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CLIMATIC CHANGE AND GEOMORPHOLOGY IN SOUTH AND SOUTH-EAST ASIA

Variations climatiques et géomorphologiques en Asie du Sud et du Sud-Est

H.Th. VERSTAPPEN*

RÉSUMÉ

Il est généralement admis, principalement sur base de témoignages d'ordre biologique et pédologique qu'un climat tropical ou subtropical humide a prévalu pendant une grande partie du Néogène. Des surfaces d'aplanissement comparables à celles de l'Afrique se sont effectivement formées dans l'Asie du S et du SE mais avec une extension limitée, leur altitude étant fortement influencée par la tectonique. Il n'est nullement prouvé que les conditions climatiques et les formes de terrain qui en dérivent aient été fort différentes de celles qui ont prédominé dans les autres régions tropicales.

Le Sud et le Sud-est asiatiques se trouvent dans une situation particulière quant à la répercussion de climats plus secs et plus froids sur la morphogenèse au cours du Quaternaire. En effet, les variations climatiques dans le domaine des Moussons concernent non seulement la température et les précipitations mais également la direction des vents. Des plates-formes continentales anormalement larges ont joué un rôle important lors de l'abaissement du niveau de la mer pendant les périodes glaciaires et ont ainsi renforcé le caractère particulier de la région.

Les variations de température ont plus particulièrement affecté l'évolution géomorphologique des parties montagneuses telles que l'Himalaya, la Chaîne Centrale de la Nouvelle-Guinée, certains volcans isolés. L'abaissement des neiges permanentes et des différentes limites de végétation s'est produit en même temps que la glaciation des zones les plus élevées, la cryoclastie dans des parties plus basses ainsi que la formation de dépôts grossiers et anguleux dans certaines zones du piémont.

Les variations de précipitation ont eu un effet marqué sur la morphogenèse, principalement dans les parties basses où des surfaces de piémont se sont développées sous l'effet du ruissellement en lame des phases sèches, alternant avec les processus linéaires des phases humides. Des séquences sédimentaires, des champs de dunes et des profils pédologiques témoignent de ces périodes sèches antérieures.

Les fluctuations du niveau de la mer ont eu un effet prononcé sur la morphologie côtière, notamment au cours de l'Holocène et même dans des périodes très récentes. Au cours du Pléistocène, l'évolution des côtes a été liée, dans une grande mesure, à l'émergence des parties peu profondes de la plate-forme continentale pendant les périodes glaciaires. Les récifs coralliens de ces parties ont dépéri tandis que ceux des parties plus profondes continuaient à se développer. Les récifs soulevés

* International Institute for Aerospace Survey and Earth Sciences (ITC). P.O.B. 6 - 7500AA Enschede, The Netherlands.

ont gardé les traces de ces changements du niveau de la mer. Le rétrécissement de la zone récifale a probablement constitué un facteur de moindre importance.

Les fluctuations dans les directions de vent ont également eu un effet prononcé sur le développement des côtes et des îlots coralliens dans l'Asie des Moussons, plus spécialement au cours de l'Holocène et des périodes récentes.

Les effets de ces quatre genres de changements climatiques sont illustrés par des exemples dans les différentes régions de l'Asie du Sud et du Sud-Est. L'homme y est devenu un agent géomorphologique depuis quelques millénaires. Son action est susceptible, à l'avenir, de provoquer à son tour des changements climatiques.

ABSTRACT

It is generally assumed, mainly on the basis of biological evidence and palaeosols, that humid tropical and sub-tropical conditions prevailed in large parts of the globe during the Neogene. Tropical planation surfaces, comparable to those in Africa do occur in South and South-East Asia, though of more limited extent and their altitude varying with tectonism. There is no evidence that the climatic situation and the resulting effects on landform development in the area deviated substantially from those then prevailing in other parts of the tropics.

The South and South-East Asian region represents a rather unique situation, however, as far as the climatic effects on landform development during the generally somewhat drier and cooler Quaternary are concerned. Large tracts being affected by the monsoonal wind systems the substantial climatic changes then occurring were not only characterized by fluctuations in temperature and precipitation, but also in wind direction. The occurrence of shelf areas more extensive than in most other parts of the world is another characteristic that has played an important role during the low sea levels of the Pleistocene Glacials and thus contributed to the area's uniqueness.

Fluctuations in temperature have affected the geomorphological development particulary in the highest, mountainous, parts of the region, such as the Himalayas, the Central Range of New-Guinea, some isolated volcanoes and other high peaks. The lowering of the snow line and vegetation limits resulted in glaciation of the highest parts; frost shattering in somewhat lower areas and debris formation in some piedmont zones.

Fluctuations in precipitation had a marked effect on landform development especially in the low parts of the region where foot slopes developed under the effect of laminar processes in the drier periods that alternated with linear processes during the more humid phases. Many sediment sequences, dune fields and soil profiles still bear witness of former drier periods.

Fluctuations of sea level governed the coastal evolution to an important degree during the Pleistocene when the extensive shallow shelf seas emerged during the glacial periods. Coral reefs on the shelf died off while those in the deeper seas developed without interruptions. Long records on Pleistocene sea level changes are on record in raised coral reefs. Narrowing of the coral belt was probably a factor of minor importance only.

Fluctuations of wind patterns also had a pronounced effect on coastal development and coral cays in Monsoon Asia, especially during the Holocene and even in recent decades.

These four aspects of climatic change are illustrated by examples from various parts of the region. Man became a geomorphological agent in S and SE Asia a few thousand years ago. His global activities may become a driving force of climatic change in the future.

There is ample evidence that climate has changed substantially throughout the geological history of the earth due to gradual cooling of the earth, extratellural factors, atmospheric changes, and some other in part still unknown causes. When around 2000.10^6 B.P. oxygene was yet absent in the atmosphere reduction phenomena instead of the current oxydation processes occured. The latter only started when photosynthesis of algae had produced atmospheric oxygene. The Jurassic period (150.10^6 B.P.) was a particularly hot period while around 60.10^6 B.P. during the Tertiary gradual cooling set in although the temperatures of large tracts of the globe were still humid tropical during the Neogene.

Our knowledge of Quaternary climates has grown rapidly in recent years through studies on Pleistocene air circulation (FLOHN, 1952), ocean floor drilling (EMILIANI, 1955; CONOLLY, 1967), ice core data (BARNOLA *et al.*, 1987; GENTHON *et al.*, 1987), and last not least the increased age-range of absolute dating techniques. The Holocene record is continuously refined among others by the growing number of pollen analyses in conjunction with ^{14}C dating and in some areas, such as SE and E Asia, the study of ancient gazetteers (CHU, 1926; MURTON, 1984). Attempts at giving an overview in an earth science context are from FAIRBRIDGE (1970) and PETIT-MAIRE (1991).

Apart from - or rather superposed on - these temporal climatic changes there have always been spatial variations due to differences in latitude, altitude, exposure/aspect etc. Changing patterns of air circulation and ocean currents have, at times, resulted in increased precipitation in one area and, simultaneously, decreased precipitation elsewhere. There is a danger in undue extrapolation of palaeoclimatic data to other areas and in generalizing palaeoclimatic conditions. Part of the numerous discrepancies in dating cold/warm and wet/dry variations are not due to contamination of samples, errors in site evaluation etc., but reflect real spatial differences and thus regional individuality.

Megageomorphological phenomena such as the global distribution of continents and the position of high altitude relief zones affect the air circulation and thus the distributional patterns of climate. The coastal configuration and the submarine relief of the continents complement the wind systems in governing the ocean currents that play an important part in global heat transfer and climate. The El Niño-Southern Oscilation (ENSO) phenomenon examplifies this. The climatic conditions in their turn play an important part in the type and intensity of the geomorphological processes operating in various parts of the world and thus affect the geomorphological terrain configuration, even in mega scales when with time large planation surfaces are formed. This interaction between geomorphology and climate varies with time when either of the two changes.

Among the earth scientific factors to be considered in this context the changing distributional pattern of continents and oceans under the influence of plate tectonics ranks high. A continent changing in latitudinal position thus may travel from one climatic zone to another and the effects of this in terms of changing weathering processes then is recorded in the geomorphology and the soil development. An example is the northward drifting of India that passed across the warm humid climate of the equatorial belt and subsequently entered the warm seasonal climate of the monsoonal zone as is demonstrated by Figure 1 (DERCOURT, 1991). A review of the plate tectonic situation in SE Asia is from KATILI (1989).

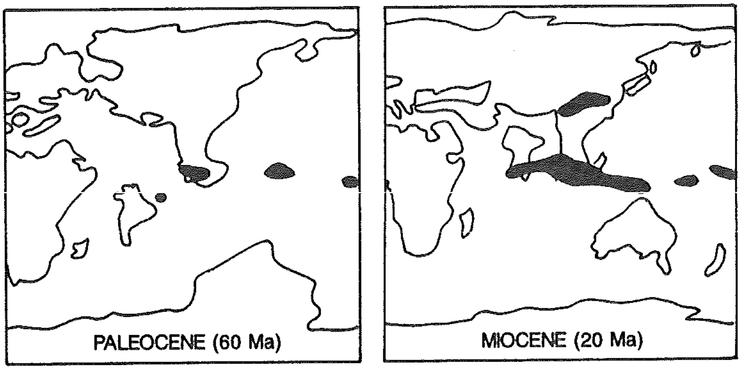


Fig. 1 : Palaeoclimates and plate tectonics. Black : high precipitation zones. India moved across the humid tropical belt into monsoonal conditions (DERCOURT, 1991).

The dominance of continental landmasses in the northern hemisphere and of oceanic areas in the southern hemisphere has become increasingly pronounced since the early Tertiary. This has a marked affect on air circulation (and sea currents !) and notably on climatic seasonality. The Intertropical Convergence Zone (ITCZ) is displaced southward during the northern winter due to the strong anticyclones developed where large continental land masses occur in the north. The ITCZ there sweeps back northward again during the northern summer when the anticyclones are weak. In contrast the ITCZ remains continously near the equator where oceans are present in both hemispheres as is evident from Figure 2 (VERSTAPPEN, 1975). The monsoonal belt of seasonal climates that thus is formed over and near the continents is best developed in South, South-East and East Asia. The asymetry of the hemispheres has caused a southward sweep of the northern hemisphere climatic belts during Quaternary cold periods. As a consequence these belts were then somewhat compressed in the southern hemisphere (VERSTAPPEN, 1989). The southward sweep of the ITCZ is cushioned to some extent in the Moluccas and nearby areas as a consequence of the nearness of the Australian continent. This is even reflected in minor climatic fluctuations of the last few decades.

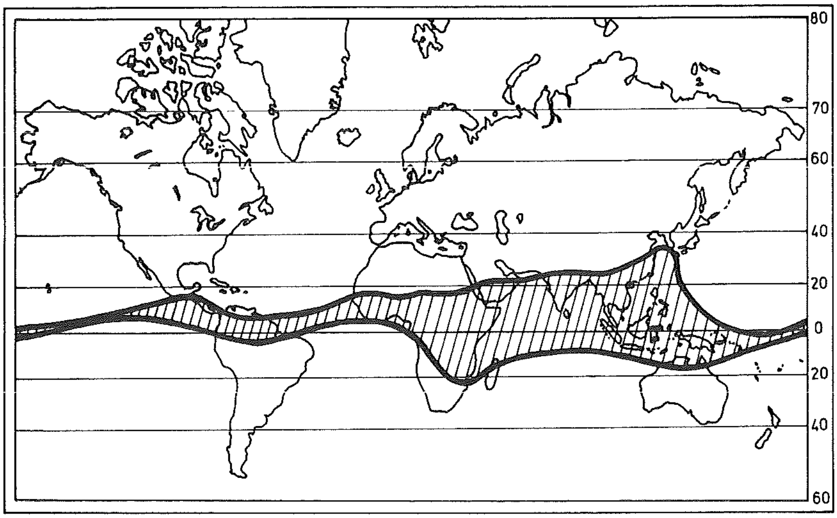


Fig. 2 : The effect of global distribution of continental masses on the annual shift of the Intertropical Convergence Zone (ITCZ) and thus on the width of the monsoonal zone (hachures).

Apart from changes in the spatial distribution of continental land masses also changes of their vertical dimensions affect the spatial distribution of climates. An effect of plate tectonics is the rise of high mountain ranges (Himalayas; Central Range of New Guinea) and highland areas (Tibet) where continental plates collide and of island arcs with connected volcanism where subducting oceanic plates occurs. The tectogene zones so formed are in strong contrast to the cratogene plates with their characteristic ancient planation surfaces and extensive shelf seas (Sunda shelf and Sahul shelf). Relics of such surfaces in tectogene areas are rare, of limited extent and topographically high as for example in Sumatra (VERSTAPPEN, 1973). Erosion in tectogene areas is rapid notably where soft Tertiary rocks outcrop. Tectogene and cratogene landforms are quite distinct also where climatic geomorphology is concerned.

