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QUATERNARY EVOLUTION OF THE CENTRAL PART OF THE BRASILIAN COAST

THE ROLE OF RELATIVE SEA-LEVEL VARIATION AND OF SHORELINE DRIFT

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ABSTRACT

The central part of the Brasilian coastline experienced considerable relative sea-level fluctuation during the Quaternary. It has been possible to identify three high marine levels. The last two of these, during which the sea-level was at a maximum 8 $\pm$ 2 m (120,000 years B.P.) and 4.8 $\pm$ 0.5 m (5,100 years B.P.) above the current level, have left substantial records, whose identification was possible due to numerous absolute datings. The fact that this coast was submerged until about 5,100 years B.P., when it emerged above water, is crucial for an understanding of Holocene littoral sedimentation mechanisms. In fact, starting in 5,100 years B.P., relative sea-level decline supplied large amounts of sand from the nearby platform. Deposited on the beach, these sands were taken up by littoral drift and moved on until they encountered an obstacle or a trap that would allow them to accumulate. It seems quite clear that waterways played an important role as obstacles (damming littoral transport), but only a secondary one in the supplying of sand. This accounts for the existence of progradation zones, whether linked or not to the mouth of a river.

1. INTRODUCTION

Until recent years, the old shorelines of so-called stable regions (for example, Brasil) were considered to be records of the world's ocean level. One of the goals of the "Sea-level" project (PICG N° 61, 1974-1982) was to determine a worldwide eustatic curve for the Holocene. However, field studies conducted all over the planet very quickly showed that this was not a realistic undertaking, and all the specialists now accept that it is not possible to define a general curve, but only local or regional curves. Thus it is evident that so-called eustatic curves, such as that of Fairbridge (1961), cannot be used as models of relative sea-level variation over recent millennia.

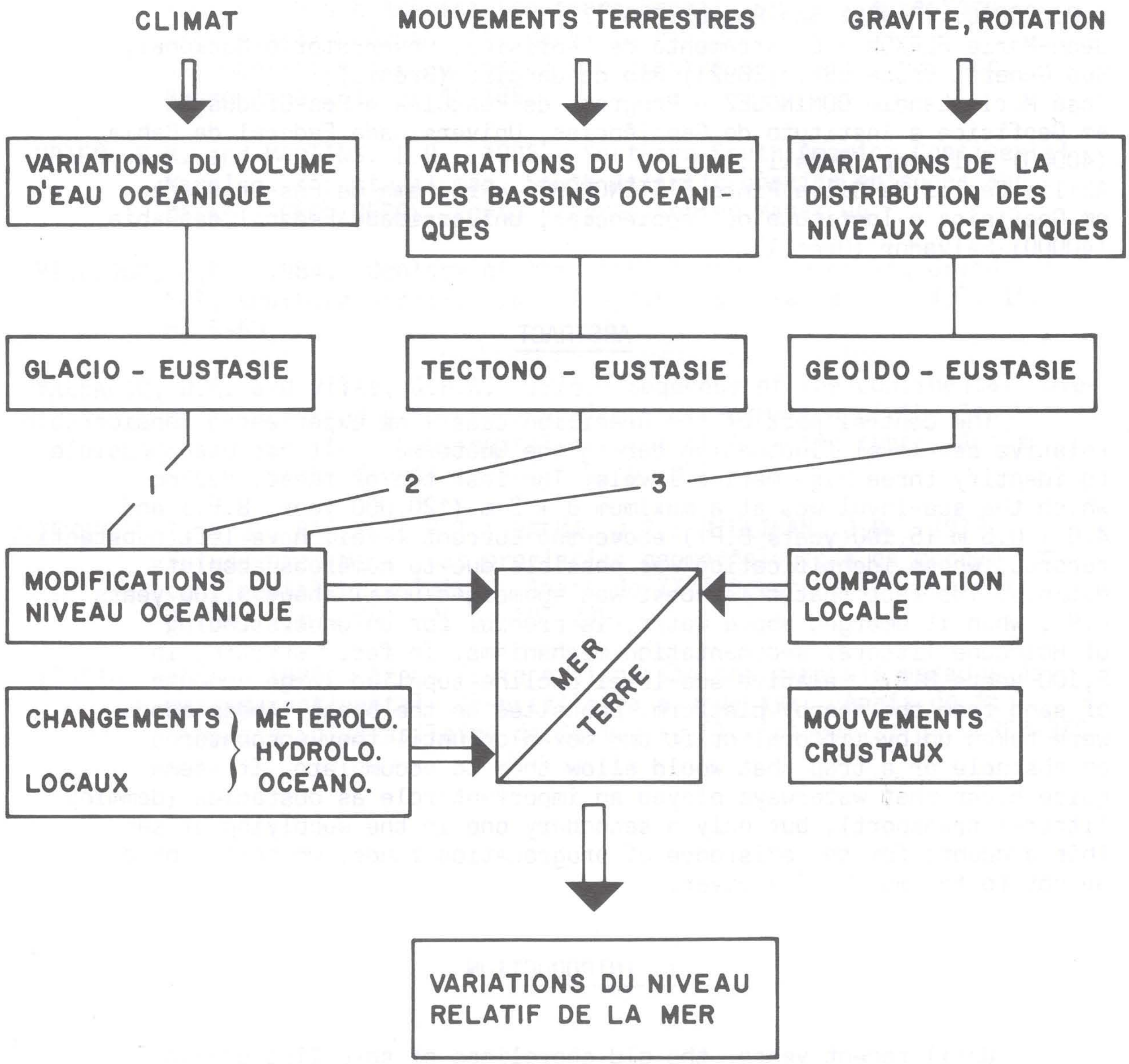


Fig. 1 - Complexity of mechanisms governing the sea-level (according to MORNER, 1980)

It is just as evident that, during the same age, coastal zones can be stable, submerged or emerged. Consequently there can be no general model of coastal sedimentation.

1.1. THE COMPLEXITY OF FACTORS GOVERNING RELATIVE SEA-LEVEL VARIATION

Relative sea-level fluctuations are the result of actual variation of the sea-level (eustatics) and of changes in the level of continents (tectonics and isostatics), as illustrated in the diagram of Figure 1. Obviously, then, whenever we reconstruct an old position of the sea--level, we are talking about a relative level.

Changes in the level of continents are controlled by:

a) tectonic movements, whether horizontal or vertical, which affect the earth's crust by mechanisms with a time scale ranging anywhere from the long-term to the instantaneous (seismic movements);

b) isostatic movements related to load variations connected with: the formation and disappearance of ice caps; the erosion of continents and accumulation of sediments in sedimentary basins; transgressions and regressions on continental platforms (hydro-isostatics);

c) deformations of the continental geoid (the latter constitutes our current reference).

Changes in the level of the ocean surface are also controlled by a number of factors:

a) fluctuations of the total volume of oceanic basins, as a consequence of plate tectonics (tectono-eustatics);

b) fluctuations of the volume of water in oceans, connected with glaciation and deglaciation (glacio-eustatics);

c) deformations of the ocean surface.

The height of the marine surface has an oceanic component and a geophysical component. The oceanic effects, capable of influencing the height of the sea, are mainly tides, major currents and associated whirlpools, and slope variations due to wind, pressure, water temperature and salinity. The overall effect of these does not exceed 1 to 2 m, and thus is quite small compared to the enormous marine surface hollows and humps caused by density heterogeneities within the planet. This geophysical component corresponds to the geoid, which is essentially the average sea-level. Starting in 1975, altimeters on board the GEOS 3 and SEASAT satellites made it possible to measure the position of the marine surface with extreme precision. This resulted in the demonstration of the existence of undulations, with very long wavelengths and with amplitudes of several dozen metres (up to 100 m to the South of India). Arguments based on the absence of correlation between these undulations and the surface topography on the one hand, and their amplitudes and wavelengths on the other, generally conclude by attributing them to

Fig. 2 - Orientation map

density differences in the lower mantle, or even at the core-mantle interface. At shorter wavelengths, the marine geoid exhibits a highly varied spectrum of anomalies.

The geoid surface is an equipotential surface of the gravity field, determined by the forces of rotation and gravity that affect the planet earth. These forces, and hence the geoid shape as well, vary as a function of the core and mantle composition (the origin of paleomagnetic crises) and of the relationship between the astenosphere and the lithosphere, but also as a function of several orbital phenomena and their interactions. It seems that geoid surface modifications can occur rapidly: MORNER (1984) quotes rates of 10 mm/yr, with gradients of several metres per km. A one milligal change in the force of gravity can deform the ocean surface by 3.3 m, and the surface of the earth's crust by 1.7 m.

Thus the ocean level at a given point on the coast is the instantaneous product of complex interactions between the surfaces of the ocean and the continent. Fluctuations of the volume of ocean basins (tectono--eustatics) and variations of the volume of oceans (glacio-eustatics) exert their effects on a global scale. On the other hand, geoid surface area changes (geoido-eustatics) and continental level changes exert their influence on a local or regional scale.

Therefore, it is quite logical that there exist inconsistencies among reconstructions of the sea-level position in the same age but at different points of the globe. This is particularly noticeable over the last 7,000 years. Indeed, before 7,000 years B.P., the rate of glacio-eustatic rise was too fast, masking components due to local or regional factors.

1.2. RECONSTRUCTION OF OLD RELATIVE SEA-LEVEL POSITIONS

THE VARIATION CURVE

To reconstruct an old relative sea-level position, it is necessary to define a marker for it in space and in time. In order to define the position of this marker in space, it is necessary to know its present altitude with respect to its original altitude; that is, to know its position with respect to sea-level at the time of its formation or deposit. In order to define the marker in time, it is necessary to know the age of its formation or deposit (isotopic, archaeological or other dating methods). A marker thus defined gives a relative position of the sea-level at a certain age. If we manage to establish a sufficiently large number of old relative sea-level positions, satisfactorily covering a certain period of time, we can then plot a variation curve for this period. It is quite obvious that only information originating from one coastal sector, where the local phenomena are always the same, can be utilised. Hence we are often confronted with the following dilemma: a) to construct a curve based on a large number of reconstructions covering the time period in question, but this often involves using data from a relatively large coastal sector, with the risk that local factors may not be the same throughout the sector; b) to consider only a restricted sector of the coast, but in this case the number of reconstructions may be insufficient to yield a precise or complete curve.

2. QUATERNARY VARIATIONS OF THE RELATIVE SEA-LEVEL ALONG THE CENTRAL PART OF THE BRASILIAN COAST

2.1. HISTORY

The important role played by relative sea-level variations in the evolution of Brasilian coastal plains was observed very early. Quite a number of authors (HARTT, 1870; BRANNER, 1904; FREITAS, 1951; BIGARELLA, 1965) have described the record of these sea-level oscillations. The evidence, mostly morphological in nature, was considered to be Tertiary by the first to study it, but was subsequently attributed to the Quaternary. Nevertheless, until the beginning of the seventies, systematic studies of relative sea-level variation along the Brasilian coast were fairly rare (SUGUIO, 1977). Only the work of VAN ANDEL and LABORAL (1964), on high Holocene levels, was based on Carbon-14 datings. Starting in 1974, relative sea-level variations in the Quaternary were studied by a group of researchers of the University of Sào-Paulo, of the Federal University of Bahia and of the national observatory in Rio de Janeiro, in conjunction with ORSTOM (French Research Institute for Co-operative Development). This team completed studies of the coastal Quaternary of the State of Sào-Paulo and the southern half of that of Rio de Janeiro (MARTIN and SUGUIO, 1975, 1976 a and b, 1978; SUGUIO and MARTIN, 1976, 1978 a and b, 1980 a and b; MARTIN et al., 1979 a and b, 1980 a), of the States of Bahia, Sergipe and Alagoas (BITTENCOURT et al., 1979 a and b; MARTIN et al., 1978, 1979 b, 1980 a and b, 1982; VILAS-BOAS et al., 1982; DOMINGUEZ, 1982; DOMINGUEZ et al., 1982, 1983 a and b), of the northern half of the coast of the State of Espirito--Santo (SUGUIO et al., 1982), and of the northern part of the coast of the State of Rio de Janeiro (MARTIN et al., 1984). Similarly, research work was carried out in the States of Parana and Santa-Catarina (SUGUIO et al., 1986; MARTIN and SUGUIO, 1986). Moreover, the same team systematically studied sedimentary deposits at the mouths of the Paraìba do Sul, Doce, Jequitinhonha, Sào-Francisco and Parnaìba Rivers, with a view to determining the role played by relative sea-level variation and by littoral drift in the construction of these coastal plains.

2.2. EVIDENCE FOR QUATERNARY MARINE LEVELS ALONG THE CENTRAL PART OF THE BRASILIAN COAST

Sedimentary evidence

Littoral sandy formations, whose peaks are very much above the present equivalent deposit zone, provide indisputable evidence for Quaternary old marine levels positioned above the present sea-level.

Thanks to some detailed cartography, together with absolute datings, it has been possible to distinguish between two principal generations of sandy terraces, recording two periods of high Quaternary marine levels. The nature of the sedimentary structures found in these terraces makes it possible spatially to reconstruct the sea-level position with fairly good precision.

Biological evidence

Along almost all the rocky part of the Brasilian coast, there is biological evidence for old marine levels higher than the present level. It generally consists of deposits of limpets (gastropoda), oysters and of sea-urchin holes, situated above the living zone of these organisms. Since the distribution zone of limpets is very narrow (0.5 m), the presence of a fossil deposit of these organisms makes it possible to reconstruct an old sea-level position with good precision. All along the north-eastern part of the Brasilian coast, there are numerous dead reefs made up of calcareous algae and coral. The position of these reefs' peaks provides evidence for old sea-levels higher than the present level. Finally, in the sandy terraces, one can find fossilised burrows of Callichirus major (marine arthropode) above the present living zone of these organisms.

Archeological evidence

Numerous "sambaquis" (artificial accumulations of shells), built by ancient inhabitants of the coastal zones, can be found in various parts of the Brasilian coast. The position of some of these "sambaquis" can be explained only by a lagoon extension significantly greater than the present one, and hence a higher than present marine level. Moreover, the "sambaquis" whose bases are below the present high tide level could have been built only at a time when the sea-level was lower than at present.

2.3. HIGH QUATERNARY MARINE LEVELS ALONG THE CENTRAL BRASILIAN COAST

The oldest known high marine level on the Brasilian coast has been demonstrated only on the coast of the states of Bahia and Sergipe. It is known by the name of Old Transgression (BITTENCOURT et al., 1979). This event is not well defined, because there are no outcrop deposits that can be associated with it with certainty. The only record we know of its existence consists of cliffs cut through continental Pliocene sediments of the Barreiras Formation, and probably of a non-outcrop reef formation in the south of the State of Bahia (CARVALHO and GARRIDO, 1966). The peak of this formation was found on the islet of Coroa Vermelha, 11 metres below present sea-level.

