Neotectonics in Northeastern Brazil

A Thesis Submitted for the Degree of
Doctor of Philosophy of the University of London

by

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This thesis is the sole work of the candidate except where acknowledgment is made in the text. The paper by Bezerra et al. (1998) cited in chapters 3 and 4 was almost entirely written by the candidate.

Francisco Hilário Rego Bezerra
To my son, Raul
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ABSTRACT

The thesis describes neotectonic deformation in the continental intraplate region of northeastern Brazil and explores its links with modern seismicity. The region, which is under E-W-oriented compression and N-S-oriented extension, shows shallow earthquake swarms which last for several years and include 5.0-5.2 m$_b$ events. Remote sensing, borehole and geophysical data, in conjunction with field structural information, indicate a continuous faulting process since the Miocene which has reactivated Cretaceous faults and Precambrian shear zones or in places generated new faults which cut across existing structures.

Three main sets of faults are recognised across the area: a NE-striking set, a NW-striking set and a N-striking set. The first and the second sets are pervasive and their cross-cutting relationships show that they locally form a conjugate set and display both a strike-slip and a dip-slip component of movement. They have generated troughs filled by as much as 260 m of Cainozoic sediments. Radiocarbon dating shows that some of the faults slipped as recently as 4,041-3,689 cal. yr BP. Although the elevation of coastal deposits is consistent with the predictions of glacioisostatic models for the area, tectonic influence can be detected notably near the Carauauais fault, where rapid emergence by at least 5 m to the east of São Bento occurred 4,080-2,780 cal. yr BP. Secondary ground failure, which includes hydroplastic deformation, liquefaction and landslides, can be seen in Quaternary alluvial sediments and is reported in the historical record.

The present data show that the potential for large earthquakes in northeastern Brazil has been underestimated. Empirical relationships using liquefaction and surface rupture point to events of at least M$_s$=6.8 compared to a maximum m$_b$ = 5.2 recorded instrumentally. The finding that NE- and NW-trending faults are favourably orientated for reactivation in relation to the current stress field is of potential value for seismic hazard assessment.
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- M = magnitude
- $M = \text{moment magnitude}$
- $M_L = \text{local magnitude}$
- $m_b = \text{body-wave magnitude}$
- $M_s = \text{surface-wave magnitude}$
- MMI = modified Mercalli intensity scale
- $S_{H_{\text{max}}} = \text{maximum horizontal stress}$
- $\delta^{13}C = ^{13}C/^{12}C \text{ ratio rel. PDB}$
- $\delta^{18}O = ^{18}O/^{16}O \text{ ratio rel. PDB}$
- SEM = scanning electron microscope
- XRD = X-ray diffraction
- ASL = above sea level
- cpm = counts per minute
- SLAR = Sideways-looking airborne radar
- CDM-RN - Companhia de Desenvolvimento Mineral do Rio Grande do Norte
- CAERN - Companhia de Águas e Esgotos do Rio Grande do Norte
- IPT - Instituto Politécnico do Estado de São Paulo
Establishing a context

Since the general acceptance of plate tectonics, there have been increasingly sophisticated studies connecting earthquakes and active faults to lithosphere mechanisms driven by plate motion and modulated by the nature of the plates themselves. Deformation and its consequences, such as earthquakes, are routinely classified into interplate (plate margins) and intraplate categories. The vast majority of earthquakes fall into the first category. Yet intraplate deformation and seismicity cannot be ignored. According to Johnston (1989), passive margins account for one third of all seismicity in intraplate settings, and more than half of all large (Mₜ > 6) intraplate earthquakes.

The causes of intraplate deformation are still poorly understood and cannot be explained solely by seismological data. The average recurrence interval for intraplate faults can range from 10,000 yr to 100,000 yr or more (Crone et al. 1997). Instrumental and historical records of earthquakes are too short and incomplete in geological terms to explain fault behaviour in continental intraplate settings. In many areas the recurrence interval of active faults is much longer than the period of historical settlement (McCalpin and Nelson 1996). Therefore, a history of these faults cannot be achieved without the help of neotectonics.

Examples of Quaternary faults in continental intraplate settings are abundant. Prehistoric surface ruptures are known to exist in many parts of the world. Eleven historical earthquakes in continental intraplate settings are known to have produced surface rupture (Table 1.1). But the international literature has no examples of coseismic rupture in the intraplate part of the South American continent and few examples of deformation (e.g. Martin et al. 1986a).
The Brazilian coast, within the South America passive margin, displays two areas of relatively high seismicity (Assumpção 1998): (a) northeastern Brazil, where the seismicity occurs onshore; and (b) southeastern Brazil, where the seismicity is concentrated offshore and is therefore difficult to investigate by direct observation. Seismicity in northeastern Brazil occurs within the first 1-12 km of the upper crust (Ferreira et al. 1997), where brittle processes are dominant and some kind of surface expression is therefore to be expected. The region experiences many earthquake swarms, including events up to $m_b=5.2$, which have been ascribed to the reactivation of basement faults or the generation of new ones (Takeya et al. 1989, Assumpção 1992, Ferreira et al. 1995, Ferreira et al. 1997, Ferreira et al. 1998). For these reasons, northeastern Brazil emerges as an ideal field area for combining neotectonic studies with seismology in order to advance understanding of intraplate deformation in general.

Table 1.1 - Historical intraplate earthquakes which produced documented surface rupture (from Crone et al. 1997).

<table>
<thead>
<tr>
<th>Location</th>
<th>Date</th>
<th>Magnitude</th>
<th>Rupture length (km)</th>
<th>Scarp height (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cutch, India</td>
<td>16 June 1819</td>
<td>$M_s=7.25-8.25$</td>
<td>&gt; 90</td>
<td>7.9</td>
</tr>
<tr>
<td>Accra, Ghana, Africa</td>
<td>22 June 1939</td>
<td>$M = 6.5$</td>
<td>9-17</td>
<td>0.46</td>
</tr>
<tr>
<td>Central Sudan, Africa</td>
<td>9 October 1966</td>
<td>$m_b=5.1$</td>
<td>6</td>
<td>0.0</td>
</tr>
<tr>
<td>Meckering, WA, Australia</td>
<td>14 October 1968</td>
<td>$M_s=6.8; m_b=6.0$</td>
<td>37</td>
<td>3.5</td>
</tr>
<tr>
<td>Calingiri, WA, Australia</td>
<td>11 March 1970</td>
<td>$M_s=5.7; m_b=6.0$</td>
<td>3</td>
<td>&lt; 0.4</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>5.7</td>
<td></td>
</tr>
<tr>
<td>Cadoux, WA, Australia</td>
<td>6 June 1979</td>
<td>$M_s=6.4; m_b=6.3$</td>
<td>28</td>
<td>&lt; 1.4</td>
</tr>
<tr>
<td>Guinea, Africa</td>
<td>22 December 1983</td>
<td>$M_s=6.2; m_b=6.4$</td>
<td>9.4</td>
<td>0.13</td>
</tr>
<tr>
<td>Marryat Creek, SA, Australia</td>
<td>30 March 1986</td>
<td>$M_s=5.8; m_b=5.7$</td>
<td>13</td>
<td>~ 0.9</td>
</tr>
<tr>
<td>Tennant Creek, NT, Australia</td>
<td>22 January 1988</td>
<td>$M_s=6.3, 6.4, 6.7$; $m_b=6.1, 6.1, 6.5$</td>
<td>32</td>
<td>1.8</td>
</tr>
<tr>
<td>Ungava, Quebec, Canada</td>
<td>25 December 1989</td>
<td>$M_s=6.3; m_b=6.2$</td>
<td>10</td>
<td>1.8</td>
</tr>
<tr>
<td>Killari, India</td>
<td>29 September 1993</td>
<td>$M_s=6.4; m_b=6.3$</td>
<td>3</td>
<td>&lt; 0.5</td>
</tr>
</tbody>
</table>
Definition of neotectonics

The term neotectonics was first proposed by Obручев (1948) to denote active geological processes and used by him to define a new branch of geoscience. At first, neotectonics was viewed as equivalent to Cenozoic or Quaternary deformation (e.g. Jennings et al. 1975). Increasingly, however, there is a tendency to abandon precise definitions of the period encompassed by neotectonics. Hancock (1986), for example, proposed that the neotectonic period in a certain region started when the present-day configuration of relevant plate boundaries and motions was established. According to Pavlides (1989) there is no definition of neotectonics which is globally valid. This is in agreement with the term ‘present seismotectonic regime’ used by Slemmons (1991) and defined as the period of time when the present stress field and tectonic processes came into place. More simply, Stewart and Hancock (1994) defined neotectonics as the branch of tectonics devoted to the study of earth movements which started in the past but still continue at the present. Some studies prefer the term ‘active tectonics’ (e.g. Wallace 1986, Keller and Pinter 1996), but as neotectonics is still widely used and its plate- and stress-related connotations well understood, it is retained for the present study.

Aims and scope of the thesis

The main objective of the present thesis is to describe neotectonic deformation in the continental intraplate region of northeastern Brazil, which concentrates most of the seismicity of Brazil, using a variety of methodologies. It tries to bridge the gap between instrumental seismicity and the geological evidence for late Cretaceous structures primarily by investigating neotectonic deformation in Cainozoic rocks along the coastal plain. The research goals include (a) the mapping of major neotectonic faults, describing their geometric and kinematic features; (b) the dating of faulting events by the radiocarbon method and by stratigraphic correlation; (c) the investigation of secondary seismic effects in the geological record such as liquefaction; and (d) correlation of the results with seismic data. The study area is located in the Rio Grande do Norte state, northeastern Brazil, between latitudes 4°55’ S and 6°25’ S, and longitudes 34°58’ W and 37° W (Figs. 2.2 and 2.3).
The present thesis is divided into three main parts. The first includes chapters 2 and 3 and provides the background information necessary for the neotectonic study, namely the tectonic setting of the study area (Chapter 2) and the Cainozoic stratigraphic units which were investigated in the field (Chapter 3). The second part (chapters 4 - 7), reviews the neotectonic findings obtained using different methodologies. Chapter 4 concentrates on the late Cainozoic coastal deposits as evidence for neotectonics, Chapter 5 describes faults mapped using remote sensing, borehole data and geophysical methods. Chapter 6 describes outcrop data for the same faults. Chapter 7 is devoted to liquefaction and hydroplastic structures seen in outcrop. The third part deals with the correlation of the neotectonic data with instrumental and historical seismicity (Chapter 8), neotectonic evolution in the light of models of deformation for the area, and the issue of seismic hazard (Chapter 9).
The target area of this research is the passive continental margin of the South American plate, whose geometry and limits are depicted in Fig. 2.1. The western and eastern edges of the South American plate are better defined than its northern and southern boundaries, whose limits are more complex (Minster and Jordan 1978; DeMets et al. 1990; Meijer and Wortel 1992, Coblentz and Richardson 1996). Along its western boundary, the Nazca Plate has been subducting along a trench almost 6,000 km long at a rate of approximately 84-90 mm/yr (DeMets et al. 1990; Dewey and Lamb 1992). The eastern boundary is well marked by shallow seismicity along the South Atlantic mid-ocean ridge, where spreading rates are roughly constant (32 to 38 mm/yr) from 5° S to 40° S, and then decrease to 25-30 mm/yr at 55° S (Solomon et al. 1975; Gordon and Jurdy 1986; DeMets et al. 1990). In contrast, the northern margin of the plate is more complicated, and even uncertain, though broadly defined by the adjoining Caribbean and North American plates (Argus and Gordon 1989, Coblentz and Richardson 1996). The southern limit lies along the North and South Scotia ridges, where the Antarctic and the South American plates converge at a rate of 21 ± 0.2 mm/yr (Minster and Jordan 1978).

Three major physiographic units may be identified within the South American plate: the Mesozoic-Cainozoic orogenic belt, the continental cratonic region (Precambrian shields and Palaeozoic to Cainozoic platforms) and a zone of oceanic crust formed along the South Atlantic mid-ocean ridge (Fig. 2.1). The first includes a volcanic arc above the subduction zone with important gaps in northern Peru and central Chile. At the rear of the orogenic belt, the Altiplano (Peru and Bolivia) and the Puna (west Argentina) are largely above 3,000 m, and are bordered on the east by the Cordillera Oriental, up to 6,500 m high and forming a fold belt up to 400 km wide. The eastern Andes slope gently towards the Amazon basin foreland and the Argentina Pampas, which are represented mainly by Precambrian rocks covered by younger sedimentary basins of Palaeozoic to Cainozoic age (Almeida et al. 1981). The zone of oceanic crust dates from the last 100 Ma (Popoff 1988).
three major tectonic units contrast not only in their evolution but also in their tectonic behaviour at the present day.

The Precambrian shields and Palaeozoic to Cainozoic platforms may be subdivided into a number of geological provinces. The Precambrian shield, where part of this research is located, includes the Guyana, Central Brazil and Atlantic shields (Almeida et al. 1981) (Fig. 2.2a). The Palaeozoic to Cainozoic platforms comprise the vast majority of the sedimentary basins on the continent and are divided into South American and Patagonian platforms (Almeida et al. 1981) (Fig. 2.2a).

In the Atlantic shield, where part of the study area is located, the crystalline regions may be grouped into Archean and Proterozoic provinces. The Archean terrains, mostly composed of gneiss-migmatite complexes and greenstone belts, occur in the Tapajós, Rio Branco and São Francisco provinces; the Proterozoic belts, which formed during the Brasiliano Orogeny (650-550 Ma), occur in the Tocantins, Mantiqueira, and Borborema provinces (Almeida et al. 1981) (Fig. 2.2b). The South American platform comprises the Amazon, Parnaíba, Paraná, and Coastal provinces (Almeida et al. 1981) (Fig. 2.2b). The former three provinces are Palaeozoic (post-Ordovician) basins composed of sedimentary covers up to 5,000 m thick in their depositional centres (Fig. 2.2b) (Soares et al. 1979). An important Mesozoic flood continental basalt event, due to the Gondwana break-up, took place in some of these troughs, as in the Paraná basin (Piccirillo et al. 1988).

The Coastal province, where the present study was carried out, overlies more than 8,000 km of coast along the Atlantic margin of South America (Fig. 2.2b). It was formed during the opening of the South Atlantic ocean in Permo-Triassic times (LePichon and Huchon 1984). Chang et al. (1992) suggested that two different margins of the South Atlantic developed during the Africa-South America break-up: the North Brazilian Equatorial margin, where strike-slip motion between the two continents led to complex shear-dominated basins; and the East Brazilian margin, where basins were formed by crustal extension (Mascle 1976). Chang et al. (1992) also suggested extension in a NE direction between both continents, and injection of NW-trending dikes. They recognised five main sedimentary stages dominated by
Figure 2.1 - The South American plate and surrounding regions (modified from Coblentz and Richardson 1996), including some directions of maximum horizontal compression $S_{\text{max}}$ (thin lines) from Assumpção (1992) and Lima et al. (1997). An E-W-oriented topographic cross-section of the plate is provided.
Chapter 2 - Tectonic setting

siliciclastic systems in southern Brazil and carbonate platforms in northern and northeastern Brazil: the continental syn-rift, the transitional evaporitic stage, the shallow marine carbonate platform of early drift stage, the open marine transgressive and the regressive marine stage. They also pointed out that along the eastern margin, normal faults have usually NNE to NE strikes, whereas transform faults show NNW to NW strikes.

In summary, the area of research is located within the Borborema province (Atlantic shield) and the Coastal province (Palaeozoic to Cainozoic platform), which is bordered by the Mesozoic-Cainozoic orogenic belt along the Andes Cordillera and by younger oceanic crust formed along the South Atlantic mid-ocean ridge. The Precambrian shield consists of Archean and Proterozoic fold belts covered by post-Cambrian sedimentary basins and the Mesozoic and Cainozoic sedimentary rocks of the Coastal province. Each tectonic province has its magmatic, sedimentary and metamorphic characteristics. The last greatest tectonic event took place during Mesozoic times, when the South American and African continents were separated.

Stress pattern

As elsewhere, intraplate stress data in South America have come from a great diversity of sources, such as focal mechanisms, Quaternary faults, breakouts, and hydraulic fracturing stress measurements. There is a lack of information on stress in large parts of both continental and oceanic regions. Focal mechanisms are by far the most common source of data, with stress measurements deduced from geological faults the second important source. Both focal mechanism and fault data come predominantly from the Andes region, and few fault data are available for the cratonic area. Some breakout and hydraulic fracture measurements are available within the plate but they are mostly concentrated along the Atlantic coast. A general picture of the stress pattern in South America is presented in Fig. 2.1.

The early study by Mendiguren and Richter (1978) proposed a primarily NW-SE oriented maximum horizontal stress ($S_{\text{Hmax}}$) for the cratonic part of South America based on focal mechanisms for five teleseismic earthquakes. Yet more recent works pointed out that $S_{\text{Hmax}}$ is
approximately constant within most of the plate, and correlates roughly with the direction of plate motion (Zoback et al. 1989; Stefanick and Jurdy 1992; Assumpção 1992). Although the intraplate stress is not uniform throughout the continent, the cratonic part of the South American plate is predominantly under a horizontal E-W-oriented compressional stress (Zoback et al. 1989; Assumpção and Suarez 1988; Assumpção 1992). Even in the Andes region, where there is a change in the trench direction from N-S (South of 18° S latitude) to NW (North of 18° S latitude), the E-W-oriented $S_{\text{Hmax}}$ is remarkably constant.

Nonetheless, significant differences in the stress field can be inferred in some regions. The Andean E-W stress field, for example, decreases in magnitude towards the East, which suggests a different stress province in the Central Amazon (Assumpção 1992). In addition, in spite of the important E-W compression in the Andes Cordillera, the high elevated areas that exceed 3000 m show stress regime dominated by N-S extension (Zoback et al. 1989; Richardson and Coblentz 1994). Lima et al. (1997) also showed that there are significant changes in the maximum horizontal stress ($S_{\text{Hmax}}$) in the Brazilian sedimentary basins and that the $S_{\text{Hmax}}$ orientation is roughly parallel to the coastline in the Equatorial and eastern margins north of latitude 15° S (Fig. 2.1 and 2.2b).

Several stress provinces have been proposed for the Brazilian shield and the South American platforms in accordance with variations in the stress pattern. Assumpção (1992) identified an Amazon-Guyana province based on N-S-oriented P axes of focal mechanisms. Between the E-W-oriented $S_{\text{Hmax}}$ in the Andes Cordillera and the $S_{\text{Hmax}}$ direction of the Amazon-Guyana province, there is a zone of very low seismicity, indicating that the transition between stress provinces in that region is marked by a decrease in the stress magnitude (Assumpção 1992) (Fig. 2.1). However, the main cause of stress rotation from the E-W-trending Andes Cordillera to the N-S-trending Amazon-Guyana Province is probably a large lateral density contrast along the Amazon rift (Nunn and Aires 1988).

The southern part of Brazil, as well as its neighbouring regions such as Paraguay, display another stress pattern. The interior of the Paraná basin is much less seismic than its border (Assumpção 1992). The reasons are not well understood. Although horizontal compression
Chapter 2 - Tectonic setting

predominates, there is no uniformity of data to the north and to the east of the Paraná basin. In that region, Saadi et al. (1991) detected $S_{\text{Hmax}}$ with a NW-SE trend in Quaternary-Tertiary sediments. Similarly, Caproni and Armelin (1990) presented hydraulic fracturing data from shallow boreholes (100-150 m deep) drilled in granite, which indicated NW-SE horizontal oriented compression. Departures from this pattern have been described in other parts of the Paraná basin. For example, the 1964 Mato-Grosso-do-Sul and 1982 Paraguay earthquakes have reverse and strike-slip focal mechanism in the western part of the basin respectively and display ENE-WSW-oriented P axes. The $S_{\text{Hmax}}$ has an E-W orientation west of the Paraná basin at the western margin of the South American plate (Assumpçâo and Suarez 1988) (Fig. 2.2b). In northeastern Brazil, there is a strong concentration of seismicity outside the Mesozoic Potiguar basin. The $S_{\text{Hmax}}$ is parallel to the E-W coast and is oriented E-W and locally WNW-ESE (Lima et al. 1997; Ferreira et al. 1998) (Fig. 2.2b).

According to Zoback et al (1989), there is a strong correlation between $S_{\text{Hmax}}$ orientation and absolute plate directions, especially in fast moving plates such as the South American. Richardson and Coblentz (1994) argued that the E-W-oriented $S_{\text{Hmax}}$ reflects far field rather than local sources. They assumed that the 3,000 m contour line in the Andes Cordillera was the boundary between extensional and compressional regimes, where nearly equal horizontal and compressional stress result in low seismicity. Their model is based on a density structure which balances the positive density contrast of the topography with a crustal roof, taking into account different lithospheric densities between the Andes region and the Precambrian shield. Assumpçâo (1992) remarked that the regional $S_{\text{Hmax}}$ presented no significant variation in areas of plate contact and far field areas. Furthermore, he suggested that significant differences between the observed orientation of the Nazca plate convergence and the $S_{\text{Hmax}}$ in the western South America meant that plate collision was not the only direct contributor to the stress field.

To summarise, the axes of maximum compression (P axis) are roughly horizontal in South America and point to an overall E-W intraplate compressive stress regime. However, important stress variations are observed within the plate. Despite poor understanding of the
Figure 2.2 - (a) Major geotectonic regions of the South American continent; the dashed line is the political boundary; (b) Structural provinces: 1, Rio Branco; 2, Tapajós; 3, São Francisco; 4, Tocantins; 5, Mantiqueira; 6, Borborema; 7, Amazon; 8, Parnaíba; 9, Paraná; 10, Coastal province and Continental margin (from Almeida et al. 1981). Black and white bars are the maximum horizontal compression ($S_{Hmax}$) after Assumpção (1992) and Lima et al. (1997).
seismology of the continent, especially in the cratonic region, most models assume that ridge push is the dominant source of stress.

**Northeastern Brazil**

The easternmost part of northeastern Brazil lies in the Borborema and Coastal provinces (Almeida *et al.* 1981) (Fig. 2.2b). The first of these represents the regional basement and is composed mainly of volcanic-sedimentary terrains and granite plutons deformed by one or more orogenic cycles (Jardim de Sá 1994). The coastal province is characterised by several basins formed during the South Atlantic opening. In the area of study, the Borborema province is composed predominantly of the Caico and Serido groups, deformed by Precambrian orogenic cycles; the Coastal province is represented by the Potiguar and Pernambuco-Paraíba basins, which formed during Mesozoic-Cainozoic times (Fig. 2.3).

A good deal of evidence indicates a complex evolution for both the Caico and the Serido groups (Jardim de Sá 1994). The oldest Precambrian rocks are found in the Caico group, which is comprised of orthogneisses (2.23 Ga to 2.18 Ga) and metavolcanics-sedimentary layers (Souza *et al.* 1993). The Serido group of early Proterozoic age is also composed by supracrustal sequences (paragneisses, quartzites, metavolcanic rocks and mica schists) which are cut across by granites (Jardim de Sá 1994). At least three orogenic cycles can be identified in the region: the Palaeotransamazonian orogeny (2.3-2.15 Ga), the Neotransamazonian orogeny (1.9 ± 0.1 - 1.95 ± 0.05 Ga), and the Brasiliano orogeny (650-550 Ma). The last of these strongly deformed and overprinted older structures throughout the area.

Intracontinental shear zones are one of the most important deformation features in the Precambrian basement (Fig. 2.3). Jardim de Sá (1994) remarked that the most striking peculiarity of the Precambrian basement orogeny is strain partitioning between domains of folding and strike-slip or oblique-slip mylonitic belts. He added that in the eastern part of the Borborema province, the Brasiliano structures display a transtensional style with negative flower structures and extensional detachments. By contrast, dextral compression is the most important deformation regime along the central and western part of the belt, where positive
flower, contractional positive duplexes and variable crustal thickening are the most common structures. The shear zones are aligned mainly E-W and NE to NNE, and reflect movements that are related to the same deformational event. The zones are generally marked by strong pervasive sub-vertical foliation and horizontal stretching lineation, and their deep roots are indicated by strong gravity anomalies (Lins et al. 1993; Jardim de Sá 1994).

The passive margin basins of the Coastal province such as the Potiguar and the Pernambuco-Paraíba and their related magmatism were formed mostly in the Mesozoic Era (Fig. 2.3). The Potiguar basin, whose age ranges from the early to the late Cretaceous, lies along the E-trending coast of Brazil and extends over a crystalline basement area of 41,000 km² both onshore and offshore (Souza 1982). Its structural framework comprises half grabens separated by intrabasinal highs related to the early Cretaceous rifting episode of the Brazilian marginal basins. Such half grabens are bounded by stable platforms to the west and to the southeast. The stratigraphic column of the Potiguar basin consists of clastic non-marine rift sediments and marine post-rift sediments. Deposition was controlled by NE-trending structures during the rift phase in the middle and late Cretaceous, giving way to its present geometry. Two main formations crop out in this region: the Açú formation, which is a clastic unit formed by an Albian-Turonian fluvial mega-cycle; and the Jandaíra formation, which represents a shallow carbonate shelf overlaying the former unit.

The Pernambuco-Paraíba basin is one of the less well known Atlantic-margin basins of Brazil (Fig. 2.3). Rand and Manso (1990) suggested that the N-S oriented coast, where this last basin occurs, was divided into several faulted blocks. More recently, Oliveira (1994) proposed a division of the eastern sedimentary zone between Recife and Natal into two sub-basins: the Cabo and the Pernambuco-Paraíba basins. He argued that the basins were separated by the E-W-striking Pernambuco shear zone (Fig. 2.3). According to his study, the Cabo basin displays major NNE-striking normal faults, whereas the Pernambuco-Paraíba basin shows NW and N-S-striking normal faults. Both groups of ruptures were affected by transfer faults with NW and NE trends. The Pernambuco-Paraíba basin described by Oliveira (1994) consists mainly of a ENE-axis half graben 250 m deep. The major boundaries between the Potiguar and the Pernambuco-Paraíba basins is still a problem. One proposal is
that the boundary between the Potiguar and the Pernambuco-Paraíba basins is marked by an important gravity anomaly located in the Goianinha valley (Rand 1977).

Many hypotheses have been presented to explain the origin of the Potiguar and the Pernambuco-Paraíba basins. Popoff (1988) argued that the Gondwana break-up reworked continental shear zones under a transtentional regime, leading to the formation of passive margin basins in both the African and the South America continents. He also stressed that the northeastern Brazilian margin represents the final link between both the Central and South Atlantic oceans. Equally, a great number of studies remarked that part of the Coastal province, where the Potiguar and Pernambuco-Paraíba basins occur, was the last part of the South American continent to be separated from Africa. Rand and Mabesoone (1982) and Mabesoone and Alheiros (1988) suggested that the Pernambuco-Paraíba basin was the last basin to be formed during the closing stages in the opening of the South Atlantic ocean during the Maastrichtian. A similar suggestion has been made for the sedimentary zone located between Touros and João Pessoa during the Lower Albian (Françolin and Szatmari 1987).

Two distinct structural origins have been proposed for the Potiguar basin. Françolin and Szatmari (1987) and Szatmari et al. (1987) postulated a sequence based on rigid plate rotation. They suggested that the rotation pole of the South America plate was located approximately at 39° W and 7° S. The position of the pole led to N-S-oriented crustal stretching east of the pole and E-W-oriented compression north of the pole during the Neocomian. At the same time, N-S-oriented compression and E-W-oriented stretching occurred to the west and to the east of the pole, respectively. As a consequence, NE-oriented Precambrian lineaments were reactivated as right-lateral strike-slip faults, and E-W-trending counterparts were reactivated as normal faults. Transpressional and transtensional regimes acted to the south and to the north of the Rio Ceará-Mirim magmatism and, once the E-W horizontal compression was over, NE-striking faults began to act as normal ones. According to Matos (1992), however, the geometry of the Equatorial basins including the Potiguar basin was strongly controlled by Proterozoic shear zones. In his view NW-SE-oriented
stretches was responsible for the NE-oriented normal faults and NW-trending transform faults.

In the Potiguar basin, the E-W-oriented dike swarm known as the Rio Ceará-Mirim magmatism (Gomes et al. 1981) (Fig. 2.3) was the most important magmatic event in the region during the Gondwana break-up. Belliene et al. (1992) argued that this igneous event is represented mainly by tholeiitic dikes of middle Jurassic (175-160 Ma) and early Cretaceous (140-130 Ma) age. They suggested that the Rio Ceará-Mirim magmatism correlates in terms of composition and tectonic setting with the Bennu trough dike swarm and that both dike swarms correspond to the early rifting events responsible for the South Atlantic opening.

