). The latter work also shows that the raised beaches and terraces were formed during short periods with relatively stable shorelines.

3.3. Eustatic changes in the Pleistocene and Holocene

In this paper, the model adopted to elucidate the Pleistocene events relies mainly on the established evidence for alternating periods of glaciations and interglaciations; by definition these are considered the principal cause for eustatic sea level fluctuations. The surrogate concept to this, the Glacial Control Theory has frequently been invoked in sea level studies (Bloom et al., 1974).

Table 3 shows that regression maxima occurred during the Mindel and Riss glaciations. Large areas of the present continental shelf were subjected to terrestrial conditions; near the old shorelines, wave action and shallow water sedimentation continued to exert influence.

Caswell (1953) and Oosterom (1988) compared the eustatic sea level changes along the Kenyan coast and the glacial events for the high altitudes. In East Africa, Leakey (1950) and Hove (1980) correlated Gunz, Mindel, Riss and Würm glacial periods to Kageran, Kamasian, Kanjeran and Gamblian pluvials, respectively. The Mindel and Riss glaciations are considered to be related to the 60 m and 45 m—MSL submarine terraces represented by the drowned valleys such as the deep Kilindini Channel and Malindi's offshore banks, respectively (Caswell, 1953; Thompson, 1956). According to Thompson (1956), a platform found at 8 m—MSL found near Watamu could have been formed during the Würm glaciation.

During the Lower Pleistocene (Table 3), the 60 m—MSL marine platform developed as a narrow fringe along the coast, indicating that events that formed it were of short duration. This platform served as the basement upon which the Lagoonal Sands and Clays, Coral Reef Complex, the North Mombasa Crag and the Cocquinas were deposited during the subsequent interglacial (Table 3). Towards the end of this episode, the sea level must have risen to about 30 m + MSL as indicated by fossil coral reef outcrops on the 30 m + MSL terrace. At a section cut into a limestone quarry south of Malindi town, lithological interpretations show that during this transgression, the eustatic rise of sea level was not constant but was marked with pauses and even with brief reversals. The quarry exposures show evidence of reef chocking represented by thin elastic beds separating coral growth. These elastic sediment intercalations may not be due to silting and hiatus related to falling sea level but merely to silting and death of reef systems. Under favorable environmental conditions the coral reef could again be rejuvenated above such deposits.

The Glacial Control Theory described by Bloom et al. (1974) states that coral growth would mainly be vertical during periods of rapid sea level rise, horizontal during slow rise or pauses. Regression periods are related to subaerial erosion of existing reefs and deposition of alluvial and aeolian sediments. Some brief reversals were also characterized by deposition of thin marine beds (Braithwaite, 1984). Estimates for the time taken for coral formation range from 10 000 years (Thompson, 1956) to 60 000 years (Caswell, 1956), representing growth rates of 9 mm/year and 1.5 mm/year, respectively.

Gregory (1921) noted a variation in height of the raised coral reefs in different parts of the coast: 24 m + MSL at Mombasa, 39 m + MSL at Kilifi and 12 m + MSL at Malindi; these irregularities were attributed to differential tectonic movements. But there are several reasons that can also explain such differences; these include subaerial erosion and lack of distinction between coral reef and coral breccia deposits during height measurements.

In probably the earliest published work on raised beaches and coastal terraces, Sikes (1930) concluded that there was only one transgression indicated by the 30 m + MSL terrace and one regression represented by the deep cutting of the Kilindini channel to 60 m—MSL. In that study, the 9 m and 4.5 m + MSL terraces were attributed to oscillations during a general sea level fall. A more recent study by Abuodha (1998) explained that the raised beaches at 4.5 m + MSL (usually associated with aeolian deposition) mark the culmination of the Climatic Optimum at about 7000 BP. Geomorphological features along the Kenyan coast seem to sustain argument for global trends which show that the culmination of the last glaciation coincided with a drastic lowering of sea level to −120 m; there was still stand at the present position of the reef platform 30 000 years ago. Thereafter, the postglacial marine transgression prevailed until about 7000 BP, when sea level rose to its present position. The Holocene events are well represented by raised beach ridge-dune complexes that show eastward progradation of the coastline.

