

THE HOLOCENE DEVELOPMENT OF THE YOUNG COASTAL PLAIN OF SURINAME¹

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ABSTRACT

Roeleveld, W. & A. J. van Loon 1979 The Holocene development of the Young Coastal Plain of Suriname – Geol. Mijnbouw 58: 21-28.

A new series of 19 radiocarbon dates from the Holocene Young Coastal plain in Suriname fits into the picture of a rapid sea-level rise in this area until about 6000 B.P. From 6000 B.P. onward the sea-level remained essentially at its present position. Indications for a higher position of the sea-level at about 6500 B.P. are inconclusive.

From about 6500 B.P. onward the marine influence in the interior part of the Young Coastal Plain diminished, but some marine oscillations were registered. The available evidence is too limited to permit pertinent conclusions about the regional occurrence of marine fluctuations. From an analysis of the younger Holocene development in the Hertenrits area it is concluded that a new subdivision has to be established for understanding of these marine oscillations.

INTRODUCTION

The present note deals with the results of radiocarbon dating of a number of peat seams which occur within the Holocene coastal succession in Suriname (South America).

The objective of the investigation was primarily to establish a more detailed transgressive-regressive chronology for the area than was hitherto available. The picture of short term fluctuations of marine influence in the Suriname coastal area was expected to be also of interest to the study of comparable phenomena undertaken in the North Sea area (cf. ROELEVELD, 1974; GRIEDE, 1978) and to serve the purpose of a better understanding of the nature of such variations in general.

At the same time the data obtained contribute to the knowledge of the evolution of the Suriname coastal area during the Holocene, and shed new light on the relative Holocene sea level movement in the area.

THE GEOLOGICAL SETTING (Fig. 1)

Suriname is situated on the northeast coast of South America. The major, interior part of the country consists of Precambrian rocks of the Guiana Shield. To the north the crys-

talline basement is overlain by Mesozoic and Cenozoic sediments. These sediments fill the coastal Guiana Basin, which rapidly deepens toward the north and northwest. For a more detailed outline of the geology of Suriname the reader is referred to KROOK (1975) and to BOSMA ET AL. (1977).

Morphologically, the sedimentary belt is subdivided into the Coastal Plain and the Cover Landscape (or Zanderij Landscape). In the latter predominantly sandy fluvial sediments of the Upper Coesewijne Formation (Pliocene) crop out. Within the Coastal Plain a distinction is made between the Young (Holocene) Coastal Plain, and the Old (Eemian and/or older Pleistocene interglacial) Coastal Plain situated further inland. The Old Coastal Plain has an elevation of 2 to 7 m above mean sea-level. The marine clays and fine sands of the Coropina Formation which make up the Old Coastal Plain have been strongly dissected, apparently during the low sea-level of the last glacial period.

The Young Coastal Plain has an elevation of predominantly 0 to 3 m above mean sea-level. To the south the sediments of the Young Coastal Plain (Demerara Formation) extend into the erosional gully system developed in the Old Coastal Plain, and at places even beyond, into the Cover Landscape.

The development of the Young Coastal Plain (Table I) has been summarized by PONS (1966) and BRINKMAN & PONS (1968). During an initial phase its development was governed by the rapid post-glacial rise of the sea-level, which lasted until about 6000 C-14 years B.P., when the sea-level approximately attained its present position. In this early phase mainly pyrite clays rich in organic matter, and some intercalated peat layers were formed in a coastal mangrove (*Rhizophora*)

Manuscript received and accepted: 1978-09-30.