The effect of these mega-relief characteristics is particulary clear in the climatic and palaeoclimatic conditions in South and South-East Asia. During the change from (northern) winter to summer conditions the westerly upper-air (500 hPa) currents suddently change their path from South of the Tibet obstacle to North of it. This causes the so-called "burst" of the monsoon in India. The relief and radiation of the Tibetan landmass has also a great effect on the palaeoclimatic conditions during the Pleistocene Glacials (FLOHN, 1952). The Central Range of New Guinea likewise acts and acted as a topographic barrier with the connected effects on rainfall. Present and (much more extensive) Pleistocene glacial and periglacial phenomena are inherent to these high areas.

The extensive Sunda and Sahul shelf areas have had an important effect on the Pleistocene climates of the region and notably of SE Asia : less evaporation and thus greater dryness occurred when they ran dry during the low sea levels then prevailing. Also the climatic seasonality increased during low sea level phases because of increased continentality. During interglacial high sea levels ocean currents contributed to the exchange of heat and moisture and thus to reduction of regional climatic differences. The water of the South East Asian seas originates from the Pacific. Part of the westward flowing North Equatorial current bends southward forming the Mindanao current. During the South-east monsoon the inflow is strong through the Celebes Sea and the Flores sea and the outflow is through the South China Sea to the Strait of Taiwan and the Luzon Strait. This trend is reversed during the North-west monsoon. Some outflow to the Indian Ocean through Strait Malacca and Strait Sunda occurs throughout the year (APRILANI, 1992).

It is evident from these considerations that the earth scientific setting has an important effect on the present climatic situation and has had an equally important part in the past on the palaeoclimates. In their turn the palaeoclimates have given rise to certain geomorphological processes and thus have affected the landform development as will be elaborated upon in the next sections of this paper. Although our knowledge about the palaeoclimates and related landforms of Africa and South America is yet better than that of the situation in South and East Asia, also in this area a wealth of information has been gathered in recent years.

Ancient, Pre-Tertiary and Tertiary, tropical planation surfaces exist in many parts of the region such as Malaysia, India and Sri Lanka. Deep chemical weathering is

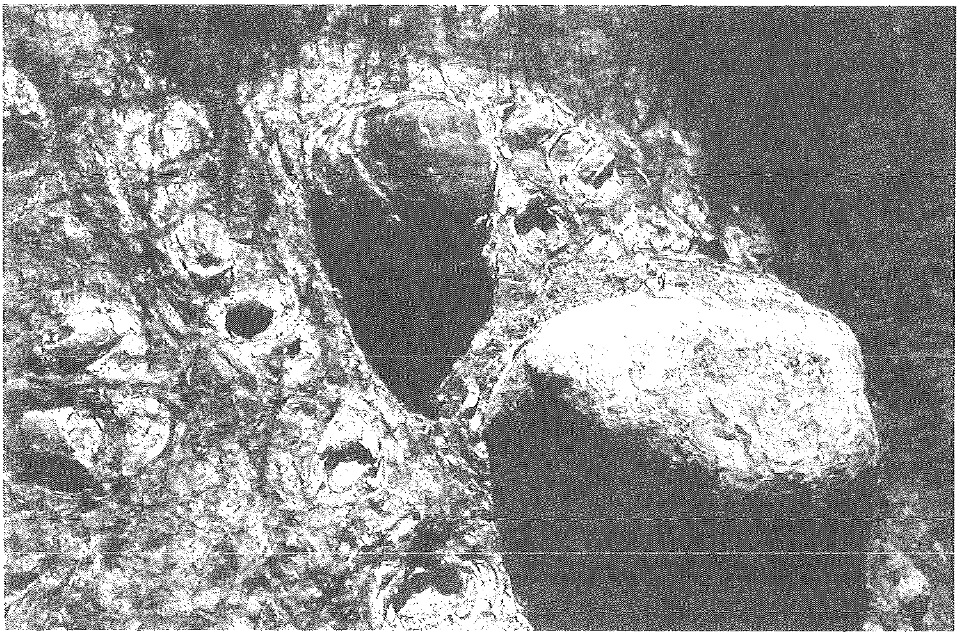


Fig. 3 : Deep humid tropical weathering in an ancient planation surface near Kandy, Sri Lanka.

Fig. 4 : Two semi-arid glacis foot slopes near Quetta, Pakistan. Beveled soft Tertiary rocks are covered by a thin veneer of angular debris.

characteristic (Fig. 3). Smaller relics of such tropical planation surfaces also occur in some tectogene zones. They are of a completely different nature than the more recent pediments that require semi-arid conditions for their formation. BREMER (1981) and SPÄTH (1981) who have studied the humid tropical planation surfaces extensively in Sri Lanka found geomorphological and pedological evidence for subcutaneous transport of colloids in the humid tropical parts of the island resulting in fairly coarse-grained and permeable residual soils. The actual weathering is restricted to the contact zone with the bed-rock. This process is an important element in the formation of the old and deeply weathered planation surfaces.

The drier conditions that marked the transition to the Quaternary have led to different geomorphological processes that resulted in the formation of foot plains like erosion glacis near Quetta in the dry parts of NW Pakistan (Fig. 4). Different processes in the less arid areas of SE Asia have led to the formation of another type of footplain (VERSTAPPEN, 1975) studied in detail in Malaysia by DE DAPPER (1987, 1989) and DEBAVEYE (1986) and in Indonesia by VAN DER LINDEN (1978) and KLOOSTERMAN (1989). BREMER and SPÄTH found evidence for drier Upper Pleistocene climates in some profiles in the so-called "dry zone" of NW Sri Lanka and in the lowlands of the island. The lateritic crusts occurring at places are considered by them as evidence for drier young Pleistocene climatic conditions. Diffuse surface wash is a dominant factor in the formation of pediments under such drier conditions although the earlier mentioned subcutaneous processes also persist.

The climatic changes can be grouped under four headings namely changes in temperature, in precipitation, in sea level and in wind patterns. Temperature changes were a dominant factor in the glaciation and the periglacial processes in the mountains while rainfall variations affected dune formation and planation in the low lands. Sea level changes have affected the coastal zone and the areas now covered by shallow shelf seas. Changes in wind patterns were/are an important factor in the evolution of low land coasts and coral cays during the Holocene and at present. The geomorphological aspects of these four aspects of climatic change are discused in the following sections.

QUATERNARY TEMPERATURE CHANGES AND HIGHLAND GEOMORPHOLOGY

In the Himalayas and in the Central Range of New Guinea the lowering of air temperatures of approximately 5-6° C during the Pleistocene glacial periods has resulted in a lowering of the snow line of approximately 1000 m and the forest line and the various altitudinal vegetation zones were lowered accordingly. As a result the glaciation expanded and glacier tongues reached down to at least to 3000 m above sea level at places. Simultaneously the periglacial zone stretching below the glaciated belt was also lower and much more extensive. The full cycle of repeated Glacial and Interglacial periods known from Europe and North America has, however, not (yet ?) been traced in these mountain ranges. Mostly only the occurrence of one (or two) glaciations have been established so far. This may be due to the paucity of data but it has been suggested by some investigators that the uplift of these mountains is so young that they only reached sufficient height for glaciation during the later part of the Pleistocene. Possibly also the extent of earlier glaciations was much smaller for this reason and have the younger glaciation(s) overridden and erased their traces.

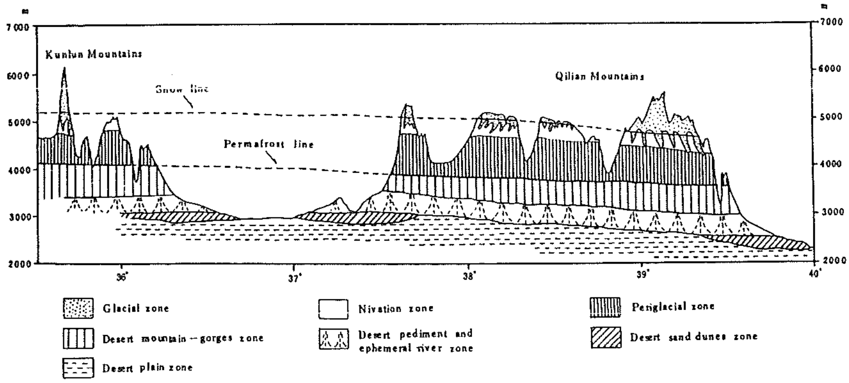


Fig. 5 : Climatic-geomorphic zones in NE. Tibet (HÖVERMANN & WANG, 1987). Large areas were covered by ice or affected by permafrost during the Pleistocene glacials.

Periglacial areas much more extensive than at present (Fig. 5) existed in Tibet and adjacent high land areas during the Pleistocene Glacials (HÖVERMANN & WANG, 1987). The extent of Pleistocene glaciation in these parts was long believed to be limited due to dryness. New investigations by KUHLE (1987), however, revealed that glaciers have covered much larger surface areas than nowadays and glacier tongues have been even found entering Central Asian dune fields. In north-eastern Tibet fluvioglacial deposits have been dated 35,000 -23,000 B.P. by HÖVERMANN and WANG which corresponds with the Würm glacial maximum. The fossil pollen spectra from the Early Holocene indicate a treeless environment not substantially different from today while the snowline from 9,500 - 6,400 B.P. also was comparable with the present situation. The rise in temperature provoking the retreat of the glaciers thus must have occured prior to 9,500 B.P. This corresponds well with the findings of SINGH and AGRAWAL (1976) who radiocarbon dated the deglaciation of the NW Himalayas at an altitude of 3120 m at 15,000 - 14,000 B.P. (See also : GELLERT, 1976, 1990, 1991).

Among the early investigators on Pleistocene glaciation in Asia DE TERRA and PATERSON (1939) and VON WISSMANN (1959) should be mentioned. The former authors already pointed out that the ice ages were associated with drier climates. Obviously the Pleistocene Glacials were in the mountains marked by strong physical weathering as a result of frost and scarcity of vegetation. When the rivers could subsequently tansport the coarse debris gravelly terraces and fans were formed e.g. in the Lesser Himalayas and at the south flank of the Central Range of New Guinea. The

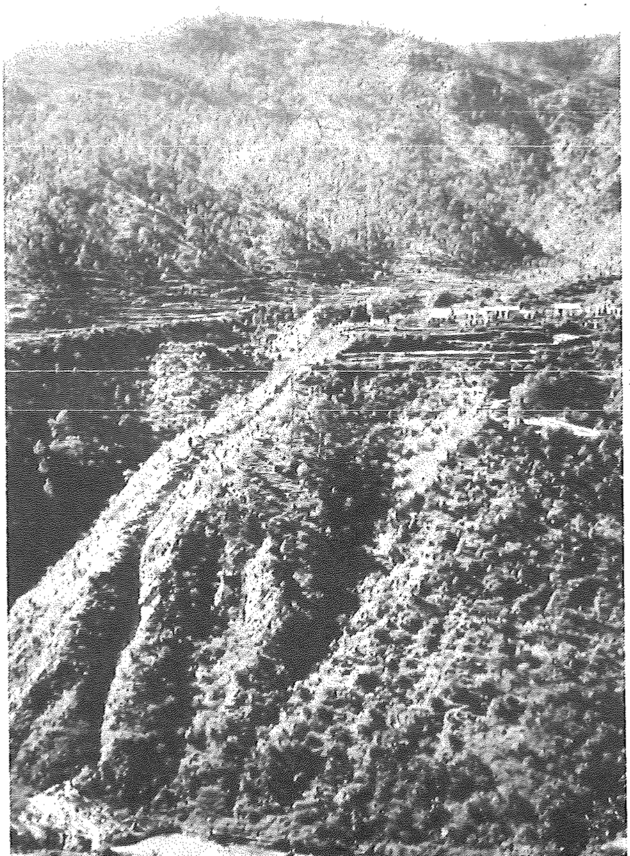


Fig. 6 : River terrace in the Kimaun Himalayas. The river gravel was deposited by high-discharge rivers during a humid phase. On top a thin layer of angular slope debris was formed during a subsequent drier period.

author could observe at several localities in the Kimaun Himalayas that these gravels are covered by angular periglacial material derived from nearby mountain slopes when aridity again invaded the area (Fig. 6).