High marine level at 120,000 years

The Old Transgression was followed by a new transgressive phase, during which the relative sea-level, about 120,000 years B.P., was

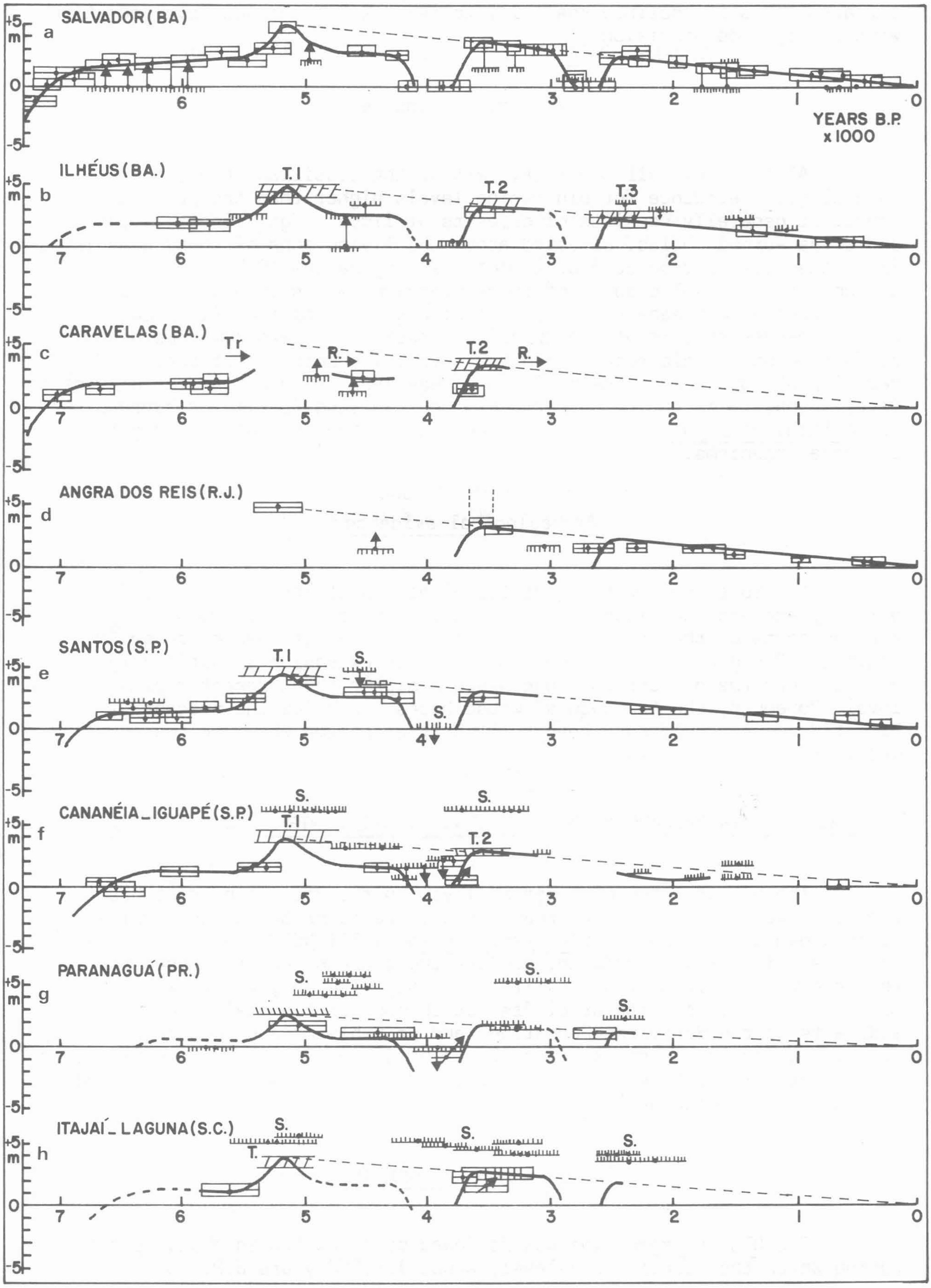


Fig. 3 - Relative sea-level variation curves for the last 7,000 years along several sectors of the Brasilian coast

at 8 ± 2 m above the present level. The age of 120,000 years was established by means of Io/U dating of 5 samples of coral (MARTIN et al., 1982; BERNAT et al., 1983). This transgression is known as the Cananeia Transgression along the Sào-Paulo coast (SUGUIO and MARTIN, 1978), and as the Second-Last Transgression on the coasts of the States of Bahia, Sergipe and Alagoas (BITTENCOURT et al., 1979). The record of this high level consists essentially of substantial sandy terraces, extending more or less continuously throughout the region in question. On the basis of sedimentary structures and fossilised burrows of Callichirus major, it is possible to reconstruct the spatial sea-level position. However, in the absence of datings (with the exception of the five Io/U determinations), it is not possible to do a temporal reconstruction, or to establish relative sea-levels in the vicinity of 120,000 years B.P., and to compare altitudes of the same age at different points of the coast.

High marine level in the Holocene

The most recent high marine level is very well known, thanks to a very large number of determinations of old relative sea-level positions in time and space, carried out on the basis of more than 700 Carbon-14 datings. Moreover, the position of a number of "sambaquis", together with dating of the shells they contain, and with the delta 13C(PDB) values of carbonates of these same shells, have given us interesting complementary information about relative sea-level oscillations over the last 5,500 years (MARTIN et al., 1985). On the basis of all these data, it has been possible to plot or at least to sketch relative sea-level variation curves for several sectors of the Brasilian coast. In order to obtain homogeneous curves, we have used data originating from littoral sectors of limited dimensions and exhibiting uniform geological characteristics.

2.4. RELATIVE SEA-LEVEL VARIATION OVER THE LAST 7,000 YEARS ALONG THE CENTRAL PART OF THE BRASILIAN COAST (Fig. 3)

Sector to the North of Salvador (Bahia)

In this sector of some fifty km length, about sixty determinations of old relative sea-level positions, covering the last 7,000 years very regularly, have been made. These data have made it possible to plot a very precise curve which shows that:

- the present zero (mean level) was exceeded for the first time in the Holocene at about 7,100 years B.P.;

- about 5,100 years B.P., the relative sea-level went through a first maximum of 4.8 ± 0.5 m above the present level;

- after this maximum, there was a rapid regression until 4,900 years B.P., slowing down until 4,200 years B.P., and speeding up again until 3,900 years B.P. At about this time the marine level passed through a minimum, probably below the present level;

- between 3,900 and 3,600 years B.P., a rapid transgression occurred, and at about 3,600 years B.P., the relative sea-level passed through a second maximum of 3.5 ± 0.5 m above the present level;

- between 3,600 and 3,000 years B.P., the relative sea-level fell slowly and regularly. Starting at 3,000 years B.P., the decline became very rapid, and at about 2,800 years B.P., the relative sea--level must have been slightly below the present level;

- between 2,700 and 2,500 years B.P., the relative sea-level rose very rapidly, passing through a third maximum of 2.5 ± 0.5 m above the present level at about 2,500 years B.P.;

- since 2,500 years B.P. the relative sea-level has fallen regularly to its present position.

This very well defined curve can be used as a reference for coastal sectors, where the number of reconstructions is insufficient to allow a complete curve to be plotted. In this type of sector, it is possible to compare the available reconstructions with the Salvador curve, and to see whether or not they fit on the curve.

Sector of Ilheus (Bahia)

In this sixty km sector, the number of reconstructions of old relative sea-level positions over the last 7,000 years is insufficient for a complete curve to be drawn. However, the available determinations exhibit no shift with respect to the Salvador curve. It has been possible to demonstrate the presence of 3 sandy terraces, witnessing the existence of 3 high level periods, positioned between 5 and 4, 4 and 3, and 3 and 2 metres above the present level. It is logical to conclude that these 3 terraces correspond to the 3 maxima determined in the Salvador sector.

Sector of Caravelas (Bahia)

It was possible to do only 11 reconstructions in this thirty km sector. However, 7 of them fit into the range between 7,000 and 5,700 years B.P., so this part of the curve has been established with good precision. All the available data are in accordance with the Salvador curve.

Sector of Angra dos Reis (Rio de Janeiro)

In this sector of some 60 km, only 17 old relative sea-level positions could be reconstructed. However, the segment of the curve running from 0 to 2,500 years B.P. is fairly well defined. We also managed to obtain indications that there were two maxima, one slightly above 3 m between 3,650 and 3,450 years B.P., and the other in the vicinity of 4.8 m at about 5,200 years B.P.

Sector of Santos (Sào-Paulo)

About thirty reconstructions, resulting in a fairly complete curve, were made in this sector of about 60 km length. It is interesting to note that the present zero was exceeded for the first time in about 6,800 years B.P., that is, considerably later than in the Salvador sector. The parallel maxima of 5,100 and 3,600 years B.P. were positioned at 4.5 $\pm$ 0.5 m and at 3.0 $\pm$ 0.5 m above the present level.

Sector of Cananeia (Sào-Paulo)

Only 10 old relative sea-level positions could be reconstructed along these approximately 100 km of coast. However, as 7 of them fell into the range between 6,650 and 5,300 years B.P., that part of the curve was drawn up satisfactorily. Moreover, "sambaquis" datings in this region, associated with delta 13C(PDB) variations of the carbonates of their shells, provided additional information. It seems that the present zero was exceeded for the first time in about 6,600 years B.P., and that the maximum of 5,150 years B.P. (whose age was established with great precision by means of the 13C(PDB) variation curve) was no higher than 4 m above the present zero.

Sector of Paranagua (Parana)

Not much good data is available from this fifty km long sector. Nevertheless, a few bits of precise information bring to light the major trends of relative sea-level variation over the last 7,000 years. For instance, in Paranagua Bay the peak of the outer part of the Pleistocene marine terrace is at 2.5 m above the present high tide level. Given that on the surface of this terrace there are traces of old Pleistocene littoral belts, it is clear that the terrace was not submerged during the Holocene, and consequently that, at the time of the maximum of 5,100 years B.P., the relative sea-level could not have been more than 2.5 m above the present level. "Sambaquis" datings provide us with interesting additional information. It seems that the Paranagua curve is comparable in shape to that of Salvador, but shifted downwards by a substantial amount.

Sector of Florianopolis (Santa-Catarina)

The information available on the Quaternary of the coast of Santa-Catarina State provides us with an understanding of the major sea-level variation trends over the last 7,000 years. The resulting curve again has the same shape as that of Salvador, and it is also shifted downwards, but certainly not as much as the Paranagua curve.

The coast of Alagoas State

It was not possible to plot a relative sea-level variation curve for the length of coast in Alagoas State, because precise reconstructions are too few in number, and because the sector in question is relatively long. However, when compared with the Salvador curve, the available data exhibit no significant differences. Thus it is reasonable to believe that relative sea-level variations along the coast of Alagoas State were much the same as in the coastal sector to the North of Salvador.

2.5. GENERAL REMARKS ABOUT THESE CURVES

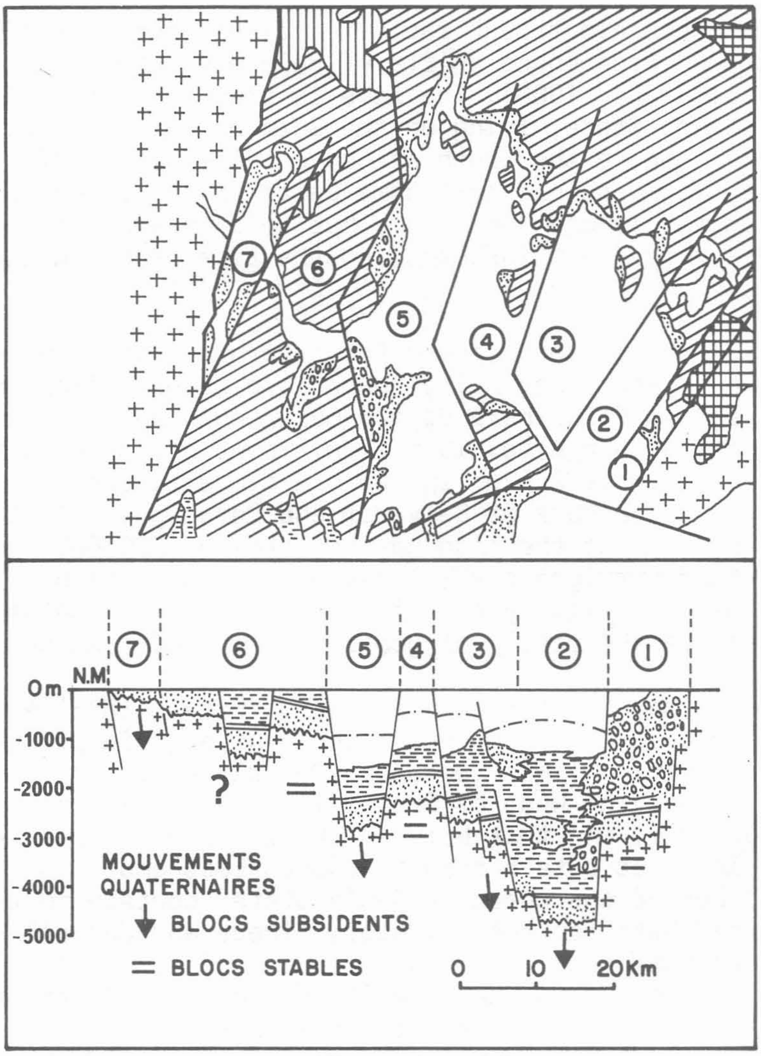
The first important fact is that, in all the sectors, the relative sea-level has been higher than the present level, with an elevation maximum always occurring at about 5,100 years B.P. Moreover, all the curves have the same general shape, but some are shifted vertically. Finally, all the sectors seem to have experienced, after 5,100 years B.P., two rapid relative sea-level oscillations of 2 to 3 metres. These are too large to be glacio-eustatic in origin.

On the Salvador curve, plotted with more precision than the others, the 17 reconstructions of old relative sea-level positions, used to establish the curve between 0 and 2,500 years B.P., fit onto a straight line segment. Moreover, 6 other reconstructions, used to establish the curve between 3,000 and 3,600 years B.P., are situated on the extension of the same straight line segment. Finally, if one extends this segment all the way to 5,100 years B.P. (age of the maximum), one obtains a relative sea-level position in the vicinity of 5.0 m above the present level. Now the experimental reconstruction yields an old relative sea-level position of 4.8 ± 0.8 m above the present level at 5,150 ± 110 years B.P. Thus we have, between 0 and 5,100 years B.P., a large number of points on the same straight line, too large, in fact, for this to be accidental. But during certain shorter periods of time, the experimental curve moves away from this straight line. It looks very much as if, since 5,100 years B.P., a first phenomenon caused a regular decline of the relative sea-level, and a second phenomenon, superimposed on the first, generated very rapid oscillations of this same sea-level.

Comparing the entire set of curves, it seems that those of Salvador, Ilheus and Caravelas are not shifted with respect to one another. On the other hand, the Anga dos Reis curve is shifted slightly downwards. This shift becomes accentuated in the curves of Santos, Cananeia and Paranagua, where it is maximal; but the Florianopolis curve, albeit shifted with respect to that of Salvador, is less so than the Paranagua curve.

2.6. NATURE OF THE PHENOMENA

In some well delineated sectors of the coast, it has been possible to demonstrate Holocene beach-line shifts, as a consequence of vertical tectonic movements. For instance, in the Bay of Todos os Santos (Bahia),



COUPE E._W. A_TRAVERS LE BASSIN DO RECONCAVO
(Filho at al ,1982)

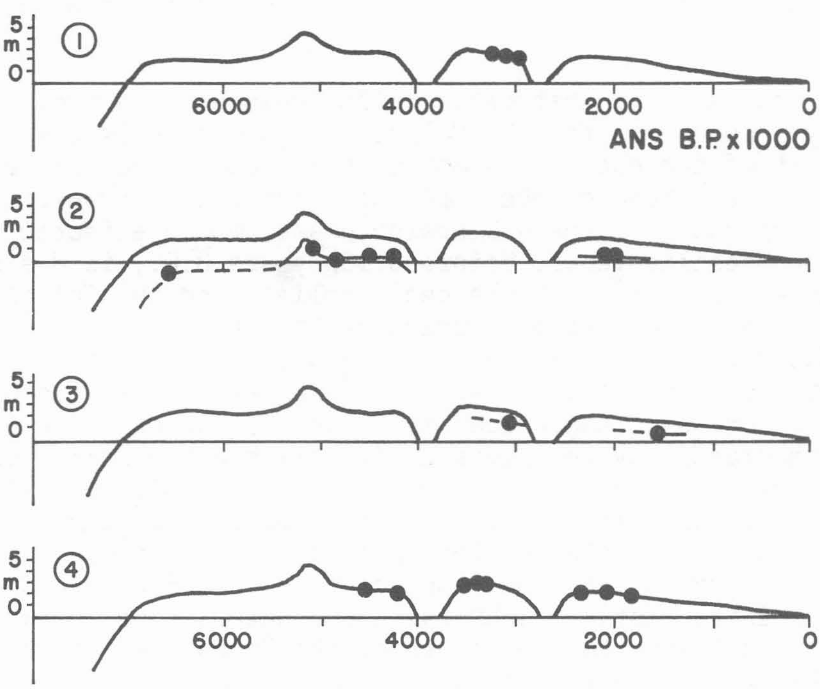


Fig. 4 - Tectonic sectioning of Reconcavo Basin. Positions of a number of reconstructions of old ocean levels in various sections of the Basin, plotted against the Salvador curve

lying on the Reconcavo trough fault, vertical movements of massifs have resulted in pronounced shifts of Holocene shorelines (Fig. 4) (MARTIN et al., 1984, 1986). The same is true for parts of the coast of Rio de Janeiro State, lying on the Guanabara trough fault (MARTIN et al., 1980), and to the South of Cap de Sào-Tomé (MARTIN et al., 1984). It is also possible that some parts of the coast were affected by a continental flexure mechanism, but this phenomenon does not seem to have had a very great influence at the time of the Holocene (MARTIN et al., 1976).

In all the sectors chosen for determination of a curve, with the exception of Angra dos Reis, there is a record of the marine terrace of 120,000 years B.P. Nowhere do the most inner parts of this terrace (of roughly the same age) exhibit significant altitude differences. If the shift of almost 2.5 m, in the maximum altitude of 5,100 years B.P., between the sectors of Salvador and Paranagua were tectonic in origin, the records of the high marine level of 120,000 years B.P. would be shifted very greatly (almost 60 metres), which is not at all the case. Thus it is likely that the shifts observed between certain sector curves are the result of geoid surface deformations.