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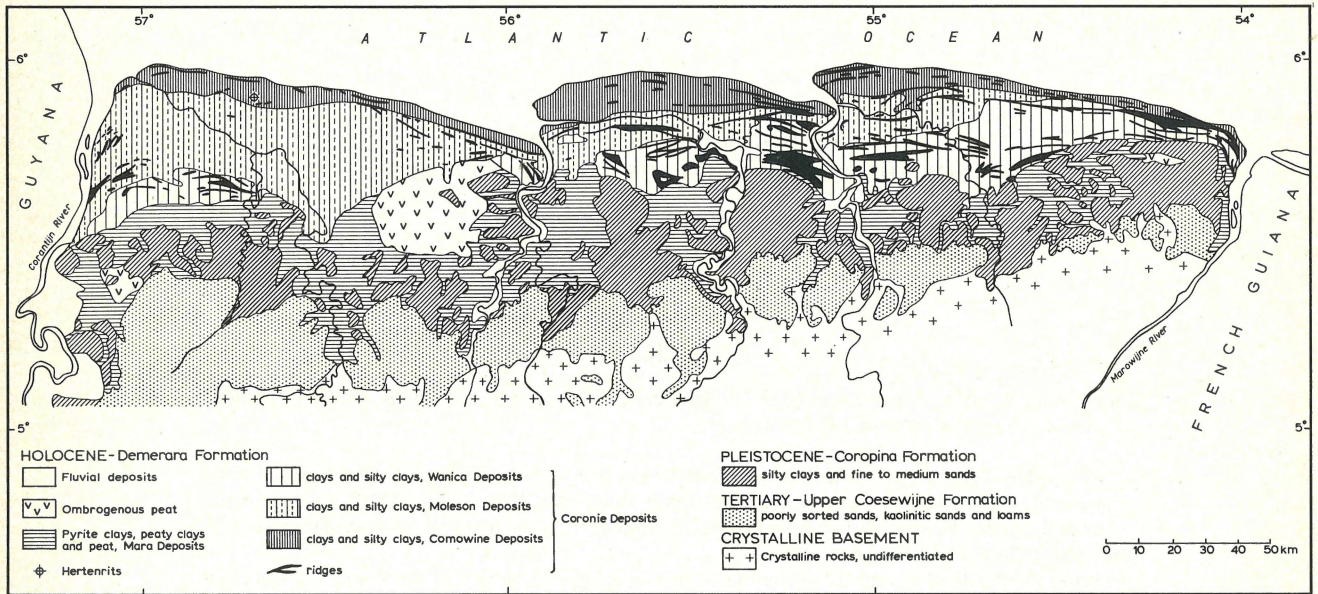


Fig. 1 Geological map of the Young Coastal Plain of Suriname (modified after Krook, 1979).

swamp in the interior part of the present-day coastal plain, and notably in the erosional gullies which penetrate inland (Mara Subformation).

This predominantly vertical development was replaced at about 6000 B.P. by a different type of sedimentation. Lateral accretion alternating with phases of erosion became predominant; mud flats built up under *Avicennia* vegetation (with a relatively low pyrite content) and sandy and shelly chenier ridges were formed (Coronie Subformation).

BRINKMAN & PONS (1968) recognize three depositional phases during this younger stage (Table I), viz. the Wanica phase (6000/5500 – 3500/3000 B.P.), the Moleson phase (2500 – 1300

B.P.) and the Comowine phase (about 1000 B.P. – present). This subdivision is based essentially on geomorphological and pedological criteria. The phases are thought by Brinkman and Pons to be separated by periods of prevailing erosion or non-deposition (3000-2500 and 1300-1000 B.P. respectively). BRINKMAN & PONS (1968, Fig. 7) suggest that these erosional phases are associated with a temporary minor lowering of the sea-level, whereas the onset of the next depositional phase should be related to a slight rise of the sea-level. During the main part of the three depositional phases, however, the sea-level is supposed to be constant.

Table I Stratigraphy of the Holocene in the Young Coastal Plain of Suriname (modified after Brinkman & Pons, 1968). The stratigraphic terminology is newly introduced for this area, in order to avoid confusion.

LITHOSTRATIGRAPHY		Age; conv. C-14 y B.P. (Brinkman & Pons, 1968)	
DEMERARA FORMATION	CORONIE Subformation	Comowine member	present 1000
		Moleson member	1300 2500
		Wanica member	3000-3500 5500-6000 6000
	MARA Subformation		

FIELD INVESTIGATIONS AND SAMPLING

During a six week campaign in Suriname in 1975 a reconnaissance survey was made in the Young Coastal Plain, in order to locate representative sections for the sampling of peat layers. On the basis of available information (personal communications, internal reports and material from the archives of the Soil Survey at Paramaribo) a number of localities were selected, also taking the factor of accessibility in consideration.

A total of 80 hand borings was performed in 47 different locations, in a zone ranging from the Marowijne River to west of the Coppename River (Fig. 2).

With a single exception (Burnside), all borings are situated in the interior extensions of the Young Coastal Plain where the most interesting sections were to be expected, and which could, moreover, be reached relatively easily. As a consequence the investigated sections are mainly concerned with the older Holocene Mara Subformation (Table I), and