The late Tertiary rocks of northeastern Brazil formed during the open ocean phase of continental drift are represented mostly by a non-fossiliferous sedimentary sequence called the Barreiras formation which crops out throughout the coastal region and in some areas as high as 700 m in the Martins, Cuité, Portalegre and Santana mountains (Mabesoone et al. 1972). Tertiary magmatic activity in the easternmost part of northeastern Brazil is represented mainly by the Mecejana volcanism in Ceará state and the Macau volcanism in Rio Grande do Norte and Paraíba states (Fig. 2.3). These rocks display alkaline features and have no counterparts in the Precambrian shield (Almeida et al. 1988). The Quaternary rocks of the region are composed mainly of continental and coastal deposits (see Chapter 3).

Seismicity and stress pattern

Northeastern Brazil has experienced many earthquake swarms with a recurrence period of approximately 4 years and each lasting for at least 6 months (Ferreira et al. 1987, Takeya et al. 1989). Within the northeastern region, the Potiguar basin boundary and its surroundings are the most important seismogenic areas. Seismic activity here has been known since 1804, when an earthquake of estimated magnitude $m_b = 4.0$ and intensity MM≥VI occurred near Açú city in the Rio Grande do Norte state (Ferreira and Assumpção 1983). The best known series of earthquakes are Pereiro (1968), whose main shock had a magnitude of $m_b=4.6$; Parazinho (1973), whose main shock had a magnitude of $m_b=4.3$; Cascavel (1980), whose
main shock had a magnitude of $m_b=5.2$; Palhano (1988), whose main shock had a magnitude of $m_b=4.9$; and João Câmara (main period from 1986 to 1989), whose main shock had a magnitude of $m_b=5.0$ (Assumpção et al. 1989).

The earthquake swarm of João Câmara is the most important seismic episode ever recorded in Brazil. Nearly 4,000 buildings were damaged in the 1980’s in João Câmara city, a small town located in the Rio Grande do Norte state, and the surrounding area. More than 40,000 events have been recorded in this area since 1986 (Ferreira et al. 1987; Takeya et al. 1989; Henderson et al. 1995). Seismic activity was concentrated along the Samambaia fault, which is defined by a narrow zone of epicentres with a strike of 040° strike, a dip of 70°, and a length of 30 km, oblique to the N-S to NNE strike of older Precambrian shear zones (Ferreira et al. 1987; Takeya et al. 1989) (Fig. 2.4). The fault zone could also be divided into two other sub-parallel faults in the southern part of the main rupture (Henderson et al. 1995).

The composite focal mechanism of the João Câmara earthquake swarm indicates that the $S_{Hmax}$ is E-W oriented in the area (focal mechanism a, Fig. 2.3). There was a second fault approximately 8 km long and 2.5-4.5 km deep and located east of the main zone near Poço Branco. Most of its seismogenic characteristics are similar to those of the Samambaia fault zone (Takeya et al. 1989). Although the seismic activity was shallow (5-10 km) and many neotectonic features have been described in the area (Lima et al. 1990; Bezerra et al. 1993; Torres 1994; Fonseca 1996), no offset related to the Samambaia fault has been detected so far.

The study of induced seismicity also shows a good match with the focal mechanisms of previous examples of natural seismicity. Ferreira et al. (1995) observed a clear correlation between water level and the number of seismic events in the Açú dam, which has a maximum depth of 31 m and a water volume of $2.4 \times 10^6$ m$^3$ and whose filling started in 1983. Ferreira et al. (1995) described a great number of small events above $m_b=1.7$, which were felt by local population from 1987 to 1994. Two major events of $m_b=2.6$ (1990) and $m_b=2.8$ (1994) were recorded by local seismographic stations. Epicentres were located inside the lake or near its border. Two main earthquake clusters were identified: the 1989 events, which were deeper
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Figure 2.3 - Tectonic framework of northeastern Brazil (modified from Schobenhouse et al. 1984 and Matos 1992). The focal mechanisms are after Assumpção (1992), Ferreira et al. (1997) and Ferreira et al. (1998): a, João Câmara; b, Açú; c, Augusto Severo; d, Tabuleiro Grande; e, Palhano; f, g and h, Cascavel; i and j, Irauçuba; k, Groafras; l, Hidrolândia.
than 3 km; and the 1990/1991 events, most of which are shallower than 2 km. The composite focal mechanism obtained by this study indicated right-lateral strike-slip movements with E-W horizontally oriented P-axes, favouring a NE-striking fault plane (focal mechanism b, Fig. 2.3). Ferreira et al. (1995) suggested that, in contrast with the Samambaia fault, the induced seismicity of the Açú dam occurred in pre-existing zones of weakness.

The Augusto Severo earthquake swarm (focal mechanism c, Fig. 2.3), from 1990 to 1991, had hypocentral depths concentrated between 0.5 and 3.0 km, and $S_{\text{Hmax}}$ consistent with the João Câmera and Açú focal mechanisms. The same pattern occurred in the Tabuleiro Gande earthquake swarm (focal mechanism d, Fig. 2.3), in 1993, with hypocentral depths between 1.0 and 2.5 km (Ferreira et al. 1997; Ferreira et al. 1998).

Other important seismic swarms have been described in the northern Ceará state. Another series of earthquakes occurred in 1980 near Cascavel, including the largest magnitude earthquake ever recorded in northeastern Brazil (magnitude $m_b=5.2$ and epicentral intensity MMI=VII). The depth of the $m_b=5.2$ earthquake was found to be 5.1 km, and its focal mechanism presented a strike-slip feature with an ESE-WNW-oriented $S_{\text{Hmax}}$ (Assumpção et al. 1985) (focal mechanisms f, g and h, Fig. 2.3). Likewise, the Palhano swarm, also located in the northern part of the Ceará state, was confined to a range depth of 2.5-5.0 km and a composite strike-slip focal mechanism with a small normal fault component (focal mechanism e, Fig. 2.3). Epicentre locations and focal mechanism data indicate that the P-axis ($S_{\text{Hmax}}$) was NW-SE oriented and strongly suggests an ESE-WNW-trending fault. Farther east, however, the 1991 Irauçuba swarm described by Assumpção (1992) had a focal depth of 7-11 km and a focal mechanism related to a right-lateral strike-slip fault, indicating that the $S_{\text{Hmax}}$ was roughly horizontal and oriented NW-SE (focal mechanisms i and j, Fig. 2.3). The Groafras and Hidrolândia earthquake swarms had a NW-SE- and E-W-oriented $S_{\text{HMax}}$, and maximum magnitudes $m_b=4.1$ and $m_b=2.4$ respectively (focal mechanisms k and l respectively, Fig. 2.3). The hypocentral depths did not exceed 12 km in either case (Ferreira et al. 1997).

More recently, Lima et al. (1997) and Ferreira et al. (1997) have pointed out that there is
good correlation between focal mechanism and breakout data in the Potiguar Basin. They concluded that the mean $S_{H\max}$ is roughly parallel to the coastline and that strike-slip is the dominate faulting mechanism.

Many seismological studies have tried to explain the relatively high seismicity of northeastern Brazil. Assumpção (1992) thinks that the asthenospheric shear stress and ridge push are not the only forces to influence the stress pattern in the region, and suggests a model combining superposition of local stress (extension caused by sedimentary loading and density contrast between continental and oceanic crust) and regional stress (ridge push and asthenospheric stress). According to his model, local and regional stresses act together to produce strike-slip seismic faults in the E-W-trending coast, whereas in the N-S-trending coast the stresses balance each other leading to a low zone of stress release and few earthquakes. Assumpção et al. (1985) and Assumpção (1992) argued that the orientation of major Precambrian shear zones play an important role in the regional seismicity. Thus there is no event greater than $m_b=4.0$ south of 6° S latitude because the majority of shear zones are oriented mainly E-W. On the other hand, north of that latitude the structural framework is mainly made up of NE-oriented shear zones which could be reactivated by E-W-oriented horizontal compression. Henderson et al. (1995) also argued that the migration of fluids, possibly caused by meteorological or climatic conditions, could be responsible for a small group of earthquakes along the Samambaia fault. According to them, such migration is not linked to the occurrence of a main shock.

From the published information, it can be concluded that all focal mechanisms of northeastern Brazil are consistent with the breakout data and indicate a strike-slip regime and a small normal component. The average $S_{H\max}$, including breakout measurements, have an orientation roughly parallel to the current coastline. The seismicity of northeastern Brazil occurs at shallow depths, usually as earthquake swarms, but no offset related to the seismogenic faults was found at any epicentral area in the region until the present, despite the fact that many studies have pointed out clear evidence of neotectonic activity throughout the region (Gusso and Bangnoli 1989; Lima et al. 1990; Saadi and Torquato 1992; Bezerra et al. 1993; Torres 1994; Fonseca 1996).
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Bento Fernandes
Cretaceous to Cainozoic sedimentary cover
Precambrian crystalline basement
Precambrian shear zone
Earthquake epicentre

Figure 2.4 - The João Câmara Epicentral area and the Samambaia seismogenic fault indicated by the epicentre cluster and arrows (modified from Takeya et al. 1989).
Chapter 3

CAINozoic stratigraphic units

This chapter describes the stratigraphic units in which the area’s neotectonics was investigated. The rock units in question form part of the Coastal province of Almeida et al. (1981). They are the Macau formation and the Barreiras formation, as well as various alluvial, aeolian and shallow-water marine sediments (Fig. 3.1). The local Precambrian and Cretaceous units have already been described in Chapter 2. The radiocarbon ages used to date and quantify deformation are presented in chapters 4 and 6. Shallow-water marine deposits which did not outcrop along the littoral zone or during low tides did not form part of the present research.

The Cainozoic units are displayed in the geological map (Fig. 3.1), which omits tectonic features for the sake of clarity. The Cainozoic stratigraphic units were mapped with the help of remotely sensed imagery (see Chapter 5). The Cretaceous Açú and Jandaíra formations, the Macau formation and the crystalline basement were derived from previous maps such as DNPM (1987a, b, c, Folhas Natal, Cabedelo and Areia Branca/Mossoró). The established stratigraphic Tertiary nomenclature was retained. No formal stratigraphic names were applied to the Quaternary units as there is a lack of consistency in the current literature, and the aeolian, alluvial and shallow-water marine deposits often cannot be distinguished as individual units.

Cainozoic and present climate

The origin of the Cainozoic sedimentary units is strongly dependent on climatic conditions. Northeastern Brazil has a warm and semi-arid tropical climate. The average temperature is ~30° C and the rainfall, which totals 600-1000 mm/yr, is concentrated mainly in this first half of the year (Nimer 1989).
Facies and pedological studies (e.g. Mabesoone and Lobo 1980; Alheiros et al. 1990) point to arid climatic conditions during the Miocene and Pliocene, with only slight oscillations in the littoral zone. During the Quaternary interglacials, the local climate was humid and warmer (Mabesoone and Rolim 1973); during the glacial periods it was much drier in the inner part of northeastern Brazil and humid tropical along the littoral zone (Damuth and Fairbridge 1970; Mabesoone and Rolim 1973). The great majority of Quaternary and middle to late Tertiary continental sediments were formed in oxidising conditions which did not favour the survival of fossils and which tended to obscure subtle differences in stratigraphy (Mabesoone and Rolim 1978).

**Macau formation**

The Macau formation consists of mafic volcanic rocks such as basalt, basanite, and ankaratrite which occur as plugs, dikes and necks in the Paraiba and Rio Grande do Norte states (Sial 1975, 1976; Almeida et al. 1988). In the study area, the Macau formation occurs south of Macau city, in the Açú delta, and cutting across the crystalline basement (Figs. 3.1 and 3.3). The Macau formation was also encountered in boreholes and indicated by geophysical studies (e.g. Salim et al. 1975).

Using the K-Ar method, Sial et al. (1981) dated 18 samples of the Macau formation from the Potiguar basin and obtained Oligocene ages of 29.0±0.9 Ma to 36.3±1.0 Ma. In contrast, Mizusaki (1989) obtained Eocene to Oligocene ages between 44.6±6.6 Ma to 29.0±0.9 Ma in the same area. In Paraiba state, south of the study area, Brito-Neves (1982) obtained late Eocene to Miocene ages of 19.1±1.0 Ma, 29.5±1.0 Ma and 37.0±3.0 Ma. An Eocene to Oligocene age has been adopted here for the unit (Fig. 3.2.).
Chapter 3 - Cainozoic stratigraphic units

Aeolian and shallow-water marine sediments (Pleistocene-Holocene)

Macau fm.-volcanic mafic rocks (Eocene-Oligocene)

Alluvial sediments (Pleistocene-Holocene)

Acu and Jandaira fms. - sandstones and limestones (Cretaceous)

Barreiras fm.-sandstones (Miocene-middle Pleistocene)

Precambrian basement (crystalline rocks)

Figure 3.1 - Geological map of the study area. Tectonic features are omitted and some units represented by one pattern for the sake of clarity.
Figure 3.2 - Simplified Cainozoic stratigraphic column for the study area.
Barreiras formation

The name Barreiras is applied to poorly lithified, siliciclastic continental sediments which have been studied since the pioneering work of Branner (1904). The Barreiras formation is the most widespread Cainozoic unit along the Brazilian coast, and outcrops from Rio de Janeiro state in southeastern Brazil to Pará state of northern Brazil. It formation is composed chiefly of sandstone, but includes layers of mudstone and conglomerate. It fossil content is limited to pollen. The formation forms plateaux as high as 200 m; outside the study area some plateaux are as high as 750 m. On the coast the formation forms cliffs up to 15 m high which rise abruptly from the foreshore zone (Fig. 3.4).

The age of the Barreiras formation has long been a source of debate. Relative chronological dating by micropollen and palaeomagnetism indicates that the Barreiras formation dates from Miocene to Pliocene times. A peat layer found in the Barreiras formation at Natal, described by Salim et al. (1975), yielded angiosperm pollen of Zonocostites ramonae. Although this species ranges from the Eocene to the Holocene, it is abundant in late Miocene sedimentary rocks. More recently, palynological studies of the upper Barreiras formation in the Potiguar basin yielded Retisteplanocolpites gracilis, indicating a Pliocene age (Lima et al. 1990). In Bahia state, 1,000 km to the south of the study area, a Pliocene age was also proposed for the base of the Barreiras formation by Suguio et al. (1986) on the basis of palaeomagnetic dating. The dates they obtained range from 4.5 to 5.0 Ma (early to middle Pliocene) near the base of the deposit to 3.0 to 3.4 Ma (late Pliocene) near the top.

Other studies have extended the age of the Barreiras formation into the Pleistocene on the basis of stratigraphic and palaeoclimatic correlations based on facies and pedological features respectively. Mabesoone et al. (1972), for example, concluded from pedological and stratigraphic correlations that the top part of the Barreiras formation near Macaíba city is of Pleistocene age. A study by Alheiros et al. (1988) based on facies and stratigraphic characteristics of sedimentary rocks supported the investigation
by Mabesoone et al. (1972) by placing the top of the Barreiras formation in the middle Pleistocene. For the purpose of the present study, the local Barreiras formation will therefore be taken to range from the Miocene to the middle Pleistocene (Fig. 3.2).

Alluvial sediments

The term alluvial sediments is used in the present study to designate all alluvial rocks which overlie the Barreiras formation. They are one of the most important stratigraphic units in the present study because they contain abundant evidence of neotectonic deformation. Although other stratigraphic units including the Barreiras formation are also composed of alluvial sediments, the term will be limited here to the post-Barreiras stratigraphic unit described below.

The alluvial unit includes not only sediments deposited in fluvial systems, but also sediments from deltas and estuaries (Fig. 3.1). The sediments are found within active river valleys such as the Açú, Potengi and Ceará-Mirim. These valleys are characterised by river gradients lower than 1% (Fonseca 1996) and by ephemeral streams, although the Açú valley became a perennial river after the impoundment of the Açú dam to the south of the study area. As in other semi-arid regions, alluvial sediments are transported by seasonal flash floods. They are usually oxidised owing to the dry conditions during deposition, and therefore contain little or no organic matter outside the deltas and estuaries (Figs. 3.5 and 3.6).

The most abundant type of sediments are gravelly and sandy braided deposits, some of which represent abandoned channels, whose grain size ranges from boulder to clay, sand and pebbles being the most common (Figs. 3.5 and 3.6). Mud and silt-rich sediments occur in some of other flood plain deposits, where the largest clasts such as boulders and pebbles are transported only during floods. Alluvial terraces are found in the Potengi and Açú valleys. Usually, the alluvial sediments interfinger with aeolian and shallow-water marine sediments along the littoral zone.
Figure 3.3 - Cabugi mountain - Tertiary basalt of the Macau formation (see Fig. 3.1 for location).

Figure 3.4 - Erosional unconformity (arrow) between the Barreiras formation (lower bed) and aeolian sediments (upper bed). Active cliff near Touros.
Figure 3.5 - Imbricated quartz pebbles in gravelly sediments (alluvial sediments), Açú valley.

Figure 3.6 - Gravel layer overlain by sandstone layer (alluvial sediments), Açú valley.
Most stratigraphic studies accept that deposits range in age from Pleistocene to Holocene (e.g. Mabesoone and Campos-e-Silva 1972; Salim and Coutinho 1973; Mabesoone and Rolim 1974; Mabesoone 1974) (Fig. 3.2). Silva (1991) obtained a Pleistocene age of 30,190 ± 370 yr and has confirmed the estimate age for post-Barreiras deposits in the Açú delta. No single stratigraphic name has been consistently used for the alluvial sediments. Local names including Moura formation have been used for deposits in the southern part of the Ceará state and along the Açú valley (Mabesoone and Campos-e-Silva 1972).

Aeolian sediments

Aeolian sediments consisting of stabilised and mobile sand dunes are concentrated in the littoral zone (Fig. 3.1). The vast majority of aeolian sediments are fixed by vegetation, ground water or small lakes. The mobile sand dunes are younger and are undergoing modification. This type is common along the coast and further inland where they have been reactivated by deforestation. Although it is easy to distinguish the two main types of aeolian sediments in the field and on satellite imagery, they cannot be distinguished on small scale maps such as Fig. 3.1.

The sediments are composed of well sorted quartz sands and, in places, of thin layers of heavy minerals such as rutile, titanite and ilmenite. Peat or other humic horizons frequently intercalate with sand. On the N-S-trending coast, the dunes extend as far as 15 km inland and display an elongated hairpin shape (Fig. 3.7). On the E-W-trending coast, the sand dunes give rise to barrier islands. Despite these differences, however, the sand dunes on both coasts have a maximum elevation of 100 m and display steep faces (Fig. 3.4). In the littoral zone between Touros and São Bento (Fig. 3.1) the top part of a regressive coastal sequence is composed of backshore rocks characterised by large-scale tangential cross-stratified and trough cross-stratified sandstone layers indicating an aeolian origin (Lima-Filho et al. 1995). The deposits display burrows up to 1 m long, mainly in the backshore facies, and teepee structures developed as a result of
calichification. The sand of the aeolian sediments originated mainly offshore by sea-
level fall, as indicated by the wind direction (Figs. 3.7 and 3.8), but alluvial and
lacustrine systems served as subordinate sources.

The aeolian sediments overlie the Barreiras formation (Fig. 3.4) and interdigitate with
alluvial and shallow-water marine sediments. A relative chronology was proposed by
Campos-e-Silva (1966), who distinguished one older generation characterised by white
sands and a younger generation with both white and yellowish sands. More recently,
Costa and Perrim (1981), identified three generations of aeolian sediment based on
colour and location of outcrop. A radiocarbon date by Silva (1991, see Chapter 4)
indicates that the upper part of some of the deposits dates from the Holocene.

**Shallow-water marine sediments**
The littoral zone of the study area includes a great variety of shallow-water marine
deposits. However, only a few were used as sea-level indicators (Chapter 4) or
displayed clear signs of neotectonic deformation (chapters 5 and 6). The shallow-water
marine sediments include a sequence of foreshore to shoreface sediments, beachrock
deposits, coral reefs, peats and tidal flats all of which overlie the Barreiras formation
and locally interfinger with the alluvial or aeolian sediments (Fig. 3.2). Generally,
exposures are too small to be mapped at the scale of the present research. Therefore,
they are represented in association with the aeolian sediments in Fig. 3.1.

**Foreshore to shoreface sediments**
Between Touros and São Bento, a calcarenite and sandstone of foreshore to shoreface
facies overlies the Barreiras formation by an erosional unconformity (Fig. 3.2) and is
overlain by aeolian sediments, which form cliffs up to 5-6 m high (Fig. 3.9).
Figure 3.7 - Mobile coastal dune showing hairpin shape to the south of Natal.

Figure 3.8 - Wind direction measured in the aeolian sediments (modified from Fortes 1982).
Lima-Filho et al. (1995) described two sedimentary facies which indicate marine regression. The lowest is composed of shoreface rocks characterised by tabular cross-stratified sandstone and conglomerate layers. The top part is represented by foreshore rocks characterised by herring-bone and swash cross-stratified sandstone layers (Fig. 3.10). These two facies contain fossils of red algae, foraminifers and molluscs. The foreshore/shoreface facies interpretation by Lima-Filho et al. (1995) for the basal and medium part of such a sequence is in agreement with the previous findings of Campos-e-Silva et al. (1964), who described shallow-water marine molluscs such as Anomalocardia sp. and Cerithium sp. in these rocks. Srivastava and Corsino (1984) have described aragonite cement, fossils such as red algae (Lithothamnium sp. and Lithophyllum sp.) and benthic foraminifers (Quinqueloculina sp., Triloculina sp., Pyrgo sp.) from these beds, which are characteristic of shallow-water marine environments. The foreshore to shoreface sediments were correlated by Srivastava and Corsino (1984) with the offshore shelf carbonate sediments of the Guamaré formation (late Cretaceous to Holocene), commonly intercalated with the coastal fan deposits of Tibau formation (late Cretaceous to Holocene) and suggested that they represent an uplifted part of the Guamaré and Tibau formations (see Chapter 4).

Beachrock

Beachrock is a common feature on the northeastern Brazilian littoral zone. The first studies were carried out by Darwin (1841) and Branner (1904). Morais (1969) investigated beachrock along the coast of Ceará state and suggested that their origin was related to ground water. Coutinho and Farias (1979) identified two main levels of beachrock along the Pernambuco coast and concluded they had formed in the foreshore zone by the dissolution and precipitation of organic fragments from submarine beach sediments. In addition, Oliveira et al. (1990) described two lines of Holocene beachrock of different ages on the Rio Grande do Norte beaches (see Chapter 4).
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Figure 3.9 - Active cliff formed by the foreshore to shoreface sediments, outcrop located ~4 km to the west of Touros.

Figure 3.10 - Herring-bone cross-stratification of shoreface facies (foreshore to shoreface sediments), outcrop ~1 km to the east of Touros.
Beachrock bodies vary widely in dimensions and shapes, but are usually parallel to the littoral zone. The only exception is the Recuado beachrock, which trends N80°W and makes an angle of 25° with the local N75°E-trending beach. In the majority of cases, beachrock bodies are elongated bodies ranging in length from a few kilometres (e.g. Perobas, Barreta, Guaraíra, Cunhauí, Jacumã, Farol de Sto. Alberto, Fig. 3.11, see Fig. 4.1 for location) to dozens of meters (e.g. Via Costeira, Guajiru, Galinhos, Fig. 4.1). Beachrock also occurs in patches (e.g. Pedra Grande, Recuado, Macau, Lagoa do Sal, Fig. 4.1). The beachrock bodies range in width from 50 cm to 3 m and present tabular sets whose boundaries are generally erosive. In a few cases, an erosional unconformity is seen between beachrock bodies and the Barreiras formation, but in general the base of the beachrock bodies does not outcrop, and they display gentle seaward-dipping bedding surfaces (<10°).

Erosional processes have modified the beachrock. In some cases, as at Touros and at São Bento, cemented beachrock bodies has been broken off into blocks which were incorporated into topographically lower beachrock of a later prograding sequence. In the littoral zone between São Bento and Touros, beachrock deposits also contain clasts of the foreshore to shoreface sediments and clearly postdate this material.

Comparisons between beachrock and modern shallow-water marine sediments were made in order to characterise the zone of beachrock cementation. In the study area, the normal tides attain a maximum of 1.0-2.0 m, whereas spring tides have a range of 3.2 m (Hayes 1979), which can penetrate into rivers as far as 20 km. Foreshore processes, however, are not affected by hurricane or tsunamis because such catastrophic events do not occur or are extremely rare along the Brazilian coast in comparison with coastal areas near plate margins (Bigarella 1972).

Beachrock formed by mesotidal beaches are strongly influenced by the tidal range. Semeniuk and Johnson (1982) showed that in Western Australia the shoreface is
characterised by trough-bedded sand and gravel, whereas the foreshore is distinguished by parallel-bedded sand and laminated-bubble sand. Dabrio (1982) described the sedimentary structures typical of mesotidal beaches in southern Spain as cross-laminated sand and cross-bedded sands overlain by cross-bedding, and finally by parallel-bedded sands. All these studies agree that physical processes do not change sharply at low water but extend to the shoreface zone. Despite that, the transition between the lower foreshore and the upper shoreface is marked by the accumulation of the coarsest available grain size, which is therefore a sea-level indicator (Dabrio 1982, Inden & Moore 1983).

Similar structures were seen in modern beaches and beachrock of the study area (Figs. 3.12, 3.13, 3.14 and 3.15) and were, therefore, used to identify former sea level. Two beachrock facies were identified mainly on the basis of sedimentary features and comparison with modern beaches. They will be referred to as facies (a) and (b). Although there is considerable lateral and vertical variation within the two types they can usually be distinguished.

Beachrock facies (a) represents the lower foreshore and the upper shoreface zones. It is a medium to coarse, sometimes conglomeratic, sandstone. Its terrigenous constituents are quartz, limonite, marine shell and fragments of the underlying rocks. It presents a great variety of textural types and mineralogical maturity indicating different formative processes and sources. The most common sedimentary structures are trough cross-stratifications 0.2 m to 1.5 m in thickness, interpreted as the result of migration of sinuous crested bedforms (Fig. 3.12 and 3.14), which points to an important traction flow mechanism during transportation. Palaeocurrents show transport predominantly to the N-NW on the N-S trending coast as far as Touros city, where the coast bends to the W. The palaeocurrents shift predominantly to the W and to the NW on the E-W trending coast.
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Figure 3.11 - General view of the Barreta beachrock.

Figure 3.12 - Recent trough cross-stratification in the lower foreshore zone, Barreta beach.

Figure 3.13 - Recent swash cross-stratification in the medium foreshore zone, Barreta beach.
Both palaeocurrent patterns are similar to those observed on the present coastline, which are influenced mainly by longshore currents. Although the shoreface zone can extend from the low tide level to the fairweather wave base (Reading and Collison 1996), the low-water level can in places be identified and used as a sea-level indicator with a precision of ±0.5 m. The low-water level is characterised by the coarsest texture associated with trough cross-stratification, sometimes capped by facies (b) (Fig. 3.15). Facies (a) corresponds to the lower foreshore beachrock which Flexor and Martin (1979) described in the Salvador littoral zones and the upper foreshore to lower shoreface beachrock described by Oliveira et al. (1990) at Cunhau-Natal.

Beachrock facies (b) corresponds to sediments deposited on the foreshore. They are usually medium to coarse sandstones, which form tabular beds and sheets from 0.1 m to 1.0 m in thickness. Facies (b) is composed chiefly of quartz grains, heavy minerals (ilmenite, magnetite, zircon, tourmaline, staurolite, and rutile), and fragments of marine shell. The grain size increases on the lower foreshore, where the low water level can sometimes be identified. The most common sedimentary structure of this facies is seaward dipping, swash cross-stratification (parallel bedding of previous studies) (Fig. 3.15), which is a sea-level indicator of the middle to lower foreshore with a precision of ±1.0 m. Other common structures on the foreshore are ripple marks, subcritically climbing strata, aeolian waves, and thin deflation pavement.