An examination of the geoid map of Brasil (MARTIN et al., 1985) shows that the East of the country lies over a geoid protuberance, whose equal elevation lines run approximately North-South (Fig. 5). One also sees that the West of Brasil lies over another protuberance, centred on Bolivia, and that between these two protuberances there is a depression that cuts across the South-East and North coast of Brasil. The part of the coast of Bahia State, containing the sectors having provided data for the Salvador, Ilheus and Caravelas curves, and running approximately N-S, is more or less parallel to the lines of equal geoid height. On the other hand, the part of the coast containing the sectors having provided data for the Angra dos Reis, Santos, Cananeia and Paranagua curves, and running approximately NE-SW, cuts obliquely across the lines of equal geoid height. A horizontal displacement of the geoid relief, in an approximately N-S or E-W direction, would have no effect on the first three curves, but would trigger a shift of the others.

If one accepts that geoid relief changes on a regional scale are partly responsible for the Holocene high marine levels found along a large part of the coast, the shifts described above can be explained by the fact that these changes are not identical everywhere. For example, one could suppose that the submersion phase, which affected a major part of the Brasilian coast before 5,100 years B.P., is due partly to a temporary elevation of the geoid relief, and the following emersion, on the other hand, due to a lowering of the same relief. In fact, a slight displacement of the central depression's axis to the East, during the lowering of the geoid relief, could explain the shifts observed between the curves of Angra dos Reis, Santos, Cananeia and Paranagua, as shown schematically in Figure 6. If this hypothesis is correct, Holocene marine levels in the North of Brasil should be shifted with respect to corresponding levels from the Salvador region. Unfortunately, we have no numerical data for this part of the coast. Nevertheless, it is interesting to note that the coast between Sào-Luis and Belem (Fig. 2) exhibits distinct submersion characteristics: broken coast, sharp cliffs cut into the sediments of the Barreiras Formation, lower parts of waterways transformed into rias.

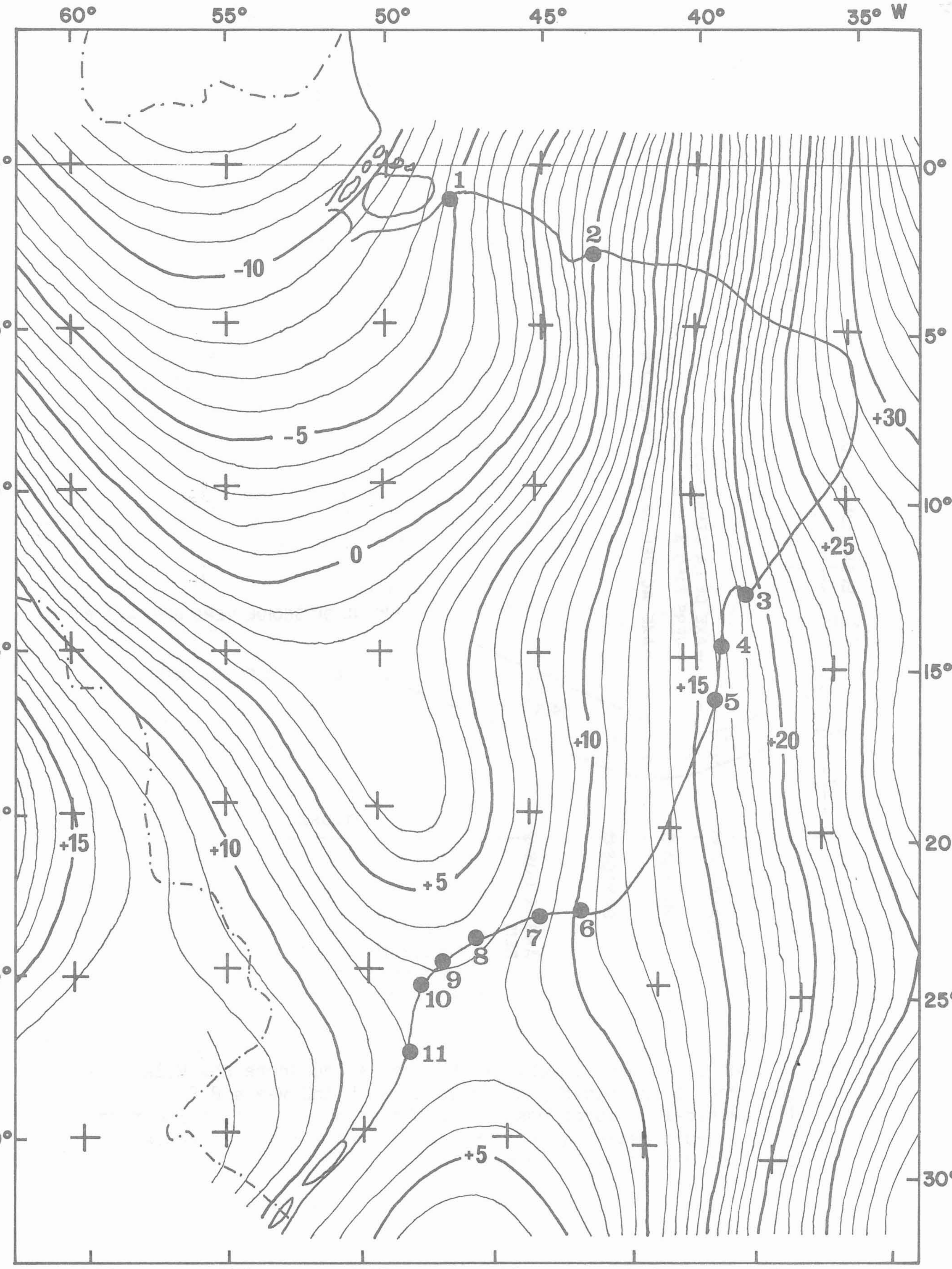


Fig. 5 - Geoid map of Brasil. 1) Belem 2) Sào Luis 3) Salvador 4) Ilheus 5) Caravelas 6) Rio de Janeiro 7) Angra dos Reis 8) Santos 9) Cananeia 10) Paranagua 11) Florianopolis

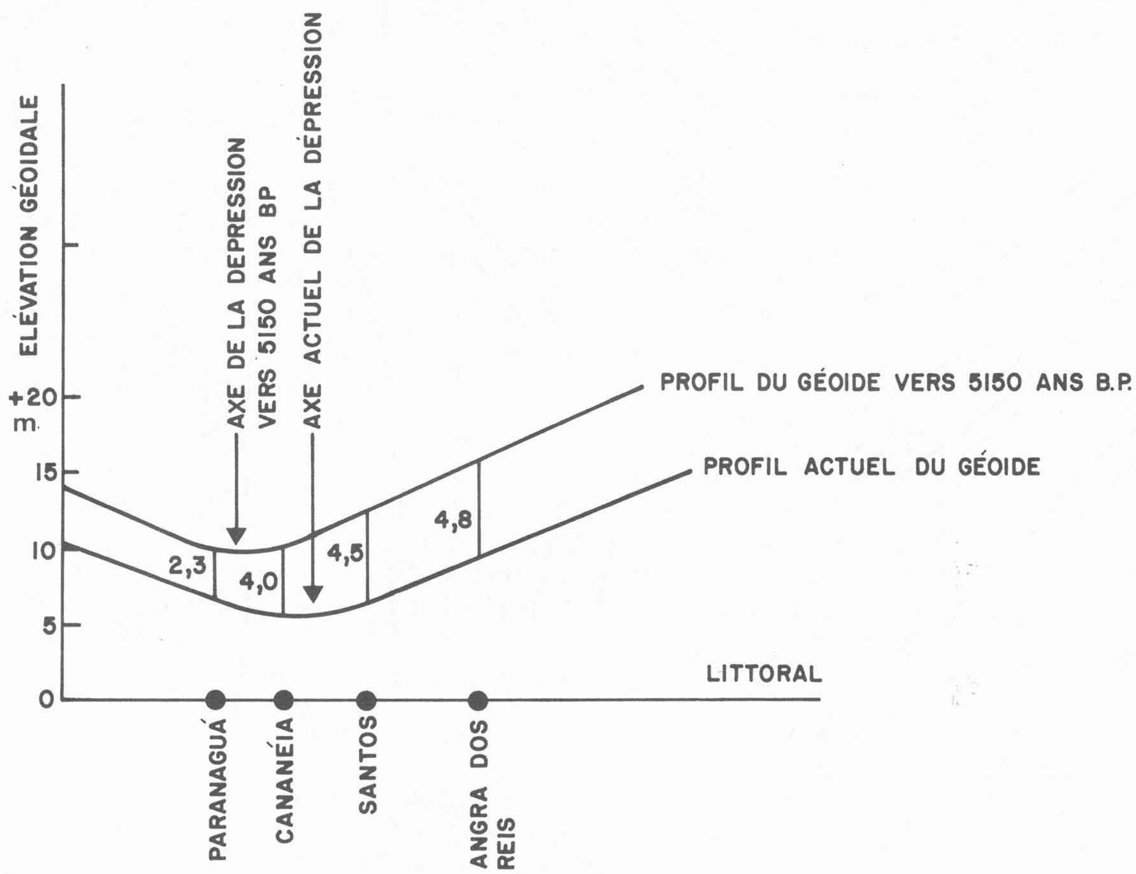


Fig. 6 - Present geoid profile between Paranagua and Angra dos Reis compared to the geoid profile in about 5,150 years B.P. The vertical dislocations can be obtained by a simple lowering of the geoid relief in conjunction with a slight horizontal displacement towards the East

To conclude, it does seem that the Holocene high marine levels of Brasil (which can be neither glacio-eustatic nor tectonic in origin) can be explained, at least in part, by a regional uplift of the overall geoid relief until about 5,100 years B.P., followed by a sinking and a minor horizontal displacement to the East. Similarly, regional sinking of the overall geoid relief, followed by an uplift on a scale of centuries, can explain the rapid oscillations that have occurred since 5,100 years B.P. (oscillations which also can be neither glacio-eustatic nor tectonic in origin).

3. CONSEQUENCES OF RELATIVE SEA-LEVEL VARIATION FOR LITTORAL SANDY SEDIMENTATION

Schematically speaking, it can be said that (whatever the causes) most of the Brasilian coast under study was submerged until about 5,100 years B.P., and, disregarding the two rapid oscillations, emerged since then. This is by no means an universal circumstance. For example, along the Atlantic Coast of the United States of America, the relative sea--level did not rise above the present level at all during the Holocene (Fig. 7). Thus it is obvious that coastal zone evolution could not have been the same in Brasil and in the United States over the last 7,000 years. Submerged coasts (the United States case) are characterised by the existence of barrier island/lagoon systems, while emerged coasts (the case of Brasil for 5,000 years) present vast sandy expanses covered by alignments of old littoral belts. A reconstruction of the situation prevailing in the Rio Doce mouth region before 5,100 years B.P. (based on Carbon-14 datings) shows a remarkable resemblance to the current appearance of Cap Hatteras in the United States (Fig. 8).

3.1. THE ROLE OF RELATIVE SEA-LEVEL VARIATION IN SANDY LITTORAL SEDIMENTATION

A sandy littoral zone has an equilibrium profile that is a function of dynamics and granulometry. The dynamics vary incessantly (tides, surge, etc.), so the profile is constantly being destroyed. However, if one considers a sufficiently long period of time, one can presume the existence of a mean equilibrium profile. It is quite obvious that a relative sea-level fall or rise destroys this equilibrium. The rule of BRUUN (1962) states that a sea-level rise destroys this equilibrium, which is then re-established by a displacement of the profile in the direction of the continent, manifested by erosion of the beach prism, and transfer of the eroded material in the direction of the outer beach. This causes the outer beach bottom to rise by a height equal to that of the sea-level rise (Fig. 9). Field and laboratory tests (SCHWARTZ, 1965, 1967; DUBOIS, 1976, 1977) have demonstrated the validity of the BRUUN rule. Even though this rule was established for the sea-level rise situation, it is logical to suppose that a sea-level fall causes destruction of the equilibrium profile, manifested by erosion of the outer beach bottom, and by transfer of the eroded sand towards the

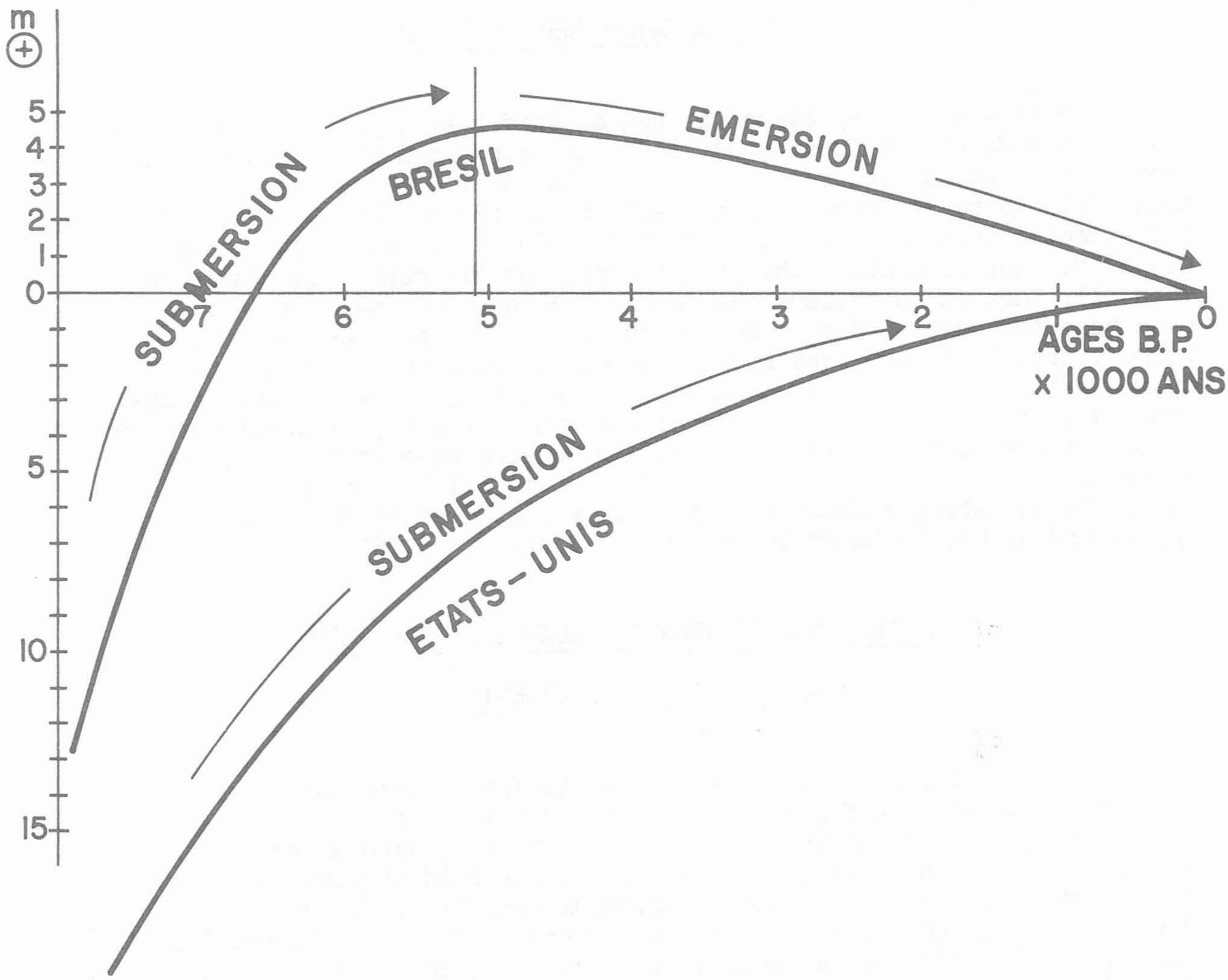


Fig. 7 - Average schematic curves of relative sea-level variation along the central part of the Brasilian coast, and along the Atlantic Coast of the United States of America, over the last 8,000 years