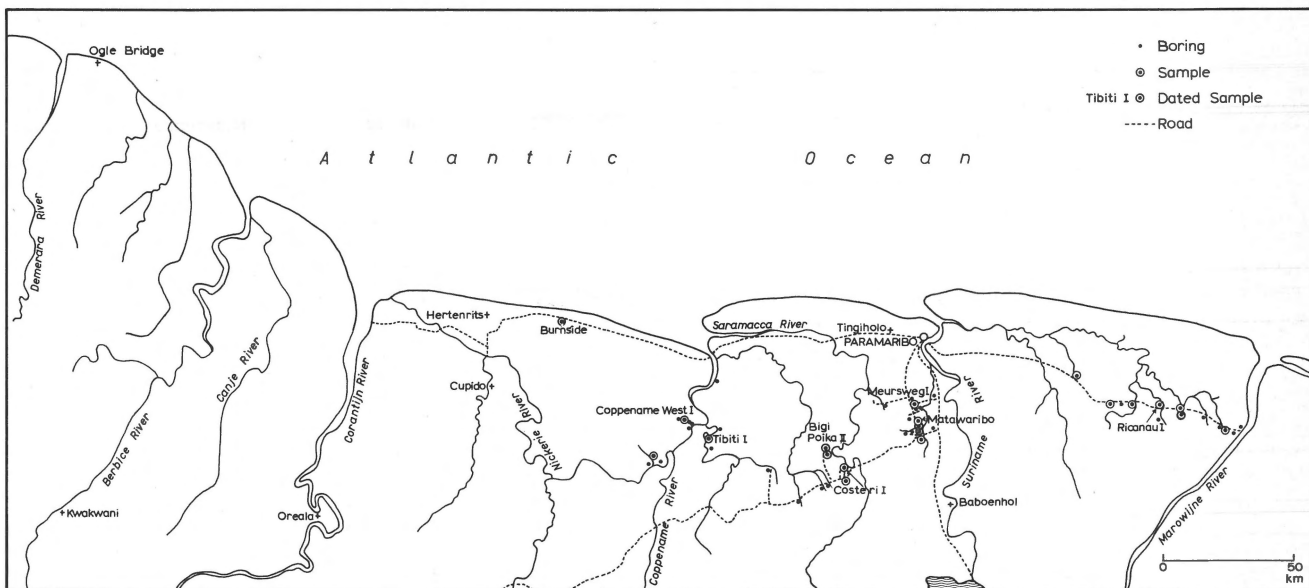


Fig. 2
Map showing sample locations and localities mentioned in the text.

only limited information on the younger Holocene history became available.

In the field 70 cores of 25 cm length and 4.5 cm diameter were collected. From these cores relevant samples for C-14 dating and pollen analysis were prepared (cf. Fig. 5). As yet 19 samples have been radiocarbon dated; the palynological results are not yet fully available.

C-14 DATES

The results of the radiocarbon determinations which were performed on behalf of the present investigation are presented in Table II. A few other relevant dates have been added to the list. All ages are expressed in conventional radiocarbon years B.P. (before 1950).

It should be kept in mind that the calculated ages of some of the more superficially situated samples may be too low, due to contamination by (sub)recent rootlets. Notably the samples Costeri I/87-91, Meursweg I/33.5-37, Meursweg I/75-79, Bigi Poika II/140-144 and Bigi Poika II/62-66 cm could have been affected to some degree by this effect. In the Groningen C-14 laboratory, however, all precautions have been taken to rule out this type of error as far as possible.

The samples collected during the 1975 campaign consist of (more or less clayey) peat, with the exception of the sample Burnside, which is considered as a vegetation horizon and consists of finely dispersed black organic matter in a clayey matrix. In the radiocarbon laboratory three successive alkaline extracts were dated to apparent ages of 1440 ± 70 , 1430 ± 70 and 1380 ± 90 B.P. respectively. This indicates the possibility of an incomplete removal of younger contamination

from the sample. Consequently, the age of 2140 ± 50 B.P. obtained for the remaining material of the sample Burnside may still be too low.

INTERPRETATION OF THE RESULTS

'Sea-level movement

In figure 3 the dated peat samples are represented in a time/depth diagram; to this series the date Ogle Bridge in Guyana (VAN DER HAMMEN, 1963), which is concerned with the early Holocene, has been added. It may be assumed on the basis of limited palynological evidence (VAN DER HAMMEN, 1963; ROELEVELD, 1969) that most peat layers have been formed in brackish coastal *Rhizophora* swamps at about mean sea-level.

Unfortunately, the surface elevation at the sampling locations could not be established, and this constitutes a significant source of error in the diagram. We have used the following estimated values for the surface elevations with respect to present-day sea-level:

Costeri I:	+ 2 m (cf. SCHULZ, 1960, Fig. 53)
Meursweg I:	+ 2 m (cf. VEEN, 1970)
Ricanau I:	+ 2 m
Bigi Poika II:	+ 2 m (analogous to Costeri)
Tibiti I:	+ 2 m
Coppename West II:	+ 1 m
Matararibo:	+ 0.75 m (ROELEVELD, 1969)
Burnside:	+ 1.5 m

For the Cupido dates (BRINKMAN & PONS, 1968) and Baboenhol

Table II
Radiocarbon dates from the Holocene in Suriname and Guyana.