Even after measuring heights of marine terraces and grouping the levels statistically into eight categories, Ase (1978, 1981) still found the assemblage of shore features difficult to depict in a chronological sequence. These results are complicated further by probable reoccupation of previous shorelines and terrestrial processes. Ase's data indicate a higher shore displacement northward to Malindi, where emergent shore features such as dunes and beach ridges are dominant, in contrast to the southern sector where creeks represent a shoreline of submergence.

3.4. Eustatic versus isostatic and tectonic factors

Several studies consider the presence of raised terraces to be of pure eustatic origin (Caswell, 1953; Thompson, 1956; Williams, 1962; Read, 1981; Braithwaite, 1984). Sikes (1930) suggested that the tropical regions were free from isostatic loading due to the weight of ice sheets and the lag-cause influence; sea level changes were therefore paramount to independent crustal movements. The influence of tectonics on the genesis and later development of the marine terraces was first advocated by Gregory (1921) and later supported by Battistini (1969), Temple (1971), Ase (1981) and Oosterom (1988).

Indications that tectonic movements were important in the development of the landscape include the following:

(1) The elevation of the Foot Plateau (higher terraces) compared to the lower terraces.
(2) The presence of parallel step faults on all but the Shelly Beach terrace.
(3) The absence of a true fore reef facies in the raised coral reefs.

Along the Kenyan coastline, the lowest shoreline of the last major Pleistocene event can be reconstructed at around 60 m—MSL (drowned river channel at Kilindini and submarine terraces) compared to the generally accepted decline to −120 m—MSL. Therefore, the coastline experienced an uplift of approximately 60 m, which means a rate of about 3 mm/year. This is almost twice as high as the elastic response of 20–30% (uplift expected from the 120 m rise) of the ocean floor caused by the postulated inflow of melt water calculated by Walcott's (1972) model.

It is therefore possible that isostatic and tectonic crustal movements, based on the following five mechanisms, can explain half of the uplift. First, the growth of coral reef to a thickness of more than 100 m could generate isostatic loading and a rise of the coastal area (Hori, 1970). Secondly, intense denudation of the hinterland could result in continental emergence relative to the sea. Thirdly, the thick fluvial deposits could cause differential downwarp of the coast; Saggerson and Baker (1965) used this mechanism to describe the downflexing of the Tana basin along Kenya's north coast. Fourthly, the doming and rifting associated with the Great Rift Valley could lead to down warp of the coastal zone (Oosterom, 1988). Fifthly, continental uplift with its hinge point east of the present-day shore (at the

Table 3
Chronology of the Pleistocene and Holocene events in the southern coast of Kenya based on earlier studies

Period	General Pleistocene terminology	East African pluvial sequence	Alpine glacial sequence of Europe	Sea level changes	Major coastal events and morphological developments
Holocene	Post-glacial	Post-pluvial	Post-glacial	Rise to present level	Silting up of Port Reitz and Port Tudor mangrove swamps; sand dune formation
Upper Pleistocene	Last glacial	Gambian pluvial	Wurm glaciation	−7.5 m	Cutting of −7.5 m terraces. Accumulation of Cocquinas (aeolian sands) and raised alluvial deposits
	Last interglacial	Third interpluvial	Last interglacial (Eemain)	4.5 m 9.0 m	Formation of 4.5 m raised beaches, caves and platform; temporary rise of sea level; cutting of 9 m marine platform
Middle Pleistocene	Penultimate glacial	Kanjeran pluvial	Riss glaciation	−45 m	Sea level regression and emergence, Sub aerial erosion; deposition of Aeolian sands
	Penultimate interglacial	Second interpluvial	Great interglacial	30 m	Growth of coral reefs and formation of 30 m terrace
Lower Pleistocene	Antepenultimate glacial	Kamasian pluvial	Mindel glaciation	−60 m	Marine regression to −60 m; cutting of marine platform on which the coral grew; cutting of Mwachi River deep channel and Kilindini; deposition of Kilindini Sands, North Mombasa Crags and aeolian sands
	Antepenultimate glacial	First interpluvial	First interglacial	60 m	Accumulation of Magarini Sands
	Early glacial	Kageran pluvial	Gunz glaciation	?	–
Pliocene	–	–	–	90 m	Deposition of Marafa Beds

Ruvu-Mobasa Fault) probably also played a role in the geomorphological development of the coast.