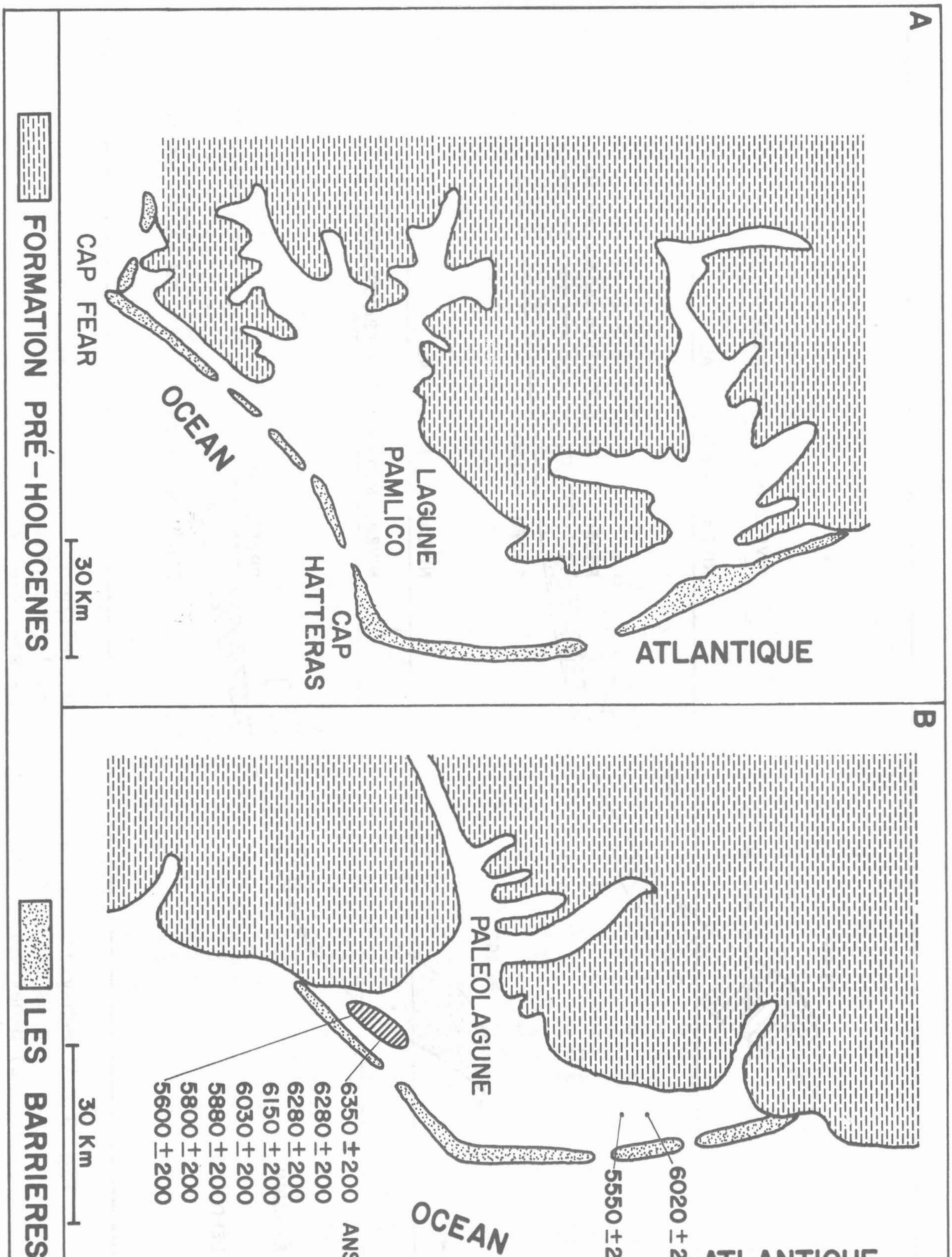


Fig. 8 - Comparison between the present situation in the Cap Hatteras region (USA) and that of the Rio Doce coastal plain before 5,100 years B.P.

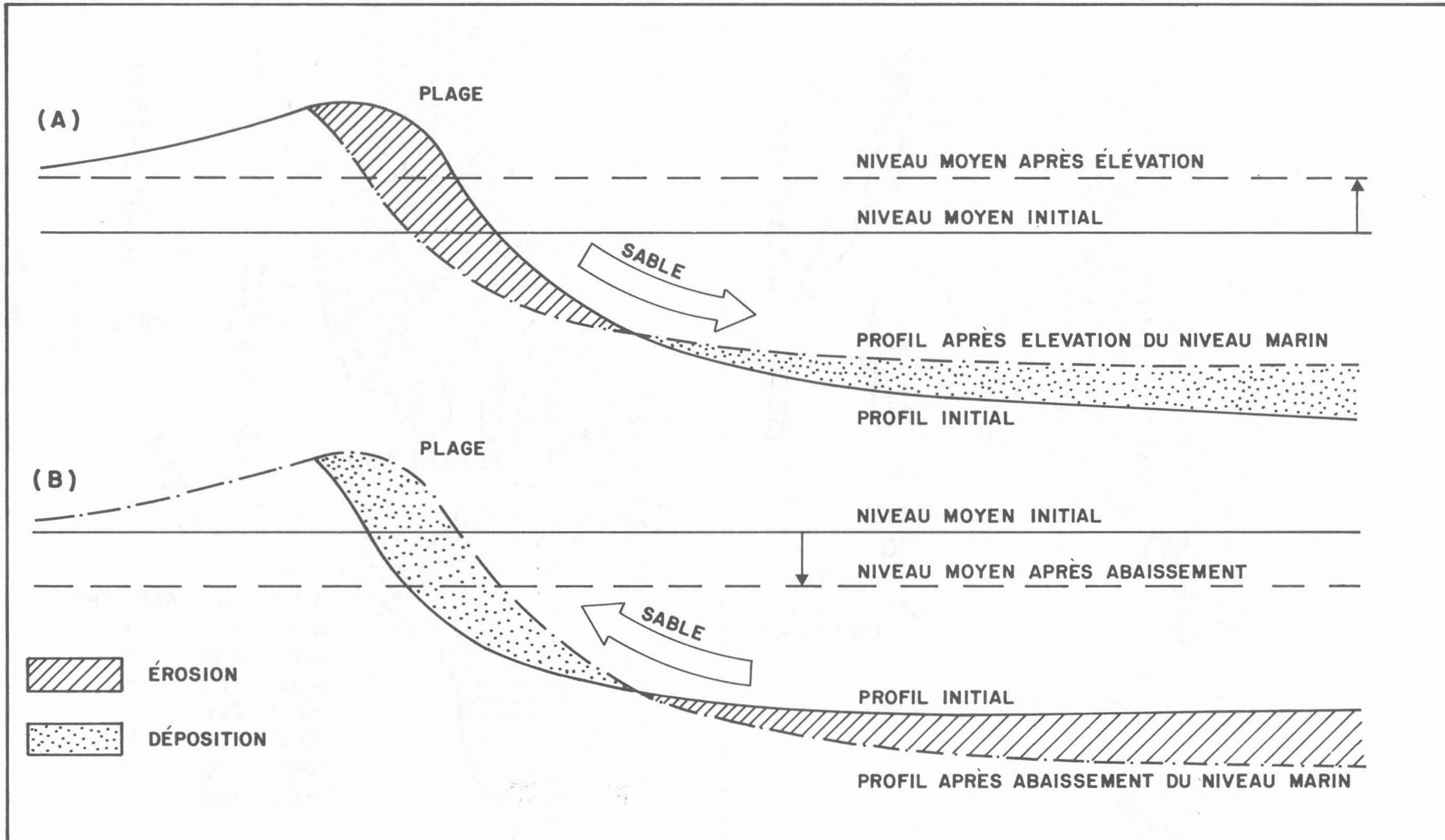


Fig. 9 - A) Behaviour of the equilibrium profile of a sandy littoral zone as resulting from a sea-level rise (BRUUN's Rule, 1962)
B) Behaviour as resulting from a sea-level fall

coast. This transfer from the outer beach towards the beach prism ends when the original depth is re-established (Fig. 9 B). This mechanism is quite similar to the one that occurs when the beach profile is re-established after a storm, by sand transfer from the outer beach to the beach prism, which is very well documented in the literature (DAVIES, 1972; KING, 1972; KOMAR, 1973; SWIFT, 1976). Moreover, this mechanism can be ideally observed during a monthly cycle of tides. During running water tides (corresponding to a small transgression), there is erosion of the beach prism and accumulation on the outer beach; on the other hand, during dead water tides (corresponding to a small regression), there is beach prism accumulation and outer beach erosion.

Hence it is clear that, on gently sloped sandy coasts, lowering of the relative sea-level generates a substantial flow of sand from the inner shelf to the beach. If the littoral transport is zero, this then results in substantial "progradation" of the coastline, by the successive adjoining of littoral belts.

3.2. THE ROLE OF LITTORAL DRIFT IN COASTAL SANDY SEDIMENTATION

As they approach the shore, waves break when the depth is not sufficient for them to continue moving. This breaking involves liberating a great amount of energy, which is taken up partly by putting sand grains into suspension, and by the formation of a current parallel to the coast, known as the longshore current. Naturally, this current appears only if the wave-fronts strike the coast obliquely. The longshore current has a slow speed, but because its action takes place in the zone where sand grains have been suspended by breaking, the volume of sand transported can be substantial. In addition, the splashes of water from breaking waves generate saw-tooth pulses of sand transport (shore casting). Obviously, the direction of the transport is a function of the direction at which the wave-fronts reach the shoreline.

It is clear, then, that during a period of relative sea-level decline, part of the sand brought in to the beach prism will be moved along the beach by the longshore current. This transport continues until the sand is taken up in a trap or blocked by an obstacle. This explains the great differences that can exist between two regions having been subjected to an equivalent lowering of the relative sea-level. Sandy deposits are small or even absent in transit regions, and very large in regions where a trap or an obstacle causes sand accumulation. Such traps or obstacles come in different forms: recesses of the coastline, islands or shallow bottoms forming low-energy zones, rocky points, mouths of waterways, etc.

3.3. BLOCKING OF LITTORAL SEDIMENTARY TRANSPORT BY THE FLOW OF A WATERWAY

Under certain conditions, the flow of water from a river-mouth can constitute an obstacle tending to block sand transport, in the manner of an artificial groyne built on a beach. These well-anchored

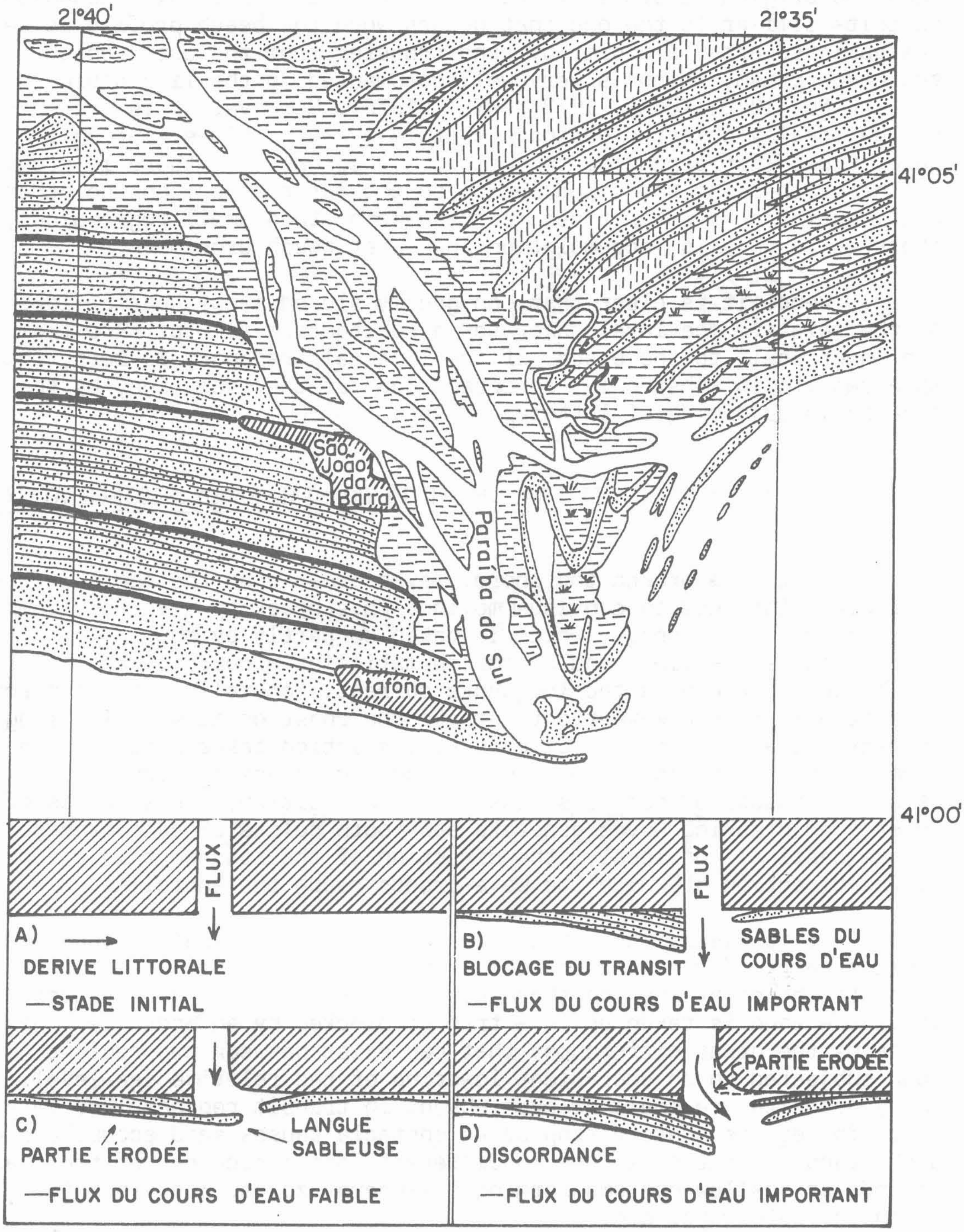


Fig. 10 - Schematic diagram of littoral transport blocking by the flow of a waterway. Map of the Rio Paraìba do Sul mouth illustrating this mechanism

works are generally built so as to extend beyond the wave breaking zone. They completely interrupt the littoral transport of sandy sediment, which leads to an accumulation on the longshore current side, so that the coastline advances there. On the other hand, on the side of the groyne lying below the longshore current, this current continues to remove sand, which leads to erosion and coastline retreat.

When there is a dominant littoral drift, the mechanisms operating at the mouth of a waterway can be described schematically as follows (Fig. 10):

a) in a period of high river discharge, the flow of water near the mouth constitutes an obstacle tending to block the transport of sand by littoral drift. This results in sand accumulation on the longshore current side, with possible erosion on the side below the current. However, this erosion is usually compensated by deposits of the waterway itself (Fig. 10 B).

b) in a period of low discharge, the obstacle formed by the waterway's flow tends to disappear, and the littoral drift causes partial erosion of the deposit formed during the preceding period. A neck of sand, tending to close off the mouth, is then formed. This mechanism is recorded in the coastal plain by lines of discordance in the alignment of littoral belts (Fig. 10 C). If the low-energy period lasts long enough, the neck of sand achieves a width that allows it to resist the following high-energy period. In some cases, only its tip is destroyed, and the dam set up by the waterway's flow is displaced in the direction of the current, with the beginning of another accumulation (Fig. 10 D). This displacement is marked by a series of stages, underscored by the discordances in the alignment of belts.

As a consequence of the groyne effect exerted by the flow of a waterway opening onto a sandy coast, one observes distinct dissymetry between the parts of the coastal plain on the two sides of the mouth. Whereas the coastline on the drift current side advances because of sand deposition by the littoral drift, the coastline below the longshore current advances mainly because of sand deposited by the waterway. The latter advance is generally less pronounced, and takes place starting from sandy points anchored on the side below the drift current, or by reshaping of mouth banks, which evolve into "moon islands" as a result of wave action. Once these sandy points and islands are constructed, protected zones form behind them, and are quickly colonised by mangroves. Thus the coastal plain on the side of the mouth that is in the drift current will consist of a fairly regular piling up of littoral belts, whose sand is not deposited by the waterway. On the other hand, the coastal plain area below the drift current will be formed by an alternation of clayey-sandy low zones and sandy zones, whose material is partly brought in by the waterway.