Important palaeoecological implications may be drawn from correlations between beachrock fossils and the modern beach fauna. There is no quantitative or qualitative difference between the fossil content of beachrock and death assemblages of marine shells which occur in modern beaches in northeastern Brazil (Maury 1934; Campos e Silva et al. 1964; Mendonça 1966). Death assemblages of shells tend to concentrate in the lower foreshore on modern beaches. In beachrock, they are similarly more abundant in the coarsest part of Facies (a). The most common species, which characterise
shallow-marine waters, in order of abundance are *Donax striata*, *Divaricella quadrisulcata*, *Tivela mactroides*, *Anomalocardia brasiliana*, *Anadara ovalis*, and *Ostrea* sp. According to these studies, the beachrock also includes *Anadara brasiliana*, *Anca campechensis*, *Arca* sp., *Trachycardium* sp., *Bulla* sp., *Lucina aproximata*, *Mulina branneri*, *Cerithium algicola*, *Corbula cf. uruguaensis*, *Codakia* sp., *Dentalium* sp., *Diodora listeri*, *Iphigenia brasiliana*, *Mactra alata*, *Neritina* sp., *Petaloconchus irregularis*, *Pitar circinata*, *Strigilla pisiformis*, and *Trivia* sp. Oliveira et al. (1990) and Bezerra et al. (1998) have proposed a diagenetic evolution for beachrock of the study area which affected facies (a) and (b). They described 5 stages related to change in the mineralogical composition of the cement. These diagenetic phases are marked by successive carbonate overgrowth and dissolution which include acicular crusts of aragonite, micrite grains of calcite, porosity reduction due to the growth of aragonite and Mg-calcite cement, crystalline calcite and porosity growth. The diagenetic phases indicate consecutive stages of foreshore uplift and subsidence, which are discussed in Chapter 4.

Other shallow-water marine sediments

Other foreshore deposits were investigated in order to secure additional evidence of sea-level changes. Among the most common shallow-water marine deposits are coral reefs, which overlie the Barreiras formation. The Pirangi and Jacumã coral-reef banks (see Fig. 4.1) occur as patches sometimes more than 1 km long and parallel to the present shoreline. They rise from the sea floor up to 5 m, 1 m of which is exposed at low tide. Exposure to sunlight and dry conditions during low tides may have killed the coral reefs: most of the reef mass exposed in both of the exposures that were sampled is made up of dead organisms, so that the original growth position may account for only a small part of the whole bank. Coral reefs are not necessary good indicators of sea level but reefs in the study area represent the minimum low-water level at the time of their death. A detailed investigation of the coral-reef fauna of the study area was presented by Kempf and Laborel (1967).
Figure 3.14 - Trough cross-stratification in facies a of the Barreta beachrock.

Figure 3.15 - Boundary (see arrow) between trough cross-stratification (facies a) and swash cross-stratification (facies b), Barreta beachrock.
Another usual type of shallow-water marine deposit are raised tidal flats, which are composed of shell-rich layers 10-30 cm thick in sand and mud. Most of them have formed over existing aeolian or estuarine-deltaic sediments. They are more frequent along the E-W-trending coast. The largest occur in the Açú delta and to the south of Galinhos, where salt flats are also found. The latter are partly covered by vegetation. The area is protected from direct wave action by a large spit to the north. Tidal flats are commonly dissected by tidal channels, some of which are tectonically controlled (see Chapter 5). They represent mean sea level with a precision of ± 1.0 m. Their fossil content is very similar to that described for the Holocene beachrock bodies.

Less common are peats of mangrove-swamp origin, which display a minimum thickness of 1.5 m and are composed mainly of wood fragments (tree branches) near the top and mud near the bottom (Figs. 3.16 and 3.17). The upper part of the peats was deposited in the middle to upper foreshore, and the bottom part in the middle to lower foreshore (Martin et al. 1996b). The peats occur locally as thin layers intercalated in foreshore sands. Shells in living position are found mostly in the bottom layer of large outcrops (Fig. 3.17), and can be used as sea-level indicator with a precision of ± 1.0 m. For example a marine fauna composed mainly of Arca sp. in growth position occurs to the east of São Bento. Its position indicates the minimum height of low-water level before its death.

In short we have in the study area Cainozoic deposits from a wide range of depositional environments many of which bear on former sea levels or can be used to measure and date deformation.
Figure 3.16 - General view of the Rio do Fogo peat overlain by aeolian sediments.

Figure 3.17 - Shell-bearing bottom layer and wood fragment bearing upper layer of the Rio do Fogo peat.
Chapter 4
QUATERNARY FAULTS: EVIDENCE FROM COASTAL EMERGENCE AND SUBMERGENCE

The primary goal of this Chapter is to investigate the height and radiocarbon age of the coastal deposits described in Chapter 3 in order to identify late Quaternary coastal emergence and submergence. Holocene sea-level changes in the study area are compared with glacioisostatic predictions for the region in an attempt to isolate tectonic movements. An attempt is also to extend the tectonic chronology into the Pleistocene.

Several sea-level studies along the east coast of South America describe changes that have taken place during the Quaternary but little information is available on the relative importance of sea level, neotectonics and isostatic rebound. Tricart (1959) observed tectonic deformation in the littoral zone near Salvador (see location in Fig. 2.3) and in the south of that city. Mabesoone and Coutinho (1970) also identified neotectonic activity on the coast of northeastern Brazil. Martin et al. (1980) described vertical movements which had displaced shorelines during the Holocene in the Guanabara graben and south of Cape São Tomé in southeastern Brazil. Martin et al. (1986a) also described vertical block movements in the Recôncavo basin, near Salvador, and claimed that it had displaced Quaternary shorelines during the Holocene by as much as 3 m. In their view several of the faults that bound the blocks are still active.

Apart from Bezerra et al. (1998), only two studies analyse neotectonic deformation in the study area with the help of coastal deposits. According to Srivastava and Corsino (1984), the foreshore to shoreface sediments can be correlated on the basis of petrography with intercalated rocks of the Tibau and Guamaré formations, which are Campanian to Recent (Araripe and Feijó 1994), and the vertical faults they display may have remained active. Silva (1991) also observed faults cutting across Cainozoic sedimentary rocks in the Açú delta and concluded that they had affected Quaternary sedimentary deposition there. He reported that Pleistocene as well as older fluvial and marine sediments in the delta had been downfaulted by more than 45 m.
The present study faced two obstacles: which sea-level curve to use as datum against which
tectonics could be evaluated; and which sea-level indicator to use along the littoral zone.
The best constrained curve for the coast of northeastern Brazil is the Salvador curve of
Suguio et al. (1985), which is based on 60 $^{14}C$ ages for intertidal organisms in living
position, marine terraces and shell middens collected along 50 km of littoral zone north of
Salvador. The Salvador curve predicts that relative sea level started to rise about 7,000 yr
BP and reached a highstand shortly before 5,000 yr BP. During the ensuing regression there
were two lowstands at 3,800 yr BP and 2,700 yr BP. The Salvador curve was used by
Martin et al. (1986a) as the reference datum for assessing deformation near Salvador.
Oliveira et al. (1990) made an attempt at a sea-level curve for the present study area but
armed with only four $^{14}C$ ages they were forced to adopt the Salvador curve.

Two glacioisostatic predictions for the Recife area were presented by Clark et al. (1978) and
Peltier (1988). They were based on the sea-level curve for the Brazilian coast of Fairbridge
(1976). These sea-level predictions have been supplanted by recent glacioisostatic
predictions; in any case the Recife area is more than 200 km south of the study area. For this
reason, the glacioisostatic predictions by Peltier (1997, pers. comm.) for the Touros area
hereafter called the Touros curve are used as the reference datum in this investigation.

The second problem is the choice of sea-level indicators. Keith et al. (1964) suggested
that dating death assemblages should be avoided because they are likely to include
specimens from a range of environment and ages. Nevertheless, Richards (1982) drew
attention to the fact that even shells in living positions could lead to erroneous
interpretation if they were not carefully identified. He went on to demonstrated that
zonation patterns can be used to support radiocarbon dating of death assemblages in the
reconstruction of sea level. In this study, death assemblages for beachrock are used for
dating and the sedimentary facies defined in Chapter 3 are used as sea-level indicators.
Although the shells sampled were not found in living position in beachrock or tidal flats,
they displayed their original colours and lacked any of the features to be expected from
transportation. In the beachrock and marine terraces of the Catanduba, Rio do Fogo
and Recuado sites, however, the sampled shells were in living position.
Radiocarbon and isotopic analyses of shells

In many shorelines throughout the world shells are the only widespread available biogenic material for radiocarbon measurements. From the earliest days of radiocarbon dating, however, shell ages have been considered by many workers as inherently undependable. For example, the first edition of *Radiocarbon Dating* by Libby (1952) lists various types of material in order of reliability. Charcoal and cloth head the list, whereas shells are at the bottom. This view owes much to poor practice. With care, shells can give dependable and reproducible results. The key is to guard against contamination by old and young carbon produced by the hardwater and reservoir effects, overgrowth and recrystallisation, and the fractionation effect.

The contamination of “old” carbon produced by the reservoir and hardwater effects takes place when a living organism takes up dissolved carbon dioxides or bicarbonates from “old” and deep ocean waters which are not in equilibrium with the atmosphere or from limestone areas. Variations caused by the reservoir and hardwater effects are problematic to correct because living organisms take different degrees of “old” carbon from the environment (Goodfriend and Stipp 1983). The hardwater effect is usually associated with continental contamination and has been described as one of the most serious problems in radiocarbon dating of shells. According to Pilcher (1991) it may add between 200 and 1,200 years to the apparent age of a sample. The hardwater effect is of course more serious in terrestrial molluscs, which are located in closed environments and can be contaminated by significant amounts of local carbonate (Hedges 1992). Goodfriend and Stipp (1983), for instance, presented strong evidence that land-snail shells from limestone areas could produce radiocarbon anomalies of around 300 yr. Shells from estuaries, lagoons and lakes may also concentrate of ‘old’ carbon (Broecker and Olson 1961). In open coastal zones, however, especially if there is no restriction in water circulation, as in the vast majority of cases in this study, marine shells are not likely to suffer from the hardwater effect.

The reservoir effect has been detected in many coasts around the world and is generated by upwelling from deep oceans which display low mixing $^{14}$C rates with the atmosphere.
(Bowman 1990). Studies such as those by Keith et al. (1964) and Taylor and Slota (1979) concluded that marine shells from open ocean environments, where upwelling and carbon isotopes ratios are known, display small variations of $\delta^{13}C$ and can provide reliable $^{14}C$ ages.

The fractionation effect is not important in marine shells as it is for certain plants such as *Claudium* which are rooted in old organic sediments and take up a significant proportion of their carbon from them (Olsson 1986). Because the correction for fractionation is approximately + 400 years (Donner and Junger 1975; Flessa and Kowalewski 1994) and the reservoir correction is commonly -400 years (Stuiver and Polach 1977), many laboratories have used the practice of assuming that the two corrections balance out. Several studies had demonstrated that, although the outer layer of shells exchanges carbon with the environment during diagenesis (e.g. Chappell and Polach 1972), the inner layer can be considered a closed system.

**Sampling and pretreatment**

A group of 29 samples were selected for dating: 18 shell samples from beachrock, mostly of facies (a); 4 shell samples from peat, tidal flat deposits and other marine sediments; 3 whole-rock samples of coral reef; one sample of live shell; and 3 additional samples from the foreshore to shoreface sediments (Fig. 4.1).

Care was taken in the field to determine the altitude and geographic location of each sample accurately. The reference datum for vertical measurements was the ‘Corrego Alegre’ Brazilian national datum. Altitudes were determined by levelling and were corrected for the local ports of Natal and Macau following the procedure recommended by the Admiralty (1996). Away from settlements location was determined using a portable GPS device.

Pretreatment was designed to detect and if necessary eliminate contamination. Visual inspection was followed by scrutiny in thin section of acetate peels under the light microscope, followed when appropriate by X-ray diffraction. Samples showing
extensive recrystallisation were rejected. The outer layer of shells selected for dating was accordingly removed by both mechanical cleaning and acid leaching. The remaining inner layer of each sample were carefully examined for contamination. Scanning electron microscopic analysis (SEM) was used to identify contamination not detected by X-ray diffraction or acetate peels and to distinguish between primary and secondary aragonite or calcite.

Most of the shells from the beachrock, peat and tidal flats display no contamination. The X-ray diffraction (XRD) analyses (Fig. 4.2) show that all the samples of marine bivalves selected consist of pure aragonite. Evidence of recrystallisation of aragonite to calcite were found in three samples of beachrock: in trace amounts in samples REC2 and BR, and significantly in sample CH2 (Fig. 4.3). In addition, all the samples from the foreshore to shoreface sediments were found to display significant amount of recrystallisation of aragonite to calcite. The results of scanning electron microscopy (SEM) confirmed the XRD analyses and showed that the aragonite samples consisted of cross-lamellar and other original structures (Fig. 4.4). A summary of the XRD and SEM results is presented in Table 4.2 together with the ^{14}C ages.

**Radiocarbon dating and isotopic analysis (**$^{\delta^{13}}$C, $^{\delta^{18}}$O**)

**Procedures**

Radiocarbon dating was carried out at University College London by the first order method (Vita-Finzi 1991), which presents good agreement with conventional $^{14}$C ages. Fig. 4.5 shows the rig currently in use at University College London to extract CO$_2$ from carbonate samples. 50% HCl is dropped on the sample and the CO$_2$ generated passed through the system for a few seconds to purge the air. The system is then linked to the trapping device and CO$_2$ bubbled through the Permafluor V-Carbosorb mixture (m) until it is saturated. Total absorption of CO$_2$ by the mixture is over in 30-40 minutes, and is recognised in three ways: by weighing the sample until the maximum gain of 1.2 g per 10 ml of Carbosorb is achieved; by appreciable cooling of the sample; and when bubbles in the mixture increase markedly in size.
Figure 4.1 - Simplified geological map of the study area and sample sites indicated by letters.
Figure 4.2 - X-ray diffraction analyses of marine shells. Results for each analysis are presented between 20 and 34 °0. A bigger scale is used for samples GR3 and REC1; (a) stands for aragonite.
Figure 4.3 - X-ray diffraction analyses of marine shells. Results for each analysis are presented between 20 and 34 °0. Different scales are used for the left and right-hand columns, (a) stands for aragonite and (c) for calcite.

The saturated samples were counted in a low level 2260XL Liquid Scintillation Analyser. The counting period was 1,000 minutes for each sample. The background activity was measured in each session of ¹⁴C dating and, together with the stability of the machine, made it possible to extend the age assessment back to 16,000 years. The age determination uses the formula:

$$t = \frac{X}{\lambda} \log_e \left( \frac{I}{I_0} \right)$$

where, $t = \text{time}$, $\lambda = ¹⁴C$ halflife (5,568 yrs), $I = \text{sample activity}$, $I_0 = \text{modern activity}$. A graph for preliminary age estimates, where sample activity is expressed in counts per minute (cpm) after subtraction of the background value, is presented in Fig. 4.6. The modern standard consists of marine shells (British Museum specimens *Pecten maximus* and *Arctica islandica*) collected live before the atomic testing era (Vita-Finzi 1983).
Figure 4.4 - Electron scanning microscope photos of cross-lamellar structure in aragonite shells: (a) sample CH2, (b) sample BR2. Note absence of overgrowth.
The ages and errors of the radiocarbon ages by Bezerra et al. (1998), as well as those by Oliveira et al. (1990) and Silva (1991), were calibrated to 2σ after the curve proposed by Stuiver et al. (1986) (Table 4.1). The activity determined for a specimen collected live (MC2), one cpm above the pre-bomb level of 7.7 cpm above background, shows that the residual effect of the atmospheric bomb tests swamps any apparent reservoir effect within the resolution of the method.

The precision presented here is based on Packard’s operation manual for the Tri-Carb 2260XL Counter. The ages and errors were subjected to a rounding off, when errors were rounded to the nearest higher multiple of ten. The final error represents 2σ percentage of the sample accumulated counts as follows:

\[ E = \frac{[(C_s - C_b)^2]^{1/2} \times 100 \times \text{age}}{C_s} \]

where \( E = \text{error}, C_s = \text{total counts}, C_b = \text{background counts.} \)

<table>
<thead>
<tr>
<th>Age (yr)</th>
<th>Error (yr)</th>
<th>Total counts (Cs)</th>
<th>Background counts (Cb)</th>
<th>Sample counts (t-b)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2,700</td>
<td>200</td>
<td>14,500</td>
<td>8,700</td>
<td>5,800</td>
</tr>
</tbody>
</table>

Isotopic analysis was carried out on the shells and beachrock both to check the reliability of the \(^{14}\text{C} \) ages and for palaeoenvironmental purposes. They were not used to normalise ages to \( \delta^{13}\text{C} = -25\% \) PDB, as it is a dubious practice when applied to marine organisms whose oceanic composition is close to 0%o (Vita-Finzi and McClure 1991). The analyses were carried out on a group of 22 samples: 19 of shells, 17 of them from death assemblages in Holocene deposits, one from a death assemblage in a modern beach and another from a live shell; and 3 samples of beachrock whole-rock samples.

The samples were run at the Department of Geology, Royal Holloway and Bedford New College (University of London), using a VG ISOCARB automatic system and a VG SIRA II triple collector mass spectrometer. The samples were reacted a 90°C with
100% orthophosphoric acid under vacuum to produce carbon dioxide gas. The gas was dried by passing through a methanol cold trap at -100°C. The purified gas was then admitted to the mass spectrometer. The precision of the system for carbon and oxygen is 0.1‰ for 2SD.

**Results and interpretation**

The result of radiocarbon analysis are presented in Table 4.2. Further ages determined by conventional radiometric dating in previous studies (Oliveira et al. 1990, Silva 1991), are presented in Table 4.3. The isotopic analyses are presented in Table 4.4.

From in Fig. 4.7, it can be seen that the shells display a small isotopic range. The δ¹³C values extend from 0.8‰ to -2.4‰, and δ¹⁸O values extend from 0.0‰ to 1.2‰. The shell assemblage plot partially in the F field (marine coastal shells of Vita-Finzi 1992) and entirely in the A and B fields (mollusc shells of transitional environment of Keith and Parker 1965). None of the marine shells from the study area are displaced towards the freshwater, continental shell fields (fields D and E in the diagram of Fig. 4.7), ruling out any continental influence in shell composition. Likewise, any reservoir effect, which was swamped by the bomb effect (see radiocarbon dating), is ruled out by the isotopic data.

There is no isotopic evidence for fractionation in the δ¹³C values.

Broadly speaking, the lowest δ¹⁸O values correspond with the highest sea-level temperatures within a single shell (e.g. Killingly and Berger 1979, Talma et al. 1992). From diagram 4.8 it can be seen that the δ¹⁸O in the analysed samples range from about -1.0 to 0.0‰. The oldest Holocene samples (CH1, CH2, FSA1, FSA2, VC, GR3) display a δ¹⁸O range between -0.8 and -0.45‰. The δ¹⁸O values increase in 5,000-4,000 cal. yr old samples from -1.0 to -0.3‰, and then fall to -0.7 and -0.8‰ for samples about 4,000 cal yr BP. The δ¹⁸O values increase to -0.2 and 0.0‰ in samples GA and PG1. They finally decrease in samples MC1 and the modern samples PG2 and MC2 (Fig. 4.8). The small range of δ¹⁸O values (0.02 to -0.94‰) in the marine shells indicates there were no great shifts in local temperature during the Holocene.
Figure 4.5 - Rig for first-order radiocarbon dating currently in use at the department of Geological Sciences at University College London.

Figure 4.6 - Log scale diagram. The modern standard used is a marine shell collected alive before the atomic testing era (Vita-Finzi 1983, 1991).
## Table 4.2 - List of $^{14}$C ages used in this study (Holocene ages are from Bezerra et al. 1998).

<table>
<thead>
<tr>
<th>Sample</th>
<th>Height (m asl)</th>
<th>Deposit</th>
<th>XRD SEM</th>
<th>$^{14}$C age (yr BP)</th>
<th>Calibrated age (yr BP at 2σ)</th>
</tr>
</thead>
<tbody>
<tr>
<td>PG1</td>
<td>+0.60</td>
<td>UCL-423</td>
<td>b A</td>
<td>2,700±80</td>
<td>2,350 ±330</td>
</tr>
<tr>
<td>MC1</td>
<td>+1.80</td>
<td>UCL-354</td>
<td>b A</td>
<td>1,600±40</td>
<td>1,550 ±90</td>
</tr>
<tr>
<td>MC2</td>
<td>+1.80</td>
<td>UCL-418</td>
<td>(live)</td>
<td>---</td>
<td>---</td>
</tr>
<tr>
<td>CDB1</td>
<td>+0.30</td>
<td>UCL-433</td>
<td>tf A</td>
<td>3,950±110</td>
<td>3,930 ±320</td>
</tr>
<tr>
<td>CDB2</td>
<td>+0.30</td>
<td>UCL-434</td>
<td>tf C</td>
<td>4,500±130</td>
<td>4,700 ±280</td>
</tr>
<tr>
<td>GA1</td>
<td>+1.10</td>
<td>UCL-416</td>
<td>b A</td>
<td>3,550±100</td>
<td>3,430 ±240</td>
</tr>
<tr>
<td>FSA1</td>
<td>-0.50</td>
<td>UCL-410</td>
<td>b A</td>
<td>6,050±190</td>
<td>6,460 ±430</td>
</tr>
<tr>
<td>FSA2</td>
<td>+0.50</td>
<td>UCL-411</td>
<td>b A</td>
<td>6,550±210</td>
<td>7,050 ±360</td>
</tr>
<tr>
<td>REC1</td>
<td>+3.90</td>
<td>UCL-397</td>
<td>b A</td>
<td>5,100±140</td>
<td>5,450 ±290</td>
</tr>
<tr>
<td>REC2</td>
<td>+5.40</td>
<td>UCL-393</td>
<td>mt A/</td>
<td>4,050±110</td>
<td>4,070 ±320</td>
</tr>
<tr>
<td>GU</td>
<td>+0.60</td>
<td>UCL-431</td>
<td>b A</td>
<td>3,050±90</td>
<td>2,760 ±220</td>
</tr>
<tr>
<td>SAL</td>
<td>+1.20</td>
<td>UCL-417</td>
<td>b A</td>
<td>3,950±110</td>
<td>3,930 ±320</td>
</tr>
<tr>
<td>PB</td>
<td>+0.10</td>
<td>UCL-420</td>
<td>b A</td>
<td>4,500±130</td>
<td>4,710 ±280</td>
</tr>
<tr>
<td>RF</td>
<td>-0.20</td>
<td>UCL-409</td>
<td>pd A</td>
<td>3,750±110</td>
<td>3,670 ±260</td>
</tr>
<tr>
<td>PJC</td>
<td>-0.40</td>
<td>UCL-424</td>
<td>cr</td>
<td>1,450±40</td>
<td>950 ±90</td>
</tr>
</tbody>
</table>

Coastal deposit code: b (beachrock); cr (coral reef); pd (peat deposit); mt (marine terrace); td (tidal flat); mfs (marine and fluvial sediments); fss (foreshore to shoreface sediments). X-ray and SEM code: (A) primary aragonite; (C) primary calcite; (sC) secondary calcite; (tr) mineral in trace quantity.
Continuation of Table 4.2.

<table>
<thead>
<tr>
<th>Sample</th>
<th>Height (m asl)</th>
<th>Lab. number</th>
<th>Deposit</th>
<th>XRD</th>
<th>SEM</th>
<th>(^{14}C) age (yr BP)</th>
<th>Calibrated age (yr BP at 2σ)</th>
</tr>
</thead>
<tbody>
<tr>
<td>JC</td>
<td>+0.50</td>
<td>UCL-413</td>
<td>b</td>
<td>A</td>
<td></td>
<td>4,950±150</td>
<td>5,280±520</td>
</tr>
<tr>
<td>VC</td>
<td>-0.70</td>
<td>UCL-430</td>
<td>b</td>
<td>A</td>
<td></td>
<td>6,300±200</td>
<td>6,570±440</td>
</tr>
<tr>
<td>PR2</td>
<td>+0.10</td>
<td>UCL-425</td>
<td>cr</td>
<td>---</td>
<td>---</td>
<td>1,150±30</td>
<td>680±60</td>
</tr>
<tr>
<td>PR3</td>
<td>-0.50</td>
<td>UCL-361</td>
<td>cr</td>
<td>---</td>
<td>---</td>
<td>950±30</td>
<td>530±40</td>
</tr>
<tr>
<td>BR1</td>
<td>+2.20</td>
<td>UCL-403</td>
<td>b</td>
<td>A</td>
<td></td>
<td>4,700±140</td>
<td>4,880±420</td>
</tr>
<tr>
<td>BR2</td>
<td>+1.80</td>
<td>UCL-404</td>
<td>b</td>
<td>A/</td>
<td>tr</td>
<td>4,500±120</td>
<td>4,710±260</td>
</tr>
<tr>
<td>GR1</td>
<td>+0.20</td>
<td>UCL-419</td>
<td>b</td>
<td>A</td>
<td></td>
<td>5,600±170</td>
<td>5,970±380</td>
</tr>
<tr>
<td>GR2</td>
<td>0.00</td>
<td>UCL-421</td>
<td>b</td>
<td>A</td>
<td></td>
<td>6,550±210</td>
<td>7,070±360</td>
</tr>
<tr>
<td>GR3</td>
<td>+0.70</td>
<td>UCL-405</td>
<td>b</td>
<td>A</td>
<td></td>
<td>5,950±170</td>
<td>6,370±370</td>
</tr>
<tr>
<td>CH2</td>
<td>+1.70</td>
<td>UCL-432</td>
<td>b</td>
<td>A/</td>
<td>sC</td>
<td>5,400±170</td>
<td>5,790±420</td>
</tr>
<tr>
<td>CH1</td>
<td>+1.50</td>
<td>UCL-414</td>
<td>b</td>
<td>A</td>
<td></td>
<td>6,550±210</td>
<td>7,070±360</td>
</tr>
<tr>
<td>SM1</td>
<td>+2.50</td>
<td>UCL-412</td>
<td>fss</td>
<td>A/s</td>
<td>C</td>
<td>&gt;16,000</td>
<td></td>
</tr>
<tr>
<td>SM2</td>
<td>+7.00</td>
<td>UCL-415</td>
<td>fss</td>
<td>A/s</td>
<td>C</td>
<td>&gt;16,000</td>
<td></td>
</tr>
<tr>
<td>SM3</td>
<td>+7.00</td>
<td>UCL-422</td>
<td>fss</td>
<td>A/s</td>
<td>C</td>
<td>&gt;16,000</td>
<td></td>
</tr>
</tbody>
</table>

Coastal deposit code: b (beachrock); cr (coral reef); pd (peat deposit); mt (marine terrace); td (tidal flat); mfs (marine and fluvial sediments); fss (foreshore to shoreface sediments). X-ray and SEM code: (A) primary aragonite; (C) primary calcite; (sC) secondary calcite; (tr) mineral in trace quantity.
Table 4.3 - List of published $^{14}$C ages used in this study.

<table>
<thead>
<tr>
<th>Sample/Source</th>
<th>Height (m asl)</th>
<th>Deposit</th>
<th>$^{14}$C age (yr BP)</th>
<th>Calibrated age (yr BP at 2σ)</th>
</tr>
</thead>
<tbody>
<tr>
<td>P20/(1)</td>
<td>+1.0-2.0</td>
<td>b</td>
<td>6,067±80</td>
<td>6,490±150</td>
</tr>
<tr>
<td>P21/(1)</td>
<td>+1.0-2.0</td>
<td>b</td>
<td>6,067±100</td>
<td>6,370±120</td>
</tr>
<tr>
<td>P14/(1)</td>
<td>+3.00</td>
<td>b</td>
<td>4,737±130</td>
<td>4,970±240</td>
</tr>
<tr>
<td>P29/(1)</td>
<td>+3.00</td>
<td>b</td>
<td>4,609±100</td>
<td>4,830±320</td>
</tr>
<tr>
<td>S-2</td>
<td>-18.5</td>
<td>mfs</td>
<td>6,060±80</td>
<td>6,940±150</td>
</tr>
<tr>
<td>BL/(2)</td>
<td>+1.0(?)</td>
<td>mfs</td>
<td>6,060±80</td>
<td>6,480±220</td>
</tr>
<tr>
<td>S-6/(2)</td>
<td>-3.0-4.0</td>
<td>mfs</td>
<td>2,340±60</td>
<td>2,350±350</td>
</tr>
</tbody>
</table>

Coastal deposit code: b (beachrock); mfs (marine and fluvial sediments).