Sample	Depth below surface	Material	Age (conv. C-14 y B.P.)	Lab. nr.	Reference
Costeri I	87 -91 cm	peat	545 ± 35	GrN 7810	Fig. 5
Costeri I	810 -814.5	peaty clay	7620 ± 60	GrN 7811	do.
Tibiti I	227 -230.5	peat	6410 ± 50	GrN 7812	do.
Tibiti I	277 -281	peat	6580 ± 70	GrN 7813	do.
Ricanau I	264.5-268.5	clayey peat	5810 ± 60	GrN 7814	do.
Ricanau I	285 -289	peaty clay	5660 ± 60	GrN 7815	do.
Ricanau I	358 -362.5	peaty clay	6090 ± 70	GrN 7816	do.
Ricanau I	410 -414	clayey peat	6580 ± 80	GrN 7817	do.
Ricanau I	529 -534	peaty clay	6845 ± 50	GrN 7818	do.
Meursweg I	33.5 -37	peat	560 ± 35	GrN 7819	do.
Meursweg I	75 -79	clayey peat	4190 ± 60	GrN 7820	do.
Meursweg I	411.5-415.5	peat	6360 ± 70	GrN 7821	do.
Meursweg I	580.5-584.5	peaty clay	7140 ± 60	GrN 7822	do.
Bigi Poika II	62 -66	clayey peat	2020 ± 45	GrN 7823	do.
Bigi Poika II	140 -144	clayey peat	6350 ± 60	GrN 7824	do.
Coppename West I	341 -344.5	clayey peat	6650 ± 70	GrN 7825	do.
Coppename West I	370 -375	peat	6680 ± 110	GrN 7826	do.
Coppename West I	510 -519	clayey peat	7040 ± 80	GrN 7827	do.
Burnside I	40 -45	fine dispersed black organic matter	2140 ± 50	GrN 7878	
Hertenrits	320	peat	1045 ± 90	GrN 845	Vogel & Waterbolk, 1964
Hertenrits	250	wood	1265 ± 60	GrN 1897	do.
Hertenrits	60	peaty	1130 ± 45	GrN 1898	do.
Ogle Bridge	2040	peaty clay	8590 ± 65	GrN 3058	Van der Hammen, 1963
Kwakwani	180	peaty clay	6490 ± 80	GrN 3103	do.
Oreala	375	peat	6470 ± 85	GrN 3109	do.
Cupido	500	peat	6360 ± 70	GrN 4517	Brinkman & Pons, 1968
Cupido	90	peat	5130 ± 70	GrN 4518	do.
Cupido		peat	6720 ± 70	GrN 4519	do.
Matawaribo	525-555	peaty clay	7240 ± 100	GrN 4847	Roeleveld, 1969
Baboenhol	430	peaty clay	6310 ± 45	GrN 5731	Veen, 1970
Tingiholo	30-50	shells	1715 ± 25	GrN 8302	Unpublished
Saramacca		shells	1850 ± 170	I 624	Delaney, 1966
Saramacca		shells	1525 ± 150	I 625	do.
Saramacca		shells	1500 ± 150	I 628	do.
Saramacca		shells	1675 ± 140	I 629	do.

(VEEN, 1970) the sample depths have been plotted according to published information.

The depth at which the sampled peat layers now occur will in many cases have been significantly influenced by the effects of compaction of the underlying soft, waterlogged (peaty) clays. Thus, the highest situated samples of a certain age must be regarded as the more reliable sea-level indicators.

Even allowing for the above-mentioned uncertainties the diagram indicates a rapid rise of the sea-level from the early Holocene until about 6000 C-14 y B.P., and an essentially stable situation since about that date. This conclusion agrees with BRINKMAN & PONS'S (1968) sea-level reconstruction.

The material from Suriname does not provide any indications for a sea-level higher than the present one at about 6500 B.P., as suggested for Guyana by VAN DER HAMMEN (1963).

BRINKMAN & PONS (1968) came to the same conclusion; they suggested that the study area of Van der Hammen had apparently been affected by very local uplift.

If one wishes to explain the discrepancy between the data from Suriname and Guyana in terms of differential tectonics, the present authors suggest that downwarping of the Suriname coastal zone (due to the effects of hydro-isostasy) is responsible, rather than uplifting of the area in Guyana studied by Van der Hammen.

We doubt, however, the necessity of such a tectonic explanation, since in our opinion VAN DER HAMMEN'S (1963) interpretation of the situation in Guyana in terms of a former higher sea-level is open to serious doubt.