In the island arc zone between the glaciated mountain ranges of the Himalayas and the Central Range of New Guinea former glaciation has only been reported from two isolated mountain peaks : Mt Kinabalu in Sabah (KOOPMAN & STAUFFER, 1968) and Mt Leuser in Sumatra (BEEK, 1982). The Kinabalu glaciation is limited to a few square kilometers above 12,500 ft. Two small glacier tongues have been reported. The longer one measures about 1 km and flows northward within the V-shaped top area. The other is a very small one flowing southward beneath the Panar Labah ice fall. The evidence from the much lower Mt Leuser (3381 m) in northern Sumatra is limited to some supposedly morainic material. The height of this mountain is just marginal for Pleistocene glaciation. The top of some strato-volcanoes in the area may have been covered by snow or ice but a proof of this can hardly be expected because of subsequent eruption activity. Evidence of former periglacial phenomena may come forward from some highland areas in the region.

The present and Pleistocene glaciation in the Central Range of New Guinea has first been studied by DOZY (1938) in the highest area, Mt Carstensz/Puncak Jaya (5030 m). An overview of the glaciation in the western part of the island, Irian Barat, is

Fig. 7 : Terminal moraines in the Star Mountains, Irian Barat bear witness of extensive glaciation around 17 000 BP when the snow line was about 1 000 m lower than present.

given by VERSTAPPEN (1952) who also studied the present and Pleistocene glaciation in the Star Mountains near the border with Papua/New Guinea (VERSTAPPEN, 1960, 1964). Several thousands of square kilometres have been glaciated in the range as is evidenced by the terminal moraines of Fig. 7 and the aerial photograph of Fig. 8. The present glaciation is limited to the top area of Mt Carstensz/Puncak, Jaya, Mt Jet Tégèl (formely Mt Idenburg, 3800 m) directly West of it and Mt Mandala (formerly known

Fig. 8 : Aerial photograph showing a terminal moraine below the present forest lime marking the extent of a ancient fjell-type glacier on a highland plateau (< 4000 m) West of the Star Mountains, Irian Barat.

as Mt Juliana, 3700 m) in the East. The persistence of the ice cap of Mt. Mandala is open to some doubt : its top area is close to the snow line. Aerial observations by mision pilots and others in the period 1989-1993 seem to indicate that a tiny ice cap still exists. The question is rather whether it is perennial. Mt Trikora (formerly known as Mt Wilhelmina, 3700 m) lost its tiny valley glaciers already around 1950, its steep slopes not favouring snow accumulation.

HOPE *et al.* (1976) give full details on the present state of research in the Carstensz/Jaya area. The beginning and the maximum of the last glacial phase is not known but it has been established that the most recent cold phase was 20,000 - 17,000 B.P. The snow line then was at about 3,550 - 3,650 m. Conditions cooler than present existed from at least 40,000 B.P. with some warmer interval around 30,000 - 26,000 B.P. Some gravel deposits found in sink holes in the Mt Jaya (Carstensz) area may be tilloid and then point to an earlier, more extensive glaciation.

Deglaciation started at 15,000 - 14,000 B.P. with two minor reversals at 12,500 and 11,000 B.P. Considerably drier and more continental climates prevailed in the cold period prior to 15,000 B.P. when dry SE trade winds blew over the southern plains then stretching up to the Aru islands. The two reversals mentioned above may be due to the flooding of the Sahul shelf as a consequence of the rising global sea level. More moisture and thus more snowfall in the high mountains then was introduced notably in the western part of New Guinea. No data exist on the extent of the ice sheet from 10,000 B.P. till an advance dated 3,000 B.P. which was followed by three other advances at 2,400, 1,800 - 1,600 and after 1,300 B.P. General retreat at Mt Jaya is on record for the last two centuries.

Further to the east the glaciation has been studied by VERSTAPPEN (1953, 1960, 1964), REINER (1960) and LÖFFLER (1977). HOPE (1976) and HOPE and PETERSON (1975) have also carried out new research about the glacial history of Mt Wilhelm and concluded that conditions warmer than present there prevailed from 8,500 - 5,000 B.P. while evidence for conditions cooler than present exists for the last 5,000 years with at least two phases of glaciation, the latter ending one or two centuries ago. It is noteworthy that the warmer period mentioned above corresponds with the millennia during which no evidence for glaciation has been found on Mt Jaya (Cartensz). HOPE and HOPE (1976), have elaborated on the palaeo-environments for man in New Guinea.

QUATERNARY PRECIPITATION CHANGES AND AEOLIAN LANDFORMS

Climatic changes in (semi-) arid areas are often reflected in differences in the dune types formed and/or in the formation of dunes in specific periods. The latter then usually indicate increased aridity, although it may happen that f.i. parabolic dunes are formed in humid periods when erosion of older dunes or other sandy deposits provides abundant material for aeolian activity during the dry season. The author (VERSTAPPEN, 1970) investigated dune development in the Thar Desert in NW India where large windward obstacle dunes occur at the SW side of isolated mountains rising from the plain (Fig. 9). These fossil dunes are strongly dissected; the sand is slightly cemented and soils have developed. They date from a period of much greater aridity than today, presumably the Early Holocene. Their subsequent dissection in a more humid period resulted in the formation of, much lower and smaller, parabolic dunes. The latter, are now also covered by scrub and no longer active, which indicates that the climate again deteriorated after formation. The Holocene climatic optimum was suggested for their age in view of the cultural context (Harappan civilization) and their relation to a so-called "Kankar" layer of calcium carbonate nodules (Fig. 10) (see also KAR, 1987).

MEDHI (1977, 1978) claims that the Late Pleistocene (30,000 - 10,000 B.P.) in Rajasthan was considerably (50 % ?) drier (following a much older humid tropical period with laterite formation). ALLCHIN and GOUDIE (1971) report the occurrence of microliths and a buried soil in a near-surface dune layer and since the Late Stone Age or Mesolithicum culture has been radiocarbon dated elsewhere in India at 11,000 B.P. this provided a minimum age for the dune formation. However, ABICHANDANI and ROY (1966) mention microliths from deeper in the dune body and confusion thus persisted. SINGH (1967, 1971) and SINGH *et al.* (1972, 1974) obtained ^{14}C dates and pollen from salt lake deposits in the dune areas of Rajasthan. These deposits are about 3,5 m thick; the lower part comprises of laminated clays (with pollen) while the upper meter consists of non-laminated silt (without pollen). The datings obtained range from 9,250 - 4,510 B.P. The deposits occur on top of thick (dune) sand deposits which thus are of early Holocene age. The following climatic phases have been distinguished (Fig. 11) :

Phase	I	Pre - 10,000 B.P.	- very arid
	II	10,000 - 9,500 B.P.	- much more humid than at present
	III	9,500 - 5,000 B.P.	- slightly more humid than at present
	IV	5,000 - 3,000 B.P.	- much more humid than at present
	V	3,000 - 1,800 B.P.	- very arid
	VI	early AD - present	- actual conditions

Fig. 9 : Windward obstacle dune near Jalor, Rajasthan, India, formed during the very dry conditions prevailing before 10 000 BP, and dissected in (a) more humid phases(s) but notably during the Harappan civilization 5 000 - 3 000 BP.

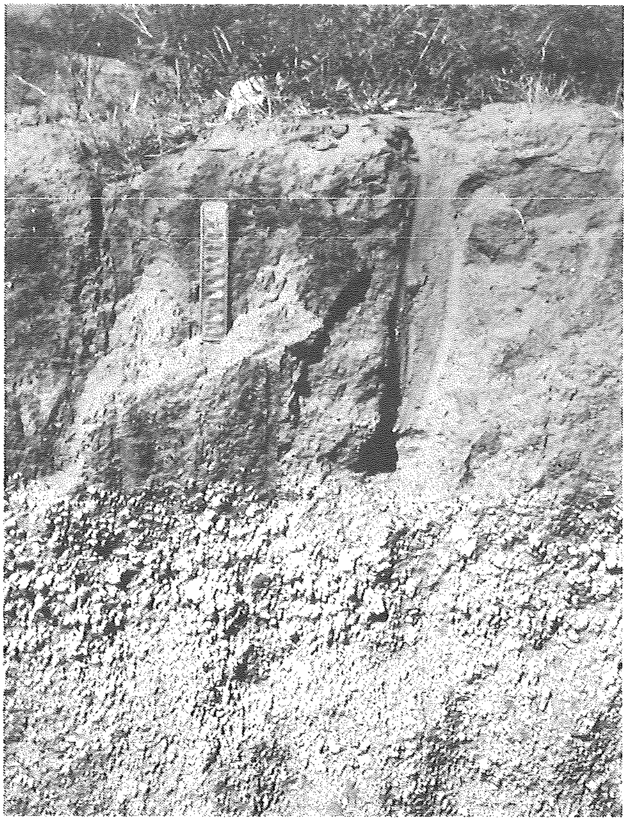


Fig. 10 : So-called "Kankar" layer of calcium carbonate nodules at shallow depth near Jalor, Rajasthan, India, related to the humid Harappan period.

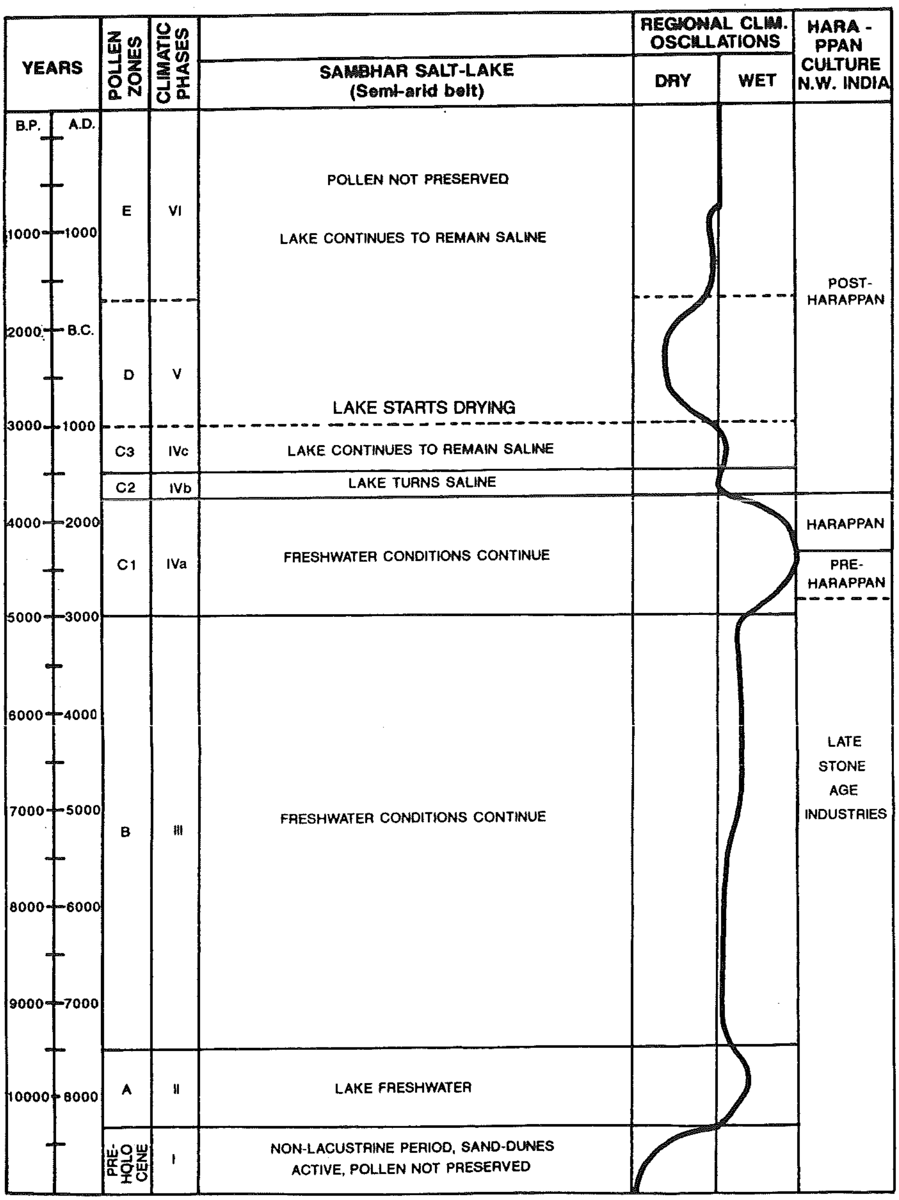


Fig. 11 : The palaeoclimates of Rajasthan based on data from Sambhar Lake (SINGH, 1974; text abbreviated).

The Harappan civilization benefitted from the humid conditions of period IV while during the later Painted Gray Ware period settlements were in the dry river beds using the dwindling water resources of period V. AGRAWAL (1975) shares this view. It also concords with the author's observations on coastal dune development in the Porali plain, Pakistan (VERSTAPPEN, 1983) where parabolic dunes preceded the barchans now being formed.The dune sequence of the Thar Desert thus supports the view of aridity being connected with the cool Glacial/Early Holocene in this part of the world.