In order to check whether this scheme is well founded, we carried out a study of the degree of rounding of beach sand on both sides of the mouth of the Rio Paraìba do Sul (Rio de Janeiro), which is a virtually perfect model of the mechanisms we have just described (Fig. 10). In the region of the Rio Paraìba do Sul mouth, there are two types of surge. The first, coming from the S-SE and especially frequent in autumn and in winter, is related to the penetration of polar air masses over

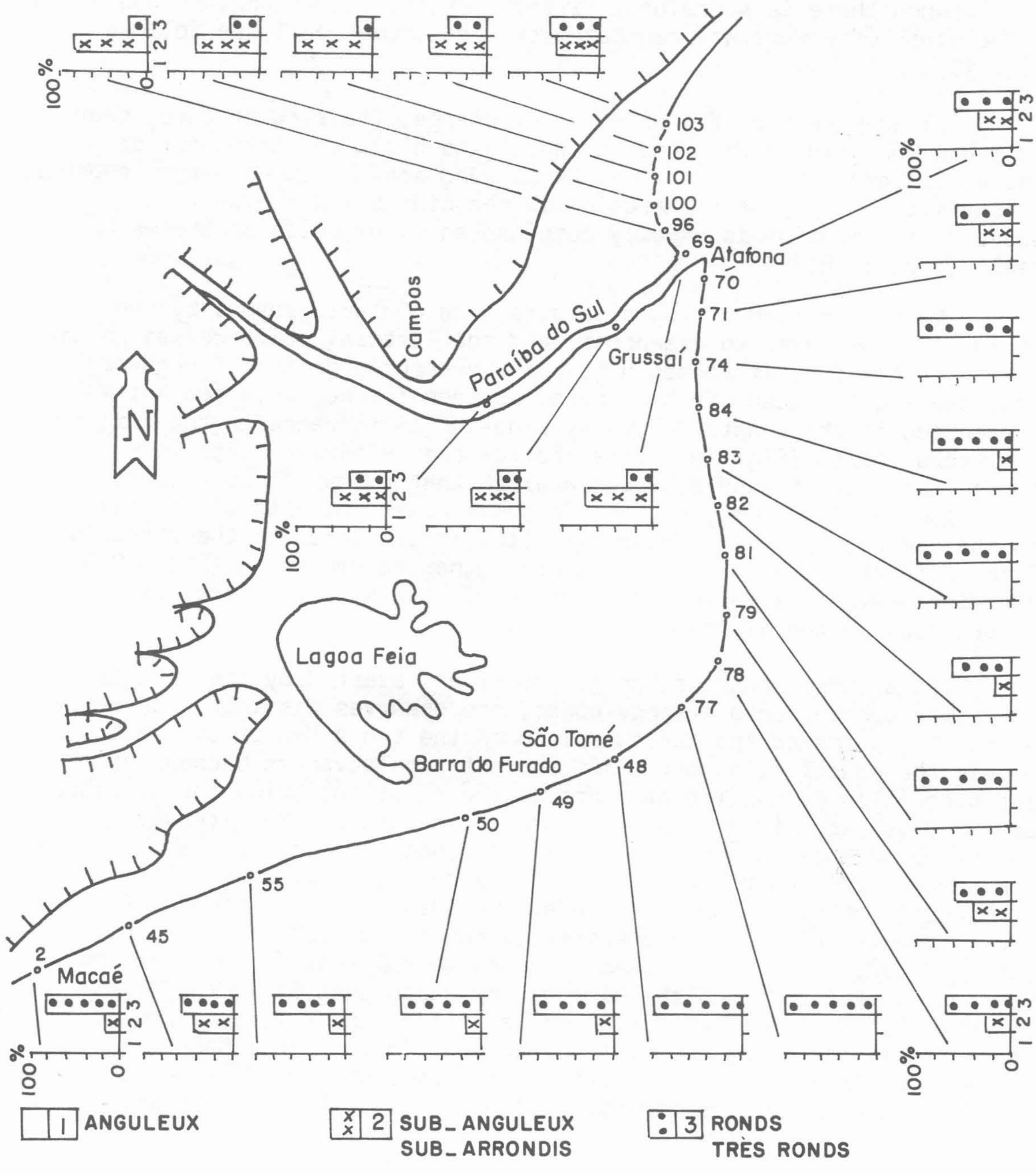


Fig. 11 - Histograms of the degree of rounding of sand grains from the present beach of the Paraìba do Sul coastal plain on both sides of the mouth, and from the present river bed

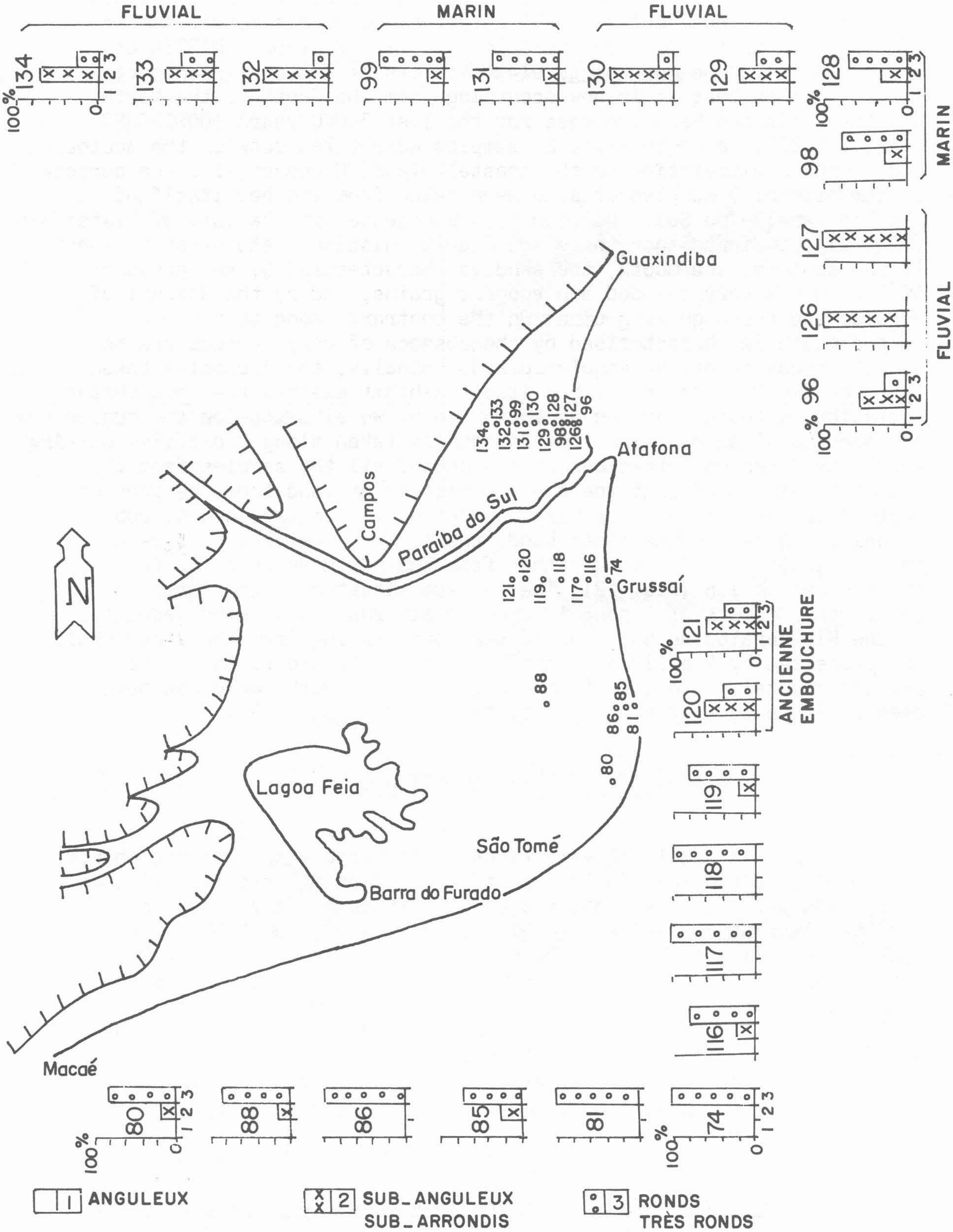


Fig. 12 - Histograms of the degree of rounding of sand grains from Holocene marine terraces on both sides of the Rio Paraìba do Sul mouth

the South American continent. The second, NE in direction, is associated with the tradewinds. But the S-SE surge, being much stronger than the NE surge, plays the dominant role in littoral transport (MARTIN et al., 1984). All the morphological indicators of littoral transport direction show that it is now occurring from the South to the North, and that this has been the case for the last 5,000 years (DOMINGUEZ et al., 1983). To begin with, 21 samples were taken between the southern and northern extremities of the coastal plain. Moreover, for the purpose of comparison, 3 samples of sand were taken from the bed itself of the Rio Paraìba do Sul. The results, represented in the form of histograms (Fig. 11), indicate that there are 2 quite distinct categories of sand. To the south of the mouth, the sand is characterised by the presence of 20% to 60% very rounded and rounded grains, and by the absence of angular and sub-angular grains. On the contrary, sand to the north of the mouth is characterised by the absence of very rounded grains and the presence of sub-angular grains. Finally, the 3 samples taken from the Rio Paraìba do Sul bed itself exhibit exactly the same characteristics as those from the northern beach. We also studied the degree of rounding of sand grains from 24 samples taken along 2 profiles cutting across terraces on both sides of the mouth. All the samples from the southern terrace exhibit the characteristics of sand from the present southern beach: presence of very rounded grains and absence of sub-angular grains. On the other hand, the northern terrace samples exhibit the characteristics of sand either from the northern beach or from the southern beach (Fig. 12). The terrace situated to the north of the mouth consists of a roughly regular alternation of sand deposited by the Rio Paraìba do Sul, and of sand originating from the inner shelf and deposited by the littoral drift. It is quite clear, then, that all the coastal plain area to the south of the mouth could not have been built up by sand deposited by the Rio Paraìba do Sul.

4. MAIN STAGES OF THE FORMATION OF COASTAL PLAINS

Relative sea-level variations, associated with climatic changes, have been the principal factors in the formation of Brasilian littoral plains. An evolutionary model has been established fairly precisely for the coast of Bahia State (MARTIN et al., 1980; DOMINGUEZ et al., 1982) (Fig. 13). This model remains valid for the entire coast lying between Macaé (Rio de Janeiro) and Recife (Fig. 2) (all this region is characterised by the presence of Barreiras Formation continental sediments). On the other hand, in the southern half of the coast of Sào-Paulo State, and along the coastline of the States of Santa Catarina and Parana, this model is only partly applicable for local reasons.

4.1. THE GENERAL MODEL

Stage 1: Deposits of Barreiras Formation continental sediments

During the Pliocene, the climate must have been hot and humid for a long period of time, which resulted in the formation of a very thick alteration mantle. At the end of the Pliocene, as the climate became drier (typically semi-arid with infrequent but driving rains),

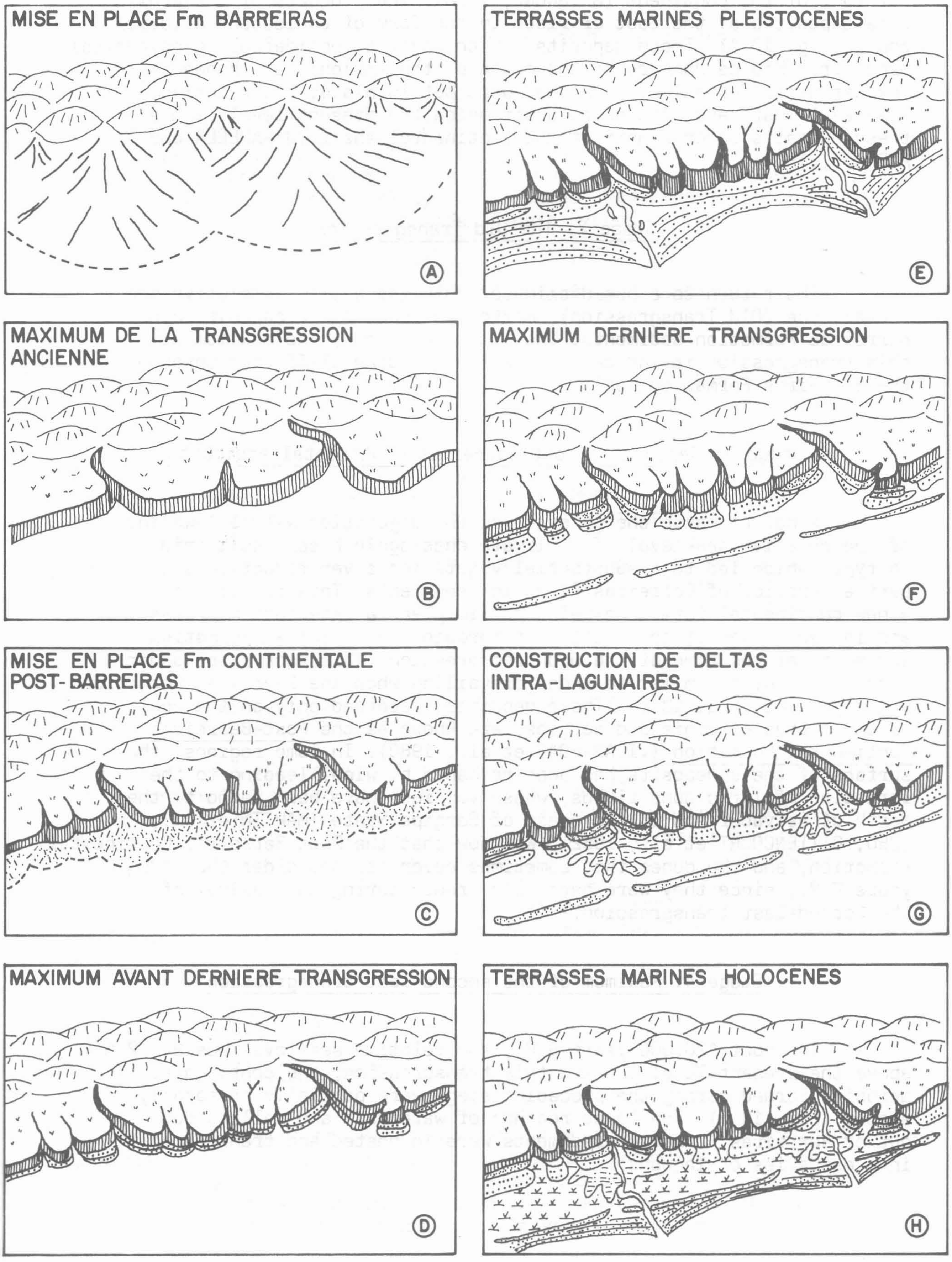


Fig. 13 - Schematised evolution of the central part of the Brasilian coast in the Quaternary

the vegetation cover tended to disappear, and the alteration mantle was exposed to erosion. This erosion was facilitated by a simultaneous upthrust of the continent (GHIGNONE, 1979). The products of the erosion were deposited at the foot of rises in the form of coalescent alluvial cones (Fig. 13 A). These deposits, which cover a considerable geographical area (from Rio de Janeiro to the mouth of the Amazon), are known as the Barreiras Formation. When they were put in place, the relative sea-level must have been quite a bit below the present level, since these deposits covered part of the continental shelf (BIGARELLA and ANDRADE, 1964).

Stage 2: The Old Transgression

The return to a humid climate, coinciding with a relative sea--level rise (Old Transgression), marked the end of the deposition of Barreiras Formation sediments. The limit reached by the maximum of this transgression is indicated by a line of dead cliffs cut through the Barreiras Formation sediments (Fig. 13 B).

Stage 3: Deposit of post-Barreiras continental sediments

A new climatic change occurred in conjunction with a lowering of the relative sea-level. The climate once again became semi-arid in type, which led to a substantial vegetation cover reduction and partial erosion of Barreiras Formation sediments. This resulted in a new continental formation being put in place at the foot of rises, and in particular of the cliffs cut through the Barreiras Formation sediments at the time of the Old Transgression. This took place under conditions quite similar to those prevailing when the Barreiras Formation was deposited (Fig. 13 C). These deposits, recorded only on the coasts of the States of Bahia and Sergipe, are known as the Post-Barreiras Continental Formation (VILAS-BOAS et al., 1980). In some regions, the surface of these deposits has been reshaped by wind, leading to the formation of large dune fields, whose vestiges have been found to the North of Salvador and on the coast of Sergipe State (MARTIN et al., 1980; BITTENCOURT et al., 1982). We know that the Post-Barreiras Continental Formation, and the dunes that sometimes cover it, are older than 120,000 years B.P., since they were partially eroded during the maximum of the second-last transgression.

Stage 4: Maximum of the second-last transgression

In about 120,000 years B.P., the relative sea-level was 8 ± 2 m above the present level. During this transgression, the continental deposits formed during the preceding stage were partially or totally eroded (Fig. 13 D). The lower reaches of waterways and valleys dug out of Barreiras Formation sediments were inundated and transformed into estuaries or lagoons.

Stage 5: Construction of Pleistocene marine terraces

A new regression started and sandy terraces covered with littoral belts were formed, either in the presence or in the absence of waterways (Fig. 13 E). Between 120,000 and 7,000 years B.P., the relative sea-level remained below the present level, in fact by about 110 m in the vicinity of 17,000 years B.P. Throughout this low sea-level period, a hydrographic network was established on the emerged part of the continental shelf, and in particular on the sandy terraces. Consequently, some rather wide and deep valleys were formed. Nevertheless, the original surface was often preserved in inter-fluvial zones, where littoral belt alignments are still more or less visible. These old belt alignments exhibit characteristics very different from those of more recent belts, which makes it possible to distinguish between them very easily on aerial photographs (MARTIN et al., 1981).