Van der Hammen bases his conclusions on the occurrence of marine-influenced sediments (according to palynological evidence), situated at rather high elevations above the pre-

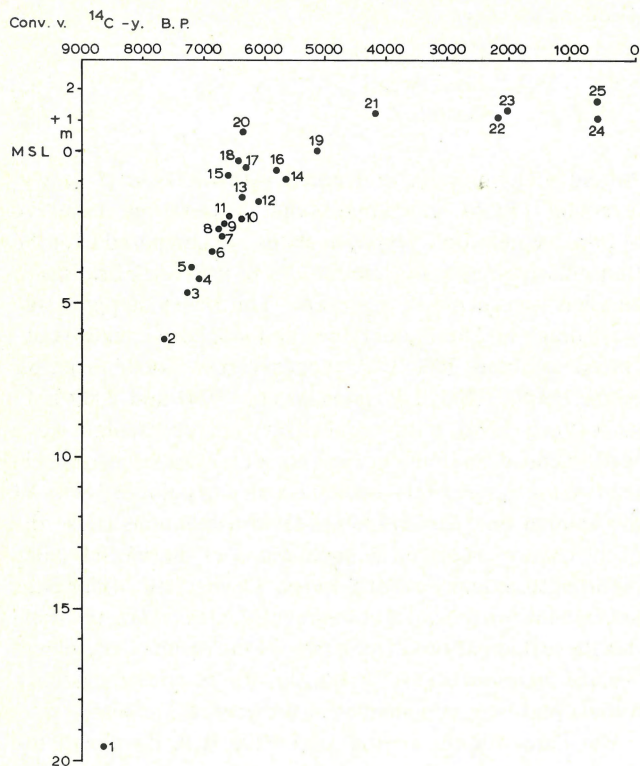


Fig. 3
Time-depth diagram of peat layers in the Young Coastal Plain; 1 Ogle Bridge (Van der Hammen, 1963); 2 Costeri; 3 Matawaribo (Roelvel, 1969); 4 Coppename West; 5 Meursweg; 6 Ricanau; 7 Coppename West; 8 Cupido (Brinkman & Pons, 1968); 9 Coppename West; 10 Meursweg; 11 Ricanau; 12 Ricanau; 13 Cupido (Brinkman & Pons, 1968); 14 Ricanau; 15 Tibiti; 16 Ricanau; 17 Baboenhol (Veen, 1970); 18 Tibiti; 19 Cupido (Brinkman & Pons, 1968); 20 Bigi Poika; 21 Meursweg; 22 Burnside; 23 Bigi Poika; 24 Costeri; 25 Meursweg.

sent-day river level (or floodplain level) along the Corantijn, Berbice and Canje Rivers, and these sediments have radiocarbon ages within the period 6500-6000 B.P. From observations made by the first author along the middle courses of the Saramacca, Suriname and Coppename Rivers, upstream of the lowermost falls (cf. BOOTSMAN & DERCKSEN, 1977; BALSEM & RHEBERGEN, in prep.; KIPS & SNEL, in prep.; DE FRETES & VAN DER GAAUW, in prep.) we gained the impression that floodplains are generally situated at rather high levels with respect to the present-day river. Even during the high discharges of the rainy season large parts of the floodplains remain at heights up to 2-3 m above the water level; they seem to be flooded in really exceptional circumstances only. Admittedly, it is unknown whether these floodplains are in fact in equilibrium with the present rivers. If, however, the rivers should be considered to have incised their beds during the more recent Holocene, it cannot be due to eustatic factors since the river stretches concerned are situated upstream of the lowermost falls which constitute local base levels.

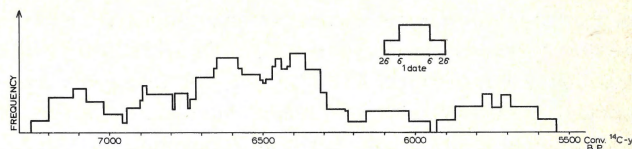


Fig. 4
Frequency histogram of 13 dated peat samples older than 5500 years.

Thus, the possibility has to be considered of whether the high-lying sediments along the Guyana rivers (thought by Van der Hammen to be representative of a former high sea-level) are not in fact in equilibrium with the present-day river level. In that case their significance for a higher position of the sea-level obviously is worthless. But even when these sediments are indeed situated above the present-day sedimentation level, it remains doubtful whether the incision (which in that case must be concluded to) is the result of a lowering of the sea-level, subsequent to a former higher position.

On the basis of the considerations we consider the evidence on the sea level movement in Suriname pictured in figure 3 as representative, and we will not take into account the possibility of a higher sea-level around 6500 y B.P.

Marine influence before about 6200 B.P.