QUATERNARY PRECIPITATION CHANGES AND PLANATION PROCESSES

The lower precipitation and notably the more pronounced dry season during the Pleistocene glacials provoked drought stress in the vegetation especially in the lowlands and caused changes in geomorphological processes. The chemical weathering was much less and physical disintegration of rocks became more important. The soil formation was affected accordingly as was already mentioned by MOHR (1922). He described black clayey soils from E Java underlying the Holocene alluvial soils and originating from weathering in an alkaline environment in areas with a particulary dry monsoon. Near Bogor (Java) deep red latosols and, in places, laterites underly considerably younger brown latosols. The former point to a more rapid decomposition of organic matter and a marked dry season where as the latter correspond better with the present more humid conditions.

The drier climate and increased seasonality during the Pleistocene Glacials also had a pronounced effect on the river regime. The monsoon forest and tree savanna conditions prevailing in the low relief areas favoured lateral riverwork and diffuse run off rather than the linear incision characterizing the humid tropical interglacial conditions. Debris formed by alteration and disintegration was transported during high discharges in the wet monsoon and with the advent of the next dry season was in part temporarily deposited in the river beds. Large tracts of the river valleys were thus characteristized by a coarse valley fill and braiding rather than meandering was the rule. At present we find this type of river regime only in SE Indonesia where it is referred to as "Timor River" after the island were it is best developed (VERSTAPPEN, 1955). Recently THOMAS c.s. (1987) have reported ancient features of similar nature from extensive tracts in western Kalimantan (Indonesian Borneo). They are radiocarbon dated as more than 40,000 B.P. and it is considered that deposition occurred whilst sea levels were low and conditions of severe seasonality prevailed. A comparison of the rainfall figures of Timor with those of West Kalimantan suggests an approximate reduction in precipitation of about 50 %. This is more than the author (VERSTAPPEN, 1974, 1980) estimated initially and in line with opinions that have since been expressed by other investigators (VON KOENIGSWALD, 1976). The investigations by BRUNNER (1975) on Pleistocene climates in Southern India also are of interest in this context.

The humid tropical climate characterizing the Interglacials and the period following the Postglacial rise in sea level resulted in a quite different environmental

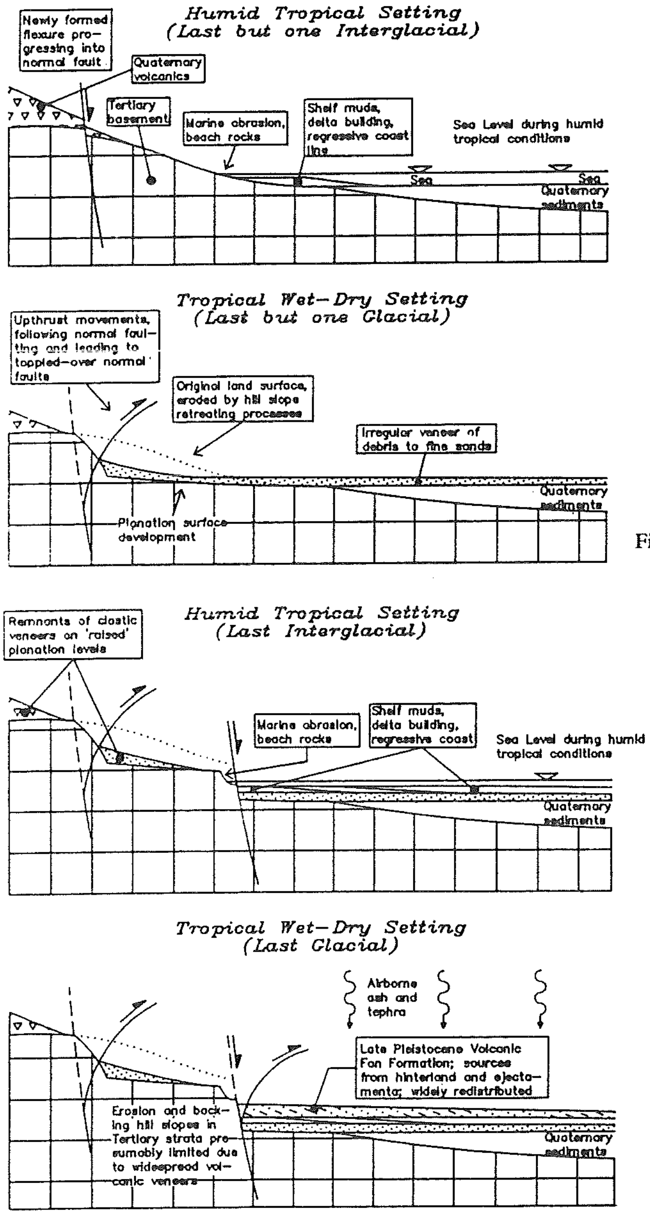


Fig. 12 : The effect of climatic fluctuations during the two last interglacial-glacial oscillations on the formation of the coastal lowlands of the north coast of west/central Java (KLOOSTERMAN, 1989).

situation with dense vegetation, linear river work and intense chemical weathering. Clays carried off in suspension by the rivers formed clay blankets off-shore on the bottom of the shelf seas. Their existence has been proved by borings and acoustic soundings (ALEVA *et al.*, 1972). KLOOSTERMAN (1989), studying the groundwater situation in the lowlands along the north coast of Western/Central Java, pointed out the role of the clay blankets in creating artesian conditions in the northward dipping layers by sealing the coarser textured aquiferous layers.

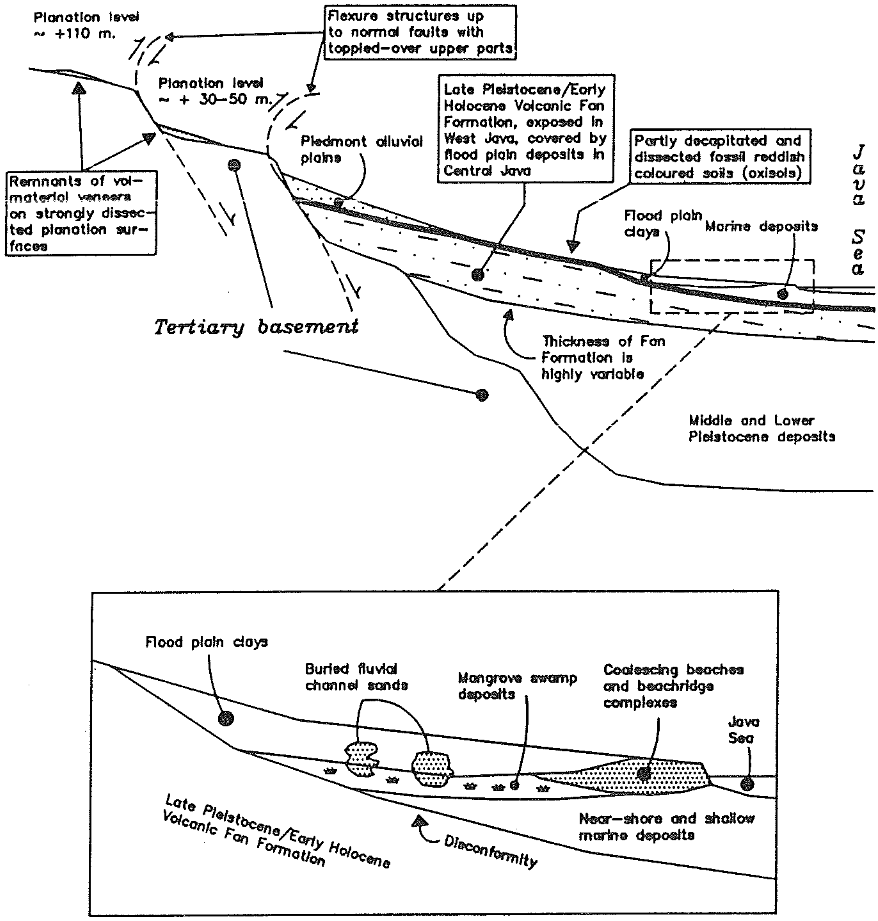


Fig. 13 : The Holocene development of the coastal lowlands along the north coast of west/central Java following on those given in figure 12 (KLOOSTERMAN, 1989).

The climatic conditions prevailing during the Pleistocene Glacials provoked the formation of planation surfaces at several localities. They are well known from S. Sumatra where the extensive so-called "peneplain" of Palembang occurs; similar landforms also exist in N. Sumatra. Evidently these features can not have been formed under humid tropical conditions with rivers confined to their beds by dense tropical vegetation and linearly incising. Coarse deposits mentioned by MOHR (1919) from the bottom of the Java sea and the "Younger planation Surface" capping the Pleistocene "Alluvial complex" of the Sunda shelf described by ALEVA *et al.* (1973) are submarine equivalents that will be discussed in the next section.

The occurrence of planation surfaces in Central Java was first reported by VERSTAPPEN (1974, 1980) and they were subsequently studied there in more detail by VAN DER LINDEN (1978) in a nearby area in the Serayu Valley. VAN DER LINDEN observed two phases of instability, the first one being a strongly erosive phase with coarse-textured deposits and the second one being less effective and resulting in residual enrichment of ferric nodules and eventually even in the formation of a stone-line. A pre-existing plinthitic Oxisol has been truncated by the processes of these instable periods. From the middle Pleistocene onward strong volcanic activity occurred in the area and Andosols were formed during a stable phase of geomorphological development. They have been destroyed and the material was transported in a later phase of less protective vegetation cover.

The most important recent contribution to the impact of Quaternary climatic changes on landform development in Java is from KLOOSTERMAN (1989). The results of his work are summarized in the Fig. 12 and 13 that relate to the Pleistocene and the Holocene development respectively. KLOOSTERMAN mentions the occurrence of extensive planantion surfaces to the landward of the alluvial plains. They are found where Lower- to Middle Pleistocene volcanic rocks are not covered by younger volcanics such as near the village of Weliri about 40 km W of Semarang. The planation surfaces belong to one gently northward sloping planation level of 110 - 200 m. Three regional levels, separated by escarpments, can in fact be distinguished, at 30 - 50 m, 100 - 200 m and 400 - <500 m respectively. The scarps are tectonic in origin and the planation thus must be older. Where young fluvio-volcanic debris is found on the planation surface the low relief and subsequent dissection of the latter is obvious. Also further to the West, near the town of Subang these planation surfaces are present. These surfaces originate obviously not from linear river work but from planation processes under drier conditions than present. An important argument in favour of the reality of these surfaces is the occurrence of volcanic pebbles and boulders and of volcanogenic rudaceous deposits within small catments that are fully situated in Tertiary clayey formations.

KLOOSTERMAN found another important indication for climatic changes in the Lower Pliocene Tapak reef limestones about 25 km SW of the town of Tegal. Exposed palaeo-karst shows terra rossa-like clayey soil remnants formed under hot humid tropical conditions before the tilting and folding of the Tapak Formation, presumably during the end of the Pliocene. It is overlain by the light-coloured regolith with many

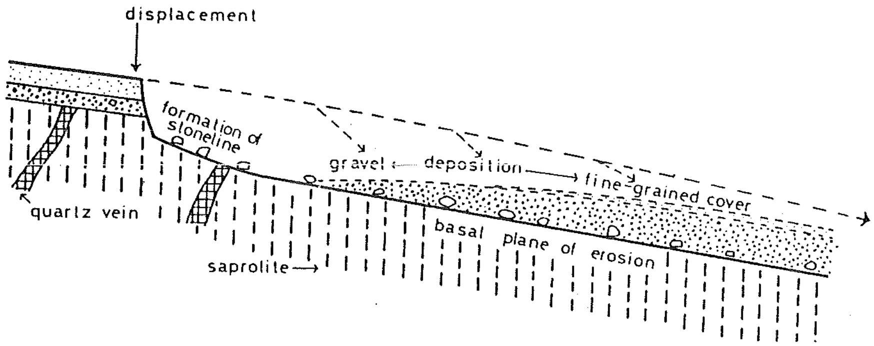


Fig. 14 : Model of formation of stone lines and of coarse and fine debris during pediment formation (DEBAVEYE & DE DAPPER, 1987).