Stage 6: Maximum of the last transgression

In about 7,000 years B.P. the relative sea-level reached the present level, and then passed through a maximum of 4 to 5 m above the present level in about 5,100 years B.P. Hence the coast was submerged right up to that date. The most common manifestation of this submersion was the formation of barrier island/lagoon systems (Fig. 13 F). Depending on the region, this barrier island/lagoon stage was of varying significance or even absent. When a waterway opened up into one of these lagoons, an intra-lagoon delta was formed, with dimensions varying as a function of the lagoon and waterway sizes (Fig. 13 G).

Stage 7: Construction of Holocene marine terraces

After 5,100 years B.P., the relative sea-level gradually fell to its present position, although two rapid oscillations did occur, between 4,000 and 3,600 years B.P., and between 3,000 and 2,500 years B.P. During the emersion stages, littoral belts clung to the outer part of barrier islands, the lagoons tended to dry up, and the waterways opening into them flowed directly into the ocean (Fig. 13 H). It has been possible in some cases to distinguish among three generations of Holocene terraces, in relation to the three emersion stages occurring after 5,100 years B.P.

4.2. MODEL VALID FOR THE SOUTH OF SAO-PAULO STATE AND FOR THE STATES OF PARANA AND SANTA-CATARINA

This part of the Brasilian coast also contains vast Quaternary sedimentary plains, and the main points of the above outline remain valid. The continental deposits corresponding to the Barreiras Formation are not very extensively disseminated, but there are several local continental deposits (Pariquera-Açu, Alexandra, Canhanduva and other

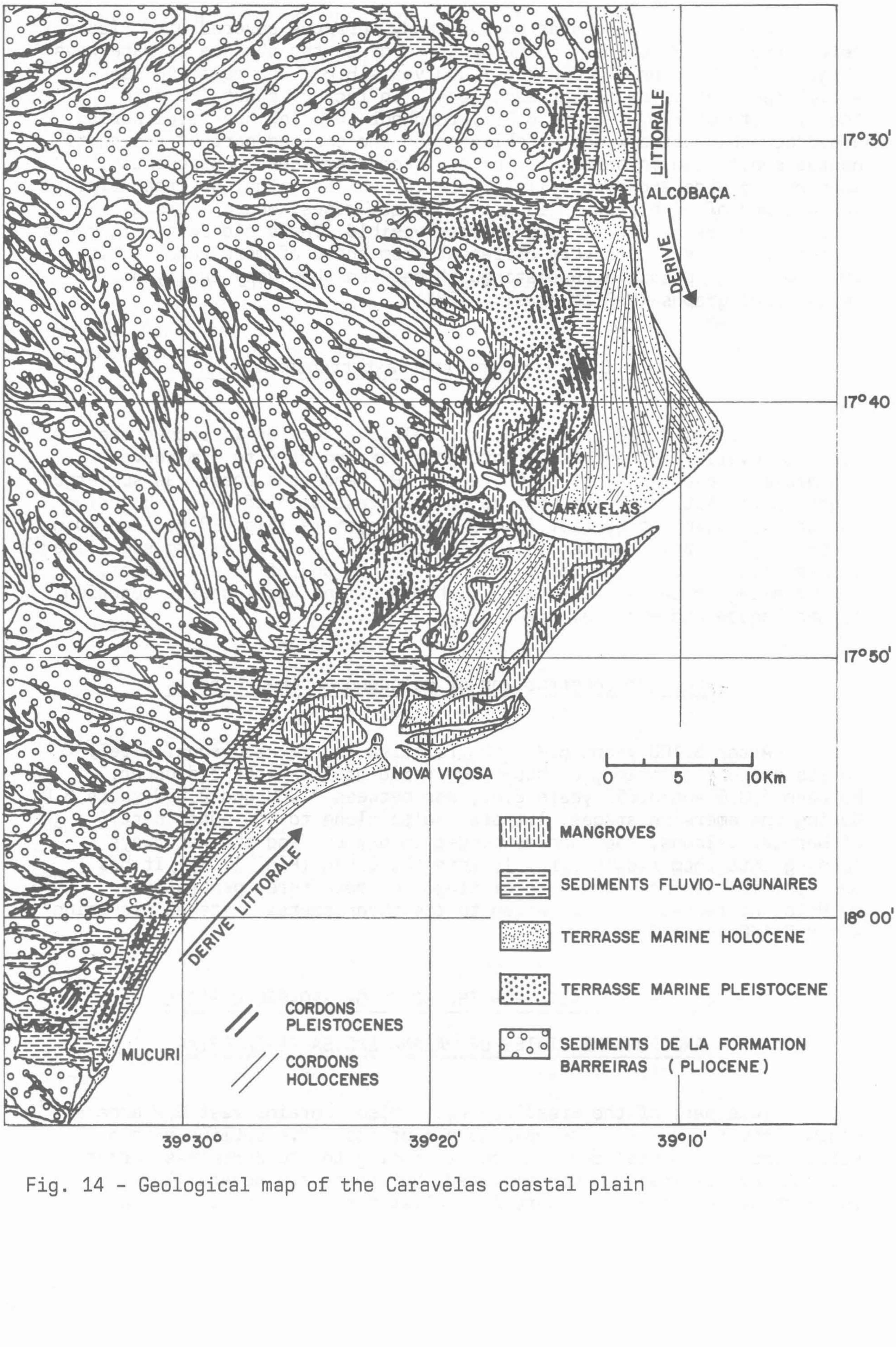


Fig. 14 - Geological map of the Caravelas coastal plain

Formations), that can be correlated with the Barreiras Formation. No records are known in this part of the Brasilian coast of the Old Transgression (Stage 2) or of the Post-Barreiras Continental Formation (Stage 3). The records of Stages 4 and 5, corresponding to the second-last transgression and the subsequent regression, are well developed. The same is true of Stages 6 and 7, corresponding to the maximum of the Holocene transgression and to the following regression. However, as the major part of drainage in this region is oriented towards the interior of the continent (Rio Parana), no substantial intra-lagoon deltas were formed, with the exception of that of Rio Tubarào, which is active to this day. Moreover, as the relative sea-level at the time of the 5,100 years B.P. maximum was lower than the level reached in the Macaé Recife sector, naturally the level then dropped less, so that markedly less sand was deposited from the nearby lower shelf, and hence Holocene terraces are considerably less extensive.

4.3. SPECIAL CASE OF SEDIMENTARY PLAINS LOCATED AT THE MOUTH OF A LARGE WATERWAY (RIOS PARAIBA DO SUL, DOCE, JEQUITINHONHA AND SAO-FRANCISCO)

In association with the mouths of the main waterways flowing into the Atlantic Ocean along the central part of the Brasilian coast, there are coastal plains that BACCOCOLI (1971), taking the definition of SCOTT and FISHER (1969) as a basis, considered to be "highly destructive deltas dominated by waves". BACCOCOLI assigned all these deltas to the Holocene age, and proposed an evolutionary scheme for the development of these deltaic plains since the maximum of the Flandrian transgression (last transgression), passing in some cases through an intermediate estuarine stage, finally to constitute typical deltas manifested by a continental advance into the sea.

However, we have seen that numerous "prograded" coastal plains, with no connection to a present or past waterway, exist all along the Brasilian coast. The most typical of these zones is that of Caravelas (Bahia), where, with the exception of fluvial deposits, one finds all the other types of sedimentary deposits existing in the "Quaternary Brasilian deltas" described by BACCOCOLI. To the point that this writer has suggested that this sedimentary accumulation could represent an old delta of the Rio Mucuri, an unexpressed waterway, located in the southern part of the coastal plain. This would then amount to a "delta dominated by waves" constructed without the presence of a waterway (Fig. 14).

The fact that "progradation" zones could be formed without the presence of a waterway immediately drew our attention. Obviously, in such a case it is necessary to look elsewhere for the source of the bulky sediments serving as raw material for the plain. Now we have seen that a relative sea-level drop of several metres can well, on a sandy coast, supply considerable quantities of sand. A study of the classical models of coastal sedimentation shows that wave energy, tide amplitude, solid fluvial discharge and others have been considered to be crucial factors for this type of sedimentation (FISHER, 1969;

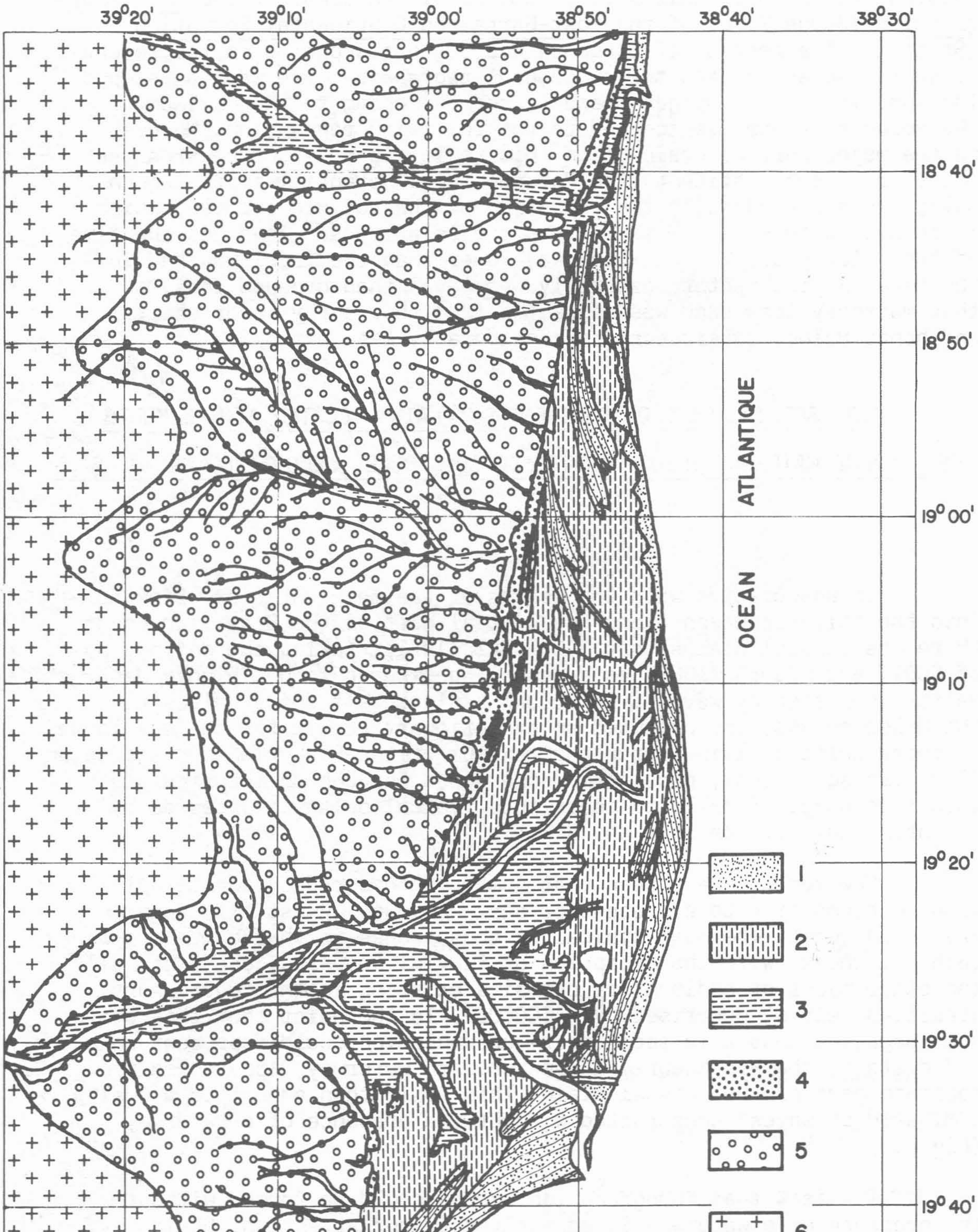


Fig. 15 - Geological map of the Rio Doce coastal plain. 1) Holocene marine terrace; 2) Lagoon sediments; 3) Fluvial sediments (intra-lagoon delta); 4) Pleistocene marine terrace; 5) Continental sediments (Barreiras Formation); 6) Precambrian formations; 7) Alignments of Pleistocene littoral belts; 8) Alignments of Holocene littoral belts; 9) Rio Doce palaeo-channels

CALLOWAY, 1975; HAYES, 1975), but strangely enough, no authors take possible relative sea-level oscillations into account. In their classical work on the subject, COLEMAN and WRIGHT (1975) analysed close to 400 parameters having an effect on the construction of sandy deltaic deposits, but they quite simply forgot to pay any attention to the most important: a possible lowering of the relative sea-level during the Holocene.
We have seen that at a river mouth, in the event that a dominant littoral drift exists, the sand brought in by the waterway can be deposited only on the side of the mouth lying below the littoral drift current. Only in the exceptional cases when there is no littoral drift (wave fronts parallel to the coast) can the sand brought in by the waterway, and reshaped by waves, be deposited on both sides of the mouth, and only in this case can a true "destructive delta dominated by waves" be formed. Now a detailed study has shown us that, in the sedimentary plains at the mouths of Rios Paraìba do Sul, Doce, Jequitinhonha and Sào-Francisco, the littoral drift has permanently, or at least for long periods of time, followed the same direction (DOMINGUEZ et al., 1983).

Coastal plains with an intra-lagoon delta

This category includes the plains at the mouths of Rios Doce (Espirito-Santo) and Paraìba do Sul (Rio de Janeiro). The Rio Doce coastal plain forms an assymetric crescent in the direction of the sea, with a maximum width of 38 km and a length of 130 km, and covering an area of 2,500 km² (Fig. 15). The Rio Paraìba do Sul coastal plain forms a lobe of 120 by 60 km, covering an area of 3,000 km² (Fig. 16). These two plains consist of Pleistocene and Holocene littoral marine sediments, and Holocene fluvial and lagoon sediments. For the coastal plain of Rio Doce, it has been possible to describe the main stages of the general model (Fig. 17 and 18). However, no record of stages 2 and 3 has been found. On the other hand, the lagoon formed during stage 6 having reached a very large size, the Rio Doce managed to build a vast intra-lagoon delta. It was also possible to find evidence for the existence of a second barrier island/lagoon system, related to the small transgressive stage that occurred between 3,800 and 3,600 years B.P. Similarly, there exists a second generation of Holocene littoral belts. It was not possible to find any evidence for the existence of a third barrier island/lagoon system, related to the third submersion stage, which occurred between 2,700 and 2,500 years B.P. This does not mean that the submersion stage did not take place, but only that it did not result in the formation of a barrier island/lagoon system. With minor differences due to local problems, the Rio Paraìba do Sul coastal plain formation runs along the same lines.

Coastal plains without an intra-lagoon delta

This category includes the plains of Rios Jequitinhonha and Sào-Francisco. For poorly determined reasons, the stage 6 (barrier island/lagoon system) lagoon did not extend in the E-W direction, and therefore remained fairly narrow; this explains why the waterways could not build, for lack of space, intra-lagoon deltas.