Sample source: (1) Oliveira et al. 1990; (2) Silva 1991.

At least 5 different stages of beachrock diagenesis, including recrystallisation and overgrowth, have been described by Bezerra et al. (1998) (see Chapter 3). As isotopic analysis of the cement was carried out on whole rock samples, only the youngest beachrock bodies which bear only the initial cement phase were analysed (Fig. 4.9). The $\delta^{18}$O values of the cement range from 0.4‰ to 0.3‰, and the $\delta^{13}$C values range from 0.7‰ to 3.4‰. Samples from the study area plot in the field of the Red Sea and Mediterranean cements, and are in accord with the general beachrock cement composition worldwide, with positive $\delta^{13}$C values and $\delta^{18}$O values between - 4.0‰ and 2.0‰.

Beachrock bodies which have undergone several stages of cementation and diagenesis may present a $\delta^{13}$C composition shifted towards negative values, i.e., $\delta^{13}$C values displaced towards the vadose zone in Fig. 4.9 (Magaritz et al. 1979, Beier 1985). From Fig. 4.9 it can be concluded that such a shift has not occurred in the youngest beachrock of the study as expected. Nevertheless the isotopic and petrographic data
suggest that beachrock cement precipitation occurred in mixed meteoric-marine conditions. The isotopic composition of one sample of beachrock cement (PG) plots in field B (Fig. 4.9, Mediterranean beachrock of Holail and Rashed 1992), whereas two other (JC and PB samples) in field A (Fig. 4.9, Red Sea beachrock of Halail and Rashed). This agreement with Mediterranean and other beachrock data supports the idea expressed in Chapter 3 that the beachrock formed in the transitional (beach) environment and not in the continental backshore subenvironment.

Table 4.4 - Isotopic analysis of shells of various coastal sediments and beachrock cement expressed in relation to the Beleminte PeeDee Bee Standard\(^1\) - PDB (\(\delta^{13}\text{C}, \delta^{18}\text{O}\)) in parts per thousand.

<table>
<thead>
<tr>
<th>Sample</th>
<th>material analysed</th>
<th>(\delta^{13}\text{C}_{\text{PDB}}) (%)</th>
<th>(\delta^{18}\text{O}_{\text{PDB}}) (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>PG1</td>
<td>whole rock</td>
<td>0.70</td>
<td>-0.44</td>
</tr>
<tr>
<td>JC</td>
<td>whole rock</td>
<td>2.98</td>
<td>0.24</td>
</tr>
<tr>
<td>PB</td>
<td>whole rock</td>
<td>3.30</td>
<td>0.23</td>
</tr>
<tr>
<td>JC</td>
<td>beachrock shell</td>
<td>0.27</td>
<td>-0.38</td>
</tr>
<tr>
<td>VC</td>
<td>beachrock shell</td>
<td>-0.50</td>
<td>-0.66</td>
</tr>
<tr>
<td>GR3</td>
<td>beachrock shell</td>
<td>-0.50</td>
<td>-0.70</td>
</tr>
<tr>
<td>MC2</td>
<td>live shell</td>
<td>-2.37</td>
<td>-0.80</td>
</tr>
<tr>
<td>CH1</td>
<td>beachrock shell</td>
<td>-2.25</td>
<td>-0.80</td>
</tr>
<tr>
<td>MC1</td>
<td>beachrock shell</td>
<td>0.48</td>
<td>-0.41</td>
</tr>
<tr>
<td>FSA1</td>
<td>beachrock shell</td>
<td>-1.23</td>
<td>-0.41</td>
</tr>
<tr>
<td>FSA2</td>
<td>beachrock shell</td>
<td>-0.73</td>
<td>-0.46</td>
</tr>
<tr>
<td>RF</td>
<td>peat shell</td>
<td>-1.53</td>
<td>-0.85</td>
</tr>
<tr>
<td>PG2</td>
<td>modern, beach shell</td>
<td>0.48</td>
<td>-0.59</td>
</tr>
<tr>
<td>BR1</td>
<td>beachrock shell</td>
<td>-1.32</td>
<td>-0.87</td>
</tr>
<tr>
<td>BR2</td>
<td>beachrock shell</td>
<td>-1.32</td>
<td>-0.94</td>
</tr>
<tr>
<td>GA</td>
<td>beachrock shell</td>
<td>0.30</td>
<td>-0.21</td>
</tr>
<tr>
<td>PG1</td>
<td>beachrock shell</td>
<td>0.54</td>
<td>0.02</td>
</tr>
<tr>
<td>SAL</td>
<td>beachrock shell</td>
<td>-0.19</td>
<td>-0.73</td>
</tr>
<tr>
<td>PB</td>
<td>beachrock shell</td>
<td>-0.05</td>
<td>-0.67</td>
</tr>
<tr>
<td>CDB1</td>
<td>tidal flat shell</td>
<td>-1.49</td>
<td>-0.78</td>
</tr>
<tr>
<td>CDB2</td>
<td>tidal flat shell</td>
<td>-1.32</td>
<td>-0.28</td>
</tr>
<tr>
<td>CH2</td>
<td>beachrock shell</td>
<td>-0.77</td>
<td>-0.71</td>
</tr>
</tbody>
</table>

\(^1\) \(\delta^{13}\text{C} = 1000 \left\{ \frac{^{13}\text{C}}{^{12}\text{C}}_{\text{sample}} - \frac{^{13}\text{C}}{^{12}\text{C}}_{\text{standard}} \right\} \left\{ \frac{^{13}\text{C}}{^{12}\text{C}}_{\text{standard}} \right\}; \) the standard is the Chicago PDB; \(\delta^{18}\text{O}\) is worked out the same way.
Figure 4.7 - Stable carbon and oxygen isotope composition of mollusc shells. Key: A, B, C - mollusc shells of transitional environment (Keith and Parker 1965); D - freshwater molluscs (Keith et al. 1964); E - freshwater gastropods (Keith and Parker 1965); F - marine coastal shells (Vita-Finzi 1992).
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Figure 4.8 - $\delta^{18}O_{PDB}$ vs. age (marine shells).
Figure 4.9 - Stable carbon and oxygen isotope composition of beachrock (bulk sample). Key: A - Red Sea beachrock cement (Halail and Rashed 1992); B - Mediterranean beachrock cement (Halail and Rashed 1992); C - Bahamas beachrock, bulk sample (Beier 1985); D - West Indies beachrock cement (Moore 1973), E - Israeli beachrock, bulk sample (Margaritz 1979); F - marine phreatic intertidal zone, bulk sample (Beier 1985); G - meteoric phreatic ground water lens and vadose zone (Beier 1985).
In summary, the isotopic data appear to rule out any fractionation, reservoir and hardwater effects. Moreover, the palaeoenvironmental characteristics of the marine shells imply that they are solely of coastal origin and were not transported from the continental shelf.

**Emergence and submergence of Pleistocene coastal deposits**

From Table 4.2, it can be seen that the foreshore to shoreface sediments SM1, SM2, and SM3 are more than 16,000 yr. BP old (Pleistocene), that is beyond the limits of the first order radiocarbon method. Furthermore, these samples display post-depositional recrystallisation of primary aragonite to secondary calcite, which suggests that their ages must be regarded as minima.

Two lines of evidence can be used to demonstrate that the foreshore to shoreface sediments correspond to tectonically uplifted Pleistocene deposits. First, it is possible to compare them with similar rocks offshore in a less deformed state. Second, the foreshore to shoreface sediments can be compared with other Pleistocene coastal deposits along the Brazilian coast and oceanic islands.

It will be recalled that the foreshore to shoreface sediments were correlated by Srivastava and Corsino (1984) with the Guamaré and Tibau formations deposited on the continental shelf and in coastal fans respectively (see Chapter 3) in the offshore part of the Potiguar basin. This implies that the emergence of both formations onshore is the product of uplift. As the Carnaubais fault system forms the western boundary of the foreshore to shoreface cliffs, it is tempting to conclude that uplift occurred along it as suggested in Fig. 4.10, in which case the uplift has amounted to some dozens of metres.

Nowhere on the Brazilian coast, there is a coastal deposit occupying a similar position to that of the foreshore to shoreface sediments. From the Pleistocene shorelines listed in Table 4.5, it can be concluded that there is no clear correlation between the foreshore to shoreface sediments and shorelines described by previous studies. The only example of former shoreline which displays a similar height for the foreshore to shoreface
sediments (± 2 m) is the Cananéia transgression (Martin et al. 1982), but it does not correlates with the prograding features of the Pleistocene sediments in the present study. This indicates that the foreshore to shoreface sediments, which form cliffs between Touros and São Bento (Fig. 4.1), were tectonically uplifted and therefore that their present height onshore cannot be correlated with other palaeoshorelines along the littoral zone of northeastern Brazil.

**Table 4.5 - Pleistocene shorelines occurring in Brazil and its oceanic islands.**

<table>
<thead>
<tr>
<th>Height asl (m)</th>
<th>Location</th>
<th>Age</th>
<th>Evidence</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>-6</td>
<td>Fernando de Noronha Is.</td>
<td>Pleistocene</td>
<td>erosion</td>
<td>Almeida (1958)</td>
</tr>
<tr>
<td>± 1</td>
<td>Fernando de Noronha Is.</td>
<td>Pleistocene</td>
<td>erosion</td>
<td>Almeida (1958)</td>
</tr>
<tr>
<td>± 2</td>
<td>Brazilian coast</td>
<td>120,000 yr BP</td>
<td>Io/U ages in raised corals</td>
<td>Martin et al. (1982)</td>
</tr>
<tr>
<td>12</td>
<td>Fernando de Noronha Is.</td>
<td>Pleistocene</td>
<td>erosion</td>
<td>Almeida (1958)</td>
</tr>
<tr>
<td>15</td>
<td>Trindade Is.</td>
<td>Middle Pleistocene</td>
<td>erosion</td>
<td>Almeida (1961)</td>
</tr>
<tr>
<td>20</td>
<td>Rachada Is. (Trindade)</td>
<td>Pleistocene</td>
<td>erosion</td>
<td>Almeida (1961)</td>
</tr>
<tr>
<td>30</td>
<td>Brazilian coast</td>
<td>Pleistocene</td>
<td>erosion</td>
<td>Bigarella and Andrade (1965)</td>
</tr>
<tr>
<td>40</td>
<td>Fernando de Noronha Is.</td>
<td>Pleistocene</td>
<td>erosion</td>
<td>Almeida (1958)</td>
</tr>
<tr>
<td>50-60</td>
<td>Brazilian coast</td>
<td>Pleistocene</td>
<td>erosion</td>
<td>Bigarella and Andrade (1965)</td>
</tr>
<tr>
<td>80-100</td>
<td>Brazilian coast</td>
<td>Pleistocene</td>
<td>erosion</td>
<td>Bigarella and Andrade (1965)</td>
</tr>
</tbody>
</table>

Code: actual sea level (asl).
Chapter 4 - Quaternary faults: evidence from coastal emergence and submergence

Figure 4.10 - Schematic cross-section showing uplift of some stratigraphic units which occur offshore (modified after Srivastava and Corsino 1984).
Emergence and submergence of Holocene coastal deposits

Bezerra et al. (1998) have used the mismatch between sea-level predictions provided by glacioisostatic models and the radiocarbon coastal chronology to define the extent of tectonic uplift or submergence. The curve by Peltier (1997, pers. comm.) denominated the Touros curve, was used as the datum against which tectonic vertical movements were calculated. The curve (Fig. 4.11) predicts two main phases. The first corresponds to a transgression which started about 7,000 cal. yr BP and reached the highstand of 2 m about 5,500-5,000 cal. yr BP. The second phase corresponds to a sea-level regression which started immediately after the highstand and fell steadily until the present.

From Fig. 4.11 it can be concluded that the Touros curve correctly predicts local sea-level evolution during the Holocene. The vast majority of coastal deposits, which represent the lower foreshore to the upper shoreface (low tide level) except beachrock facies b and tidal flats, match the curve. The ages by Oliveira et al. (1990) and Silva (1991) were not plotted on the curve because they had not been checked for contamination, overgrowth and recrystallisation.

Samples which plot slightly above or below the Touros curve may reflect low-magnitude climatic fluctuations, as proposed by Suguio et al. (1985) and Fairbridge (1992), minor changes in wind or oceanic current, described in the South Atlantic by Damuth and Fairbridge (1970), Isla (1989), and Gonzáles and Weiler (1994), or tectonic displacement. The location of some of the samples below the Touros curve (Fig. 4.11) strongly suggests local sea-level oscillation during the Holocene. The validity of sea-level oscillations can be tested by the $\delta^{18}O$ data as it gives some indication of palaeotemperatures. Fig. 4.8 shows that the lowest $\delta^{18}O$ values, which correspond to samples BR1, BR2 and CDB1, occur during the highstand at about 5,500-5,000 cal. yr BP (Fig. 4.11). In other words it was associated with warmer sea temperatures, indicating some correlation between sea level and climate in northeastern Brazil.
Figure 4.11 - Samples of the study area set against the Touros curve of Peltier (1997, pers. comm.).
The Touros curve is also in agreement with the beachrock stratigraphy and previous coastal studies. The relationship between beachrock facies a and b (Chapter 3) confirms sea-level oscillation in the Holocene. The Touros curve also supports the conclusion from the study by Oliveira et al. (1990) and Silva (1991) that there was a Holocene transgression between 7,460 (+190/-160) cal. yr BP and 5,330 (+290/-310) cal. yr BP. During the regressive phase, aeolian coastal deposits were formed by the migration of sands from the newly exposed coastal area, which is confirmed by an $^{14}$C age of 2,350 (+350/-140) cal. yr BP by Silva (1991, sample S-2, Table 4.3) of a deposit formed by sand dunes in the Açú delta.

In location such as São Bento, low-magnitude climatic fluctuations and minor changes in wind or marine current are very unlikely to explain sudden and sharp variation in sea level. Indeed, coastal emergence to the east of the Carnaubais fault system, is matched by submergence to the west (Fig. 4.12). The Galinhos beachrock (sample GA), located in the Macau high, plots on the Touros curve, indicating no uplift. A similar situation occur with the Catanduba tidal flat (samples CDB1, CDB2) and the Farol de Sto. Alberto beachrock (FSA1), both of which occur along the Carnaubais fault and plot on or slightly below the Touros curve, showing no uplift. There is a marked departure from the predicted sea level east of the Carnaubais fault after 4,000 cal. yr BP, where sudden coastal emergence is indicated by a rich marine bivalve fauna in growth position (sample REC2) dated 4,240-3,910 cal. yr BP. It lies 5.5 m above mean sea level and 1.5 m above the Recuado beachrock (sample REC1, Fig. 4.13) dated 5,600-5,290 cal. yr BP, and 5.0 m above the Guajiru (sample GU) beachrock dated 2,910-2,740 cal. yr BP. Samples REC2 and GU indicated sharp coastal uplift of about 5 m during a maximum 1,730 yr period. This is in contrast with the smoother sea-level regression which according to the Touros curve took place in the area after 5,500-5,00 cal. yr BP.

Studies by Caldas et al. (1997) and Caldas (1998) support tectonic emergence to the east and tectonic submergence to the west of the Carnaubais fault. First, a geoelectric sounding carried out by Caldas et al. (1997) across the Carnaubais fault (see Fig. 4.12 for survey location) concluded that the base of a sedimentary sequence extending
the Neocomian to the Holocene (correlated to the Tibau and Guamaré formations) and about 120 m deep had been downfaulted by 60 m in the Umbuzeiro graben. The offset, according to the study, dies out upwards (Fig. 4.14). Second, a topographic survey by Caldas (1998) concluded that the local morphology is controlled by the Carnaubais fault: the east block present heights of more than 20 m, whereas the west block present mainly heights between 20 m and 0 m. In short, the studies suggest that the movement were operating even before the Quaternary.

**Figure 4.12** - Geological map of the São Bento littoral zone showing the location of dated samples, Fig. 4.10 and Fig. 4.14 (modified from Bezerra et al. 1998).
In summary, both Pleistocene and Holocene sedimentary deposits along the E-W-trending littoral zone east of the Carnaubais fault have been tectonically uplifted. The Carnaubais fault, of Cretaceous age, was reactivated in the late Quaternary, with a vertical offset about 5 m in the Holocene and of more than 60 m since the Neocomian.
Chapter 4 - Quaternary faults: evidence from coastal emergence and submergence

Figure 4.14 - Interpretation of geoelectrical soundings across the Camaubais fault to the southwest of São Bento (after Caldas et al. 1997).
Chapter 5

TERTIARY TO QUATERNARY FAULTS: EVIDENCE FROM REMOTE SENSING, BOREHOLE AND GEOPHYSICAL DATA

The Cainozoic faults that have been already recognised are poorly known. This chapter deals with major faults of Tertiary to Quaternary age mapped using remote sensing, borehole and geophysical data. Outcrop information is presented separately in Chapter 6, where all the information is used to evaluate their geometry, kinematics and age.

In order to facilitate analysis of the area’s tectonic evolution, it was divided into 5 smaller units: the area from Natal to Baia Formosa, the area from Ceará-Mirim to Rio do Fogo, the area from João Cândido to Touros, the area from São Bento to Jandaíra, and the Açu valley area (Fig. 5.1). Each of the areas has a different density of borehole and geophysical data. In addition, LANDSAT images could not be used for most of the Natal to Baia Formosa area owing widespread cloud cover, and Sideways-looking airborne radar (SLAR) and aerial photos were used instead.

Several studies have already shown that Cainozoic tectonism in northeastern Brazil is significant. For example, Miranda and Srivastava (1984), inferred from the geometry of N-S-oriented tidal channels in the Açu delta that the area is divided into blocks by extensional Quaternary faults which have produced block tilting and westward migration of the tidal channels. Hackspacker et al. (1985) recognised the Afonso Bezerra fault (Fig. 5.1) as a brittle NW-striking structure formed during an extensional event, which affects the Eocene-Miocene volcanic rocks of the Macau formation (Sial 1975). Oliveira et al. (1993) pointed out that right-lateral movement of the Afonso Bezerra fault took place in the early Tertiary or during the Macau volcanism phase and was caused by N-S-oriented compression. In addition, Saadi and Torquato (1992) proposed Cainozoic uplift in the Central part of Ceará state, about 300 km to the west of the study area, on the basis of geomorphological, sedimentary and structural evidence derived mainly from topographic maps and field correlation. They described transpressional features such as positive flower structures in the sediments of the Miocene Camocim formation as well
Chapter 5 - Tertiary to Quaternary faults: evidence from remote sensing, borehole and geophysical data

as in younger alluvial terraces. They also identified strike-slip, right-lateral 040°-060°-striking faults and normal, 320°-striking faults, and proposed that the Cainozoic faults resulted from the reactivation of Precambrian shear zones. More recently, Fonseca (1996) proposed that the Açu valley had been split into several crustal blocks bounded by NE- and NW-striking faults which were generated or reactivated in the Cainozoic. In his view, block tilting induced river migration eastwards. These studies were unanimous in demonstrating the existence of Cainozoic tectonism in the region.

Others have concentrated mainly on subsurface information from borehole and geoelectrical sounding. Costa and Salim (1972), Salim and Coutinho (1973) and Salim et al. (1975) used borehole data to identify the downfaulted Natal and Parnamirim grabens. The former is a NE-oriented trough located along the Potengi and Jundiaí rivers where the Barreiras formation presented vertical offsets of up to 140 m. The latter is a NW-oriented trough 2 km to the north of Parnamirim city. Geoelectrical survey by Queiroz et al. (1985) along the Canguaretama valley identified a graben filled by the Barreiras formation and younger alluvial sediments and coincident with the valley. Using vibracoring and auger-drilling data, Silva (1991) concluded that faults play an important role in the Quaternary evolution of the Açu delta and that they cut across the limestone of the Jandaíra formation (Cretaceous), basalts of the Macau formation (Eocene-Oligocene), and fluvial and marine sediments of Pleistocene age. Geoelectric studies carried out by Caldas et al. (1997) across the Carnaubais fault to the south of São Bento revealed the reactivation of one of these faults in the Umbuzeiro graben which affected a sedimentary sequence 120 m thick extending from the Neocampanian to the Holocene (see Fig. 4.14). The base of the sequence displays a vertical offset along the Carnaubais fault up to 60 m which gradually dies out upwards.

Materials and method
The present study is based mostly on remote sensing and borehole data, complemented by published geophysical data. Remotely sensed images used in the current study comprise hard copies of SLAR produced by the Brazilian government (Projeto Radam) in the 1970’s; and digital images of TM Landsat, which include row data of the
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quadrants 214_064a and 214_064c of 6 August 1993, and scene 215_064 of 2 August 1989. The digital images were processed on ER-Mapper 5.2 run on a UNIX workstation at the Centre for Remote Sensing of Imperial College London. Further information was extracted from the topographic maps of SUDENE (scale 1:100,000) and aerial photos (1979, scale 1:70,000). Borehole logs were supplied by the Company of Mineral Resources of the Rio Grande do Norte state (CDM-RN), now closed, and the Company of water and swages of the Rio Grande do Norte state (CAERN), and reinterpreted geoelectric soundings from Queiroz et al. (1985) and the Instituto Politécnico do Estado de São Paulo (IPT) (1982).

Several processing techniques including directional filtering (ER-Mapper’s Sunangle) were applied and used for field checking, but only ratio and RGB (Red Green Blue composition) images, which gave the best results, are discussed in this Chapter. Ratio images used the 3rd and 1st TM-Landsat bands in order to increase contrast between areas of high and low reflectance of iron oxides and hydroxides. An enhancement histogram equalize contrast was eventually applied to the Ratio 3:1. Two types of RGB images were processed: the first RGB composition was a normal colour composition of bands 3(red), 2(green) and 1(blue); the second RGB composition was a combination of six bands of the TM Landsat derived from the ratios 3:1 (red), 5:3 (green), and 4:2 (blue). Both compositions bring out the valleys and related faults, which were also enhanced by histogram equalization.

Boreholes and geoelectrical soundings have been concentrated mainly in the area from Natal to Baia Formosa and were used to map subsurface faults. Most of the data are concentrated in the Cainozoic units and the top of the Cretaceous stratigraphic units, but a few bear on the crystalline basement. The gravity maps by Moreira et al. (1990) and Bezerra et al. (1993), located in the area from João Câmara to Touros and part of the area from Natal to Baia Formosa respectively, were used to highlight the correlation between shallow faults and structures in the Precambrian crystalline basement.

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General features of major faults

Three main fault sets were recognised: a 040°-060°-trending set, called the NE-trending faults from now on; a 300°-320°-trending or NW-trending faults; and a 350°-010°-trending set or N-trending faults. The NE- and the NW-trending faults are the dominant sets across the study area and the N-trending faults a subordinate system which assumes importance in places such as the area from Natal to Baia Formosa. The NE- and the NW-trending faults are commonly segmented and display systematic cross-cutting relationships, suggesting they are contemporaneous and have acted as conjugate faults especially on the N-S-trending coast and in the Açú delta.

There is evidence to suggest that the faults discussed here are either new Tertiary to Quaternary faults or pre-existing structures such as Precambrian shear zones and Cretaceous faults which have been reactivated in the Cainozoic. The main evidence for Tertiary to Quaternary faulting is the variation in the thickness of Quaternary and Miocene to Pleistocene sedimentary deposits over the main bounding faults, significant and sharp variations in the depth of Quaternary and late Tertiary deposits, Quaternary fault scarps and fault-line scarps, tectonic lineaments in Quaternary and late Tertiary deposits, and strongly oriented, fault-controlled drainage patterns.

The evidence of neotectonic deformation is common across the study area. For example, the NE- and NW-trending faults are strongly related to rectangular and parallel drainage patterns. Moreover, the morphology along the coast, specially along the N-S-trending coast, reflects some kind of tectonic control. The Barreiras formation forms plateaux which usually occupy horsts. The plateaux have been dissected in the uplifted blocks and capped by alluvial terraces, sand dunes or both along the downfaulted blocks.

Deformation was partitioned between normal and strike-slip movement. The majority of cases show vertical offsets (Figs. 5.4, 5.5, 5.6 and 5.18), but there is also evidence for strike-slip movement from remote sensing and field data. Although the horizontality of most of the sedimentary strata makes it difficult to establish strike-slip movement, the mismatch of bed thickness and facies across some of the faults and the straight surface
geometry of the vast majority of fault planes point to a strike-slip component. This strongly suggests that in some cases the vertical offset observed in cross-section may be the result of major oblique-slip or strike-slip movement.

Several of the faults display evidence of surface rupture to produce normal fault scarps marked by steep slopes and abrupt topographic breaks. The topographic breaks associated with fault scarps attain 30-40 m in height in places and represent the result of cumulative late Tertiary to Quaternary slip. Most of them are degraded and capped by debris slopes, vegetation and soil, indicating that they have not been active recently, despite their Tertiary to Quaternary age. Free faces are relatively uncommon and the shape of fault scarps as well as fault plane dips depicted in cross section give a rough representation of the general topography.

**Natal to Baia Formosa**

The first area to be analysed consists of the Canguaretama, Guaraíra, Trairi, Parnamirim and Potengi grabens and surrounding areas (Figs. 5.1 and 5.3). It is endowed with a high concentration of borehole and geophysical data and is affected by normal and strike-slip NE-, NW- and N-trending faults, which cut across all the stratigraphic units, including the crystalline basement. The faults provide the main boundary between the Miocene-Pleistocene Barreiras formation and Quaternary alluvial sediments, and are buried in places by Quaternary to Recent aeolian sediments.

Evidence of multiple offset along the faults is derived from borehole and geoelectric sounding (see Chapter 6 for evidence from outcrop data). The amount of strike-slip offset was difficult to assess owing to the lack of markers and the flat-lying nature of the sedimentary layers. Vertical offsets were easy to estimate in cross sections and their high average values indicate multiple normal faulting events.
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Figure 5.1 - Geological map of the study area indicating major fault zones and area of detailed investigation. Fault key: AB, Afonso Bezerra; BC, Boa Cica; BQ, Bolqueirão; CA, Caraúbas; JU, Jundiaí; PI, Pium; QL, Queimado Lake; SB, Samambaia.
The SLAR image in Fig. 5.2 illustrates the topographic features associated with the fault-bounded valleys and enables all three main sets of faults (NE-, NW- and N-trending faults) to be identified. It shows a strong contrast in surface roughness between the crystalline basement, located in the west half of the image, and the sedimentary terrain to the east. The relatively dissected and rough surface of the crystalline basement displays a high density, dendritic pattern, which is, in places deflected along Precambrian shear zones. The sedimentary terrain presents a smooth surface and a rectangular drainage pattern associated with right-angle shaped lagoons and lakes which are controlled by faults.

In the sedimentary terrain, the fault-bounded valleys are characterised by topographic breaks between plateaux of the Barreiras formation and alluvial terraces, which correspond to uplifted and downfaulted blocks respectively (Figs. 5.2 and 5.3). The best examples of such topographic breaks are seen in the Canguaretama and Guaraira valleys. Faults in these valleys cut across the flat-lying beds of the Barreiras plateaux and form scarps which are marked by bright lines on the SLAR image (Fig. 5.2 and 5.3). Such scarps are degraded by small tributaries of the main rivers and in places form triangular facets.

The Canguaretama valley originated as a graben which is bounded by two 060°-trending faults with normal offset (Fig. 5.3). The existence of strike-slip movement is indicated by the linear geometry of faults and low-rake striae (see Chapter 6). Although the main rivers which flow along the valley in places form meander loops, the valley runs almost entirely at 060° along NE-trending faults (Fig. 5.2). These river valleys are deflected in places by 320°-striking faults near the shoreline producing a tectonically controlled rectangular pattern (Fig. 5.2).

The faults that bound the Canguaretama graben are displayed in the cross section A-A' (Figs. 5.3 and 5.4), which is a combination of part of the geoelectric sounding by Queiroz et al. (1985) and new borehole data presented by this study. The study by Queiroz et al. (1985) concluded that the central block, which ranges in altitude between
2 m and 30 m asl, was downfaulted along two NE-striking faults, whereas the upper blocks (footwall) have altitudes up to 70 m asl. In addition, geological mapping by Queiroz et al. (1985) showed that the normal fault in the southern part of cross section A-A' caused silicification in the Barreiras formation.