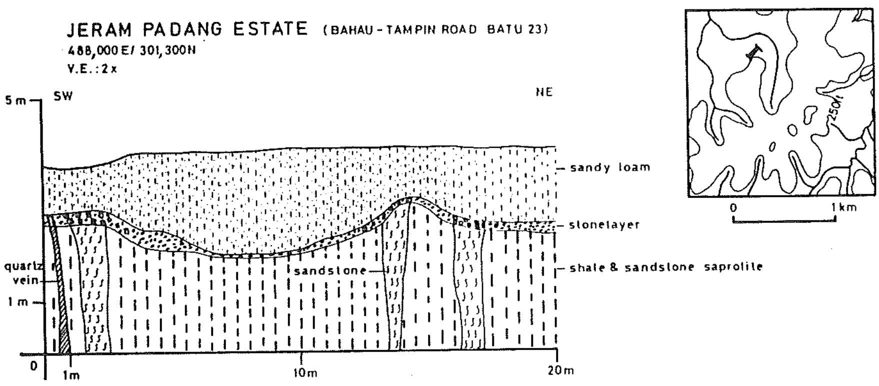


Fig. 15 : Longitudinal section across the lower/middle part of a pediment in Jeram Padang Estate, Kuala Pila, Malaysia (DE DAPPER, 1987, 1989).

angular limestone fragments that originated as talus during the Pleistocene when rock disintegration was dominant under a tropical-seasonal or more likely semi-arid climate. Tectonics continued during the formation/deposition of the mantle rock.

The sequence of events can be summarized as follows. At the turn of the Pliocene/Pleistocene the humid tropical climate gradually began to change to cooler and drier conditions. At the same time sea level lowering began. As volcanic activity during the Lower Pleistocene was fairly low, the influx of coarse materials remained limited and the sediments thus may have been predominantly fine-grained. The active tectonism and volcanism of the Middle Pleistocene resulted in massive sedimentation in the basins. Large fluvio-volcanic fans were formed that stretched far northward during the dry glacial periods of low sea level. Remnants of the planation levels should be searched for under the Notopuro and Linggopodo beds. Newer planation surfaces were formed during the glacial periods of the Upper Pleistocene. The large fluvio-volcanic fan underneath the coastal lowlands of West Java was formed during the last glacial period. The Late Pleistocene/Early Holocene fan formation is much more extensive than formerly assumed. The sedimentary environment is fluviatile and eolian, suggesting drier savannah-like conditions with strongly reduced vegetative cover allowing for planation by sheet erosion and wind removal of fines. The fans formed a land surface during the last Glacial period and since were partly covered by Holocene sediments with rising sea level. The position of the marine deposits points to a sea level at 4,500 B.P. that was 4 - 6 m higher than present.

Quaternary deposits known as "Old Alluvium" known from peninsular Malaysia and Serawak have long been interpreted as marine formations and a staircase of Quaternary higher sea levels (70, 30-15 and 5-2 m) thus has been assumed. Only fairly recently a fluvial origin has been established by HAILE (1969, 1971), SIVAM (1968) and NEWELL (1971). Postulating higher sea levels is no longer necessary if the concept of wash deposits and colluvials formed independently of sea level under (tree) savanna conditions is accepted. The rock-cut terrace described by WALKER from the Kinta Valley and the foot plains mentioned by NOSSIN (1964) fit well in this concept.

Investigations of particular interest have been carried out in peninsular Malaysia by DE DAPPER (1986, 1989), DEBAVEYE *et al.* (1986) and by DE DAPPER *et al.* (1988). They describe fairly extensive pediplanes dissected to low hills and surrounding the hihland cores. The mechanism of the pediment formation by scarp retreat, the formation of stone-lines and the separation of fine and coarse material as seen by these authors is given in Fig. 14, an example of the longitudinal profiles so developed is shown in Fig. 15.

The chronosequence of landscape development comprises of :

1. Highland cores (remnants of the oldest surface)
2. low hills, considered as remnants of an Older Peneplain/Pediplain (R.O.P.) at the foot of the Highland cores
3. a complex surface comprising a rock-river terrace (T2) and younger pediments (P2) developed at the foot of 1 and 2 and grading into T2
4. younger T1 terraces, cut-and-filled or covering T2 and grading into the coastal plain where appropriate
5. the actual river system (To) rock-cut/cut-and-filled in T1.

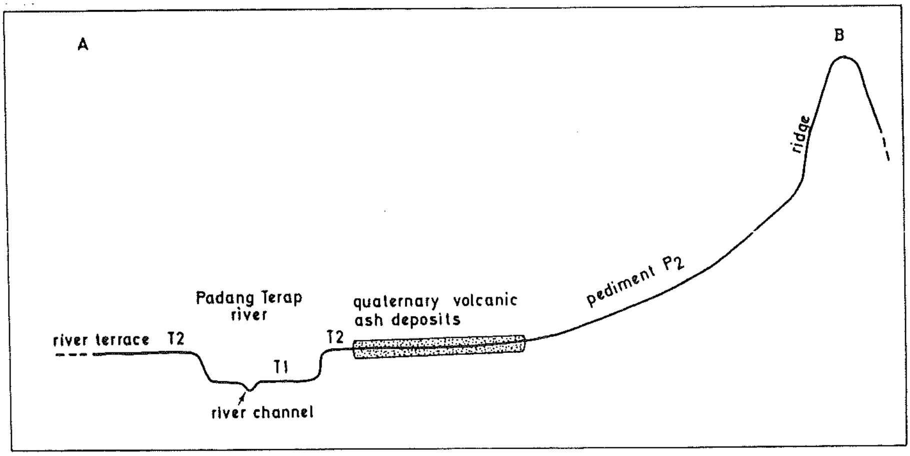


Fig. 16 : Geomorphological position of Lake Toba ashes on the T2 terrace, linked with pediment level P2, in the Padang Terap area, Malaysia (DEBAVEYE *et al.*, 1986).

An overview in tabular form of the landscape development is given in Fig. 16. The planation surfaces are related to the drier and cooler periods of the (late) Pleistocene Glacials : The superficial layers of P2 testify of extensive pedimentation under fairly open vegetation. In granite terrains most of the T2 sediments are very coarse reflecting a supply from rather unprotected slopes.

The occurrence of Quaternary ashes of a Lake Toba eruption on the T2/P2 level in the Padang Terap River (Kedah) observed by DEBAVEYE *et al.* rendered absolute dating of this surface possible (Fig. 17). On the basis of fission track dating of zircon (NISHIMURA & STAUFFER, 1981) and K-Ar dating of glass shards (DEBAVEYE, 1986) it can be correlated with the Toba eruption of 75,000 B.P. or of 30,000 B.P. Due to the tectonically more stable environment the pediments in Malaysia are obviously much better preserved than on Java and in other parts of the island arc zone.

SOIL PARENT MATERIAL AND PRESENT LANDFORM				CLIMATE	VEGETATION	TIME	PRESENT SOIL
Formed by degradation processes on shale		Formed by aggradation processes					
Present landform	Forming processes	Present landform	Forming processes				
1. Pediplaine 1.1. Old pediplaine	Landscape formation Ferralitic weathering and soil formation Incision			humid ↓ very humid ↓	dense ↓ dense ↓	Late Pleistocene	Acrorthox Haplorthox
1.2. Slope pediment	Landscape formation Ferralitic weathering and soil formation ↓	1. Old alluvium (T2/D2) Rhyolitic ash deposit	Landscape formation Ferralitic weathering and soil formation Incision	dry ↓	open ↓	30,000 B.P. (1) 36,000–10,000 B.P. (2)	Paleudult Plinthudult
2. Hills and ridges	(Landscape and soil formation) Ferralitic weathering and soil formation	2. Young alluvium (T1)	Landscape formation Ferralitic weathering and soil formation Incision	very humid ↓ humid	dense dense	Holocene 8,000–4,000 B.P. (3) 4,000 B.P. until present (4)	Tropudalf Tropudult Tropaquult

(1) Stauffer et al. (1980); Aldiss and Ghazali (1984) –
(2) Lowering of the sea level to 40–60 m below the present day level (Geyh et al., 1979) –
(3) Increase of the sea level to 5 m above the present day level (Geyh et al., 1979) –
(4) Lowering of the sea level to the present day level (Geyh et al., 1979).

Fig. 17 : Tentative Late Pleistocene and Holocene chronology of geomorphological and soil development in the Padang Terap area, Malaysia (DE DAPPER *et al.*, 1988).

There is ample evidence from the Sunda and Sahul shelf areas for the existence of coherent drainage systems when these areas had emerged during the last Glacial period (MOLENGRAAFF, 1922; FAIBRIDGE, 1953). Also the offshore mining of placer tin in the Bangka and Belitung area and near Malaysia is confined to former river courses. Fig. 18 gives the coast line and main drainage lines during the last Glacial period.

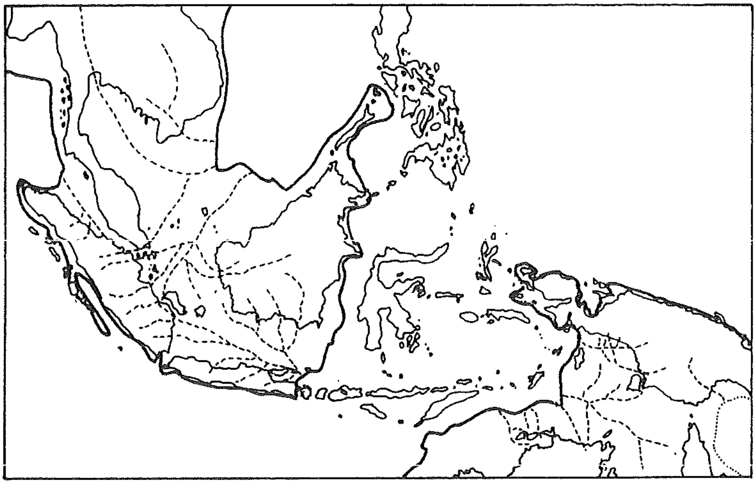


Fig. 18 : The emergence of the Sunda and Sahul shelfs during the low sealevels during the Pleistocene glacial periods. (VERSTAPPEN, 1980).

Similar observations on ancient drainage lines below present sea level are on record also from other parts SE Asia such as the Gulf of Thailand. SAWAMURA and LAMING (1974) report the existence of submerged deltas at 30 m and 45 m below present sea levels respectively. The connexed ancient beach ridges marking parts of the coast line in those periods and ancient lagoonal deposits have also been found by them (Fig. 19).

The most elaborate data, including also the deeper strata and Quaternary planation surface are those obtained during the earlier mentioned acoustic survey by ALEVA *et al.* (1972) of the bottom of the shelf sea between Bangka, Singkep and the Karimata Islands. The following stratigraphic sequence could be derived from their acoustic profiles (Fig. 20) :

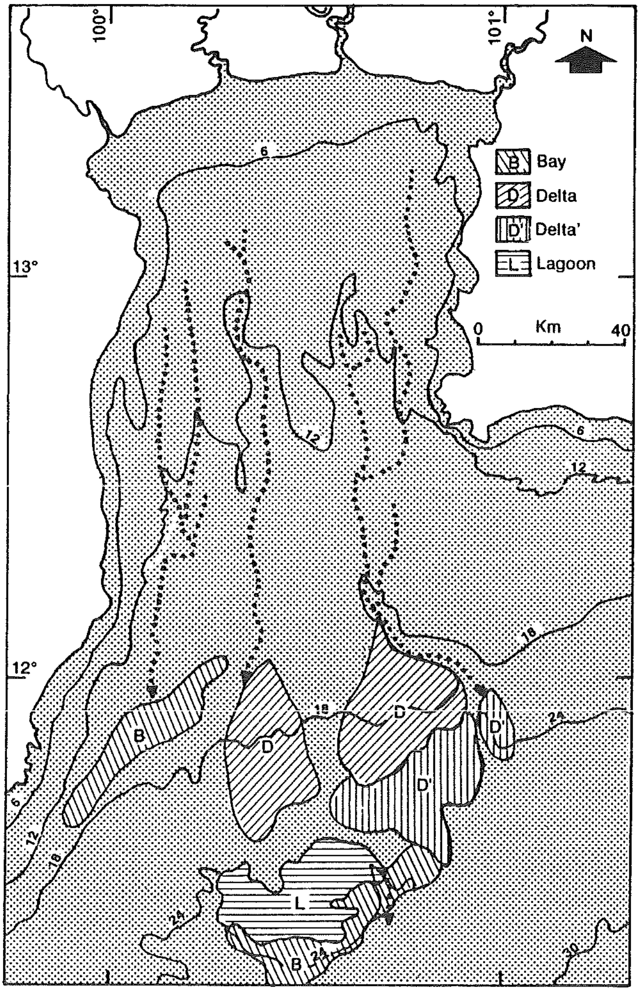


Fig. 19 : The Bay of Bangkok and adjacent parts of the Gulf of Thailand with two levels of submerged deltas (D) and bay shores (B) and submerged lagoonal deposits (L). The interval of the depth contours is 6 fathoms (SAWAMURA & LAMING, 1974).