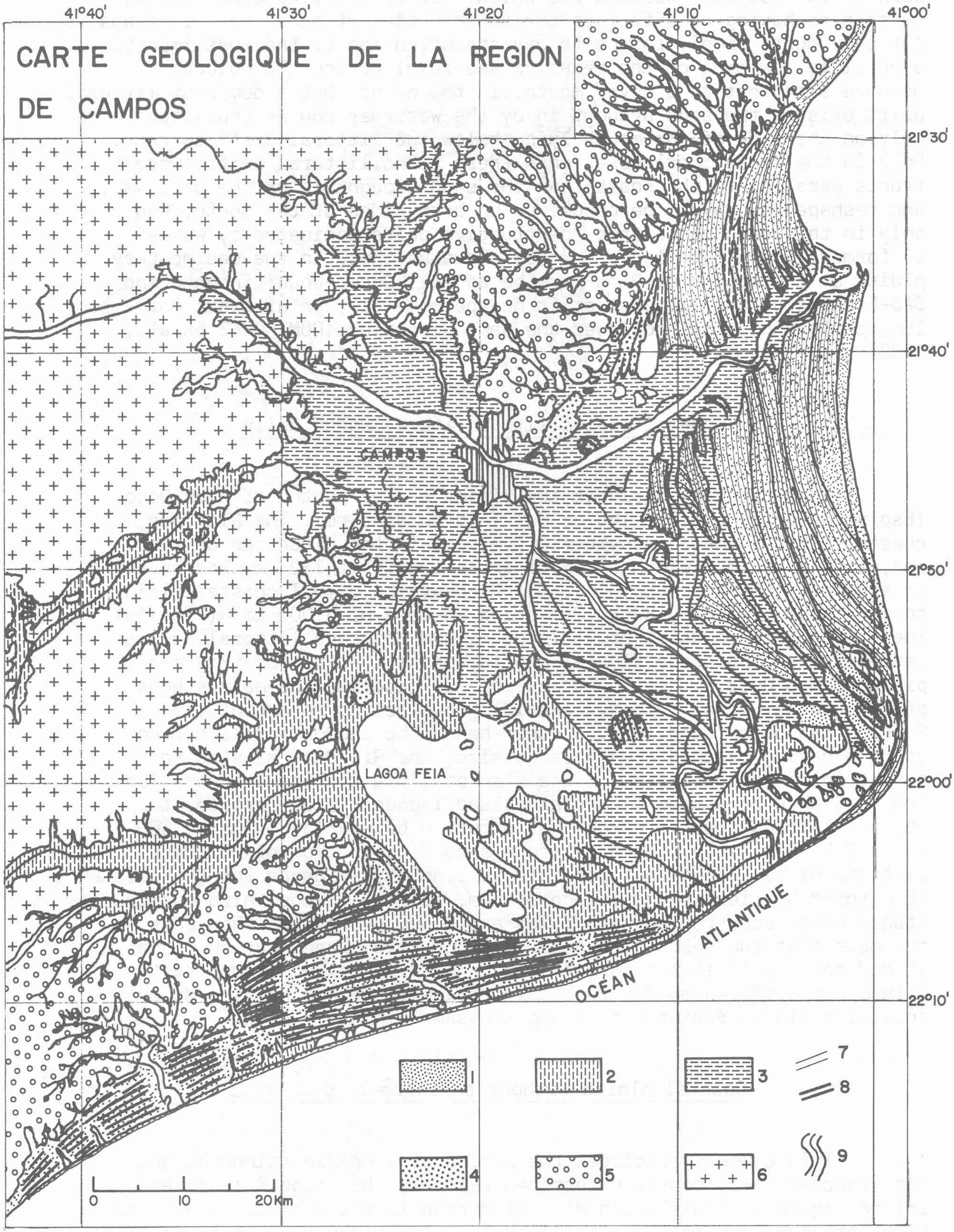


Fig. 16 - Geological map of the Rio Paraìba do Sul coastal plain.
1) Holocene marine terrace; 2) Lagoon sediments; 3) Fluvial sediments (intra-lagoon delta); 4) Pleistocene marine terrace; 5) Pliocene continental sediments (Barreiras Formation); 6) Precambrian formations; 7) Alignments of Holocene belts; 8) Alignments of Pleistocene belts; 9) Palaeo-channels of the Rio Paraìba do Sul

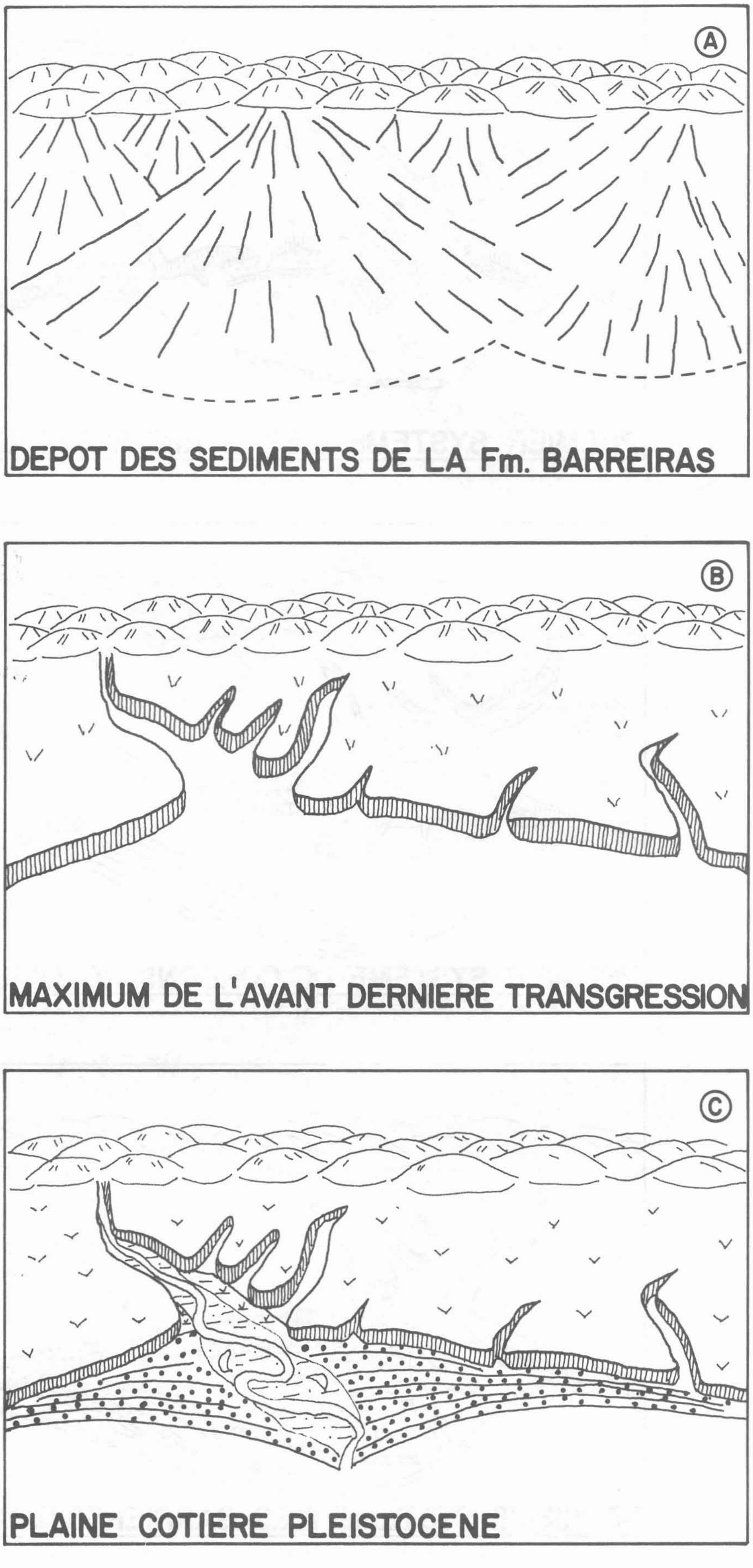


Fig. 17 - Schematised evolution of the Rio Doce coastal plain in the Pleistocene

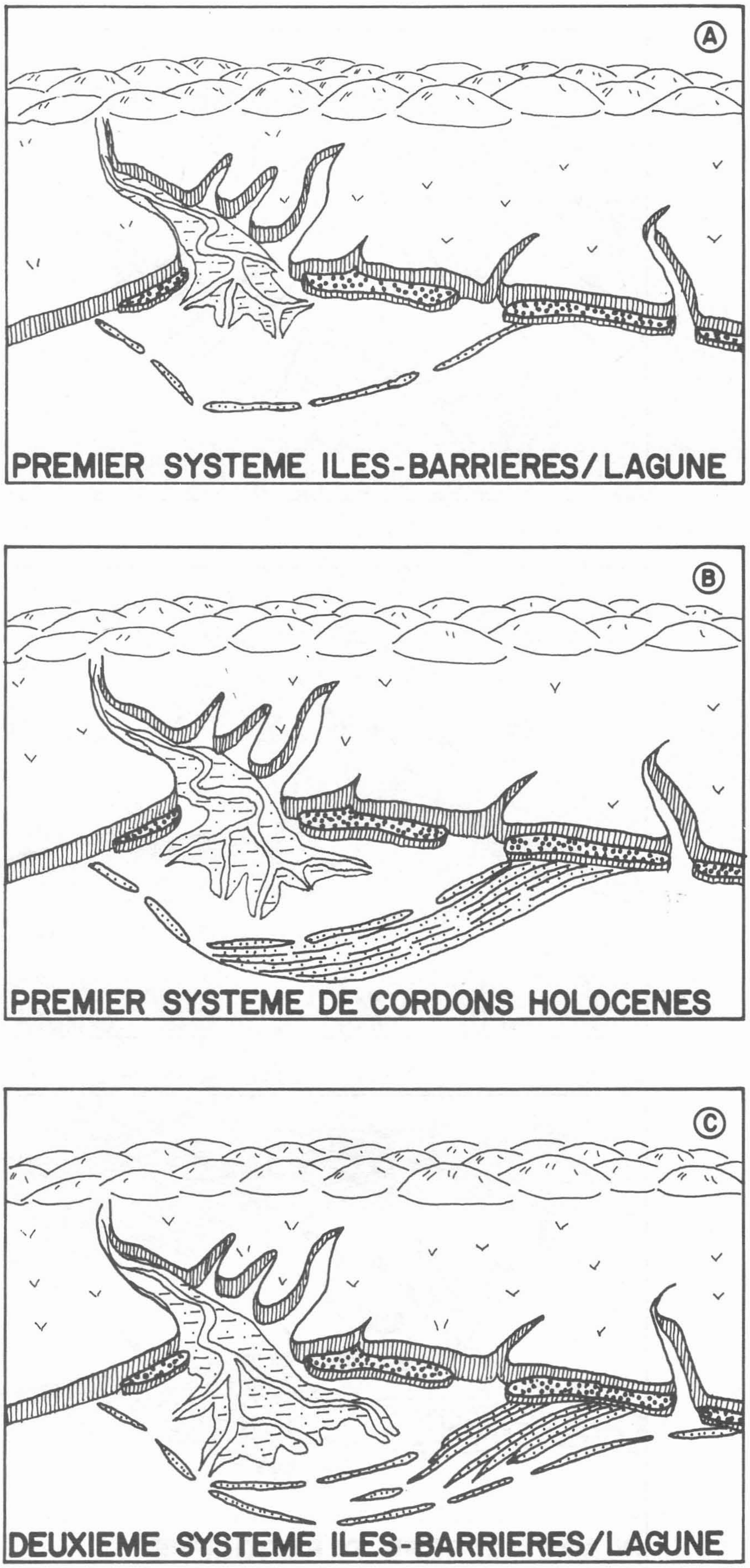


Fig. 18 - Schematised evolution of the Rio Doce coastal plain in the Holocene

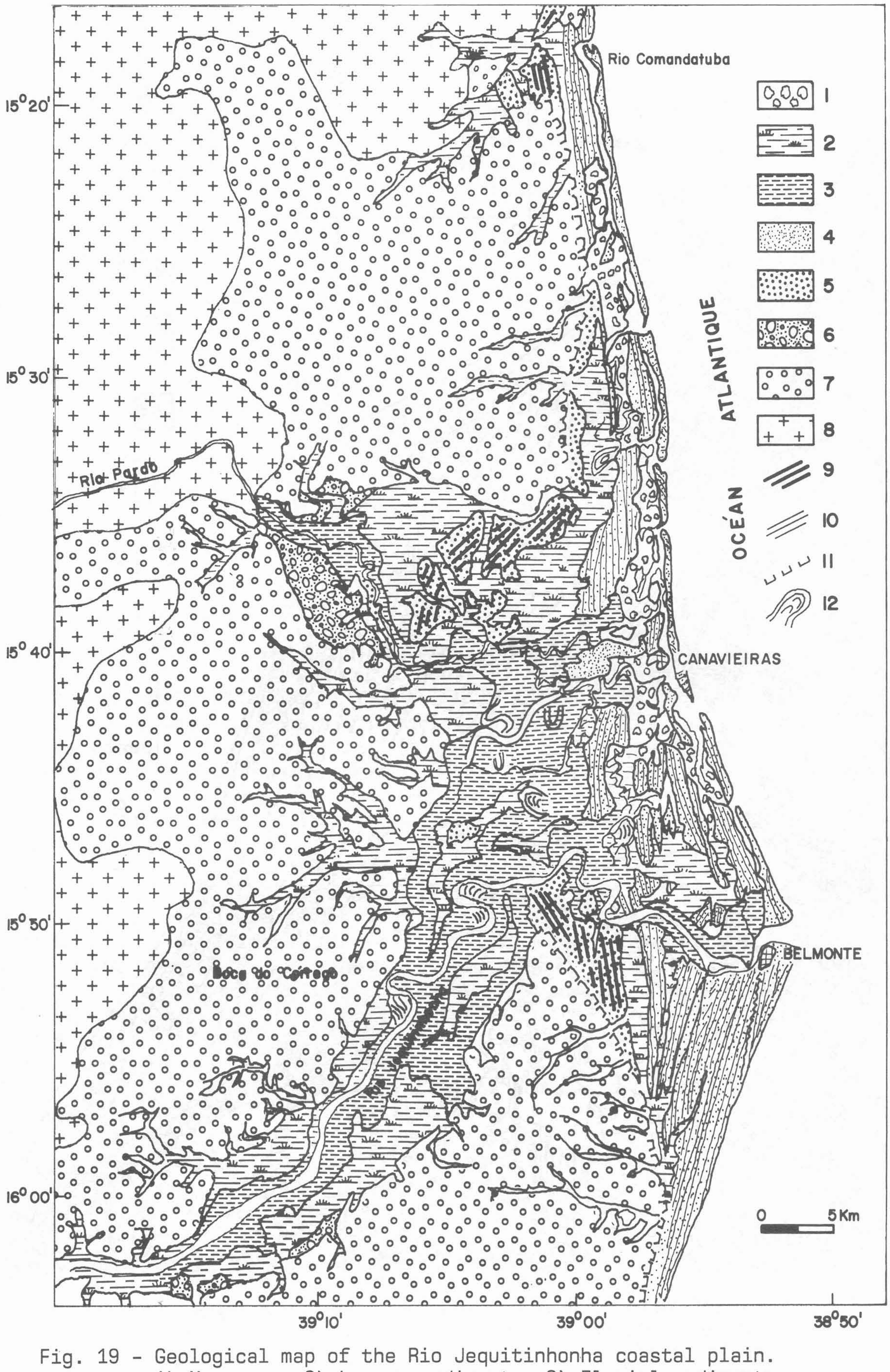


Fig. 19 - Geological map of the Rio Jequitinhonha coastal plain.
1) Mangrove; 2) Lagoon sediments; 3) Fluvial sediments; 4) Holocene marine terrace; 5) Pleistocene marine terrace; 6) Post-Barreiras continental sediments; 7) Pliocene continental sediments (Barreiras Formation); 8) Precambrian formations; 9) Alignments of Pleistocene belts; 10) Alignments of Holocene belts; 11) Dead cliffs; 12) Palaeo-channels of the Rio Jequitinhonha

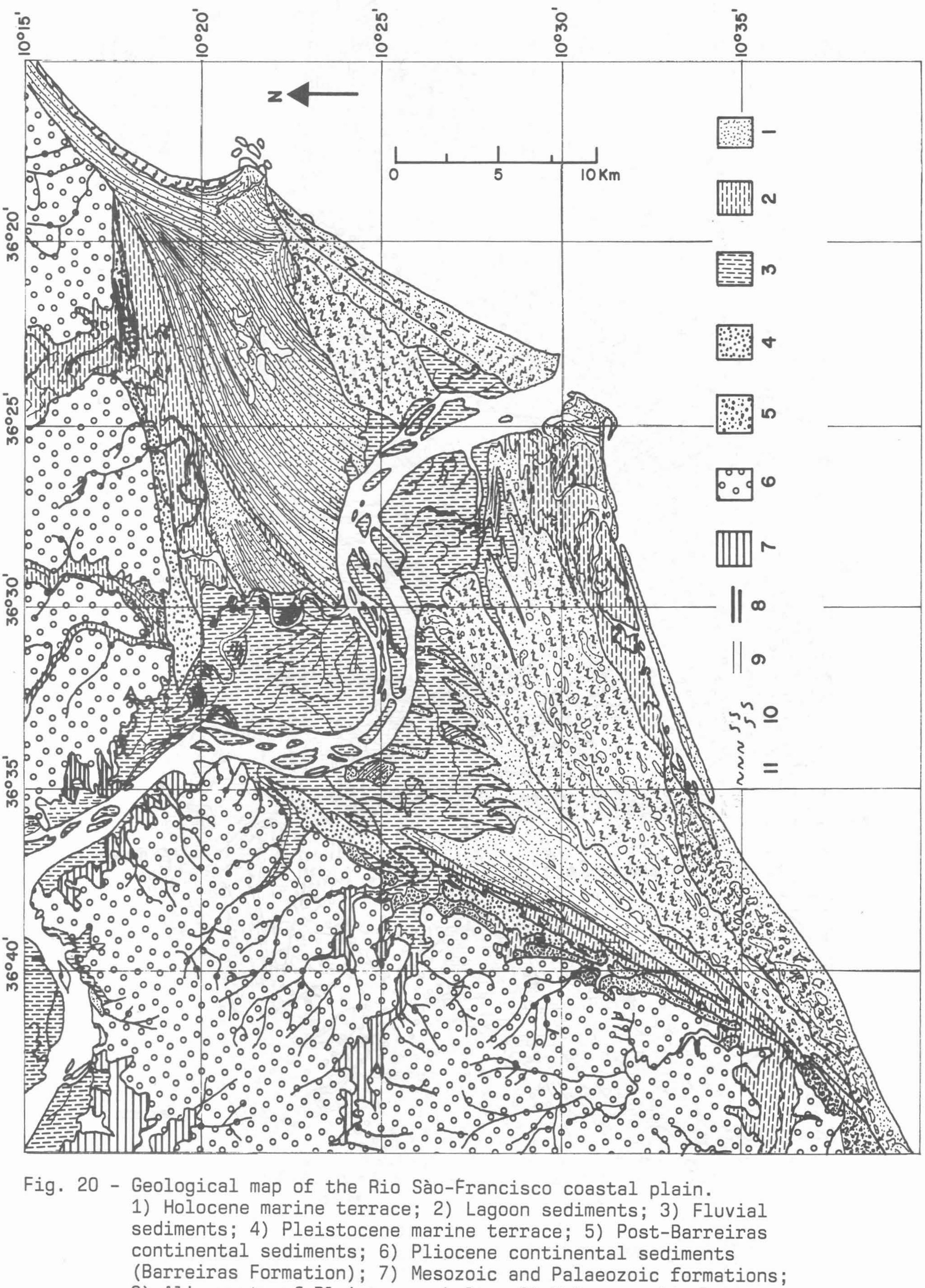


Fig. 20 - Geological map of the Rio Sào-Francisco coastal plain.
1) Holocene marine terrace; 2) Lagoon sediments; 3) Fluvial sediments; 4) Pleistocene marine terrace; 5) Post-Barreiras continental sediments; 6) Pliocene continental sediments (Barreiras Formation); 7) Mesozoic and Palaeozoic formations; 8) Alignments of Pleistocene belts; 9) Holocene alignments; 10) Fixed dunes; 11) Active dunes