Of our 18 peat samples from the interior part of the Young Coastal Plain (Table II and Fig. 5) 13 are older than 5500 years. This result is hardly surprising, considering the fact that we are primarily concerned in the sample area with the older Holocene Mara Subformation (cf. Figs. 1 and 2).

Plotted in a frequency histogram (Fig. 4), the majority of these 13 dates appears to occur in the 6700-6300 years interval, roughly around 6500 B.P. Apparently the marine influence in the interior extensions of the coastal area, which until that time had been present under the form of deposition of (humic) clays, decreased at about 6500 B.P. and (mainly *Rhizophora*) peat formation became more frequent. It is logical to relate this decrease in marine influence to the termination of the Early-Holocene sea-level rise at about this time. In our opinion the onset of lateral progradation of the coast line as a result of the declining sea-level rise would be important in limiting the marine influence in the interior zone, by enlarging the distance to the shore. Further, it may be surmised that the first significant ridge complexes evolved in this period, protecting the interior area from the sea.

It should be noted that also at about 6500 B.P. the farthest inland penetration of marine influence was registered along some of the main rivers of Suriname and Guyana. At Kwakwani on the Berbice River VAN DER HAMMEN (1963) dated a „marine transgression”, viz. a peat layer showing brackish influence in its pollen content and situated on top of pre-Holocene deposits, at 6490 ± 80 y B.P. (GrN 3103). From Oreala

at the Corantijn River Van der Hammen mentions a comparable peat layer, dated at 6470 ± 85 (GrN 3109). VEEN (1970) mentions a date of 6310 ± 45 B.P. (GrN 5731) for a peaty layer containing 60% *Rhizophora* pollen, intercalated in fluvial sediments of the Suriname River at Baboenhol.

A comparable date from the coastal plain proper is given by BRINKMAN & PONS (1968): 6360 ± 70 (GrN 4517) for the lower part of a peat layer on top of Pleistocene Coropina clay at Cupido.

Thus, the evidence on the one hand points toward declining marine influence around 6500 B.P. (peat growth instead of deposition of clays), but on the other hand reveals that the marine influence was still penetrating inland and upward at that time. This picture emphasizes that we are concerned with the final stages of the Early-Holocene sea level rise at about 6500 B.P.

Of the dates older than about 6500 B.P., Ricanau I/529-534 (6845 ± 50) and Costeri I/810-814,5 (7620 ± 60) are concerned with intercalated peat beds within the (peri)marine succession. They represent temporary decline in marine influence at those localities. Whether these dates are representative of regional tendencies cannot be decided on the strength of the available material.

The dates Meursweg I/580.5-584.5 (7140 ± 60) and Copename West I/510-519 (7040 ± 80) are concerned with the basal part of peat layers which probably directly overlies older, Pleistocene, deposits, although in both cases the top of the underlying Coropina clay seems to be slightly reworked. The dates are representative of transgressions occurring at the respective depths, and they reflect the effects of the general sea level rise rather than secondary oscillations in marine influence.

Summarizing, the available material does not permit conclusions to be made on the possible occurrence of secondary oscillations of the sea-level in the period of the Early-Holocene general sea level rise.

Marine influence after about 6200 B.P.

The dated profiles (Fig. 5) clearly reveal that sedimentation did occur at all investigated locations sometime during the period 6200 B.P. – present. This fact should be stressed, since it is not generally acknowledged that the inwardly situated Mara subformation (which is sea-derived, but probably not full-marine) has subsequently been covered by younger sediments.

In general, sedimentation during this younger period was rather limited, and at many places peat formation alternated with deposition of clastic sediments. The dates available are too few to show possible regional regularity in the occurrence of marine oscillations in the period concerned. The relevant dates, in chronological order, are:

6090 ± 70	} Ricanau I (considering their position the two younger dates are mutually contradictory)
5810 ± 60	
5660 ± 60	

4190 ± 60	Meursweg I
2020 ± 45	Bigi Poika II
560 ± 35	Meursweg I
545 ± 35	Costeri I

All dated samples are concerned with the base of peat(y) layers (cf. Fig. 5), which means that they are representative of local 'regression'. When the dates are compared with the transgressive-regressive chronology of the North Sea Basin, parallels are seen to be present. The onset of regressive oscillations in The Netherlands and adjacent countries occurred at about 5800 B.P. (ROELEVELD & MOOK, in prep.; INQUA, 1978), 4200 B.P. (ROELEVELD, 1974) and 2000 B.P. (ROELEVELD, 1974). If this parallelism were representative of a real synchronism in the occurrence of marine oscillations in the two areas, then this would be a strong point in favour of the opinion that worldwide sea-level oscillations cause the occurrence of short term fluctuations of marine influence reported from many coastal zones. The second author does not exclude (on the basis of sedimentological characteristics) that these fluctuations in majority are the result of lateral and vertical facies changes. Obviously, the restricted evidence forbids pertinent conclusions at the present moment.