1. Deeply chemically weathered basement rocks of Triasic (and older ?) age with Jurassic (?) granite intrusions
2. Old erosion surface at 30-100 m below present sea level of Cretaceous (?) age (Sunda land planation)
3. Older sedimentary cover of terrestrial (?) sands and peat presumably dating from early Tertiary lower sea levels

4. Alluvial complex of clay, sand layers and peat deposited in valley systems dating from upper Tertiary to Pleistocene
5. Young marine abrasion surface now at 20-30 m below sea level presumably from the Riss-Würm Interglacial and covered by red clays from the Würm Glacial
6. Young sedimentary cover of mostly marine clays formed during the Holocene and up to present.

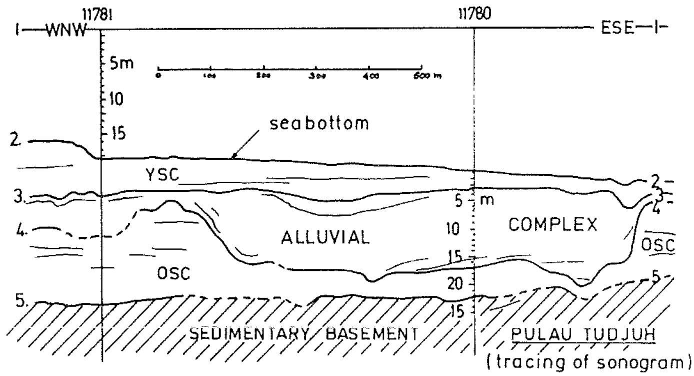


Fig. 20 :Sonogram data of the Sunda shelf deposits near Pulau Tudjuh, Indonesia, reflecting sea level changes (ALEVA *et al*., 1973).

The author adheres to Aleva's views but questions the interpretation as a marine abrasion platform of the surface mentioned under heading 5 in view of the prevailing low energy coastal environment. Its origin as a pediment or pediplane in the sense of DE DAPPER and DEBAVEYE seems more likely.

SMIT SIBINGA (1952, 1953) was the first who associated the alternation of marine and terrestrial Pleistocene sediments in Java, Sumatra and Kalimantan with glacio-eustatic sea level changes. Earlier investigators of the Quaternary geology of the region, not taking palaeo-climatic factors into consideration, were forced to invent complicated oscillating tectonic movements as to explain the situation. Smit Sibinga stressed the alternation of clayey Interglacial deposits with coarser textured Glacial deposits but he related this to sea level changes only and, not surprisingly in those days, gave no thought to climatic changes. His work is of great importance, however. He mapped for instance the coastlines during the Mindel and Günz regressions and the intervening interglacial period in north-eastern Java (For the Late Neogene climatic changes in this area see GORSEL & TROELSTRA, 1981). The studies by BATCHELOR (1979) on the patterns of discontinuously rising sea levels in Malaysia also should be mentioned in this context.

There is thus ample evidence that the shelf areas of SE Asia ran dry several times during the Pleistocene. The sea level during the last Pleistocene Glacial(s) was 90 - 100 m lower than at present. TJIA (1970, 1987) mentions drowned river mouths, fluvial and marine terraces, coastal sediments and other sub-marine geomorphological features from the Sunda shelf area to testify this. He mentions drowned shorelines at various depths ranging from 82-90 m to 7 below present sea level. The occurrence of ancient sea levels substantially higher than 5-6 m above the present level has been postulated by many earlier investigators in Malaysia. As already mentioned earlier in this paper it has since been established beyond doubt that the deposits of the so-called Old Alluvium found at these levels are not marine but fluvial in origin. Subaerial planation by pedimentation processes and at places also tectonism adequately accounts for these levels according to modern views.

A Holocene sea level of 5 m above present level is of widespread occurrence in SE Asia. It has been radiocarbon dated 6,000 - 5,000 B.P. and it thus corresponds with the now almost classical so-called "Daly" level of the climatic optimum. Its height and dating may vary to some extent but it undeniably exists. In the past doubt has been thrown on its reality since the level has not been found in some other parts of the world.

On the basis of newer scientific findings it is now clear that several factors tend to preclude universality of sea levels.Sea levels changes depend on a combination of several factors among which the volume of water - fluctuating with glaciation and water temperature - and the volume of the ocean basins - fluctuating with isostatic processes, sedimentation rates, tectonism, etc. - have a universal effect. Changes in the shape of the geoid are a third group of factors to be considered, however, and they may cause regional differentiation in sea level change. WALCOTT (1972) assumes, for instance, that at low latitudes the elastic response of the earth' crust to the growing post-glacial water column resulted in a lowering of the ocean floor that compensated 20 - 30 % of the sea level rise. Further work is by KUTZBACH (1981) and MARTINSON *et al.* (1987). TARAKANOV *et al.* (1992) have used marine terraces at various latitudes as indicators of geoidal changes. They maintain that a considerable latitudinal deformation of the geoidal surface occurred between 40,000/30,000 B.P. and the present as demonstrated by Fig. 21. Gravity anomalies are another factor to be considered as they are the cause of an irregular pattern of geoidal deformations all over the world (MÖRNER, 1976). Fig. 22 pictures the situation in the South and South-East Asian seas. It is believed that the pattern may change with time. Radar and laser altimetry from satellites is an important new means of investigation.

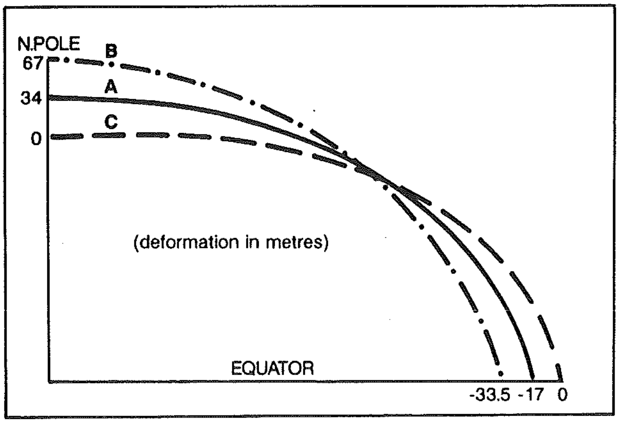


Fig. 21 : Latitudinal deformation of the geoidal surface 40-30 000 yrs BP due to glacial-induced mass redistribution based on marine terrace data (A) as compared to the actual hydrostatic ellipsoid (B) and geoidal surface (C) (TARAKANOV *et al.*, 1992).

Local and regional differences in sea level up to several metres may also result from differences in tidal range as the high tide mark is normally taken as the coastline. For the same reason differences in wave energy level due to exposure to wind or otherwise and differences in high tide mark related to the coastal configuration (e.g. estuaries) also may lead to erroneous conclusions with respect to former sea levels. Great care therefore is required in the translation of indicators of former coastlines into former sea levels, especially when the height differences are small as is usually the case for Holocene sea levels.

NOSSIN (1964) mentions Holocene sea-levels in Malaysia. TJIA (1975, 1977, 1985, 1987) proved the existence of a staircase of sea levels post-dating and lower than the earlier mentioned 5,000 - 6,000 B.P. level. They indicate minor rises in sea level at 4,000, 2,900 - 2,600 and 200 B.P. with intervening periods of lower sea levels at 5,800, 3,400 and 1,700 B.P. In a recent publication he compilated (TJIA, 1992) all data available from Malaysia and arrived at the sea level curve given in Fig. 23. Leaving apart the minor sea levels given above, it is evident that a gradual and probably step-wise regression occurred after the maximum (+ 5 m) level of 6,000 - 5,000 B.P. A distinct higher level around 1,700 - 2,500 B.P. is separated from the earlier high phase and the higher sea levels of the last millennium by lower sea levels around 3,200 B.P. and 1,100 B.P.

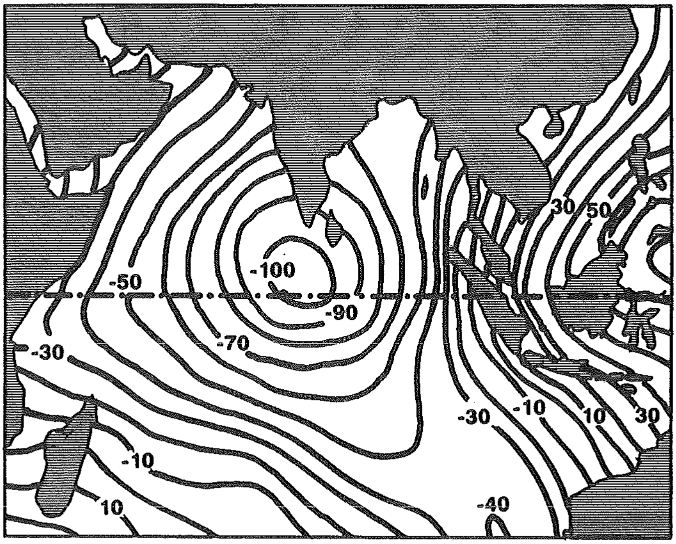


Fig. 22 : Deformation of the geoidal surface in South and South East Asian seas as a factor in spatial differences of marine terrace levels (MÖRNER, 1976).

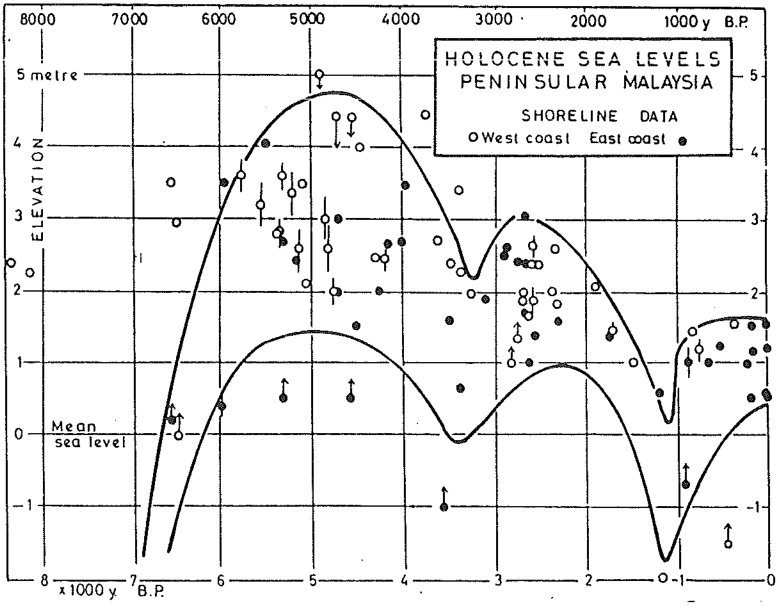


Fig. 23 : Holocene sea levels in peninsular Malaysia (TJIA, 1992).

VERSTAPPEN (1987) studied the evolution of the lowland coast of western Srilanka and obtained radiocarbon dates indicating higher (approx. 5 m) levels around 5,000 B.P. and a more recent period of strong coastal changes probably accompanying a rise in sea-level around 2,500 B.P. This corresponds with the findings of Tjia. The 2,500 B.P. period, also observed in the Langkawi islands, in Malaysia (1972) represents Fairbridge's (1961) Abrolhos submergence. Fig. 24 elucidates the situation. The + 5 m level of 5,000 B.P. is situated to the landward of the Puttelam Lagoon (DE GRAAFF, 1989). The Kalpitiya spit to the South of the town is of a complex nature and several phases of development can be distinguished. The beginning of the spit development was around 2,700 B.P. as is indicated by a ^{14}C dating 2,670 ± 50 B.P. of phase I. Of the younger stages only one has been dated so far (1,790 ± 50 B.P.). Wether these phases relate to sea level changes and/or changes in wind pattern as described in the next section, is as yet unclear. Most of their levels are not very different. Clearly a major change of long duration in coastal development has occurred since 2,700 B.P. The part of the spit situated to the North of Kalpitiya has another character : it is much younger, lower and narrower and it is at least for the greater part related to present sea-level. Its formation began following the chopping-off of the earlier spit, presumably during a period of renewed sea-level rise. It is very mobile and subjected to variations in wave energy resulting from wind pattern changes as is evident from a comparison of its appearance on a VOC map of 1753, on various maps made since 1902 and on a satellite image of 1982 with the available wind data. The role of minor recent sea-level changes is uncertain. The development of the spit clearly concords remarkably well TJIA'S sea-level curve of Malaysia given in Fig. 23.

DE KLERK (1983) found traces of elevated shorelines in his study area in SW Sulawesi (Celebes) that correspond with those mentioned by TJIA. The highest Holocene sea level (+ 5 m) was reached at 4,500 B.P. while after a stillstand phase the level dropped around 3,000 B.P. to + 3 m and subsequently to + 1 m around 2,000 B.P. He mentions a small rise between 2,000 and 700 B.P. (+ 2 m); a lowering between 1,500 and 1,000 B.P. (+ 1 m) which is separated by a stillstand phase from a period of further lowering even to below present level starting at 700 B.P. and followed by a rise continuing up to the present. The study area is considered by him tectonically comparatively stable.