The present work has brought to light additional information on the tectonic evolution of the Canguaretama graben. The limestone bed, probably Cretaceous, which occurs in the uplifted block between 5.5 km and 7.0 km from point A, is absent from the central downfaulted block and in the other uplifted block between 1.5 km and 5.5 km along cross section A-A' (Fig. 5.4). On the uplifted blocks the sedimentary thickness is greater than on the downfaulted central block, which is more dissected. The vertical offsets of the two faults that bound the Canguaretama valley range from ~60 m in the southern part of the valley to ~15 m in the northern part (Fig. 5.4).

The general picture of the Canguaretama graben is the opposite to that expected for grabens with a simple evolution. Several explanations are possible for the available data. The thickness of the Miocene to Pleistocene Barreiras formation and the lower, probably Cretaceous, limestone and sandstone beds in the downfaulted central block is difficult to explain by lateral thickness variations of these stratigraphic units. It is more probably related to a combination of several uplift/submergence phases and strike-slip movements. Some of the normal component may in fact be the product of substantial strike-slip faulting.

The Guaraira valley, in the graben of that name, is bounded by two 050°-trending faults (Fig. 5.3). The bounding faults show a vertical offset of more than 40 m in the Barreiras formation (cross section B-B' in Fig. 5.4) and some indication of strike-slip movement (see Chapter 6). The shape of Guaraira lagoon, parallel to the NE- and NW-striking faults (Fig. 5.3), also indicates tectonic control. As in the Canguaretama graben, faults in the Guaraira graben are the result of Tertiary to Quaternary reactivation of Precambrian shear zones which occur underneath these valleys and outcrop to the west in the crystalline basement (Fig. 5.1).
Figure 5.2 - Sideways-looking airborne radar image (SLAR) of part of the area from Natal to Baia Formosa. Note that the water shows up as dark tones. The flight direction during the survey was N-S. The area displayed is 42 km wide.
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Figure 5.3 - Geological map of the area between Natal and Baia Formosa showing location of geological cross-sections. The city of Natal is omitted for clarity.
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The Trairi valley occupies a NW-trending fault-bounded graben limited by two 345°-striking normal faults. The faults are parallel to NW-trending Precambrian shear zones to the west (Figs. 5.1 and 5.3). The margins of the larger lakes, the northern part of the Guaraíra lagoon, as well as the river channels in the downfaulted block, are oriented parallel to the NW-trending faults, indicating tectonic control. A further cluster of lakes in the northern uplifted block display 060°- and 320°-oriented margins, especially in Bonfim lake (Fig. 5.3), also suggesting tectonic control of the geomorphology. The Barreiras formation has a thickness up to 80 m in the downfaulted block and 20-30 m in the uplifted ones owing to erosion; its base is offset up to 40 m in one of the faults (cross section C-C', Fig. 5.4).

A NW-trending graben is found to the north of Trairi valley within the Parnamirim graben of Costa and Salim (1972). It is bounded by two 320°-striking faults truncated, in places, by NE-trending faults. The 320°-striking faults display vertical offsets that range from 30 m (cross section D-D', Fig. 5.5) to 35 m (cross section F-F', Fig. 5.5). Quaternary to Recent aeolian sediments cap part of the downfaulted and uplifted blocks, including the fault scarps (Fig. 5.3 and 5.5). Another small downfaulted block, where the base of the Barreiras formation is offset by 20 m, also occurs to the north of the Parnamirim graben (Fig. 5.3 and cross sections E-E' and F-F' in Fig. 5.5). This study locates the centre of the Parnamirim graben to the southwest of the Pium river (Fig. 5.3 and 5.5), whereas Costa and Salim (1972) place it northeast of the river.

The Potengi half-graben, which correspond to the Potengi graben of Costa and Salim (1972), extends for at least 20 km from the shoreline to Macaíba city and is bounded by the 060°-trending Jundiaí fault located on the right margin of the Potengi and Jundiaí rivers. It most probably continues offshore but neither the aerial photos nor the satellite images were good enough for underwater investigations at this site. The Jundiaí fault, truncated in places by N-trending faults, cuts across the Barreiras formation, a limestone and a lower sandstone probably of Cretaceous age, and the crystalline basement. The main topographic expression of the Jundiaí fault is a topographic break about 40-60 m high, which is frequently covered by Quaternary alluvial sediments.
Figure 5.4 - Cross-sections of the area from Natal to Baia Formosa (see Fig. 5.3 for location). The geoelectrical data of cross-section A-A' is from Queiroz et al. (1985).
Figure 5.5 - Cross-sections of the area from Natal to Baia Formosa (see Fig. 5.3 for location).
As can be seen from cross sections G-G' and H-H' (Fig. 5.6), up to 350 m of sedimentary rocks cap the crystalline basement and fill the Potengi half-graben, showing sharp changes across the Jundiaí fault. The limestone and the lower sandstone thickness were inferred for both cross sections from information of one geoelectrical sounding in cross section H-H', but the subsurface geometry of the Barreiras formation and the top of the limestone bed are well constrained in both cross sections from several borehole data. The thickness of the Barreiras formation changes from an average of 100 m in the uplifted block to more than 250 m in the downfaulted block (cross-section H-H') and its base displays a vertical offset of about 250 m. A less abrupt change in thickness is seen upstream, where the Barreiras formation is 75 m thick in the uplifted block and 80-85 m in the downfaulted block (cross section G-G', Fig. 5.6), which can indicate either preferential uplift and erosion upstream or seaward thickness variations in the Barreiras formation. All the characteristics presented above rule out the previous idea of Costa and Salim (1972) and Salim and Coutinho (1973) that the Potengi trough is a central downfaulted block bounded by two major normal faults. The Potengi half-graben is bounded by the Jundiaí fault, which acted as a master fault, showing both normal and strike-slip components of movement (see also Chapter 6). Emergence near Macaíba and subsidence near the Potengi and Jundiaí rivers estuary may be also related to the rotational component of faulting.

The geometry in cross section and surface characteristics of the alluvial sediments to some extent indicate that sedimentation in the Potengi half-graben is controlled by a master fault (see also outcrop data in Chapter 6). The alluvial sediments across the valley which includes the Potengi and Jundiaí rivers, for example, occupy a zone less than 1 km wide on the right margin of the Jundiaí river, and at least 5 km wide on the left margin of the river, where palaeomeander loops are present (Fig. 5.3). This suggests tilting and lateral migration of the river channels towards the Jundiaí fault, consistent with a half-graben structure.
Figure 5.6 - Cross-sections across the Potengi half-graben (see Fig. 5.3 for location).
In conclusion, the geometry of the alluvial sediments are strongly controlled by major faults along the Canguaretama, Guaraira, Trairi, and Potengi grabens (cross sections A-A’, B-B’, C-C’, G-G’ and H-H’. Figs. 5.4, 5.5 and 5.6). The cross-sections indicate that the faults which bound the troughs have had a major role in post-Barreiras deposition. It can also be seen that the thickness of the Miocene to Pleistocene Barreiras formation increases in the downfaulted block, suggesting that the faults were active during the deposition of the Barreiras formation and thus that their activity extends back into the Miocene.

The conclusions presented above are consistent with an earlier geophysical study by Bezerra et al. (1993) from the Trairi graben to the Potengi half-graben (Fig. 5.7). Major faults in the sedimentary cover are parallel with steep Bouguer gradients, which roughly coincide with the major Precambrian shear zones shown in Fig. 5.1. The qualitative analysis made by Bezerra et al. (1993) showed that the majority of gradients depicted in the Bouguer map of Fig. 5.7 are the result mainly of density contrasts in the crystalline basement, as vertical offsets of about 50 m or less in the sedimentary cover are not enough to produce Bouguer anomalies of up to 8 mgal. Some of the troughs parallel to Precambrian shear zones in the area from Natal to Baia Formosa coincide with the negative Bouguer anomalies of Fig. 5.7. One of the best examples is the Pamamirim graben, which correspond to a -4 mgal Bouguer anomaly, and the main faults that bound the trough, as well as others to the north, are parallel to important Bouguer gradients.

In the area of the Potengi graben, however, Precambrian shear zones are NW-oriented (Fig. 5.1), whereas the Bouguer anomaly is NE-oriented. The Bouguer map of Fig. 5.7 is consistent with vertical offsets along the Jundiaí fault as it displays a gravity high which reaches 3 mgals in the uplifted block, decreasing to -2 mgal in the downfaulted block where the sedimentary cover is much thicker (cross sections F-F’ and G-G’, Fig. 5.6). This indicates that the Bouguer anomaly is associated with the Jundiaí fault and is influenced by the vertical offset of ~ 350 m where the crystalline basement and sedimentary cover are placed side by side.
Figure 5.7 - Bouguer gravity map of the area from Natal to the Trairi valley after Bezerra et al. (1993). The major faults are marked on the gravity map. Contour interval 2 mgal.
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The Bouguer map of Bezerra et al. (1993) (Fig. 5.7) conflicts with the geological map (Fig. 5.3) in a few places. The 8 mgal Bouguer anomaly immediately to the north of the Trairi valley is orthogonal to the horst where the Bonfim and other small lakes are located. According to the vertical offsets seen on cross section C-C', such a 8 mgal anomaly should mark not the density contrast between the crystalline basement and the sedimentary cover, but the occurrence of deep mafic-ultramafic rocks (Rand and Manso 1990). Otherwise, the gravity features show a good correlation with structure.

**Ceará-Mirim to Rio do Fogo**

Remote sensing was an important tool for the recognition of faults north of the Potengi graben between Ceará-Mirim to Rio do Fogo. The RGB compositions display valley borders controlled by faults and the contrast between the Barreiras formation (light green to gray and white in the images) with younger alluvial sediments (dark green in the images). The satellite images show that conjugate NW- and NE-trending faults which occur across the area are usually marked by a rectangular drainage pattern (Fig. 5.8).

From a detailed RGB image of the Ceará-Mirim and the Extremoz valleys in Fig. 5.9, it can be seen that both valleys are controlled by NW- and NE-trending faults. The northern border of the Ceará-Mirim valley is bounded by fault scarps which mark a set of NE-trending faults. The faults form a right stepover system and the fault scarps result in topographic contrast of up to 20 m between the Barreiras formation and the alluvial sediments. In contrast, the southern border of the Ceará-Mirim valley is characterised by the absence of faults and a much smoother topography. The Ceará-Mirim valley, to judge from the imagery and field data (see Chapter 6), also presents a geometry of half-graben. But unlike the Potengi half-graben, displays a series of en echelon faults rather than a single master fault on its northern border (Fig. 5.10).
Figure 5.8 - TM RGB (bands 3,2,1) image of the area from Ccará-Mirim to Rio do Fogo. The area shown is about 30 km wide.
Figure 5.9 - TM RGB (bands 3,2,1) image of the Ceará-Mirim and Extremoz valleys. Note fault scarps in the northern part of the Ceará-Mirim valley (arrows). The area covered is about 24 km wide.
Two other NW-trending faults occur to the north of the Ceará-Mirim half-graben. Both major NW-trending faults are marked primarily by a tectonically controlled drainage pattern (Figs. 5.8 and 5.10). Satellite imagery strongly suggests that they have a normal component which originated in graben structures (Fig. 5.8). Note from Fig. 5.8 and Fig. 5.10 that there are geological similarities between these faults and the NW-trending faults in the area from Natal to Baia Formosa. In both areas, the faults mark a sharp and linear, NW-trending boundary between the Barreiras formation and the aeolian sediments. In the Natal to Baia Formosa area, they mark downfaulted blocks filled by aeolian sediments. A similar situation is probably found in the Ceará-Mirim to Rio do Fogo area.

*João Câmara to Touros*

The area to the east of João Câmara city (Fig. 5.1) has been affected by extraordinary seismic activity, notably the earthquake swarm of 1986-1988. The activity was concentrated along a 040°-trending linear cluster of epicentres mostly 5-10 km deep (the Samambaia fault) which cut across the Precambrian shear zones to the south of João Câmara (Ferreira *et al.* 1987, Takeya *et al.* 1989, see Fig. 2.4). But no surface rupture linked to what come to be called the Samambaia fault has been found so far. In addition, there is no significant contrast in topography across the fault.

A second 040°-striking fault can be detected on the satellite image of the coastal plain between João Câmara and Touros cities (Fig. 5.11). The image, which corresponds to the ratio between bands 3 and 1 of TM Landsat, emphasizes several characteristics of the terrain, mostly the high reflectance (light tones) and poor reflectance of iron oxides and hydroxides such as limonite (dark tones). The fault along the Boqueirão lake, hereafter called the Boqueirão fault, splits the subarea into two sedimentary domains: first, the Barreiras formation in dark gray tones in the northwest half of the image; and second, the post-Barreiras sedimentary cover chiefly composed of alluvial sediments in light gray tones in the southeast half of the image. Although there is no abrupt topographic contrast across the Boqueirão fault, the northwest block is slightly higher than the southeast block (40-73 m and 30-66 m respectively). Furthermore, the lake
Boqueirão, lake Coelhos, and Touros canyon (see Fig. 5.12 for location) form a continuous NE-trending lineament which cuts across the Cainozoic stratigraphic units. It indicates that the fault has been active during the Quaternary and strongly suggests its continuation offshore.

The relationship between the two blocks across the Boqueirão fault is the opposite to that observed in the seismogenic Samambaia fault, where the fault plane dips to the northwest and the uplifted block is found southeast of the fault plane. If the Boqueirão fault plane has the same attitude as the Samambaia fault plane at depth, then the former passed through a phase of left-lateral strike-slip in the past associated with a reverse component which allowed the northwest block to be uplifted.

Other small faults are associated with the and the Samambaia faults between João Câmara and Touros. A well oriented subparallel swarm of small karst caves relating to minor faults, with no consistent oversteps, is seen to the southwest of the Boqueirão fault in the Cretaceous Jandaira formation (Fig. 5.12). The perfect parallelism between all these faults points to a common origin. The rectangular drainage pattern to the east of the Boqueirão fault, where the Tatu river and its tributaries form a pattern similar to the one observed in conjugate faults, is further geomorphological evidence for faulting also involving the NW-trending fault set (Fig. 5.12).

The present data do not allow us to determine whether (a) the Bolqueirão fault is a continuation of the Samambaia fault which was not active during the 1986-1988 earthquake swarm; or (b) the Samambaia and the Bolqueirão faults are linked by a stepover or a transfer fault. The thick alluvial sediment cover makes it difficult to see how these two faults or fault segments are linked. The hypothesis of a transfer fault between them is favoured by the presence of a NW-trending limestone scarp parallel to the NW-striking faults common in the area (Fig. 5.12). But a stepover or a bridge cannot be completely ruled out. In any case, the parallelism between the Samambaia and the Boqueirão faults and the Quaternary age of the latter make it very probable that they have a common origin.
Figure 5.10 - Geological map of the area between Ceará-Mirim and Rio do Fogo. The dashed box marks the area of Fig. 5.9.
Figure 5.11 - TM-3/1 ratio image of the Boqueirão lake and surrounding area showing a NE-trending lineament. The area shown is about 19 km wide.
Figure 5.12 - Geological map of the area between João Câmara and Touros. The dashed box marks the area of Fig. 5.11.
Despite the fact that Precambrian shear zones at the surface do not coincide with the Samambaia fault, the gravity survey by Moreira et al. (1990) displays a strong gradient in the Bouguer map along the fault. According to their Bouguer map (Fig. 5.13), the Samambaia fault runs between the limit of positive (+2 mgal) and a negative (up to -14 mgal) Bouguer anomalies. It suggests that although there is no coincidence between the Samambaia seismogenic fault and Precambrian shear zones at the surface, the strong Bouguer anomalies indicate that some kind of 040°-trending Precambrian structural or lithological discontinuity it to be found at a deeper crustal level and that this discontinuity may have been acted as a weakness zone responsible for the present seismicity. The Bouguer map of Moreira et al. (1990) strongly suggests that the geology at the surface may not correspond to the tectonic and lithological framework at crustal levels as deep as 5-12 km, where the earthquakes related to the Samambaia fault have been concentrated.

**São Bento to Jandaira**

The most striking tectonic feature in the area near São Bento is the Camaubais fault system, which is a Cretaceous fault (Matos 1992). The evidence derived from remote sensing supports the proposal made in Chapter 4 that the southeast block of the Camaubais faults has undergone emergence and the northwest block has undergone submergence during the Pleistocene and Holocene. The RGB image using the ratios red-3:1, green-5:3, blue-4:2 (Fig. 5.14) shows that the boundary in the coastal area between the Barreiras formation (blue) and the Quaternary coastal sediments composed by sand dunes and tidal flats (red) lies along the Camaubais fault. The shoreline immediately to the west of São Bento runs at 060° exactly above the continuation of the Camaubais fault offshore (5.14). In addition, despite the absence of a marked topographic contrast across the Camaubais fault, the drainage flows roughly north-south and is deflected along the fault trace towards the northeast.

Another indication of faulting can be seen in the lower right corner of the RGB image (Fig. 5.14). The topography between the Barreiras formation (blue) and the Jandaíra formation (bluish, greyish and greenish colours) is marked by a limestone scarp up to 15
m high, which is roughly parallel to the Camaubais fault as far as the right-hand margin of the image (see Fig. 5.15 for scarp location). It probably indicates that erosion in the area follows zones of weakness related to faults.

All these items, combined with those presented in Chapter 4, point to the submergence of the Umbuzeiro graben across the Camaubais fault (Fig. 5.15) during the Tertiary-Quaternary. They also indicate that tectonics has greatly influenced Quaternary deposition in the area.

The Açú valley

The Açú valley is the biggest river valley in the study area. It experienced both historical seismicity (Ferreira 1983) and recent induced seismicity (Ferreira et al. 1995). The drainage patterns and the geometric form of Quaternary sedimentary deposits brought on by RGB image using ratios red-3:1, green-5:3, blue-4:2 (Fig. 5.16) indicate the occurrence of NE- and NW-trending faults. In the Açú delta, rivers and some channels reveal a combination of anastomosing and rectangular forms produced by a combination of coastal sedimentary processes and tectonics. Small lakes and rivers form prominent NW-trending lineaments to the south of the Açú delta and cut across the Barreiras formation (blue) and part of the Quaternary alluvial terraces (red and blue). The straight geometry of a fault over more than 100 km strongly suggests an important strike-slip component of movement.

Another clear indication of tectonics is seen in the relationship between the Açú river and its Quaternary alluvial sediments. The river has an anastomosing channel which flows along the right margin of the valley, and abandoned terrace loops are seen in the left margin, fairly indicating tilting and river channel migration eastwards (Fonseca 1996).

The map in Fig. 5.17 displays the most important stratigraphic units and Tertiary to Quaternary faults across the Açú valley and its surrounding area. Some of these faults are Cretaceous in age but show signs of Cainozoic reactivation.
Figure 5.13 - Gravity Bouguer map of the João Câmara epicentral area (after Moreira et al. 1990); contour interval 4 mgal.
Figure 5.14 - TM RGB ratio colour composite derived from ratio images 3/1 (red), 5/3 (green), and 4/2 (blue) of the São Bento area. White arrows mark the Carnaubais fault and limestone scarp. The area shown is about 50 km wide.
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Figure 5.15 - Geological map of the area between São Bento and Jandaíra. Dashed box marks the area of Fig. 5.14.
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The NE-trending faults such as the Pataxós fault and other faults upstream, which cut across all the stratigraphic units including the Cainozoic ones, form conjugate sets with NW-trending faults such as the Afonso Bezerra and the Queimado lake faults. Both fault sets are seen to control the geometry of river valleys and stratigraphic units including the alluvial sediments.

The most striking features in this area are the Afonso Bezerra and the Queimado lake faults. The 320°-trending Afonso Bezerra fault, described by Hackspacker et al. (1985) as a normal fault which offsets Tertiary rocks (e.g. Eocene to Oligocene Macau formation), have produced tectonically controlled geomorphological features such as the Gangorra river, the Vargem de Cima lake, and an abrupt topographic break in a Barreiras plateau to the west of the Açú valley where altitudes up to 222 m asl occur at Serra do Mel and drop sharply to about 50 m near the littoral zone. The long and straight geometry of the faults again suggests that there was an important strike-slip component in the fault movement.

The Queimado lake fault, also described by Hackspacker et al. (1985) strongly influences the shape of Queimado lake, and the Cavalos and Conchas rivers, and forms much of the boundary between the Barreiras formation and the Quaternary coastal sediments (Fig. 5.17). The shoreline from the Açú delta as far as Ponta do Mel is also NW-oriented and coincides with the Cretaceous normal fault described by Hackspacker et al. (1985), suggesting further Quaternary reactivation in the area.

The cross-section K-K’ suggests that both the Afonso Bezerra and the Queimado lake structures have moved primarily as strike-slip faults since the Miocene. This is indicated by their straight and long shape, as well as offsets as low as 40 m in the Barreiras formation, compared with more than 240 m in the Potengi half-graben (cross-section K-K’, Fig. 5.18).
Figure 5.16 - TM RGB ratio colour composite derived from ratio images 3/1 (red), 5/3 (green), and 4/2 (blue) of the Açú valley. The area shown is about 47 km wide.
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Figure 5.17 - Geological map of the Açú valley and surrounding area, including locations of cross-sections. The dashed box marks the area of the satellite image in Fig. 5.16.
Figure 5.18 - Geological cross-sections in the Açú valley area.
In contrast, the NE-trending faults, though smaller than the NW-trending faults, control Quaternary sedimentation along the Açú valley, and also display a normal component. From cross-section J-J', it can be seen that the alluvial sediments in the Açú valley fill a small trough more than 20 m deep which is defined by the NE-trending faults.

Cross-section I-I', which roughly runs parallel to both NW-trending faults, indicates that the Barreiras formation forms the Serra do Mel high, which locally rises to 222 m. Near the sea (northwest part of Fig. 5.17 and cross-section I-I' in Fig. 5.18) there is a fall of more than 160 m to sea level over 8 km, but the lack of clear horizons within the Barreiras formation makes it difficult to detect any faulting that could account for the break.

In summary, three Miocene to Holocene sets of faults are seen in the study area. The NW- and NE-trending faults in places form a conjugate set indicating an important strike-slip component. But important normal offsets are also found in cross sections across the area. Both sets cut across Precambrian shear zones and Cretaceous faults, but also reactivate these earlier structures where they are NE- and NW- oriented. N-trending faults occur in places and are less pervasive. They also present strike-slip and normal components of movement.
This chapter describes brittle deformation at a mesoscopic scale, i.e., faults of a few centimetres to a dozen metres and major faults observed in natural outcrops, road cuts and quarries. Fault features such as geometry, kinematics, and fault rocks were investigated in the same areas as those discussed in Chapter 5. The description of mesoscopic structures concentrates on Cainozoic stratigraphic units, but Cretaceous and Precambrian units were also analysed in the search of syn to post-Miocene deformation. In each area, the attitude of mesofaults and striae is represented by equal-area, lower-hemisphere stereographic projections. In some sites, such as the granite quarries near Macaíba city in the area between Natal and Baia Formosa (Fig. 6.1), deformation features are presented in great detail because of the quality and quantity of available data. The results were then correlated with those of Chapter 5.

Attempts to understand the Tertiary to Quaternary deformation of the study area using outcrop data have been carried out by Hackspacker et al. (1985), Gusso and Bagnoli (1989), Torres et al. (1990), Oliveira et al. (1993), and Fonseca (1996). The study by Hackspacker et al. (1985) on the NW-trending Afonso Bezerra fault (Fig. 5.16), which crosses the Açú valley, concluded that the structure had been reactivated since the Miocene and produced fault scarps. They associated the Afonso Bezerra fault with a volcanic doming which produced NW-oriented tension. Oliveira et al. (1993) also analysed in detail an outcrop of the fault near the Afonso Bezerra city and concluded that the fault affected Cretaceous rocks such as the Açú and Jandaíra formations and underwent right-lateral strike-slip movement which generated transpressional flower-like structures.
The study by Gusso and Bagnoli (1989) focused on a peat near Rio do Fogo village (Fig. 5.10) which displayed left-lateral N-trending faults with 60 cm of offset. Their most important finding was a right-lateral 040°-trending fault which they associated with the 1986-1989 seismic swarm along the Samambaia fault. Throughout the peat faulting had caused tilting of flat-lying beds by as much as 50°.

Torres et al. (1990) presented the results of a research project carried out by the Brazilian Geological Survey (CPRM) in the João Câmara epicentral area and surrounding areas in the aftermath of the 1986-1989 earthquake swarm on the Samambaia fault. In their view the stress field had changed three times during the Cainozoic. During the Oligocene to Miocene there was NE-SW-directed maximum extension and subordinate NW-SE extension. There followed NE-SW-directed compression and NW-SE-extension possibly from the Pliocene to the Holocene. The stress field attained maximum NW-SE-oriented compression and minimum NE-SW-oriented horizontal compression in the Holocene. The latter resulted in a Riedel system produced by right-lateral movement along E-trending megashears called the Fernando de Noronha and the Patos lineaments. 000°, 090° and 060°-070° were the main directions of active faulting.

Fonseca (1996), who concentrated his study along the Açú valley, concluded that the deformation observed in the alluvial sediments was mainly extensional and characterised by both normal and strike-slip movements. He proposed on the basis of kinematic criteria that the Afonso Bezerra fault experienced left-lateral slip and the Camaubais fault right-lateral slip during the Holocene. Microtectonic data suggested that the maximum horizontal compression ranged in that valley from NW-SE to N-S, which partly agrees with the conclusion of Torres et al. (1990) and Lima et al. (1997), but not with focal mechanism data (see Chapter 2).
Chapter 6 - Tertiary to Quaternary faults: evidence from outcrop data

General features

All the Cainozoic stratigraphic units are affected by faulting, but a great part of the study area is blanked by unconsolidated sediments where outcrops are rare and the few sections are usually degraded and covered by soil and vegetation. Nevertheless, outcrops close to the major fault zones provided enough data to reconstitute their deformation since the Miocene.

The three main fault sets identified from satellite imagery, and borehole and geophysical data were also identified in outcrops. The most pervasive are the NE- and the NW-trending sets, although sometimes only one of them is present.

The geometric and kinematic features of the mesoscopic faults are roughly constant across the study area. Fault planes are relatively smooth polished surfaces, but sometimes also irregular or listric. Fault planes often display dips of more than 80°. Kinematic investigation based on fault strike, plunge of striae and sense of slip was carried out mainly in clay-rich faulted rocks because coarse sandstones and conglomerates or unconsolidated sediments usually do not display measurable slickensides. The polarity of movement (right-lateral, left-lateral, normal or reverse) was identified in some faults on the basis primarily of displaced layers, fault terminations and Riedel fractures. Some of the main fault sets display evidence of multiple displacements. The relative chronology of movement was deduced from kinematic analysis in stratigraphic units of different ages or from overprinted striae on fault planes. This led to the separation of strike-slip and normal component of faulting events since the Miocene which matched the results presented in Chapter 5.

The faulting processes that affect the Cainozoic stratigraphic units occurred within the top ~4 km, well above the depth of ~12 km proposed by Sibson (1977, 1986) for the brittle-ductile transition zone. This is indicated by fault zones marked by non-cohesive
fault breccia and clay gouge as well as clear evidence of surface rupture such as soil disruption and alluvial infillings.

Further evidence of faulting at shallow crustal levels is indicated by the mineral paragenesis observed in fault zones affecting the Cainozoic stratigraphic units. The paragenesis of quartz and kaolinite, which indicates faulting at temperatures below 270°C and at less than 2 km deep (Power and Tullis 1989), was seen in the slickensides of sediment-filled faults near Macaíba (Fig. 6.1). Elsewhere, slickensides have developed in planes already affected by weathering, which are usually marked by iron oxides and hydroxides. Silicification, which can take place at shallow depths, is common in some other fault zones such as those that border the Canguaretama valley.

Faulting at depths of less than 4 km also took place in the Cretaceous limestone of the Jandaíra formation and in the Precambrian crystalline basement. In all the fault zones affecting the former and in some of the fault zones in the latter, the fault rocks are usually composed of non-cohesive fault breccia and gouge. Additional evidence of shallow faulting in the Precambrian crystalline basement is provided by pseudotachylite, which was described by Sibson (1977, 1986) as the friction melt product of localised seismic slip at shallow crustal level (~ less than 4 km).