Our dates for the period after 6200 B.P. do not fit the tentative chronology presented by BRINKMAN & PONS (1968). The end of the Wanica and Moleson sedimentation phases are dated by these authors at about 3000-3500 and 1300 B.P., respectively; and these figures do not occur within our set of data.

It should be noted in this respect that the terminal dates for the Wanica and Moleson phases mentioned by Brinkman & Pons are largely based on C-14 dates of shell material from chenier ridges. Thus, these datings are primarily related to the evolution of the coastal morphology, and this ridge development is not necessarily related simply to transgressive/regressive oscillations.

Concerning the nature of the classification presented by BRINKMAN & PONS (1968) and its relation to our data, as well as to the occurrence of transgressive/regressive oscillations in general, we confine ourselves in this note to the relatively well-documented situation at the Hertenrits.

The Hertenrits (Fig. 2) is a raised prehistoric Amerindian settlement in the vicinity of Wageningen in western Suriname. According to BRINKMAN & PONS (1968) and VAN DER HEIDE (1973) this artificial mound is built on top of sediments of the Moleson phase, and partly embedded in marine sediments of the Comowine phase. Radiocarbon dates on organic material from the mound give ages of 1265 ± 60 (GrN 1897), 1130 ± 45 (GrN 1898) and 1045 ± 60 (GrN 845); see VOGEL & WATERBOLK, 1964). According to LAEYENDECKER-ROOSENBERG (1966) the older date of 1265 y B.P. must, on the basis of palynological evidence, be considered representative of the start of a transgressive phase, and in view of the descriptions by BRINKMAN & PONS (1968) and VAN DER HEIDE (1973) this must be the Comowine phase.

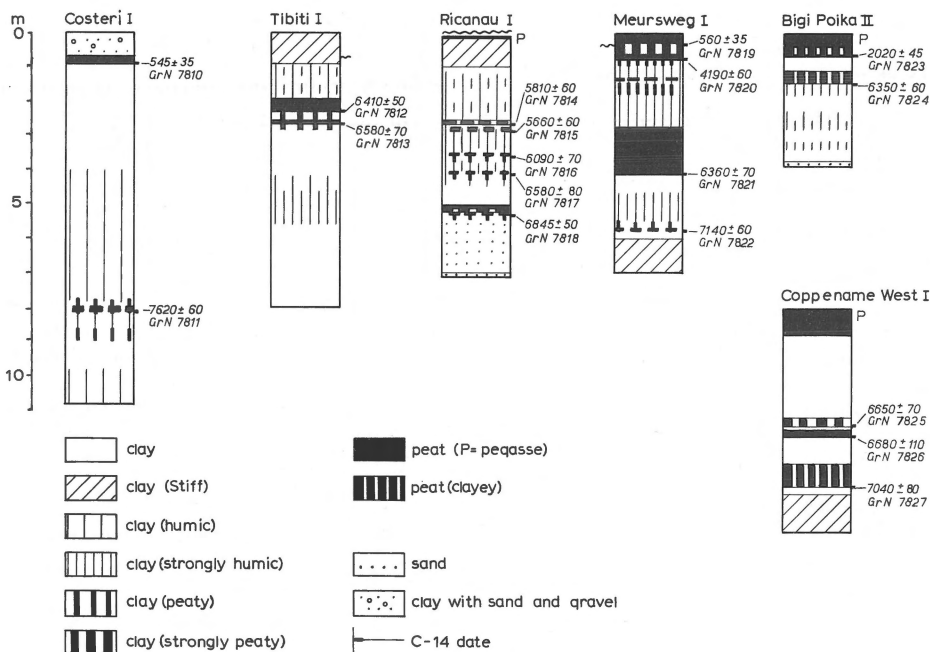


Fig. 5
Radiocarbon dated sections; the section Burnside, which is omitted, consists of a vegetation horizon at a depth of 40-45 cm in a clayey profile.

BRINKMAN & PONS (1968) interpret the radiocarbon dates differently. They assume that the construction of the mound had been largely completed at the beginning of the Comowine phase, and they date the onset of the Comowine phase correspondingly as shortly after the younger date of 1045 B.P. The date of 1265 ± 60 (GrN 1897) is regarded by Brinkman & Pons as a terminus ante quem for the ending of the Moleson depositional phase. Based on this evidence and supported by datings of shell material from ridges in the Saramacca district (DELANEY, 1966) which these authors consider to form part of the Moleson deposits, Brinkman & Pons date the end of the Moleson phase at around 1300 B.P.