Also in the tectonically unstable island arc areas new dating methods have resulted in reliable dating of Pleistocene shorelines. Interesting investigations have recently been carried out by PIRAZZOLI *et al.* (1991) on the island of Sumba, Indonesia where a flight of six major and numerous smaller coral reef terraces occurs at Cape

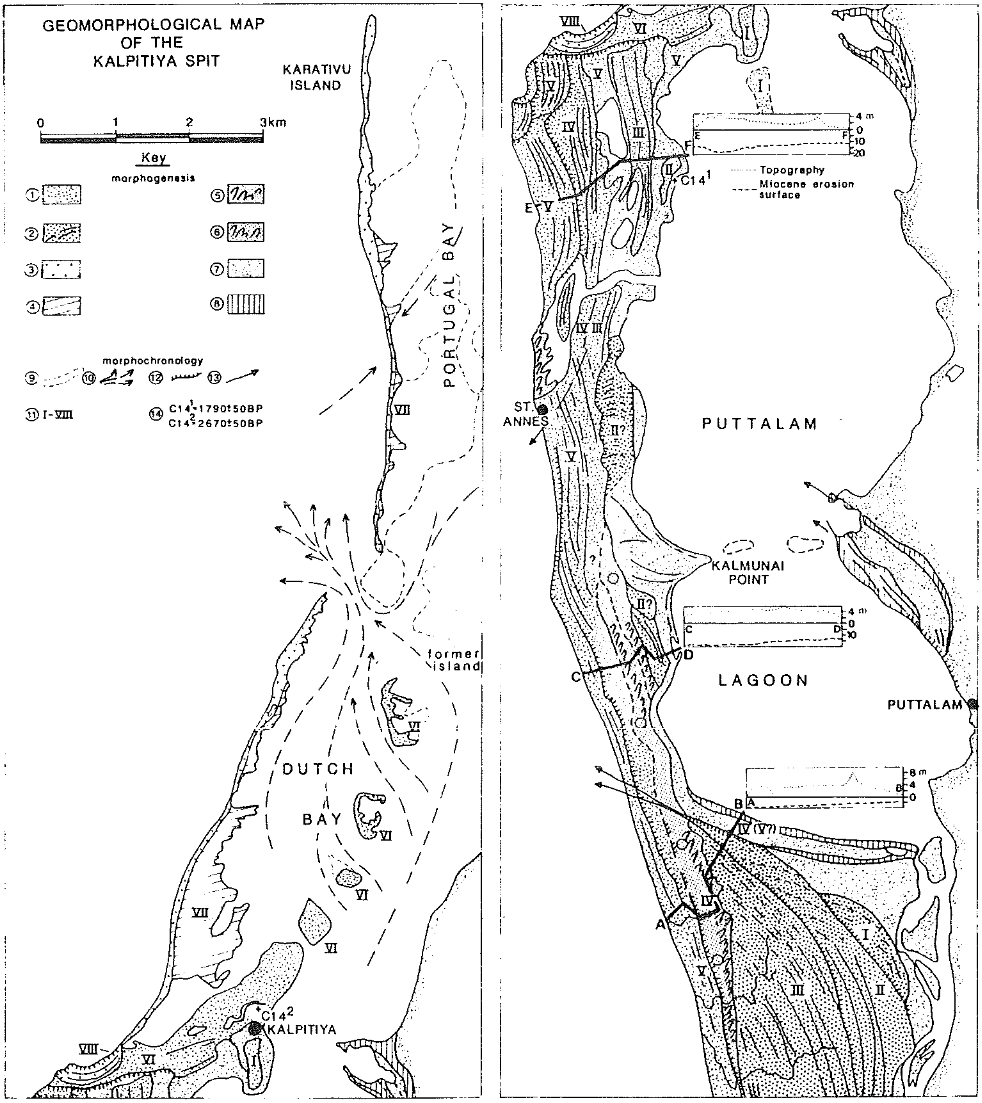


Fig. 24 : Coastal changes of the Kalpitiya spit, Sri Lanka, since its formation around 2700 years BP (VERSTAPPEN, 1987). key : 1. sandy beaches and barriers, general; 2. old barrier bars; 3. young sandy spits; 4. fine wash deposits; 5. active dunes closing ancient outlet; 6. older dunes; 7. lagoonal deposits; 8. low-energy spits and bars; 9. shallow, submerged areas; 10. tidal currents/channel; 11. sequent sets of coastlines; 12. unconformity in sequence; 13. old coast lines cut off by present one; 14. C_{14} dating. Profiles : dashed line Miocene substratum; dotted line surface of bars/dunes.

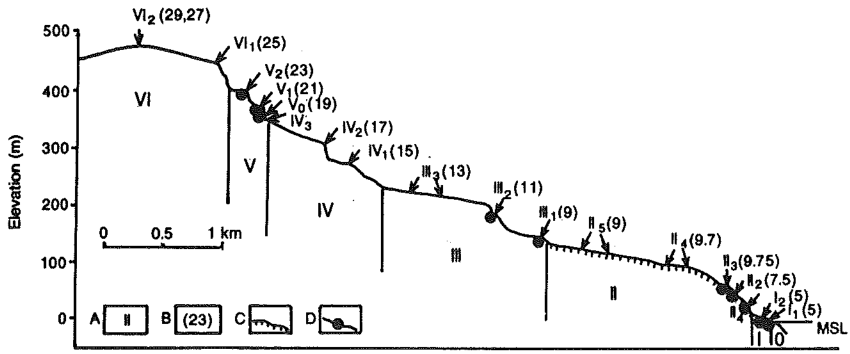


Fig. 25 : Staircase of uplifted coral terraces at Cape Laundi, Sumba, Indonesia. Legend : A terrace identification; B. inferred O isotope stage; C. abraded surface; D. crest of raised reef. (PIRAZZOLI *et al.*, 1991).

Laundi (Fig. 25) up to a height of 475 m above sea-level. Dating by electron spin resonance and uranium-series (Th-U and He-U) dating methods indicate that the oldest reef terrace formed 1 million years ago which gives an average uplift rate of 0.49 mm per year. The reef terraces, although of polycyclic origin, appear to correspond with high sea-levels of specific Interglacial stages. The sea returned to the same level repeatedly and the corals of the reef crest, found in growth position on the terrace surfaces formed during interglacial high sea-levels, are underlain by corals representing other periods. The scatter found between the various samples of the same level points to a polycyclyc origin; for example :

terrace II_2 :	height :	51 m;	275 - 117 ka
terrace II_3 :	height :	62 m;	344 - 228 ka
terrace III_1 :	height :	145 m;	397 - 322 ka
terrace IV_1 :	height :	275 m;	603 - 584 ka

Pleistocene sea-level changes were taken into consideration and the astronomically calibrated benthonic records from the Ocean Drilling Program as well as palaeomagnetic data were also used for comparison.

WIND PATTERN CHANGES AND COASTAL DEVELOPMENT

The latitudinal shift of the average position of the ITCZ during the Pleistocene alternation of Glacial and Interglacial periods has been mentioned already in the first section of this paper and its effect on the distributional pattern of precipitation has been elaborated upon. No geomorphological evidence has as yet been found for the associated changes in average wind direction that must have accompanied them. The secular fluctuations in the position of the ITCZ since the beginning of the meteorological observations in Jakarta have been studied by SCHMIDT and SCHMIDT-TEN HOOPEN (1951) in the context of the related rainfall patterns. Maximum precipitation values occur along the ITCZ while at some distance to the south and particulary also to the north lower values are recorded. As a consequence above average values occur in Java when, as a result of anticyclonal situation over Asia, the ITCZ lingers over southern Indonesia while below average rain fall values then are found in northern Indonesia. Inversely higher rainfall in the northern parts of Indonesia is accompagnied by below average values in the South.

This mechanism has been used by HOPE (1973, 1976) to explain geomorphological evidence for Quaternary climatic changes in north-east Queensland, Australia. He maintains that these were minor only but with wetter conditions during the early phase of the last Interglacial and drier conditions during the later phase that continued into the low-sea level phase. He claims wetter conditions during the last Glacial phase not only because of lower temperatures but also because of a southward shift of the ITCZ of 100 km during the (southern) summer. Dry conditions again prevailed from the later part of the Pleistocene (at least 15,000 B.P.) to the mid Holocene period. (4,000 - 5,000 B.P.). The present climate takes an intermediate position.

This sequence does not concord with that in some other parts of Australia which is comprehensible. COLHOUN (1975) and DE GEER (1988) maintain that Tasmania was not or only lightly forested in the cold period 25,000 - 10,000 B.P. notably in the plain lands while warmer/moister conditions rapidly began between 11,500 and 8,500 B.P. COLHOUN is of the opinion that the waxing phase of the last glaciation was colder and moister and the waning phase drier. The full sequence in eastern Tasmania is as follows :

75,000 - 55,000 B.P. cold and frosty conditions followed by a period of dominant aeolian activity

55,000 - 25,000 B.P. stable land surface with soil development

25,000 - 12,000 B.P. cold and frosty conditions followed by a period of dominant aeolian activity

12,000 - present stable land surface with soil development.

He also found evidence for warmer and drier conditions in the Holocene between 7,000 - 3,000 B.P. with a return to cooler and moister conditions after 3,000 B.P. (See also : GENTILLI, 1961).

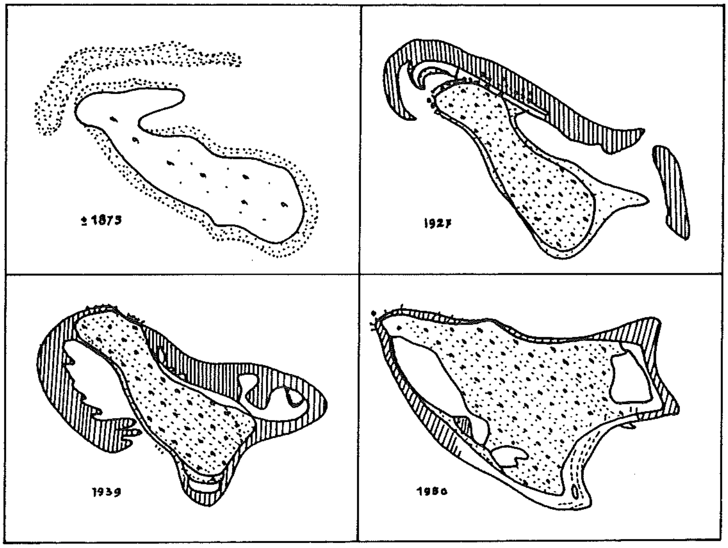


Fig. 26 : Shingle ridge development around the coral cay of P. Njanuk Besar situated in the Bay of Jakarta indicating a change in the dominant wind direction from the NW through the NE and W to the SW (VERSTAPPEN, 1953).

VERSTAPPEN (1953, 1954, 1968) was the first to establish the fluctuations in wind patterns associated with the secular shifts in average position of the ITCZ in the last 100 years and to elucidate their causative role in the evolution of coral cays in the Bay of Jakarta and on the coastal development. These fluctuations proved to be considerable : up to 500 % for some wind directions. Every minor increase in the force of onshore winds results in the formation of a coral single ridge at the side of the sand cay then most exposed to the wind and where thus wave action is strongest. Fig. 26 illustrates this. Similarly in the coastal plains every period of above average onshore winds will result in the formation of a sandy beach ridge. The repetitive occurrence of numerous minor fluctuations may ultimately lead to a "striping" of the coastal plain as is examplified by the aerial photograph of Fig. 27 showing the delta of the Noil Mina River on the island of Timor. When the fluctuations include periods of longer duration these may be reflected by larger beach ridge or spit bodies. An example is the oblique aerial photograph of Fig. 28, showing the Gilimanuk spit in western Bali that dates back to approximatively 3000 B.P. Five growth periods of the spit can be recognized (SOEJONO, 1975).