Fault characteristics point to shallow seismic slip rather than creep during faulting processes. First, faults such as those in the area between Natal and Baia Formosa, generated scarps indicating surface faulting, which is usually a seismic process. Second, slickensides in sandstone, generally those rich in clay, are shiny and reflective, which according to Hancock and Barka (1987) is related to seismic slip. Third, the vast majority of slickensides contain no evidence of fiber growth, which is symptomatic of aseismic creep (Hancock and Barka 1987). Finally, the presence of non-cohesive fault breccia and gouge is a function of the slip rate and is favoured by fast seismic slip rather than creep.
Fault age was inferred from stratigraphic and geomorphological features as well as radiocarbon dating. The recency of slip along some fault zones, such as the Boqueirão and Jundiaí zones and those limiting the Canguaretama graben, is indicated by Quaternary physiographic features such as fault scarps and topographic breaks. In addition, most of the mesoscopic faults seen in outcrop of the Barreiras formation (Miocene to Pleistocene) cut across layers previously affected by weathering, which according to Mabesoone and Lobo (1980) took place in northeastern Brazil during the Pliocene and again in the Pleistocene-Holocene. Finally, some of the faults, including the sediment-filled faults near Macaíba and those to the south of Parazinho, disrupt the soil profile.

Radiocarbon dating made it possible to date two sites (Table 6.1), where Cainozoic stratigraphic units are affected by faults. These dates, in addition to those already presented in Chapter 4, will be discussed in detail bellow.

**Natal to Baia Formosa**

All three of the main fault sets cross the coast between Natal and Baia Formosa, but their relative frequency changes in each graben. The NE-trending set is dominant along the Potengi, Guaraira and Canguaretama grabens, whereas the NW-trending set is more pervasive in the Bonfim lake horst and Trairi graben areas, which suggests that the predominance of mesoscopic faults with specific trends is related to the influence of major structures such as the master faults in grabens.

The Potengi half-graben displays a relatively high concentration of outcrops. The Jundiaí fault, which is its master fault, is exposed in quarries near Macaíba. Parts of the footwall are also well exposed along cliffs south of the Jundiaí river (Fig. 6.1).
Figure 6.1 - Geological map of the area between Natal and Baia Formosa showing attitude of faults measured in outcrops: stereonets are equal-area lower hemisphere; S1-first striae generation and S2-second striae generation; s-striae; f-pole to fault; fol-pole to foliation. The letter of each stereonet is plotted on the top right-hand side and the numbers of measurements on the lower right-hand side of each net.
The Jundiaí fault is exposed as sediment-filled fractures, indicating shallow deformation in several granite quarries at the boundary between the Precambrian granite and the Cainozoic sedimentary cover south of Macaiba (Fig. 6.1). They are good indicators of palaeoearthquakes (Wallace 1986, McCalpin 1996). Although the vast majority of sediment-filled faults die out upwards, some can be traced to the ground surface (Figs. 6.2 and 6.3). The sediment-filled faults often include planar surfaces, but curvilinear planes were also mapped, some of them dozen of metres long and at least 15 m deep (Fig. 6.2d). The gaps which filled by sediments widen upwards and range in width from 2-3 cm to 2 m (Fig. 6.3). Sediment-filled faults are capped by a 3-10 cm thick weathered zone of granite (Fig. 6.4), where the faults planes have been subjected to circulation by meteoric water.

The attitude and location of the sediment-filled faults also indicate that they represent the shallow reactivation of major faults. First, the majority of sediment-filled faults have the same strike as the Jundiaí fault zone (060°) or with the N-trending fault nearby (350°-360°) (Fig. 6.1). Second, they are located in the vicinity or continuation of major faults, as indicated by stereonets a and b (Fig. 6.1).

The material which fills these faults and overlies the Barreiras formation is an unsorted mixture of alluvium and soil dated 4,865-4,566 cal. yr BP (Table 6.1). Sidewall clasts of granite from scarp-free faces are rare.

Fault rocks along sediment-filled faults point to shallow seismic slip. Non-cohesive fault breccia is common along the fault zones, sometimes associated with rock fragments and unconsolidated gouge composed of quartz and kaolinite. The crushed pebbles in the infilling sediments display friction slides. All these features, together with the lack of fault veins and fiber striae, imply shallow seismic slip.
Figure 6.2 - Sediment-filled fault affecting alluvial sediment dated 4,865-4,566 cal. yr BP and Quaternary granite saprolite (quarry about 1 km southeast of Macafba, see Fig. 6.1 for location).
Table 6.1 - Supplementary radiocarbon ages for continental and coastal sediments.

<table>
<thead>
<tr>
<th>sample/ lab. No.</th>
<th>dated material</th>
<th>site</th>
<th>δ¹³C (%)</th>
<th>conventional age (yr BP)</th>
<th>calibrated age at 2σ (yr BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>RF/Beta 104723</td>
<td>wood</td>
<td>peat near</td>
<td>-28.9</td>
<td>3,540 ± 60</td>
<td>4,041-3,689</td>
</tr>
<tr>
<td>S-MAC/Beta 110907</td>
<td>charcoal</td>
<td>quarry near</td>
<td>-26.4</td>
<td>4,200 ± 50</td>
<td>4,865- 4,566</td>
</tr>
</tbody>
</table>

Ages from the peat and the charcoal was assayed conventionally at Beta Analytic Inc. (USA), and calibration was by reference to Pearson *et al.* (1986).

Cross-cutting relationships in sediment-filled faults point to multiple faulting and filling processes in the Holocene. The slickensides of the NE-trending faults contain as many as two different sets of striae such as striae at a low pitch to the fault plane (S1 in stereonet a, Fig. 6.1) and striae at high pitch to the fault plane (S2 in stereonet a, Fig. 6.1), implying strike-slip and normal movement respectively. To judge from the Riedel (R) fractures synthetic to the main fault plane, as well as the geometry of the releasing and restraining bends along the fault zones, the NE-trending sediment-filled faults first moved by right-lateral strike-slip. These fault planes probably presented a pull-apart geometry which opened and allowed the first episode of sediment and soil filling. Normal reactivation followed the first movement, as indicated by striae at a high pitch to the fault plane (S2) overprinted on striae at low pitch to the fault plane (S1), and brecciation of sediment and soil of previous filling. Normal reactivation is also indicated by R and T (tension) fractures affecting the sediment-filled faults themselves, the granite and its weathered layers (Fig. 6.5). On the other hand, the N-trending faults moved as right-lateral strike-slip structures only, as revealed by R fractures on fault zones, although a pull-apart effect allowed sediment and soil filling here too.

None of these steep-dipping sediment-filled faults represents reactivation of previous planes of anisotropy in the Precambrian granite. A careful examination of the granite foliation (Fig. 6.1, stereonet c), which is marked by the orientation of stretched quartz
and mafic minerals, indicates that it is subhorizontal, in agreement with data presented in Chapter 5 which depict the Jundiaí fault as a structure that cuts across the Precambrian shear zones and regional foliation (Fig. 5.1).

The lack of correlative horizons or markers such as xenoliths, veins, dikes, and igneous layering in the granite makes it difficult to evaluate the offset. But fault slip must have been important in both kinematic phases to produce well developed slickensides. Normal offsets of the NE-trending faults were measured with relative accuracy when compared with strike-slip offsets. In the alluvial and soil layers (Fig. 6.2) the former range from dozen of centimetres to 1 m. But they should be regarded as minimum offsets, as bulk displacement in the granite is unlikely to be transmitted fully to the overlying unconsolidated alluvial sediments and soil.

Figure 6.3 - N-trending sediment-filled faults in Precambrian granite (quarry ~ 1 km to the southeast of Macaíba city).
Figure 6.4 - Sediment-filled fault in Precambrian granite. Note that the sediments are locked in transtensional zones (releasing bend, see white arrows), whereas transpressional zones (restraining bends) are deprived of such sediments. Note also a thin layer of weathered granite (white colour) along the fault plane: (a) general view; (b) detailed view of sediment-filled fault (quarry ~ 1 km to the southeast of Macaiba; hammer and lenscap for scale).
Away from the Jundiaí master fault this pattern of dominant NE-trending faults seen in granite quarries near Macaíba changes completely in the Potengi half-graben footwall. The best example of such a change of pattern occurs in the littoral zone, where a dominant normal N-trending set, which shapes the present shoreline, is seen affecting the Barreiras formation (Fig. 6.1, stereonet d).

Moving on to the south, in the area of lakes clustering around the lakes Bonfim and Nísia Floresta (Fig. 6.1), the pattern of mesoscopic faults reflects major NW-trending faults which control the lakes and the valley. In the Bonfim area, both the NE- and the NW-trending sets of faults with a normal component were mapped. The latter is more pervasive (Fig. 6.1, stereonets e and f). A similar pattern occurs around lake Nísia Floresta, where there is a relatively high concentration of NW-trending faults of mainly 320° strike (Fig. 6.1, stereonet g).
The deformation caused by such NW-trending, predominantly normal, faults is better seen along the shore, where there are outcrops of beachrock some kilometres long. The best example of a NW-trending normal fault occurs 5 km to the north of the Guaraíra lagoon, where echosounding by Amaral (1997, pers. comm.) indicates that a beachrock dated to 6,190-5,840 cal. yr BP (sample GR1, Table 4.2) to 7,250-6,800 cal. yr BP (sample GR2, Table 4.2) has been downfaulted by 4 m prior to the accumulation of a younger, undisturbed beachrock dated 4,840-4,520 cal. yr BP (sample BR2, Table 4.2) to 5,110-4,810 cal. yr BP (sample BR1, Table 4.2) (Fig. 6.6), which implies a vertical offset of 4 m in a maximum of 2,730 years. The fault is the continuation of a major NW-trending inland fault (Fig. 6.1). It caused tilting (<10°) of the older beachrock as indicated by topographic levelling. Most of the fault scarp is underwater and is partly exposed during low spring tides. Downfaulting is confined to a block 10 km wide (depicted in cross-section D-D') bounded by two major NW-trending faults which are parallel to Precambrian shear zones to the west (Fig. 5.1).

Further to the south, the Guaraíra and Canguaretama grabens display a more scattered pattern of mesoscopic faults owing to the occurrence of NE-, NW- and N-trending faults. But there is a clear dominance of the normal component in all three sets (Fig. 6.1, stereonets h, i, j, k, l, m, n, and o). In the Guaraíra valley, both normal NE- and NW-trending faults are found. Outcrops near the master faults that bound the graben display widespread strike-slip faulting as well as late normal faulting affecting the Barreiras formation and Quaternary alluvial sediments, where vertical offsets range from a few metres to a few centimetres (Fig. 6.7).

Identical behaviour is seen in the Canguaretama graben, where faults have a range of strikes, but with a concentration of striae at a high pitch to the fault plane indicating normal faulting (Fig. 6.1, stereonets j, k, l, m, n, and o). In the northern footwall of the Canguaretama graben, there is an array of normal conjugate sets either between NE- and NW-trending faults or NW- and N-trending faults (Fig. 6.8a). In the southern footwall, normal listric faults are also seen (Fig. 6.8b). Both cases mark the normal faulting event associated with the formation of the Guaraíra and Canguaretama grabens.
Figure 6.6 - Normal NW-trending fault cutting across an older beachrock further seawards dated 7,250-5,840 cal. yr BP and capped by a younger beachrock dated 4,520-5,110 cal. yr BP (see Fig. 6.1 for outcrop location).
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Figure 6.7 - Faulted rocks near the border of the Guaraira graben: (a) plan view of right-lateral strike-slip fault in the Barreiras formation showing C (main shear plane), R (Riedel) and T (tension) fractures (outcrop located in the northern border of the Goianinha valley, about 7 km to the northeast of Goianinha city); (b) cross-section of normal fault in the Quaternary alluvial sediment (road cut 1 km to the south of Goianinha).
Figure 6.8 - Normal faults on the edge of the Canguaretama graben affecting Barreiras formation: (a) lateral view of conjugate N- and NW-trending normal sets (northern border of the graben about 7 km to the east of Goianinha); (b) lateral view of N-trending normal fault affecting Barreiras formation but capped by younger soil (cliff about 500 m north of Baia Formosa).
Ceará-Mirim to Rio do Fogo

This area has a concentration of outcrops in the northern part of the Ceará-Mirim half-graben and in the shoreline near Maxaranguape and Rio do Fogo villages (Fig. 6.9). In the northern part of the Ceará-Mirim half-graben, the main NE-trending high-angle faults and subordinate strike-slip NW-trending faults are seen in outcrop (Fig. 6.9, stereonets a, b, and c; Fig. 6.10). To the north of Maxaranguape, mesoscopic faults on cliffs constitute a NE-trending set, which contrasts with the major NW-trending fault nearby (Fig. 6.9, stereonets d and e). In both places, normal faults sometimes show a listric geometry, whereas strike-slip faults in places display a flower-like geometry (Fig. 6.10). These faults affect both the Barreiras formation and the Quaternary alluvial sediments. In both slickensides rarely formed owing to the coarse nature of rocks, but the faults are marked by small offsets of a few centimetres to 1-2 m and imbrication of quartz pebbles.

A peat deposit, which outcrops for more than 1 km in the intertidal zone near Rio do Fogo, presents magnificent features of Holocene deformation. The peat is dated 4,041-3,689 cal. yr BP (sample RF, Table 6.1) and overlies the Barreiras formation (see Chapter 3). The faults commonly have straight or gently curved surfaces, which are generally high-angle planes. Mesoscopic faults with varied orientation show strike-slip movement with a few instances of oblique-slip and normal movement (Fig. 6.9, stereonets f, g, h, and i). The apparently random distribution of poles to faults on the stereonet is produced by the occurrence of the three main fault sets together with a few E-trending faults. The two NE- and NW-trending faults produce elongated blocks of peat (Fig. 6.11). Both right-lateral strike-slip and normal faults display offsets which exceed a few centimetres. However, the best example of displacement was documented by Gusso and Bagnoli (1989), who measured a 60 cm strike-slip offset of a shell layer in a N-trending fault.
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Figure 6.9 - Geological map of the area between Ceará-Mirim and Rio do Fogo and attitude of faults measured in outcrops (stereonets are equal-area lower hemisphere).
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Figure 6.10 - NW-trending positive flower, strike-slip structure in Quaternary alluvial sediment (outcrop located in northern part of the Ceará-Mirim half-graben).

Figure 6.11 - Elongated block of peat produced by the erosion of sediments along NE-trending faults (outcrop in the present intertidal zone near Rio do Fogo).
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The peat is also affected by two sets of joints: a $75^\circ/140^\circ$ and $55^\circ/175^\circ$ set and an $89^\circ/000^\circ$ and $89^\circ/090^\circ$ set (Fig. 6.12). Both sets are confined to the peat and were not seen in Barreiras outcrops nearby. The former set forms shear joints indicating local vertical maximum compression, possibly associated with peat burial during its formation. The latter set forms orthogonal joints and is more pervasive in the peat deposit. This joint set is kinematically compatible with a N-trending fault mapped by Gusso and Bagnoli (1989) and may have been generated by the same event. Both sets probably have only a local expression but certainly indicate coastal deformation recently and at very shallow depths.

João Câmara to Touros

The principal structure visible in outcrops in this area has been defined as the Boqueirão fault (Chapter 5), as it is not certain whether it represents the shallow continuation of the Samambaia fault described by Takeya et al. (1989). The Boqueirão fault is marked by a wide zone of fault breccia in the Barreiras formation (Fig. 6.13), but it is not clear that it presents a surface rupture. The fault breccia fragments are usually surrounded by an unsorted reddish matrix affected by strong oxidation. Although it was difficult to work out the amount of displacement at mesoscopic scale owing to the homogeneity of the sandstone of the Barreiras formation (a problem also in satellite imagery, Chapter 5), the sense of movement was found to be right-lateral with a normal component owing to the Riedel fractures (R) associated with the main brittle shear (Fig. 6.13, stereonet a; Fig. 6.14).

Outcrop data also present evidence that the Boqueirão fault zone continues to the northeast and southwest of the Boqueirão lake. Some scattered outcrops of alluvial sediment overlie the Cretaceous Jandaira formation to the southwest of the Boqueirão fault zone and display evidence of right-lateral strike-slip movement (Fig. 6.1, stereonet b). This is indicated by R and T fractures interpreted on the basis of their angular relationship with the main shear plane (Fig. 6.14).
Figure 6.12 - Joint systems in the Rio do Fogo peat: (a) lateral view of a conjugate joint set (75°/140° and 55°/175°); (b) plan view of an orthogonal joint system (89°/000° and 89°/090°).
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Figure 6.17 - Petrographic and rheologic evidence of faults in the Cretaceous succession of the Parana Basin.

Figure 6.13 - Geological map of the area between João Câmara and Touros and attitude of faults (stereonets are equal-area lower hemisphere).
Fault sets were also seen along cliffs of the Barreiras formation and foreshore to shoreface sediments in the littoral zone northeast of the Boqueirão lake near Touros. NE-trending high-angle faults dominate, but they show a wide range of striae at high to low pitch to the fault planes (Fig. 6.13, stereonets c, d). Because of the horizontal bedding and the lack of markers, only vertical offset could be measured. They do not exceed 2 m (Fig. 6.15). Faults affect the Barreiras formation and foreshore to shoreface sediments but are usually capped by the Quaternary aeolian sediments (Fig. 6.15 and Fig. 6.16). As in the fault breccia in the Boqueirão lake, strong oxidation post- and pre-deformation was observed. Slickensides usually display a combination of scratches and debris streaks. As there is no evidence of dynamic recrystallisation or grain size reduction and because a few faults can be traced to the ground surface where they disturb the soil, it can concluded that burial was shallow during the faulting process. Joints are also common in the Barreiras outcrops nearby Touros. The vast majority of them are NE-striking high-angle planes.

Figure 6.14 - Fault breccia affecting the Barreiras sandstone displaying C (main shear plane), R (Riedel), T (tension) fractures and right-lateral fault slip (outcrop located about 500 m east of lake Boqueirão).
Figure 6.15 - Normal, mainly NE-trending faults in cliffs in Touros area (see Fig. 6.13 for cliff location).
Deformation which is geometrically and kinematically similar to that of the Boqueirão fault is seen in gravelly Quaternary alluvial sediments about 15 km to the south of Parazinho (Fig. 6.13). The main structure is a high-angle NE-trending fault with a right-lateral strike-slip component followed by normal movement (Fig. 6.13, stereonet e). The vertical offset in the sandstone layer is about 1 m, but it does not have any topographic expression on the surface. The central part of the fault is a zone of fault breccia 4 m wide, where rod-shaped fragments are common (Fig. 6.17).
Thus the NE-trending fault set is pervasive in the João Câmara to Touros area and it displays a dominant right-lateral and a minor normal fault component. However, no faults mapped by this work, despite their geometric and kinematic similarities with the seismogenic Samambaia fault, can be safely described as a rupture of the 1986-1989 earthquake swarm. But it does not mean that the Samambaia fault and the Boqueirão and other NE-trending faults necessarily have a common origin. Indeed, the Samambaia fault may be a seismic and recent expression of a faulting process active since the Miocene.

Figure 6.17 - NE-trending fault zone showing strike-slip and normal component of movement and affecting post-Barreiras alluvial sediments (quarry about 15 km south of Parazinho, see Fig. 6.13 for location).
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São Bento to Jandaira

There is a great extent of beachrock and foreshore to shoreface sediments along the shoreline, which provide evidence of tectonic uplift along the east block of the Carnaubais fault (see Chapter 4). Apart from that, outcrops are concentrated in three sites in the Jandaira formation limestone near Jandaira, Caldeirões farm and Pedra Grande, all located on the Touros platform. Different fault sets, mapped both in outcrop and in caves, are marked by a highly fractured zone where carbonate dissolution has taken place (Fig. 6.18).

Four principal fault strikes were seen at the three sites. From Fig. 6.18, it can be seen that the most important and pervasive fault strike is 350°-015° (stereonets a and f), which is, in part, slightly oblique to the N-trending fault set (350°-360°) defined in Chapter 5. Three other subordinate fault strikes (040°, stereonets b and d; 330°, stereonets c and e; and 90°, stereonet g) were also mapped.

At the Jandaíra and Caldeirões farm sites, cave geometry is closely related to faults and helps to define their attitudes (Fig. 6.19). The caves are 80 m to 340 m long, including all the galleries, and oriented 000°, 040° and 330° (Fig. 6.20), in agreement with the major fault strikes.

The kinematics of movement could be investigated mainly in the 350°-015°-trending faults (Fig. 6.18, stereonet a). At all three sites investigated, there are limestones layers which are essentially horizontal and plunge about 2°-3° seawards. Because this flat-lying position shows no indication of offset, normal movement must have been at a minimum, although kinematic indicators such as horse-tail termination and Riedel fractures indicate both right-lateral and left-lateral slip, which could not be quantified in the absence of marker beds.
Figure 6.18 - Geological map of the area between São Bento and Jandaira and attitude of faults in limestone outcrops, which are represented in equal-area lower hemisphere stereonets.
Although there is no evidence of surface rupture, faults in the Jandaira formation developed at a crustal level less than 12 km deep. This is suggested by the widespread development of indurated fault breccia which consists of angular, unsorted fragments of limestone in a matrix of carbonate gouge containing recrystallised calcite. Apart from the 350°-015° fault set, this deformation is geometrically and kinematically identical to the deformation of Miocene to Holocene age described in other areas, which suggests it may have the same age. However, the deformation observed in the Jandaira limestones could equally well be related to the pre-Miocene deformation described in the Potiguar basin (e.g. Oliveira et al. 1993). The 350°-015° set does not have a correspondent set affecting the Miocene to Holocene stratigraphic units and is very probably related to pre-Miocene deformation.

Figure 6.19 - Cave formed by the dissolution of limestone along a strike-slip fault about 1 km northwest of Jandaira.
Figure 6.20 - Plan view of caves developed along fault planes in the Cretaceous Jandaíra formation (limestone): a, b, and c are caves near Jandaíra; d is a cave in the Caldeirões farm; grey represents galleries up to 10 m deep and broken lines represent galleries more than 10 m deep (after Cruz et al. 1993).
Finally, the Jandaira formation allows the investigation of post-Cretaceous and pre-Miocene deformation which could not be observed in the investigation of the Cainozoic stratigraphic units. The main 005°-015°-trending faults are slightly oblique to the N-trending set defined in Chapter 5. Despite this possible early Cainozoic deformation, the collapse of soil and surface sediments into dolines and karst depressions indicates that the dissolution processes along fault planes at least is recent.

Açu valley

Brittle deformation seen in outcrops along the Açu valley is controlled either by the NE- or by the NW-trending fault sets. The most important faults which also display clear signs of deformation at outcrop scale are the NE-trending Pataxós and the NW-trending Queimado lake and Afonso Bezerra faults (Fig. 6.21).

The Pataxós fault cuts across Precambrian crystalline rocks such as granites and gneisses, but it is also seen on LANDSAT images to affect the Cretaceous Açu formation and Quaternary alluvial sediments. A cross-section perpendicular to the Pataxós fault to the east of the Açu river, in the crystalline basement, points to right-lateral strike-slip faulting in brittle and shallow conditions (Fig. 6.21, stereonet a). The fault zone is about 20-30 m wide and different degrees of deformation are visible. There is a sharp transition between the various types of fault rock (Fig. 6.22). The centre of the fault consists of a 10 m wide non-cohesive gouge-rich zone, where crushing processes were strong, indicating burial by less than 4 km. Such NE-trending faults are still pervasive in alluvial sediments west of the Pataxós fault, where the conjugate NW-trending set was also seen (Fig. 6.21, stereonets b and c).

Further east, about 15 km to the southwest of Açu, the alluvial sediments display mainly NW-trending reverse faults as indicated by displaced layers of gravel and sandstone (Fig. 6.21, stereonet e). Offsets are tiny (less than 30 cm) when compared with strike-slip and normal movements. This is the only place in the study area where reverse movement were seen to deform Quaternary rocks.
Figure 6.21 - Geological map of the Açú valley and surrounding area and attitude of the faults represented on equal-area lower-hemisphere stereonets.
Figure 6.22 - The Pataxós (right-lateral, strike-slip, NE-trending) fault cutting across Precambrian orthogneissess (roadcut located ~ 4 km southeast of Açú): (a) general view of the fault zone showing more than 10 m of faulted rocks; (b) detailed view of the sharp transition (white arrows) between noncohesive coarse fault breccia (left) and thin breccia rich in gouge (right).
In the area between the Afonso Bezerra and Queimado lake fault, structures observed in outcrops are strongly controlled by NW-trending strike-slip movement (Fig. 6.21, stereonet f) with a left lateral component as already discussed by Fonseca (1996). The same pattern occurs to the north, in the Açú delta (Fig. 6.21, stereonet g), where the predominant NW-trending faults and subordinate NE-trending faults affect deltaic sediments as already reported by Silva (1991) and discussed in Chapter 5. Low pitch to fault striae were seen in one NE-trending fault only, but, according to borehole data reported by Silva (1991), normal offsets are also important.

In short, outcrop data confirm the three main fault sets described in Chapter 5. There is kinematic and geometric continuity in the structures affecting the Barreiras formation and Quaternary stratigraphic units. Kinematic criteria indicate that the NE-trending set has alternated between right-lateral strike-slip movement and normal movement since the Miocene. The NW-trending set has moved as left-lateral strike-slip and normal strike slip faults. Both sets have acted as conjugate faults during strike-slip movements. In the Holocene, some of the faults have slipped as much as 4-5 m in a maximum 2,730 year period, suggesting high slip rate, and as recent as 4,041-3,689 cal. yr BP. Fault rocks and mineral paragenesis indicate shallow faulting at crustal levels of less than 4 km, and striae suggest that slip on various fault planes was seismic.
Instrumental studies span too short a period for satisfactory analysis of regional seismotectonics or the assessment of recurrence intervals. Historical and geological investigations are consequently being used to extend the record back far beyond the period of observation. Additional information may be recovered from earthquake-induced structures such as hydroplastic deformation, liquefaction, and landslides, which represent secondary ground failure as opposed to primary ground failure (Vittori et al. 1991) of which surface faulting is the obvious example. Secondary ground failure can result in building settlement or tipping, ground cracks, dam instability, road embankment failures and many other kinds of damages (Lowe et al. 1990).

An increasing number of instances of secondary ground failure have been recovered in intraplate settings, mainly in North America and Europe, from historical reports or geological evidence. Abundant examples have been observed in central USA, including the New Madrid seismic zone (e.g. Obermeier 1996a,b), and the Atlantic seaboard of North America (e.g. Amick and Gelinas 1991). Other studies have reported liquefaction events in the Quaternary of western and central Europe (e.g. Davenport and Ringrose 1987).

Hydroplastic deformation, liquefaction, fluidization and seismite are key terms applied for the deformation of unconsolidated sediments. The term liquefaction, originally proposed by Casagrande (1936, in Allen 1984), is generally limited to cases where there is failure of earthquake origin of sedimentary deposits caused by liquidization. Youd (1973) defined liquefaction as the transformation of a solid-state granular material into a liquefied state caused by the application of shear stress due to the build up of pore-water pressure. Lowe (1975) classified the deformation behaviour in unconsolidated
sediments as (a) hydroplastic, (b) liquefied, and (c) fluidized. During hydroplastic
deformation, sediments are not completely liquefied but loose cohesion and deform as
folds. During liquefaction and fluidization, the loosely packed grain framework is broken
down and grains become temporarily suspended in the pore fluid or grains are lifted and
the grain framework destroyed. Primary sedimentary structures are preserved during
hydroplastic behaviour and partially or not preserved during liquefaction and
fluidization. The term hydroplastic deformation is used in the present study to refer to
folds, whereas liquefaction is used to describe liquefied and fluidized behaviour (Lowe
1975) which produces structures such as dikes, pockets and pillars. The term seismitie,
originally proposed by Seilacher (1969) to describe liquefied mud at Elwood beach,
California, is commonly used in the literature to describe both hydroplastic structures
and liquefaction features and is also used in this study.

Liquefaction is one of the most common earthquake-induced phenomena, and depends
on factors related to earthquake characteristics, such as intensity, magnitude, distance
between deposit susceptible to liquefaction and earthquake source, seismic attenuation,
hypocentral distance, duration of seismic shaking, amplitude of cyclic shear stress, and
number of loading cycles (e.g. Allen 1984). Other important factors are related to
sediment characteristics itself such as weak grain-to-grain bounding, loose sediment
packing, good sediment sorting, high permeability, low viscosity and density of water-
sediment mixture, the absence of clay minerals, as well as gas bubbles and organic
matter (e.g. Obermeier 1996a,b).