Kent (1974) noted that minor crustal movements are supposed to have taken place from the Upper Pliocene to early Pleistocene, but Oosterom (1988) suggested that such morphogenesis of the coast probably continued throughout the Pleistocene period.

Alexander (1968) and Read (1981) have advanced the view that the Kenyan coast was stable during the late Pleistocene. However, their interpretation was based on doubtful results showing constant elevation and continuity of terraces and the absence of either faulting or tilting. The reason why the levels of shoremarks show an increasing altitude towards the north (Ase, 1981) is most likely a manifestation of erosional and depositional processes.

The patterns of faults indicate that tectonic movements occurred up to almost the youngest episode in the development of the shore zone. The age of lower shore elements, e.g. the Shelly Beach terrace, has variously been placed at late Pleistocene (Oosterom, 1988) and Holocene (Ase, 1981). Although $^{230}Th/^{234}U$ datings support a late Pleistocene age, the metachronous features in this zone point to a re occupation of a former shoreline during the Holocene. This can be attributed to the maximum Holocene submergence (Sikes, 1930; Walcott, 1972; De Graaff, 1989) connected with the Flandrian transgression at 4000–5000 BP. Ase's (1981) analysis of the levels of medieval buildings and beach ridges seems to suggest a slight sea level drop of about $\frac{1}{3}$m. In this review, the present reef platform is considered to have been cut 30 000 years ago, when sea level paused in its present position; landward of the reef platform are cliffs which are a prominent feature of the Mombasa shoreline and these probably indicate tectonic influence.

3.5. Drawbacks in correlation

The problems of correlation of the terraces and other sea level indicators have engendered many uncertainties in attempts to reconstruct the Pleistocene and Holocene development of coastlines. The main problem is whether they are purely eustatic, isostatic, tectonic or a combination of these. The rate and timing of these mechanisms are still debatable.

In addition, very few absolute age datings are available. Furthermore, radiometric dating is subject to errors as it is difficult to obtain representative samples from such diagenetically active coastal environments; assessment of sea level changes based on this method is usually inconclusive (Carter, 1991). For instance, Oosterom (1988) observed that the use of radiocarbon dating of the Kenyan coast gave indications of very high uplift rates compared to tectonically active regions—with no other field evidence to support this occurrence.

The problem of re-occupation of former shorelines and implications of tidal range on the effect of wave attack, as well as geoidal changes have also been cited as factors that affect correlations (Ase, 1978; De Graaff, 1989).

Nonetheless, when the levels of the various terraces are viewed against their supposed age as shown in Table 2, the validity of the chronological correlations can be checked.

4. Conclusions

Literature on chronological correlations of marine terraces is scanty and it is therefore envisaged that this review will instigate more enquiry into factors that were involved in landscape formation along the Kenyan coast. Existing correlations in literature of terrace heights are confusing, and are made more complicated by frequently omitting the reference datum. This problem is further compounded by the inconsistent nomenclature used for marine terraces and lack of a systematic dating of the terrace sequence.

Nevertheless, in testing the validity of previous correlations, a fair reconstruction of the Pleistocene events is possible. First, it is only the four lower levels that are considered to have formed during the Pleistocene. Secondly, some of the deposits are contemporary with, or related to, terraces. Thirdly, when the elevation of terrace levels is viewed against records of climatic and related eustatic changes, a more vivid interpretation of the order of events emerges. This suggests that global sea level fluctuations were an important control over coastal development even in tropical areas. The presence of raised coral reef is mainly attributed to isostatic loading of the continental shelf during marine transgressions caused by Polar ice melting. The growth of thick coral reef systems and fluvial sedimentation could generate isostatic loading and a rise of the coastal area. Theories of glacio-eustatic control on coral reef growth must be applied with caution since as shown in the Malindi quarry section discussed above, reefs could die due to silting without necessarily involving a retreat of sea level.

Whereas a number of authors believe that stable tectonic conditions have prevailed since Early Pleistocene times, there is evidence that tectonic movements occurred up to the youngest episode in the development of this coast. It is suggested here that neotectonic movements involving continental uplift and downwarp of the coast were involved in the geomorphological development of the study area.

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