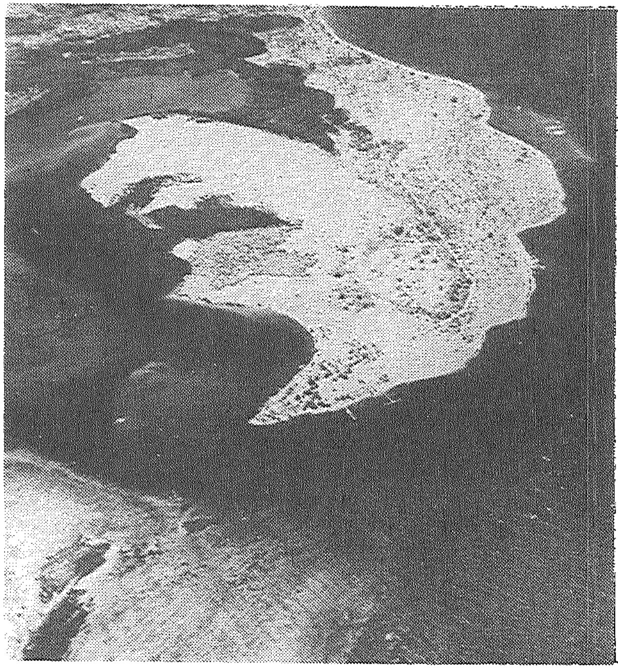
More recently (VERSTAPPEN, 1988) followed up his earlier studies in Jakarta Bay. It became clear that since the 19 fourties a marked anomaly in the wind pattern has

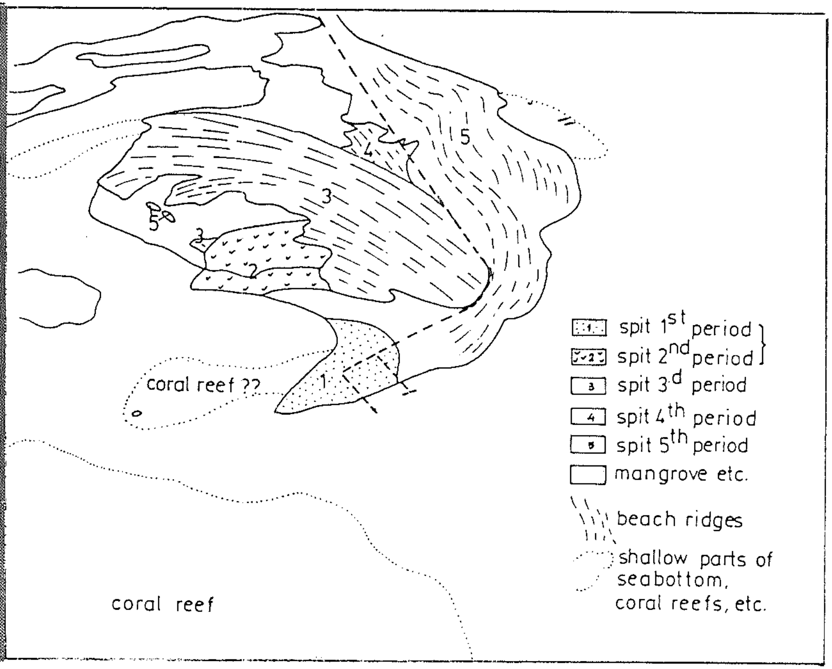
Fig. 27 : Vertical aerial photograph of the Noil Mina River mouth, along the south coast of Timor, Indonesia, with numerous beach ridges, each representing a period of strong onshore winds

occurred that led to a strong increase of northerly, onshore winds. This has - in combination with human interference with the coastal environment - resulted in strong abrasion of the coast and of the sand cays in the bay. At the same time the precipitation dropped to the lowest values on record. Apparently the period of Sahelian drought in Africa of the 19 seventies has also affected Indonesia. The lower precipitation since the 19 fourties is further substantiated by tree ring measurements in the G. Leuser area, Sumatra (BEEK, 1982) : a sudden decrease of growth rate then occurred (Fig. 30).

VERSTAPPEN (1987) also analysed the meteorological data of the last 100 years about rainfall and wind patterns along the West coast of Sri Lanka in the context of the evolution of the Kalpitiya spit already mentioned. Marked rainfall fluctuations occur in

Fig. 28 : Oblique aerial photograph of the Gilimanuk spit at the western end of Bali, Indonesia. Five growth period can be recognized, one of which is C_{14} dated (see text).





the wet zone of the island notably during the wet (Yala) season when the south-west monsoon prevails. They are in phase with those in Indonesia. The wind patterns yield an interesting result : when SW - onshore - winds are strong the S winds - blowing parallel to the coast - are weak. There thus is a clear secular variability in the energy of the waves attacking the west coast and the spits situated there. The evolution of the Kalpitiya spit (Fig. 24) thus is affected not only by the sea-level changes since about 2,700 B.P. but also by the climatic variations during that period.

Direct proof of the relation between wind patterns and coastal development can only be obtained from the period of meteorological observations. Extrapolation to earlier periods is possible to some extent, however, using data on rainfall anomalies from sources such as tree ring measurement in Java (DE BOER, 1951), records of drought/famine and floods in India (MURTON, 1984) (Fig. 31) old Chinese records (CHU, 1926), etc. These rainfall data then should be interpreted in terms of anticyclonal development over Asia and the associated average position of the ITCZ as to assess the wind patterns then prevailing. The longest record, however, can be found in the beach ridges of the alluvial plains. Research in this area may yield important results about the climates that prevailed in SE Asia during the last 5,000 years !

EPILOGUE

The climatic conditions of past geological periods have had an important impact on the geomorphological development and on the environmental situation in South and South-East Asia. They are, in turn, also strongly affected by megageomorphological factors. The important and comparatively rapid climatic variations of the Quaternary have left particulary distinct traces in the landforms. The fluctuations during the Holocene are of interest for explaining the present situation and for assessing potential near-future changes.

The use of pollen data in the study of climatic change in SE Asia is complicated by anthropogenous disturbances. FLENLEY (1988) found evidence for pre-historic land use changes in Sumatra and New Guinea. Sites in Sumatra show forest disturbances starting 4,000 B.P. or earlier while in the Highlands of New Guinea anthropogenic disturbances of swamps started 9,000 B.P. and evidence of burning exist 10,000 B.P. Man has thus changed his environment also in this part of the world already millennia ago. In the past two centuries man has become a geomorphological factor of growing importance (VERSTAPPEN, 1988) and he is now on his way to become a driving factor in climatic change. The study of palaeoclimates and climatic geomorphology thus has

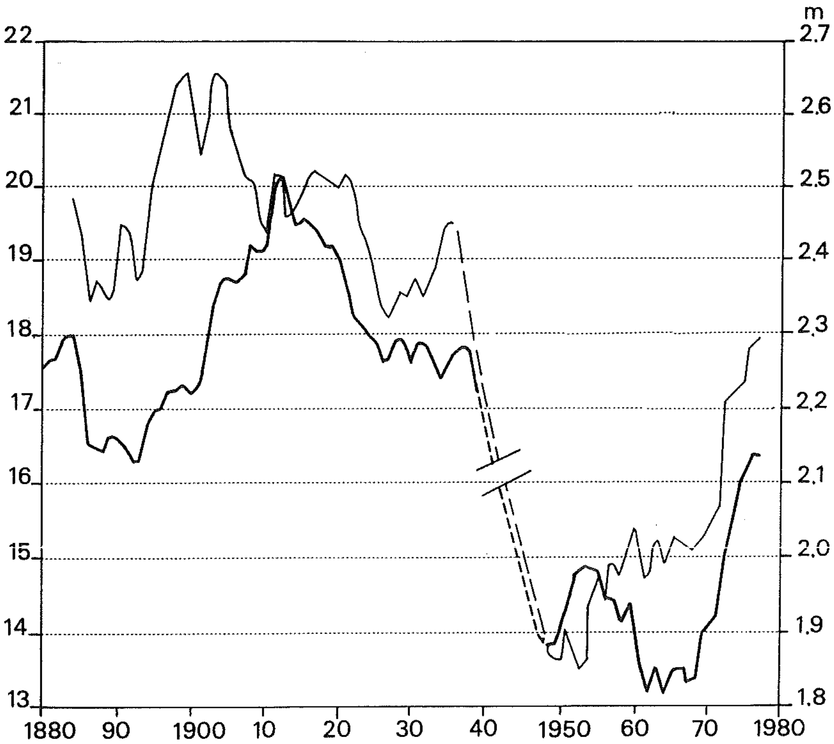


Fig. 29 : Rainfall fluctuations at Jakarta, Indonesia since 1880. The early fluctuations were followed by fluctuations at a substantial lower precipitation level in later decades when also the Sahelian drougth occurred.

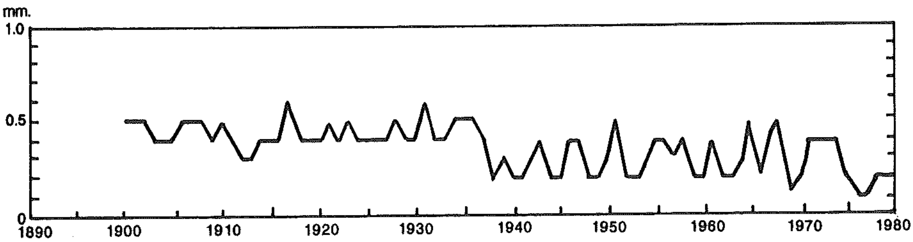
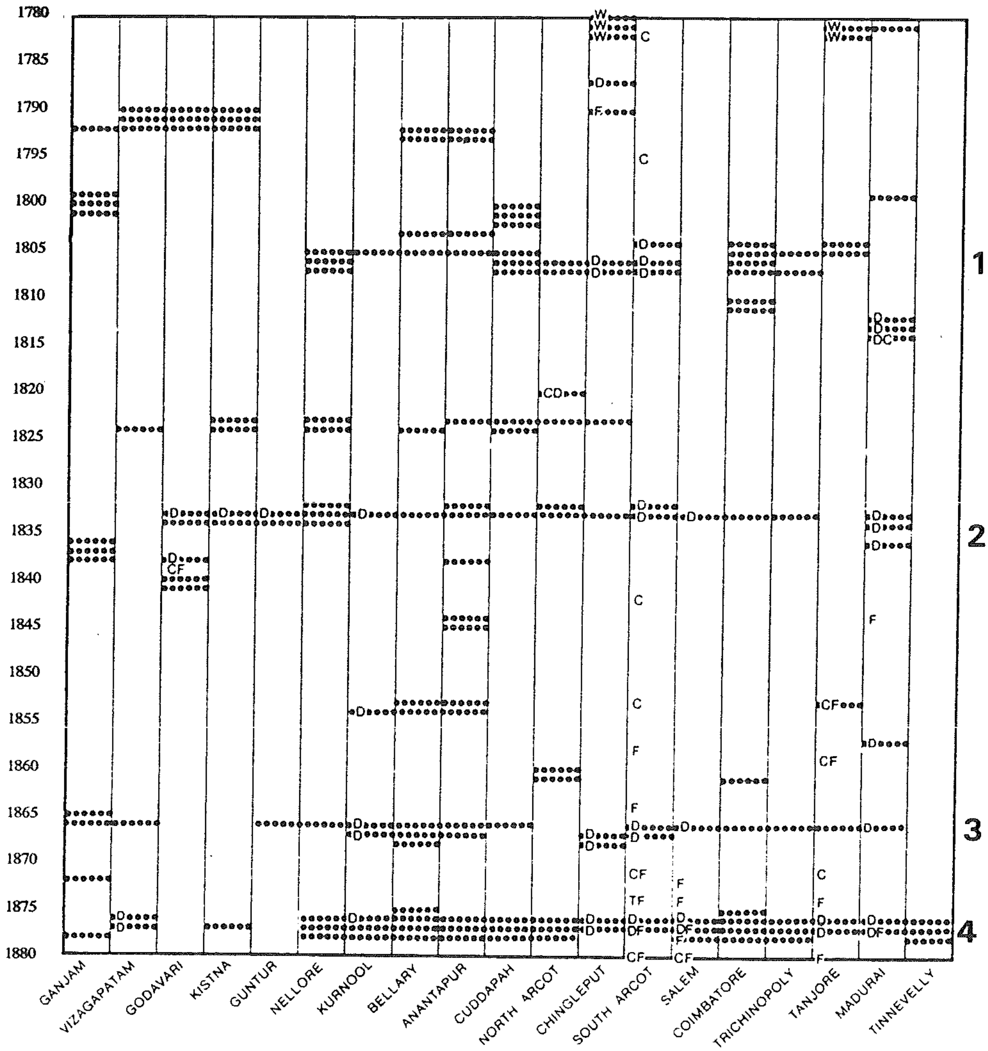


Fig. 30 : Tree ring measurements at Mt Leuser, N. Sumatra showing a marked decline in growth rate since about 1940 when the drier conditions given in fig. 29 began (VAN BEEK, 1982).

got a new dimension and urgency in the context of global environmental change. Much work still has to be done in this field in South and South-East Asia before the available data can be considered adequate.



Datted line = Famine or Scarcity, F = Flood, D = Drought,
C = Tropical Cyclone, T = Tornado, W = Warfare.
Note : The pattern of disaster frequency is incomplete for all districts.

Fig. 31 : Disaster occurrence in southern India 1780-1880 based on gazetteers and statistical accounts of the districts of the Madras Presidency. Four main periods of drought (D) can be traced (MURTON, 1984).

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