The most thoroughly studied coastal plain is that of Rio Jequitinhonha (DOMINGUEZ, 1982; DOMINGUEZ et al., 1982). Installed in a zone hollowed out from Barreiras Formation sediments, it occupies a sector of almost 100 km of coast, covering an area of 800 km² (Fig. 19). As for the Rio Sào-Francisco coastal plain, it forms a triangle of 50 by 25 km, covering an area of 750 km² (Fig. 20). The Rio Sào-Francisco has its outlet in a hollowed zone of the Barreiras Formation, which is indisputably tectonic in origin (PONTE, 1969). In the coastal plain of Rio Jequitinhonha, all the stages described in the general model have been demonstrated with very great precision (Fig. 21, 22 and 23). Stage 7 has even been sub-divided into three sub-stages, corresponding to the three emersion periods occurring in the ranges 5,100-3,900, 3,600-2,900 and 2,500-0 years B.P. Moreover, two sudden displacements of the mouth have been correlated with the two short transgressive periods of 3,80063,600 and 2,700-2,500 years B.P. The coastal plain of Rio Sào-Francisco was formed according to the same model. However, it was not possible to find evidence for several generations of Holocene sandy terraces, in connection with the different emersion stages occurring after 5,100 years B.P. This may be due to the fact that, for tectonic reasons, the lower reaches of the Rio Sào-Francisco could not shift as easily as those of the Rio Jequitinhonha, at the time of the sudden relative sea-level rises occurring after 5,100 years B.P.

5. CONCLUSIONS

The existence of large Quaternary coastal plains is one of the characteristics of the central part of the Brasilian coast. These plains are either situated at the mouth of a large waterway, or have no connection with a present or previous waterway. A second characteristic of this coast is that it was, as opposed to other regions of the world, submerged until about 5,100 years B.P., and on the average above water since then. A third characteristic of this region is that it is a high--energy coast, where the littoral drift current plays an essential role in the transport of coarse sediments. The submersion period occurring before 5,100 years B.P. often resulted in the formation of barrier islands, which isolated lagoons of varying size from the open sea. When these lagoons were sufficiently large, the waterways opening into them would build up intra-lagoon deltas. The emersion period occurring after 5,100 years B.P. was reflected in a tendency for the lagoons to dry up. It was only from that time on that the waterways opening into the lagoons could flow directly into the sea. Coastal plains situated at the mouths of large waterways, such as the Rios Paraìba do Sul, Doce, Jequitinhonha and Sào-Francisco, were classified until now in the category of "highly destructive deltas dominated by waves". It is clear that application of the term delta to these plains is an exaggeration, to say the least. In fact, we have seen that coastal plains with very similar characteristics bore no relationship at all to present or previous waterways. Moreover, we have managed to demonstrate that those situated at the mouth of a large waterway were built up for a large part from coarse sediments supplied by lowering of the relative sea-level. This is corroborated by the fact that, throughout the entire lagoon stage, sediments transported by the waterway were trapped in the lagoon, and hence could not have contributed to the edification

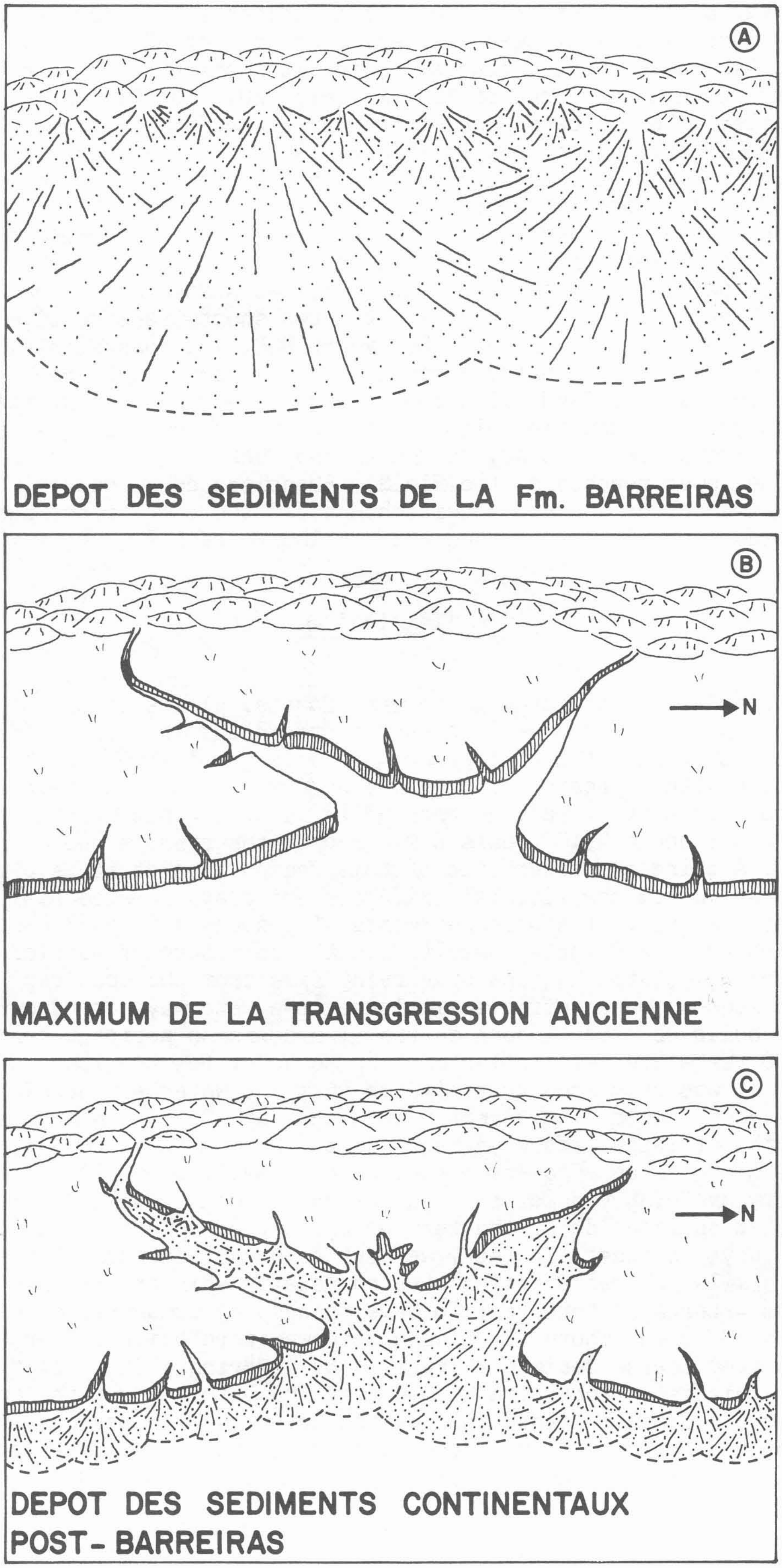


Fig. 21 - Schematised evolution of the Rio Jequitinhonha coastal plain in the Pleistocene

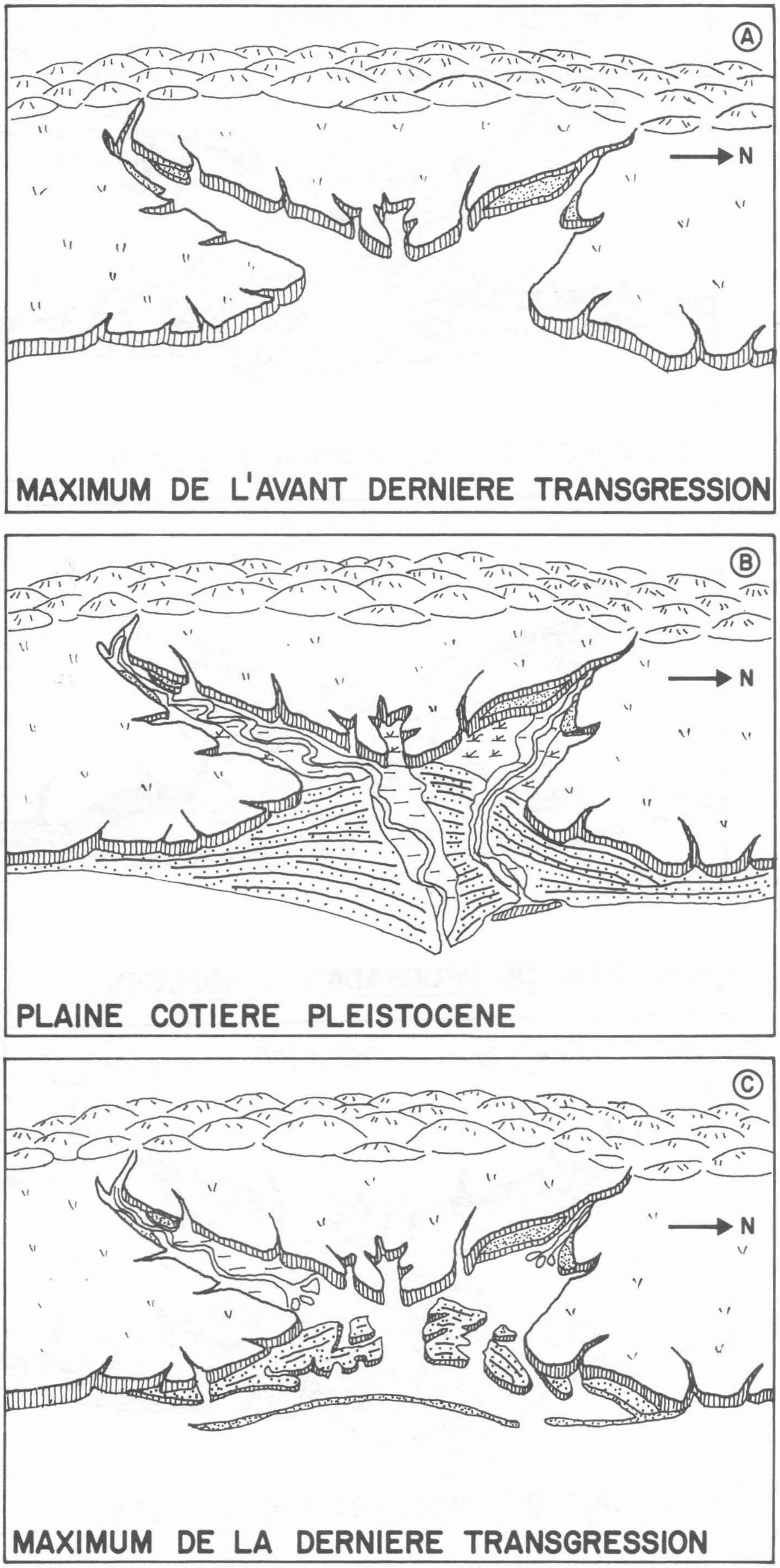


Fig. 22 - Schematised evolution of the Rio Jequitinhonha coastal plain:
A) before 120,000 years B.P. B) after 120,000 years B.P.
C) before 5,100 years B.P.

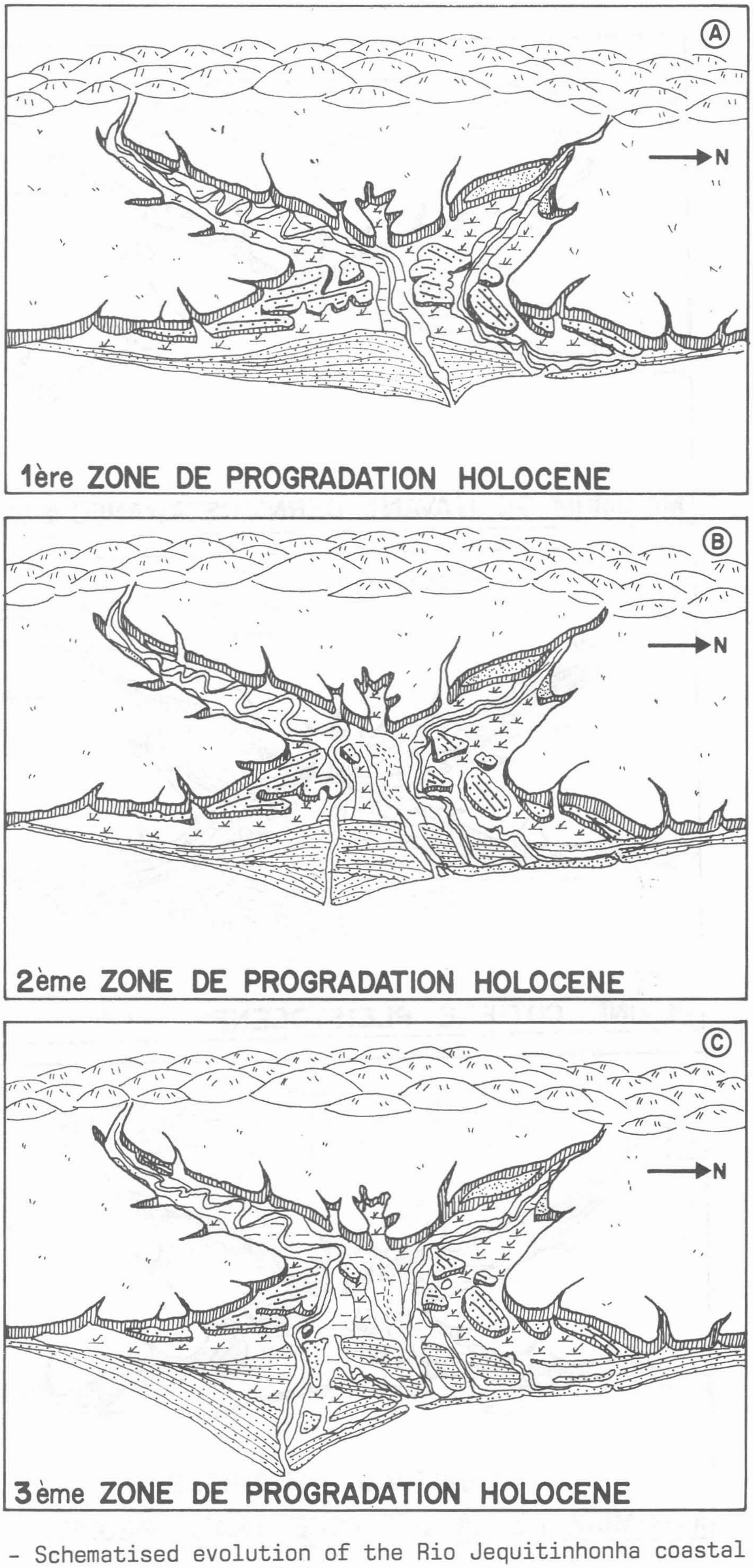


Fig. 23 - Schematised evolution of the Rio Jequitinhonha coastal plain after 5,100 years B.P.

of the sandy terraces that were formed on the outside of the barrier islands. It was only once the waterway flowed directly into the sea that it started to play an important role in the construction of the coastal plain situated at the river mouth. For in this case, the flow of water blocks littoral transport and causes an accumulation of sand in the part of the plain lying in the drift current. On the other hand, sand transported by the river is deposited only on the side of the plain located below the littoral drift current.

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