VAN DER HEIDE (1973) considers it probable that the older date of 1265 B.P. is representative of the initial phase of raising of the mound. In his opinion the initial inhabitation on the natural marsh surface at the spot may be considerably older. According to Van der Heide, the date of 1265 B.P. represents the onset on the Comowine phase, and this agrees with LAEYENDECKER-ROSENBERG's (1966) view.

Our own field observations make it probable that the vegetation horizon which was dated at Burnside correlates to the regression surface situated at the base of the Hertenrits which is considered to form the top of the Moleson Deposits. The date of (at least) 2150 ± 50 y B.P. (GrN 7828) at Burnside (Table II) indicates a much earlier termination of the Moleson phase than is surmised by Brinkman & Pons. This supports VAN DER HEIDEN's (1973) assumption that the initial inhabitation at the Hertenrits site may be considerably earlier than the oldest radiocarbon date of 1265 B.P.

In our opinion the evidence points towards a non-marine

phase in the Hertenrits area between c. 2150 and 1250 B.P. During this phase fresh-water sedimentation and peat formation occurred in some places (LAEYENDECKER-ROSENBERG, 1966), whereas in other places a relatively well-drained marsh existed with prevailing non-deposition (Burnside).

At the same time (about 1850-1500 B.P.: DELANEY, 1966) chenier ridges were formed in the Saramacca area (cf. Table II), and these ridges are included by BRINKMAN & PONS (1968) in the Moleson phase. This fact illustrates that Brinkman & Pons combine data from quite different environments in their definition and subdivision of depositional phases. Obviously, this procedure seriously affects the usefulness of their classification, notably where the significance of marine oscillations is concerned.

The synchronous occurrence of a fresh-water phase in the Hertenrits area on one hand and ridge formation in the Saramacca district on the other hand might be considered coincidental. We suggest, however, that a causal relationship between the two may well exist. Extensive ridge formation will significantly reduce marine influence in the coastal swamps, which may become completely fresh, especially when sufficiently supplied with river water. So it may be assumed that during the fresh-water phase at the Hertenrits (2150-1250 B.P.) the area was protected from the sea by a ridge system situated seaward of the present-day coast line, and that the Maratakka-Nickerie river system supplied large quantities of fresh water to the coastal swamp. The formation of such ridges would roughly coincide in time with the ridge formation in the Saramacca district mentioned above. It is noteworthy that the ridge complex to the north and northwest of

Table III
Schematic representation of the variations of marine activity during the younger Holocene as suggested by the present ^{14}C -dates.

Phase (Brinkman & Pons, 1968)	character	age
COMOWINE	decreasing marine influence in the interior; ridge formation	550 B.P.
	increasing marine influence in the interior	
MOLESON	decreasing marine influence in the interior; ridge formation	1250 B.P.
	strong marine influence	1800-2100 B.P.

Paramaribo (Fig. 1) probably dates from the same period; although BRINKMAN & PONS (1968) ascribe this ridge complex to the (younger) Comowine phase, a recent radiocarbon date on shells from underneath the prehistoric Amerindian settlement at the Tingiholo ridge (Fig. 2) gives an age of 1715 ± 25 B.P. (GrN 8302). Thus, it seems that extensive ridge formation occurred over a large area at the same period. Possibly the transition from clastic sedimentation to peat growth at around 2000 B.P. at Bigi Poika is also related to this phenomenon.

The renewed marine influence at the Hertenrits (Comowine phase) since about 1250 B.P. may be the result of (partial) erosion of the previously formed protective ridges. It seems plausible that in other places in the coastal area marine influence also increased at this time. The renewed peat formation at around 550 B.P. (Meursweg and Costeri) may reflect the building up of a new system of (Comowine) ridges along the coast from about that time.

The suggested development is schematically presented in Table III. Due to lack of data we deem extension of the scheme to the earlier post-Mara phases of the development to be too speculative.

ACKNOWLEDGEMENTS

The field work of the authors in Suriname was made possible through the financial support of the Netherlands Organization for the Advancement of Pure Research (Z.W.O.).

We wish to express our gratitude to the Directors of the Geologisch Mijnbouwkundige Dienst (Geological and Mining Service) and the Dienst Bodemkartering (Soil Survey) at Paramaribo for kindly putting all technical support at our disposal, and to Ir. P. J. Melitz and Drs. R. V. van Lissa for their personal assistance.

Prof. Dr. L. J. Pons and Ir. B. Goedendorp gave us va-

uable information on the geological and pedological conditions in the study area. Drs. A. H. Versteeg of the Surinaams Museum collected the C-14 shell sample at Tingiholo.

All radiocarbon dates were performed in the Groningen C-14 Laboratory, under the supervision of Dr. W. G. Mook.

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