Sediment depth and water table, combined with overlying sediment features, are other
elements which may influence the liquefaction of a sedimentary deposit. Empirical
observations show that the optimal depth for liquefaction is between 2-10 m (Obemeier
1996a,b) and is rarely deeper than 20 m in depth (Seed 1979). Susceptibility to
liquefaction decreases with low water saturation and with increasing depth to the water
table (Obemeier 1996a,b). It increases when an impermeable or semi-impermeable layer,
which may elevate pore fluid pressure, caps the liquefied sediment (Lowe and LoPiccolo 1974). The age of sediment is an indirect factor which influences liquefaction as it affects the looseness of cohesionless sediment and the depth of the ground water table (Tinsley et al. 1985).

Many models such as the one by Nichols et al. (1994) and field observation (Obermeier 1990) have indicated that a fine impermeable cap is necessary for liquefaction because it helps to increase fluid pressure after earthquakes and allow liquefied material underneath to escape once it is broken. But examples of liquefaction without an impermeable cap necessary to increase the confining pressure have also been described in the literature (Thorson et al. 1986).

According to Berardi et al. (1991), liquefaction can be identified in historical seismicity from information about the outflow of water-sand or water-mud mixtures from cracks in the soil, soil subsidence or soil collapse, and soil sinking followed by building tilting. In the same way, landslides can also be recognised in the historical record, mostly by rock fall/slides and disrupted soil slides (Keefer 1984).

There is little reported evidence of secondary ground failure triggered by earthquakes in northeastern Brazil. The catalogue of historical seismicity by Ferreira (1983) presents a few examples of liquefaction and landslides. As regards the geological record, clastic dikes and convolute folds have been described by Saadi and Torquato (1992) in the Camocin formation of inferred Miocene age in Ceará state. More recently, Fonseca (1996) described hydroplastic and liquefaction structures induced by earthquakes in Quaternary alluvial terraces along the Açú valley. He recognised folds, liquefaction pillars and pockets as the main features affecting gravel and sand. This study did not link secondary ground failure to active faults, estimate magnitude or seismic intensity, or set the evidence in its regional context.
Chapter 7 - Secondary ground failure

The present Chapter reviews evidence of Quaternary liquefaction along several river valleys in the study area. It includes review of the oral accounts and newspaper reports noted by Ferreira (1983), a reinterpretation of the data obtained by Fonseca (1996) along the Açú valley, and a description of new sites of secondary ground failure along other river valleys.

Seismic-induced liquefaction and landslides from historical data

The historical record of northeastern Brazil is short and patchy, because colonization dates back only five centuries, settlement has tended to be concentrated on the coast, and newspaper coverage has been poor. Nevertheless a few valuable reports of earthquake-induced liquefaction and landslides have been recognised in the catalogue of historical seismicity by Ferreira (1983).

Liquefaction occurred during at least two major past earthquakes (Fig. 7.1). The Itaparica-Bahia state earthquake of 22 March 1911 (modified Mercalli intensity - MMI VII), located about 3.0° south of the study area, was followed by soil liquefaction. According to Sampaio (1916, in Ferreira 1983), there has been localised subsidence on the coast of Itaparica, and soil collapse was observed in the epicentral area. The Araticum-Ceará state earthquake swarm in April and March 1969 also caused liquefaction. Several local newspapers described soil collapse and associated landslides; some small streams were filled by disrupted soil (Ferreira 1983).

Landslides were observed during at least three other events (Fig. 7.1). The Caboto-Bahia state earthquake of June 1815 destroyed part of a hill near the village of Cotegipe (Sampaio 1916, in Ferreira 1983). Signs of landslides were noticed during the Pereiro-Ceará state earthquake of 2 January 1968. Local and national newspaper reported boulder falls from hills after the event. Landslides also occurred in the epicentral area of the Senador Pompeu-Ceará state earthquake of 23 February 1968 (MMI VII and
This event dislodged blocks of crystalline rock from steep slopes (Ferreira 1983).

Figure 7.1 - Historic liquefaction and landslide sites interpreted from the catalogue of historical seismicity by Ferreira (1983).
Seismites in the Quaternary record

Although no landslides have been identified in the local geological record so far, there is widespread evidence of hydroplastic deformation and liquefaction induced by earthquakes. The seismites are limited to Quaternary alluvial deposits located mainly in the Açu (Fig. 7.2), Ceará-Mirim (Fig. 7.3), Trairi, Guaraíra and Canguaretama valleys (Fig. 7.4). The sites were found in road cuts and quarries, some of them over 200 m long. The seismites occur in gravel and gravelly sand, but sometimes in muds. All the seismites features were found at depths of between 1 and 5 m.

Two main types of seismites were observed. The first include features related to hydroplastic deformation, such as convolute folds. The second are water-escape structures due to liquefaction such as pillars, sand dikes, pockets and cusp-like structures. The first type of seismite represents soft-sediment folds, whereas the second group represents soft-sediment mixing bodies and soft-sediment intrusions as defined by Lowe (1975).

Convolute folds are usually associated with mesoscale graben-like features in gravel and gravelly sand along the Açu, Ceará-Mirim, Potengi, Trairi, Guaraíra and Canguaretama valleys, whereas the mud-bearing convolute fold beds are limited to the Açu river delta (Fonseca 1996). These folds were formed in loose and wet sediment. There is no clear bounding fracture limiting graben-like structure where the folds occur. Instead, there is a steady continuation of layers into the centre of the trough which indicates that the sediments were tilted without brittle failure (Figs. 7.5 and 7.6). In these examples, intercalated layers of gravel and sand collapsed into the centre of the trough owing to local horizontal extension and as a consequence wavy bedding and convolute folds were formed (Fonseca 1996). Bedding ‘boudinage’ occurred in places (Fig. 7.6) Undeformed layers commonly bound the deformed ones.
Figure 7.2 - Geological map of the area from Açú to Macau. Location of seismsites are indicated by numbers discussed in the text.
Chapter 7 - Secondary ground failure

Figure 7.3 - Simplified geological map of the Ceará-Mirim valley and location of seismites.
Figure 7.4 - Simplified geological map of the Natal to Baia Formosa area and location of seismites.
Figure 7.5 - Convolute folds in gravel in the middle part of the Açú valley: (a) gravel; (b) material collapsed from the quarry wall; (c) folded bedding (seismite 8 in Fig. 7.2).
Liquefaction pillars (Lowe and Lopiccolo 1974) are the most common type of water-escape seismite. They are particularly common along the Açu valley (Figs. 7.7, 7.8 and 7.10) but are also seen along other valleys such as the Ceará-Mirim (Fig. 7.9), Trairi and Canguaretama (Fig. 7.11). The vast majority of cases occur in gravelly sediments underlain and overlain by undeformed beds (Figs. 7.8, 7.9 and 7.10). They are generally marked by elongated vertical columns of pebble concentration which die out upwards or slightly bend to the horizontal-bedding position on the top and normally narrow downwards. Some pillars present the kind of ‘pear drop’ shaped disturbance described by Scott and Price (1988) in Plio-Quaternary sediments in southwestern Turkey (Fig. 7.7). The average pillar height is 1.0 m but some of the pillars exceed 2.0 m; pillar width ranges from 20 to 50 cm.

Two other type of pillars occur in the study area, although less frequently than the type described above. One of them occurs at the contact between the Barreiras formation and overlying alluvial terraces, being the result of liquefaction of both units (Fig. 7.11). This is one of the few cases where brittle or semi-brittle structures were used as a conduit for water escape in pillars. The other case comprises cuspate-shaped pillars formed by the same liquefaction process. In the Açu valley, such structures are marked by pebbles which sank into the liquefied sand underneath. Their final shape is identical to the detrital wedges described by Estévez et al. (1993) in Miocene sediments of southeastern Spain, which formed during seismic-induced liquefaction (Fig. 7.12).

As in the model proposed by Nichols et al. (1994), the liquefied sediment mixture in pillars of the study area was vented upwards by a vertical channel which then filled rapidly with coarse clasts. The coarse particles were replaced by finer grains towards the top. In some cases, there is a textural-upwards zonation, represented by pebbles at the base which grade upwards to gravelly sand and coarse sand. The flow within pillars destroyed sedimentary bedding nearby. Pebbles and other clasts were reoriented during resedimentation to form a well organised fabric parallel to the pillar walls. Multiple flow
occurred in places such as the Ceará-Mirim valley, where superposition of pillars is common (Fig. 7.9). Although some pillars are filled by wall clasts, there are no faults or brecciation processes associated with them, indicating that the whole phenomenon occurred when the sediments were still wet and unconsolidated. Only few examples of pillar-bearing beds display an overlying impermeable cap presumably thanks to erosion.

Intrusive bodies such as sand dikes occur alongside pillars and are also evidence of water escape during liquefaction. Sand dikes are common along the Ceará-Mirim, Canguaretama, Guaraíra, Trairí, Ceará-Mirim and Açú valleys. They are vertical and roughly planar bodies 30 cm and 50 cm wide, and about 2 m long, which cut across the gravel bed. The dikes are composed chiefly of loose, unsorted sand vented from underlying sand layers. Similarly to the bearing-clast sand dikes described by Obermeier (1996a,b), the sand dikes in the study area also display pebbles and clasts from host sediments. Both of the examples of Figs. 7.13 and 7.14 are overlain by an undisturbed thick bearing-clay sand layer more than 2 m in width. Although they vented large quantity of material, there is no clear evidence that liquefied sand reached the palaeosurface. Pockets are less common liquefaction features (Fig. 7.15). They occur in gravel sediments and consist of a bowl-shaped structures filled by clasts (usually pebbles), similar to those described by Postma (1983).

A seismic-induced origin is the most probable cause for secondary ground failure in the study area. Convolute folds and water-escape structures may be produced by gravitational loading during rapid sedimentation or on steep depositional slopes such as those of fan deltas (e.g. Lowe 1975). However, the terrace slopes under review are negligible which rules out the possibility of gravitational sliding of material. Convolute bedding in terraces which overlie carbonate rocks, like those in the central part of the Açú valley, are the result of alluvial collapse over caves, but in northeastern Brazil it does not provide the answer as convolute bedding is not limited to deposits over limestones but also occur in terraces which overlie sandstones and the crystalline
basement. The possibility that the pillar structures and convolute folds associated with graben-like structures were generated by extension-filled fissures can also be ruled out owing to the lack of fractures or mesoscopic faults connected with these features in the study area. Even when brittle structures occur, they predate or postdate the seismites. Localised artesian conditions can be ruled out as the origin of the pillars and dikes not only because artesian conditions are absent but also because there is no evidence of rhythmic sand boils in the clastic dikes. Moreover, the sand dikes cut across sedimentary strata younger than the sand-dike source, which excludes the possibility of syndepositional processes. As both hydroplastic and liquefaction features in the study area present characteristics consistent a seismic origin, occur at multiple locations, sometimes as clusters, are overlain and underlain by undeformed beds, and occur near Quaternary faults, it is reasonable to conclude that these deposits were generated by seismic shaking. The fact that not all seismites are capped by an impermeable layer does not rule out a seismic origin as seismites without confining pressure are also possible (Thorson et al. 1986).

Semi-arid to arid conditions throughout the Quaternary (Mabesoone and Coutinho 1970, Mabesoone and Rolim 1973, Damuth and Fairbridge 1970) led to intense but episodic rainfall which promoted sediment removal and deposition of coarse clastic deposits. A poor vegetation cover contributed to these type of deposit. Under such conditions alluvial valleys are among the few locations where the height of the water tables satisfies the conditions required for liquefaction. The Quaternary seismites are indeed confined to stream banks, flood plains, and deltas. On the other hand, the topography of the coastal plain does not favour landslides, which are accordingly absent from the geological record and reported in historical sources only in the areas of crystalline basement, where steep slopes are present.
Figure 7.6 - Convoluted fold in graben-like structure. Note steep layer bedding in the middle of the picture where a sand layer displays ‘boudinage-like’ feature (hammer measures about 30 cm, seismite 9, Fig. 7.2).
Figure 7.7 - Liquefaction pillar showing 'pear-drop' shape; the arrows mark alignment of displaced pebbles (hammer measures about 30 cm, seismite 1 in Fig. 7.2).
Figure 7.8 - Liquefaction pillar in gravel sediments: (a) gravel, (b) sandstone, (c) material collapsed from the outcrop; the arrows mark alignment of displaced pebbles (hammer measures about 30 cm, seismite 5, Fig. 7.2).
Figure 7.9 - Superposition of pillars in gravelly sediments in the Ceará-Mirim valley. Note that the larger pillar (centre of the picture) is superposed on another smaller pillar (right-hand corner) (GPS device measures about 15 cm, seismite 8, Fig. 7.4).
Figure 7.10 - Liquefaction pillar in gravelly sediment (a) capped by mud-bearing sandstone (b) and soil (c); (d) represents road pavement. Note that both layers above the liquefied gravel are undisturbed (hammer measures about 30 cm, seismite 5, Fig. 7.2).
Figure 7.11 - Liquefaction pillars in the Curimatau valley (seismite 1, Fig. 7.4) formed between the Barreiras formation (a) and a younger alluvial terrace (b). Liquefied sand was vented from (a) to (b) along (c) (the arrows mark direction of displaced pebbles) along minor faults and joints (d).

Figure 7.12 - Cuspade-shaped pillars along the Açú valley (seismite 2, Fig. 7.2, hammer measures about 30 cm).
The strength of palaeo earthquakes

Several empirical relationships between earthquake size and secondary ground failure have been developed in the last two decades. Some of the studies have associated hydroplastic deformation, liquefaction and landslides with earthquake magnitude, others with seismic intensity. Obermeier (1996a,b) insisted that not all methods for palaeoearthquake estimation are useful, and concluded that associations between seismic intensity and secondary ground failure are crude because these phenomena do not always reflect earthquake size. He suggested that methods which take into account earthquake magnitude are more accurate because they make it possible to estimate distance from epicentre and minimum magnitude. But methods that relate ground failure to intensity can provide a first order evaluation of the palaeoseismicity and this approach has been followed bellow.

Secondary ground failure can of course result from repeated small events. Ambraseys and Sarma (1969) pointed out that liquefaction will change dramatically if a sedimentary deposit is subjected to several cycles of low stress before failure. Testing a variety of soils exposed to cycle loading, Lee and Seed (1967) and Peacock and Seed (1968) concluded that a gradual build up of pore pressure may result in failure under small stress. More recently, Dobry (1989, in Obermeier 1996a) concluded that the critical shear strain for liquefaction can be as small as 0.04% for earthquakes characterised by long duration and many cycles. Earthquake swarms in northeastern Brazil, lasting for several months or even years and composed by events equal or below $m_l=5.2$ (Ferreira et al. 1987, Takeya et al. 1989, and Ferreira et al. 1997) favour liquefaction and landslides.

Liquefaction is produced by earthquakes as low as moment magnitude $M=5$ but it becomes common in magnitudes $M\geq5.5-6.0$ (Ambraseys 1988). All seismites recognised in the study area are located no more than 6 km and some of them less than 1 km from a
Quaternary fault. On the assumption that the observed hydroplastic and liquefaction effects were induced by the nearest fault, equations and diagrams from empirical studies were used to work out the minimum magnitude associated with the features. Of course if seismites were induced by distant faults, the associated magnitudes must have been greater than suggested by calculations.

The empirical relationships for sedimentary response of Kuribayashi and Tatsuoka (1975) were used by Allen (1986) to derive the following equation:

$$M = 0.499 \ln \left( \frac{X}{3.162 \times 10^5} \right),$$

where $M$ is the moment magnitude and $X$ is the maximum epicentral radius in kilometres. Applying the maximum distance between seismites and their nearest possible source in the study area (~ 6.0 km), a magnitude $M \geq 5.4$ was obtained. Ambraseys (1988) has proposed the following relation as the lower bound for the moment magnitude ($M$) and maximum epicentral distance from liquefaction effect ($R_{\text{max}}$, in cm):

$$M = -0.31 + (2.65 \times 10^8 R_{\text{max}}) + (0.99 \log R_{\text{max}}).$$

Using this last equation, $M \geq 5.4$ is the threshold magnitude for an epicentral radius of 6.0 km, which is the maximum distance between seismites (seismite 1 in Fig. 7.2 and seismite 8 in Fig. 7.4) and a Quaternary fault in the study area.

An empirical relation between surface-wave magnitude ($M_s$) and maximum epicentral distance where liquefaction has been observed was developed by Youd and Wieczorek (1982). From Fig. 7.16a it can be concluded that the minimum magnitude for seismites less than 1 km is $M_s \geq 5$ and for those 6 km from the fault plane it is $M_s \geq 5.9$. Another relation between moment magnitude ($M$) and maximum epicentral distance to liquefaction effect (Fig. 7.16b), proposed by Munson et al. (1995, modified from Ambraseys 1988 and Obermeier et al. 1993), gives magnitudes of $M \geq 4.6$ and $M \geq 5.5$ to liquefaction effects less than 1 km and 6 km respectively from a related contemporaneous fault.
Figure 7.13 - Sand dike in the Ceará-Mirim valley (seismite 8, Fig. 7.4): (a) gravel, (b) sand dike, (c) bearing-clay sand bed, (d) small pillars associated with the major dike.

Figure 7.14 - Small sand dike cutting across gravel layer and also capped by a bearing-clay sand bed (seismite 8, Fig. 7.4).
Figure 7.15 - Liquefaction pocket in gravelly sediments (for site location see seismite 2, Fig. 7.2): (a) sandstone, (b) gravel, (c) road pavement.
These relationships between magnitude and maximum epicentral distance from liquefaction effects do not take into account the texture, sorting, and composition of liquefied deposits. Several studies have in fact concluded that the liquefaction of gravel and gravelly sand, as in northeastern Brazil, requires a much higher threshold magnitude than deposits chiefly composed by sorted sand and which do not contain mud, gravel or organic matter (Tinsley et al. 1985). According to Obermeier (1996a), the high gravel content of a sedimentary deposit increases the internal friction resistance, which makes liquefaction difficult. Liquefaction of gravel in modern-day earthquakes, including the 1988 Armenian Earthquake (M_s=6.8, Yegian et al. 1994), the 1983 Borah Peak-Idaho Earthquake (M_s=7.3, Youd et al. 1985) and the south-central Indiana Earthquake (M=6.9, Munson et al. 1995), is less common than liquefaction of sorted sand. Valera et al. (1994, in Obermeier 1996b) stated that the threshold magnitude to induce liquefaction in gravel is 7, whereas it is reduced to about 5.5 in sand deposits. If a 30 cm impermeable layer caps the deposit, liquefaction is favoured at lower ground acceleration (Yegin et al. 1994).

It is therefore possible to conclude that the minimum magnitude associated with liquefaction features in the study area ranges from $M \geq 4.6$, for epicentral distances about 1 km, to $M \geq 5.4$, for epicentral distances of about 6 km, using various empirical studies. These minimum magnitudes increase to $M_s \geq 6.8$ if the gravel granulometry of sediments in the study area is taken into account. In every case, the minimum magnitude for induced seismites is much greater than in the current instrumental record. Similar conclusions can be drawn for the two historical earthquakes which have caused liquefaction, which had moment magnitudes of at least $M \geq 4.6$.

Another approach is to associate liquefaction with seismic intensity. Berardi et al. (1991) concluded that liquefaction is limited to earthquake intensity of historical Italian earthquakes MMI $\geq$ IX. According to Sims (1975) an MMI (modified Mercalli intensity) of about VI or over is the threshold intensity for liquefaction, but the National Research
Council (1985) later insisted that liquefaction features become common at MMI intensities VII or greater. Davenport (1994) also proposed an empirical relation between seismic intensity and hydroplastic deformation/liquefaction. He concluded that MMI=VI is the threshold intensity for hydroplastic deformation, and MMI=VII is the threshold intensity for liquefaction.

The map of maximum intensities of northeastern Brazil (Ferreira et al. 1990, Fig. 7.17), which uses modern and historical macroseismic data clearly shows that the E-W- and NW-SE-trending littoral zone has experienced at least MMI=III. MMI=VI, the threshold for hydroplastic deformation, occurs in eleven different places, most of them within the Potiguar basin. MMI=VII, the threshold intensity for liquefaction, has been recorded at four different places, again mainly within the Potiguar basin. According to these intensities, it is possible to conclude that both palaeohydroplastic deformation and palaeoliquefaction are compatible with local instrumental and historical seismicity.

Moderate earthquakes can also trigger landslides. The threshold magnitude necessary, according to Keefer (1984) is MMI=IV, although he admitted that the lowest ever reported intensity associated with landslides is MMI=V and that the usual value is MMI=VI. Keefer (1984) also noted that MMI intensities for landslides distribution are one to five levels lower than those shown by explicit criteria on the MMI scale. In northeastern Brazil, historical landslides characterised by rock fall, rock slide, soil fall, disrupted localised artesian conditions soil slide, soil slump and soil block slide would correspond to a minimum MMI=IV or MMI=V according to Keefer (1984). From the map of maximum intensities (Fig. 7.17), the lowest threshold intensity for landslides (MMI=V) is often attained in the study area.
Figure 7.16 - Empirical relationship between earthquake magnitude and epicentral distance from liquefaction effect: (a) after Munson et al. (1995) (modified from Ambraseys 1988 and Obermeier et al. (1993); (b) after Youd and Wieczorek 1982. Grey rectangles stand for minimum epicentral distance observed for some seismites in the study area.
Figure 7.17 - Map of maximum intensities of northeastern Brazil (modified from Ferreira et al. 1990).
Individual soft-sediment deformation sites may have various interpretations. For example, the cuspade-shaped pillars in the Açú valley (Fig. 7.2) may alternatively be interpreted as sedimentary features caused by overload such as flame structures and load casts associated with passively deformed beds (e.g., Allen 1984). Similarly, convolute folds like those north of Açú (Figs. 7.5 and 7.6) may be related to gravity sliding. The liquefaction pocket presented in Fig. 7.5 can also be interpreted as a synsedimentary palaeochannel. However, the collective occurrence of liquefaction structures in a variety of lithological, sedimentological and topographic conditions, as already pointed out by Fonseca (1996), strongly suggests a palaeoseismic origin. In addition, structures such as the sand dikes in the Ceará-Mirim valley (Figs. 7.13 and 7.14) have been interpreted as unequivocal proof of seismically-induced features (Obermeier et al. 1990, 1996) and their spatial and stratigraphic association with other soft-sediment structures indicates that they were generated by palaeoeartquakes.

To sum up, historical and palaeoseismic data point to past earthquakes much bigger than those observed instrumentally in the last three decades. Historical data reliably indicate liquefaction and landslides in the past 200 years. In every case, they were limited to epicentral areas. The absence of historical seismicity in places where the geological record reveals seismites is probably due to a lack of historical records rather than the stability of the region. The seismicity experienced in the last 200 years in northeastern Brazil has been at a level sufficient to trigger hydroplastic deformation, liquefaction and landslides. Empirical relationships between magnitude and epicentral distance from liquefaction effects (seismites) indicate that palaeoeartquakes are not only bigger than those observed recently but also at different epicentres associated with Quaternary syn-sedimentary faulting.
This chapter has three objectives. The first is to establish what should be considered neotectonic deformation in northeastern Brazil and to describe the neotectonic evolution of the region. At a preliminary level, the concept of neotectonic deformation in northeastern Brazil is extended to most of the Brazilian passive margin, on the basis of the geometry of the South American plate, the stress field, and the local stratigraphic units that can serve to trace the neotectonic deformation.

Second, it compares the neotectonic data presented in the previous chapters and historical and instrumental seismic data from the published literature in an attempt to explain neotectonic evolution and current seismicity of the region. The correlation between both neotectonic and seismic data is based mainly on the geographic distribution of earthquakes and focal mechanisms and their association with the age, kinematics and geometric attributes of neotectonic faults.

Third, it presents a brief review of the main hypotheses advanced to account for the neotectonics and seismicity of northeastern Brazil. The data gathered during the present research are used in order to test and refine these models.

Saadi (1993) made an early attempt to integrate all the available data in the literature in order to define the beginning of the neotectonic period in Brazil. Having correlated neotectonic deformation in the intraplate part of South America with orogenic events in the Andean Cordillera, he placed the onset of the neotectonic period in Brazil during the Incaic II tectonic event in the Andes, which Frustos (1981) placed in the Eocene-Oligocene.

In northeastern Brazil, the last part of the South American plate to be separated from the African plate (Szatmari et al. 1987), it is not yet clear whether, following the South America-Africa breakup, there was a clearly defined, widespread manifested tectonic event in the late Cretaceous to Tertiary. But several studies have revealed late
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Cretaceous to Tertiary tectonism associated with some kind of tectonic uplift and faulting.

According to Françolin and Szatimari (1987) and Szatimari et al. (1987), northeastern Brazil was subjected to N-S-oriented compression during the late Cretaceous which affected several stratigraphic units of the Potiguar basin, including the Açú and Jandaíra formations. Such N-S-oriented compression lasted from the Campanian until the early Tertiary and was caused by the clockwise rotation of the South American continent in relation to the African continent.

Cremonini (1995) has described two major tectonic events which affected the Potiguar basin and were caused by heat from the oceanic plate during the late Cretaceous (Mesocampanian) after the formation of the Potiguar rift. The first and stronger event was responsible for regional erosion during the Mesocampanian as well as extensive uplift and reverse faulting. The second event, during the Tertiary, resulted in E-W folding.

An instance of late to post Cretaceous deformation affected the Potiguar basin. Oliveira et al. (1993) has described the NW-trending Afonso Bezerra fault, which affected the Jandaíra formation, and concluded that early Tertiary movement occurred along the fault because of N-S subhorizontal compression. He suggested that the tectonism was related either to Campanian erosion or to the magmatism that generated the Macau formation during the Eocene-Oligocene. Oliveira et al. (1996) have since described two tectonic events which affected the Jandaíra formation immediately southwest of the study area, the first a ESE-WNW extension and the second a phase of NNW-SSE compression.

Evidence of late to post-Cretaceous deformation leading to uplift and reverse faulting has also been observed offshore. Gomes (1997) has found evidence in some seismic surveys for inversion tectonics on the east coast of Brazil which affected the oceanic crust and overlying Cretaceous to post-Cretaceous sediments but not pre-Miocene age volcanic rocks (12.3 My) in oceanic islands at low latitudes.
Despite all this evidence of late Cretaceous to Tertiary deformation, there is no consensus on how many tectonic events took place and what caused them. Nevertheless, the deformation described in Chapters 4, 5, 6 and 7 shows geometric and kinematic continuity since the start of the Barreiras formation deposition in the Miocene.

The continuity is consistent with the E-W-oriented maximum compression and N-S-oriented extension (Assumpção 1992; Lima et al. 1997; Ferreira et al. 1998) implicit in the shape and direction of movement of the South American plate since the Miocene (Fig. 8.1).

It is therefore reasonable to propose that the beginning of neotectonic deformation in northeastern Brazil coincided with deposition of the Miocene to Pleistocene Barreiras formation. This unit extends from the Amazon region (~ 2° S) to southeastern Brazil (~ 20° S), making it a good reference datum for deformation along most of the Brazilian passive margin. At least in northeastern Brazil, there is no other widespread, post-Cretaceous stratigraphic unit which records Tertiary to Quaternary deformation so well.

**Neotectonic evolution of northeastern Brazil**

It was widespread faulting in the Miocene that allowed deposition of the Barreiras formation in fault-controlled troughs. Several tectonic pulses favoured the development of strike-slip and normal faulting, but no great kinematic or geometric change in the fault pattern has occurred since then. Comparison between the major faults and the mesofaults also indicates pervasive brittle deformation.

The most important structures formed since the Miocene have been the NE- and NW-trending faults, which assume in places the characteristics of conjugate faults and have reactivated pre-existing structures including Cretaceous faults and Precambrian shear zones. Both sets of faults are characterised by normal and strike-slip faulting and sometimes by oblique slip faulting since the Miocene. The cumulative offset along both fault sets has made possible the deposition of several stratigraphic units including the Barreiras formation in the Miocene to middle Pleistocene, and alluvial and aeolian
sediments from the Pleistocene to the Holocene. Several troughs have accumulated as 
much as 260 m of Cainozoic sediments. On occasion, however, neotectonic faults, such 
as the Jundiaí fault, cut across the Precambrian fabric, where only fractures and 
mesoscale faults are filled by sediments.

The reactivation of previous structures doubtless occurred when they were oriented at 
30° to the maximum horizontal compression (Marshak et al. 1982). This is the case for 
the NE- (~60°) and NW-trending (~300°) Cretaceous faults and shear zones. When slip 
occurring on faults under brittle conditions, however, the empirical law which governs 
faulting assumes the form of the Coulomb criterion (Byerlee 1978, Marshak et al. 
1982). Should the maximum compression axis temporarily come to be vertical, normal 
faulting is favoured. Sibson (1977) showed that vertical strike-slip faults lying at 22° to 
32° to the maximum compression, which corresponds to 58°-72° and 292°-302°-trending 
faul.ts in the study area, are favourably oriented for frictional reactivation. Faults resist 
reactivation when their orientation departs by more than ±15° from the above values. 
Sibson (1977) also pointed out that, for nearly pure normal and strike-slip faulting, 
planes are regarded as favourably oriented when the acute angle between the maximum 
compression and the fault plane varies from 12° to 42°, corresponding to the faults that 
trend 48°-78°- and 282°-312°.

Of course the N-trending faults may have been formed by other mechanisms, such as 
local stress regime which temporarily superseded the ongoing E-W-oriented 
compression, or by the reactivation of faults which arouse during the South America-
Africa breakup. Whatever their origin, these faults have acted in several places as 
transfer faults linking the more pervasive NE- and the NW-trending faults.

The area’s tectonic evolution since 30,000 yr BP can be described more precisely thanks 
to the radiocarbon data ages obtained for the present study as well as by other workers. 
Although there is clear continuity with earlier events, the record is detailed enough to 
reveal the interplay between sea-level change, coastal sedimentation and faulting. The 
influence of sea-level changes, as we have seen, can be divided into two major phases.
Figure 8.1 - Reconstruction of the South America plate in the early Tertiary (65My), Oligocene (36My), Miocene (23My), and Pleistocene (< 1.6My). Squares indicate the study area and the white arrows indicate the main direction of plate movement, which generates most of the stress field, in northeastern Brazil. Plate reconstruction are from the Ocean Drilling Stratigraphic Network Web site (qdsn.de/odsn/services/paleomap/paleomap.html).
From 10,000 cal. yr BP sea-level rose, passed the current level about 7,000 cal. yr BP, and reached a 2 m highstand at 5,000-5,500 cal. yr BP. The second phase is characterised by steady fall. During the regression phase, significant coastal sedimentation and faulting took place. Strong winds eroded the exposed shelf, redepositing a large quantity of sand onshore and fault reactivation of several NE- and NW-trending faults encouraged aeolian sedimentation in downfaulted blocks mainly along the N-trending coast between Natal to Baia Formosa and Ceará-Mirim to Rio do Fogo.

In the Açu delta, Holocene sedimentary evolution was strongly controlled by shallow faults (Silva 1991) which coincide with Cretaceous faults of the Potiguar basin and in the crystalline basement previously mapped by Fortes (1987). Another example of Holocene fault reactivation is provided by the NE-trending Carnaubais fault, which moved in the late Pleistocene and again at 4,240-2,740 cal. yr BP. The latter movement amounted to 4-5 m and displaced Holocene coastal deposits as well as deposits of the Tibau and Guamaré formation, which were concentrated in the crustal block between São Bento and Touros. The NE-trending Jundiaí fault moved 4,865-4,566 cal. yr BP and offset alluvial deposits onshore. The NW-trending Boa Cica fault was active after deposition of the Guaraíra beachrock and before the deposition of the Barreta beachrock, between 7,430 and 5,620 cal. yr BP, when the older beachrock was downfaulted by 4-5 m.

**Correlation between neotectonics, historical and instrumental seismicity**

Seismological knowledge has made great progress in recent years in Brazil. The historical record for northeastern Brazil only goes back to the last century and is thus too short to provide a realistic evaluation of the past seismic activity. Not surprisingly, most of the data (Ferreira 1983) are concentrated along the coast where colonization of Brazil took place. The same can be said about instrumental seismicity, which is concentrated in the coastal zone where the density of instrumental studies has been greater than inland. When compared with other coastal areas, the area which comprises
the Potiguar basin and its margins displays usually high level of historical and instrumental seismicity.

Fig. 8.2 depicts the historical and instrumental seismicity of northeastern Brazil down to magnitude $M = 2.0$ after the catalogue of the seismic bulletins of the Brazilian Geophysical magazine (Revista Brasileira de Geofísica, volumes 1-13) and Berrocal et al. (1984) superimposed on a simplified geological map. Besides the coastal concentration there are epicentres within the areas underlain by the Mesozoic-Cainozoic sedimentary cover and crystalline basement. The focal depths, usually between 5-12 km, indicate that the origin of the vast majority of earthquakes is in fact place within the crystalline basement below the sedimentary cover (Ferreira et al. 1997, Ferreira et al. 1998).

The focal mechanism data in Fig. 8.2 show that the maximum horizontal compression is E-W-oriented and bisects the acute angle formed between the NE- and NW-trending faults generated since the Miocene. The NE- and NW-faults trends match the fault orientation and the auxiliary planes of most of the focal mechanisms. If the angular relationship between the NE- and NW-trending faults and the stress field are taken into account, it can be inferred that both faults sets are likely to have undergone reactivation under the present stress condition.

This was confirmed by Ferreira et al. (1997) when analysing the relationship between focal mechanisms and the major Cretaceous faults which mark the boundaries of the Potiguar rift (see Fig. 8.2). They drew attention to the parallelism between the seismogenic Samambaia fault (focal mechanism a) and the orientation of the rift margins offshore, to the northeast of Touros city and showed that the focal mechanism solution in the Açú area (focal mechanism b, b') coincides with the main direction of the Potiguar rift nearby. It is worth adding that such fault plane solutions are not consistent with the direction of shear zones in the area. There seems to be no direct relationship between the focal mechanism a and shear zones nearby. It suggests that although the Samambaia fault is parallel to some NE-trending Cretaceous faults nearby as well as the NE-
trending Cainozoic Boqueirão fault, it cuts across the N-trending shear zones which are not in the optimal direction for reactivation. This is a clear case of where seismic and Cainozoic structures are primarily influenced by zones of weakness which are susceptible to reactivation by E-W maximum compression and N-S extension.

Another case where the focal mechanism solution does not agree with previous structures is provided by the Tabuleiro Grande focal mechanism (c in Fig. 8.2). In this case, the fault plane solution is NNW-oriented (strike 347°), whereas the major Precambrian shear zones nearby trend NE (strike 25°-40°).

Finally, Ferreira et al. (1997) concluded that the E- and NE-trending faults at Palhano and Cascavel (focal mechanisms d and e, e’, e’’ respectively) correspond to the main faults that bound the Jacaúna graben and the Potiguar rift offshore (Fig. 8.2). In their view these two directions may have been reactivated onshore. The present study shows that the NE-trending faults corresponding to three Cascavel focal mechanisms are matched by faults in the geological record onshore. However, the E-W-trending fault of the Palhano focal mechanism, although parallel to the Ceará-Mirim dike swarm, has no counterpart in the Miocene to Holocene sedimentary record in the study area. In conclusion, the most common fault trends seen in the majority of focal mechanisms correspond to neotectonic faults bar the E-trending faults which are not represented in the geological record.

Moving on to the relation between the seismic and aseismic nature of neotectonic faults and their connection with seismic data, it is possible to make some observations and propose a few hypotheses. Focal mechanisms point to the dominance of strike-slip faults, whereas neotectonic data point to an important, sometimes predominant, component of normal faulting from the Miocene to the Holocene. Strike-slip focal mechanisms may dominate instrumental seismicity because this represents seismic slip and because strike-slip movement may correspond to the vast majority of displacements which take place in the area, whereas normal faulting caused by neotectonic deformation could be caused by aseismic movement by creep.
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Figure 8.2 - Relationship between historical and instrumental seismicity of northeastern Brazil and geological features. Epicentre locations are from the seismic bulletin of the Brazilian geophysical magazine (Revista Brasileira de Geofisica, volumes 1-13) and from the catalogue of Berrocal et al. (1984), whereas focal mechanisms are from Ferreira et al. (1997). Cretaceous faults of the Potiguar basin are after Matos (1992).
Processes related to creep have been described both at deep and shallow crustal depths. Several studies show that dislocation in some minerals such as quartz is important at temperatures greater than 300°C (Sibson 1983, 1985). However, aseismic creep, though rare in active faults, commonly occurs at very shallow crustal levels in the sedimentary cover of the Salton trough, California (USA) in poorly consolidated rocks (Scholz 1990).

In northeastern Brazil, aseismic slip on normal faults may also be restricted to the upper crustal levels in sedimentary stratigraphic units, whereas strike-slip faulting by seismic slip could predominate at greater depths in the crystalline basement. This inference apparently clashes with the evidence for liquefaction described in the Quaternary alluvial sediments (Chapter 7), which indicates strong seismicity in shallow sediments. On the other hand such liquefaction could have been caused by strong earthquakes in the crystalline basement underneath the sedimentary cover at depths as great as 5-10 km.

Few inferences can be made about the recurrence of earthquakes from neotectonic or seismological data. If we accept, with Muir-Wood and Mallard (1992), that a fault which has moved in the current tectonic regime should be considered active, all the faults described in Chapters 4, 5 and 6 should be regarded as active. But there are insufficient stratigraphic or radiometric ages to establish the frequency of earthquakes.

Moreover, palaeoseismic investigations in Australia, India and North America have shown that seismic activity in intraplate settings could be at intervals of 10,000 to 100,000 years (e.g. Crone et al. 1997; Muir-Wood and Mallard 1992). The brevity of the period covered by historical and instrumental investigations, coupled with the lack of radiocarbon ages appears grossly inadequate for any secure inferences about seismic recurrence.

Nevertheless, although the historical and instrumental data suggests that surface rupture has not taken place in northeastern Brazil in the last 200 years, the neotectonic evidence indicates much stronger seismicity in the past causing not only surface rupture but also
coastal coseismic uplift and widespread liquefaction. According to Bonilla (1988) and dePolo (1994), the threshold of surface rupture is local magnitude $M_l = 5.5$ and $M_s = 5.6$, which has not occurred in the last 200 years. In the Macaíba quarries, however, there is clear evidence for surface faulting as recently as 4,865 to 4,566 cal. yr BP. Moreover, the amount of liquefaction and hydroplastic deformation in alluvial sediments, and of surface rupture and coseismic uplift at several other locations, all point to higher levels of stronger seismicity in recent millennia.

The origin of neotectonic deformation and current seismicity in northeastern Brazil

Several hypotheses have been proposed to explain the present and former seismicity of northeastern Brazil. One of the first attempts in the framework of plate tectonics was made by Sykes (1978). He linked the distribution of intraplate earthquakes with that of igneous rocks postdating continental rifting in many intraplate parts of the world, including northeastern Brazil. One of his examples brought together the Macau formation, the southern margin of the Potiguar basin, the abrupt change in strike of the continental margin from easterly to southerly at northern Brazil, and the Fernando de Noronha fracture zone (a major E-trending transform fault located offshore about $4^\circ$ S latitude). This group of geological features was regarded by him as a major weakness zone in northeastern Brazil, which could explain both former and present-day seismicity.

This idea of E-trending weakness zones was pursued by other studies based mainly on neotectonic data. The best example is the model proposed by Torres et al. (1990) and Torres (1994) to explain both neotectonic evolution and seismicity of the study area. They proposed two E-trending right-lateral mega shear zones extending offshore which would give rise to a distinctive stress pattern and a complex system of seismogenic faults. Gomes (1997) also proposed that great oceanic fractures in the Brazilian platform might behave as zones of weakness where intraplate seismicity could be concentrated.

The evidence obtained during the present study does not support these early proposals. First, there is no Tertiary to Quaternary pervasive faulting in the study area consistent
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with the E-trending mega shear or weakness zones proposed by them. Despite the presence of geological features such as the E-trending Ceará-Mirim dike swarm of middle Jurassic to early Cretaceous age (see Chapter 2), no reactivation onshore in this direction was observed.

Second, there is no evidence from plate movements to indicate differential slip of the South American plate along transform zones. The spreading rate between the South American and African plates varies from 34.1 mm/year at latitude 6° S to 30.8 mm/year at latitude 54.6° S (DeMets et al. 1990). But no significant difference in spreading rates was observed across several transform zones in the South Atlantic, specially at low latitudes. This fact does not favour reactivation of transform faults or their behaviour as zones of weakness.

Third, there is no indication from the neotectonic data or plate geometry since the Miocene that the stress field has changed appreciably in the late Tertiary and Quaternary in response to the E-trending mega shear movement proposed by Torres et al. (1990) and Torres (1994), although some kind of neotectonic, and even present seismic deformation, along such “weakness zones” cannot be ruled out. An example is the clear correlation between historical and instrumental seismicity and the Potiguar basin margins (see Fig. 8.2).

The model by Assumpção (1992) favours the superposition of regional stress caused by ridge push and asthenospheric drag, and of local stress fields caused by density contrast between the continental and oceanic crust, as well as flexure due to the sedimentary load on the continental shelf. Such a superposition would generate strong E-W-compression and N-S-extension which would trigger earthquakes in northeastern Brazil. The concentration of epicentres along the E-W-oriented coast and the relative lack of seismic activity to the south of longitude 8.5° S would favour this hypothesis. A numerical model by Conblentz and Richardson (1996) led to the conclusion that the orientation and magnitude of the stress field is changed significantly by continental margins and topography. But they also concluded that lateral density variations associated with the
Brazilian continental margin were not sufficient to support the observed change in stress field suggested by Assumpção’s (1992) model, and alternatively, they suggested that other factors such as lithosphere thickness or mean crustal density values were likely to control the stress field in the region. Ferreira et al. (1998), however, pointed out that Coblentz and Richardson’s (1996) model has generated a uniform maximum horizontal compression in northeastern Brazil which does not correspond to reality. In addition, they note that the model does not include the effects of local flexural stresses caused by the sedimentary load of marginal basins. They therefore support the hypothesis proposed by Assumpção (1992) that the local and regional superposition of a stress field is primarily responsible for the relatively high seismicity in the E-W-trending coast of northeastern Brazil.

The present investigation indicates that NE- and NW-trending faults, driven by E-W-oriented compression, have been active in northeastern Brazil since the Miocene and display a level of tectonic intensity much greater than the one currently seen in historical and instrumental data. The present investigation also confirms strong neotectonic activity at shallow levels in the Potiguar basin and a close correlation with instrumental seismic activity in the Precambrian crystalline basement at depths of 5-10 km.

This relationship between Tertiary to Quaternary faults and the stress field is valid for the E-W-trending coast where the stress field is roughly constant. It should change west of the study area, in the western Potiguar basin, where the coast and the stress field bend NW-SE.

The model by Assumpção (1992) is an elegant explanation for the relatively high seismicity in the E-W-trending coast and low seismicity in the N-S-trending coast of northeastern Brazil and would also explain the pervasive faulting process that has taken place in the study area since the Miocene. However, it does not provide a full explanation of why the seismicity is not strong in other parts of the E-W coast west of Fortaleza, where similar processes of superposition of local and regional stress field
might be expected. One possibility is that sedimentary load of the coastal basins such as the Potiguar basin, is here insufficient to trigger seismicity.
CONCLUSIONS AND SEISMIC-HAZARD IMPLICATIONS

The background to neotectonic deformation in northeastern Brazil can be summed up as follows. Northeastern Brazil is located on the Brazilian passive margin within the intraplate part of South America. The region displays a Precambrian crystalline basement deformed mainly by shear zones, overlain by Cretaceous sedimentary basins formed during the South America-Africa breakup and by Cainozoic stratigraphic units.

Although the orogenic evolution of the Andes Cordillera influences, in some degree, the evolution of other parts of the South American plate, it is not sufficient to trigger faulting processes at distant intraplate regions. In other words, the neotectonic period in northeastern Brazil to some extent responded to local parameters. Strong faulting started in the Miocene, when, to judge from the geometry and orientation of the South American plate, the current stress field was established. Since then, the region has been under E-W-oriented compression and N-S-oriented extension.

There are three main sets of faults in the region, trending NE, NW and N. The maximum horizontal compression deduced from focal mechanisms is oblique to the NE- and the NW-trending faults in a compressive direction, which favours right-lateral and left-lateral strike-slip respectively. This obliquity is about 30° for the majority of the fault planes. The NE- and NW-trending faults have the same kinematic pattern as that found in focal mechanisms by seismological studies. In particular, there is a clear association between mapped faults and seismicity in the João Câmara epicentral area, where the earthquake swarm of 1986-1989 (the Samambaia fault) is parallel to a fault which cuts across the Barreiras formation and continues offshore.

Although strike-slip offsets could not be quantified, vertical offsets indicate significant movement along major faults since the Miocene. The base of the Miocene-Pleistocene Barreiras formation was displaced by as much as 260 m in the Potengi half-graben. Pleistocene foreshore-to-shoreface deposits in parts of the littoral zone were uplifted very probably by the Carnaubais fault by some dozen metres. Holocene coastal deposits
to the east of the Carnaubais fault were uplifted by 4-5 m in excess of the glacioisostatic predictions. The uplift is matched by downfaulting of coastal deposits to the west. The uplift was probably associated with coseismic movement and is sufficient to disturb the sea-level record.

The geological record of northeastern Brazil is characterised by successive phases of fault reactivation. The vast majority of Tertiary to Quaternary faults represent reactivation of pre-existing zones of weakness such as Cretaceous planes or basement-controlled structures, which are in the range of optimum orientation for reactivation, suggesting this should be an important condition for deformation of previous structures in the region. But important exceptions have been found such as the Jundiaí and Pataxós faults which cut across existing structures.

Dating of deformed and undeformed beachrock bodies, as well as permitting comparison of beachrock age and altitude with glacioisostatic prediction, helped to bracket the age of the Boa Cica and the Carnaubais faults. Vertical offset and radiocarbon ages in these two examples indicate minimum slip rates of 2.8 mm/yr and 1.3 mm/yr respectively, which are one order of magnitude lower then some slip rates observed in active faults at plate margins.

The pervasive faulting has controlled sedimentation and geomorphological features across the region. The Barreiras formation usually occurs in troughs controlled by NE- and NW-trending faults. This pattern has been already observed in the Cainozoic record of southeastern Brazil, where sediments occur usually in small grabens (Saadi 1990; Saadi and Pedrosa-Soares 1990).

Geomorphological features seen along the coastal plain have a tectonic origin. Valleys are filled by alluvial and aeolian sediments which overlie the Barreiras formation, and horsts are formed by dissected Barreiras plateaux. Extensive bodies of aeolian sediments which penetrate as far as 15 km inland are formed by the combination of downfaulted blocks, sea-level fall, and wind directions which facilitated their deposition along
grabens. But some geological features at the surface may not correspond to faulting processes at deeper crustal levels, as indicated by the gravity study of Morreira et al. (1990), who suggested that although the Samambaia fault is oblique to Precambrian shear zones at the surface, it is parallel to strong Bouguer gradients reflecting the alignment of a deeper structural and lithological framework.

Miocene to Holocene brittle faulting at shallow crustal levels (less than 4 km) is usually associated with seismic slip and is recognised in Quaternary alluvial sediments composed chiefly of gravel and gravelly sand and in Barreiras sandstones. The empirical relationship between magnitude and epicentral distance from liquefaction effects, and comparisons with seismites in gravel, reinforce this view and indicate palaeoearthquakes of \( M \geq 6.8 \). The absence of magmatism associated with neotectonic activity in northeastern Brazil may indicate that neotectonic deformation is limited to the upper crust and is not associated with the deep crustal structures that allowed previous magmatic events such as the Rio Ceará-Mirim and Macau volcanism to occur.

The methods used during this study, including remote sensing, borehole logging and geophysical investigation, were useful for locating and determining the geometry of Tertiary to Quaternary faults. Outcrop study combined with stratigraphic and radiocarbon dating made possible kinematic and temporal analysis of faulting. The \(^{14}\)C ages analysed by the first-order method of Vita-Finzi (1991) are consistent with conventional results presented by Oliveira et al. (1990) for local beachrocks. The palaeoseismic evidence of surface faulting, coseismic uplift, liquefaction and soil disruption indicates events much stronger than those in the instrumental and historical record. The present study suggests that, whereas most of the seismicity is accommodated by strike-slip faults, aseismic slip may affect normal faults in the upper crust, mainly in loose sedimentary rocks.

Although some seismological studies indicate that the E-W coast in northeastern Brazil is more seismic than other coastal areas along the Brazilian passive margins, there is no reason to believe that the neotectonic evolution and pervasive faulting observed here is
not repeated in other coastal areas on the eastern part of the South American plate. The lithology and age of the Precambrian crust of northeastern Brazil are not the most important controls of fault reactivation and seismicity: rather, it is the stress field orientation, its magnitude, and the orientation of existing faults which appear to determine Tertiary to Quaternary seismicity.

Seismic hazard

One of the most important applications of neotectonic investigation is the assessment of seismic hazard. The recognition of strong and pervasive seismicity in the neotectonic record of northeastern Brazil thus has far-reaching implications.

Most surface-faulting earthquakes in intraplate settings have taken place in “unexpected locations”, regarded as essentially aseismic or considered to display low seismicity (Crone et al. 1997). However, earthquakes in northeastern Brazil are known since the beginning of the last century and have occurred at shallow depths, producing noise locally named as *estrondos* (bursts). Earthquakes as great as 5.0 to 5.2 m$_s$ in the 1986-1989 earthquake swarm along the Samambaia fault, for instance, damaged more than 30,000 buildings and caused significant panic in the population which was displaced during the events.

Even so, public awareness regarding active faults and earthquakes in northeastern Brazil, as in many other intraplate settings, is relatively low. Because intraplate earthquakes, such as the ones in northeastern Brazil, are less frequent than those that occur at plate margins, both the population and the structures are less prepared to cope with them. Another reason why intraplate earthquakes can cause serious damage is that they may produce ground motion larger than comparable-size interplate earthquakes (Crone et al. 1992, Hanks and Johnston 1992). Moreover, McGarr (1984), has shown that strike-slip earthquakes, such as the ones in northeastern Brazil, may produce stronger ground motion than normal earthquakes of comparable size and depth. This is a fact of concern, as northeastern Brazil displays a great concentration of dams, used for both electricity generation and irrigation, and some important pipelines and oil fields. In addition, the
coastal area is densely populated, including urban areas such as Recife with more than 2.8 million inhabitants.

An essential step in seismic hazard assessment is the recognition of active faults. According to the U.S. Regulatory Commission (1982), an active fault is one which has moved during the last 10,000 years. From this point of view, the Jundiaí, Carnaubais, Boa Cica, and Samambaia faults should be considered active. But if we concede that a fault should be designated active or extinct in terms of local parameters (Muir-Wood and Mallard 1992) several other faults across the region which are optimally oriented for reactivation should also be regarded as potential zones for earthquake generation.

Several approaches can be used to estimate earthquake size generated in the past or likely to occur along active faults and faults which have optimum orientation for reactivation. Assessing the seismic hazard where the historical record is short is a matter more of judgment than of knowledge (Johnston and Nava 1990). Therefore, this approach is likely to yield controversial results. Likewise, the instrumental record is not sufficient for dependable seismic-hazard analysis, although the lack of both historical and instrumental seismicity along some active faults or those likely to be reactivated may in fact indicate that they are accumulating stress which could be released by seismic events in the future.

It follows that the most reliable approach for seismic-hazard assessment in northeastern Brazil is one based on neotectonic evidence, as this can easily be integrated with historical and instrumental data. One of the most common palaeoseismic methods uses palaeosurface ruptures for earthquake estimation. It has been applied to many intraplate regions across the world (McCalpin 1996), but, despite clear evidence of surface breaks in northeastern Brazil, the length of the ruptures is difficult to measure with confidence as weathering and erosion along fault scarps make it impossible to quantify the size of palaeoruptures. Erroneous assessment could mean that the magnitude of palaeoearthquakes is in error. For example, if scarps generated by more than one event
are assigned to a single palaeoearthquake, the maximum event could be grossly overestimated.

Yet, although it is not often possible to quantify the size of palaeoearthquakes using rupture length, some estimate of their minimum size can be made. In the historical data set of Wells and Coppersmith (1994), the minimum surface rupture associated with normal faults is about $M=5.2$, whereas in strike-slip faults it is $M=5.8$. These may provisionally be taken as the minimum moment magnitudes for northeastern Brazil in the Holocene.

Methods of palaeoseismic estimation using secondary ground failure such as liquefaction and landslides are less developed than direct methods using fault scarp length or displacement. But empirical studies clearly indicate that earthquakes which cause liquefaction in gravelly sediments, such as the Quaternary alluvial deposits in the study area, have magnitudes $M_s \geq 6.8$ (Yegian et al. 1994). There is no reason to rule out events as strong as those in northeastern Brazil in the future.

Finally, some kind of assessment can be made by comparison with examples in similar settings elsewhere (dePolo and Slemmons 1990). Like the passive margin of North America, the passive margin of South America is characterised by areas of relatively high and areas of relatively low seismicity. It may be that some areas of low seismicity are underrepresented because of inadequate monitoring. The relatively high seismicity of northeastern Brazil is not exceptional: several other intraplate regions across the world show similar or even higher activity. Earthquakes as high as $m_b 6.0-6.5$ have occurred in continental-rift environments under compression. Examples include the Rhine graben (Europe), the Cawbay and Godava grabens (India), the St. Lawrence rift (North America), the Adelaide and Fitzroy troughs (Australia), and the Sirte graben (Africa) (Johnston and Nava 1990). But one of the most striking parallels with the tectonic setting in northeastern Brazil is found in west Africa, where the 1983 Guinea earthquake ($m_b = 6.4$) produced at least 9 km of surface rupture. The region was part of the South America plate before the South Atlantic opening, and the earthquake took
place in Precambrian crust which is a continuation of crust in northeastern Brazil, which was subjected to similar Cretaceous evolution, and which is in a NW-SE-compressional stress field.

The above, taken in conjunction with the neotectonic data presented in this thesis, would seem to indicate that the potential for large earthquakes in northeastern Brazil has been underestimated largely because it is much greater than that represented in the historical and instrumental record.

**Suggestions for further work**

Despite the advances in seismological research discussed above, and the neotectonic studies presented in the current thesis, much uncertainty still remains. Consequently, some suggestions for further investigations are presented below.

The present thesis presented outcrop data and related radiocarbon analyses along the littoral zone. But farther inland little has been done. Trenching and further numerical dating along some of the fault zones identified in the current research may establish more precisely the timing and progress of faulting, which could in turn lead to recurrence and slip rate estimates.

The Samambaia fault has not yet produced a recent surface break, which makes it unsuitable for the use of the surface-rupture seismic hazard approach. However, the rupture-area method (Wells and Coppersmith 1994) can be used here in order to estimate the potential for large earthquakes. This approach would benefit from precise microseismic studies of aftershocks like those carried out by Takeya *et al.* (1989). Moreover, the use of the frictional-fault-length technique of Slemmons and Chung (1982) for strike-slip faults in intraplate settings can allow the determination of the total system length that could rupture during individual events.

To judge from instrumental seismicity, surface rupture is unlikely anywhere in the area. But several epicentral areas, such as João Câmar (m<sub>0</sub> 5.0, maximum earthquake),
display intermittent lakes whose sediments are usually wet and loose. They provide ideal sites for trenching in order to check any probable liquefaction effect caused by earthquakes as large as 5.2 m\(\text{b}\) (the highest instrumental event in the region) or smaller ones combined in swarm events. Seismological studies such as those that followed the main shocks in several earthquake swarms in northeastern Brazil (e.g., Ferreira et al. 1987, Takeya et al. 1989) should in future include postfaulting geodetic monitoring in order to detect possible aftershock slip along fault zones.

All these procedures need to be combined with structural modelling. There are no seismic building codes for northeastern Brazil and for that matter anywhere in Brazil. The recognition of a real seismic hazard in the region could at the very least lead to measures which can mitigate the impact of seismicity in the more densely populated areas.
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Holocene coastal tectonics in NE Brazil

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Abstract: We have identified two beachrock facies in NE Brazil which can be used as reliable sea-level indicators. Facies (a) represents the lower foreshore and the upper shoreface, being mainly characterized by coarse unsorted sandstones with trough cross-stratification, whereas facies (b) corresponds to sediments deposited on the middle to lower foreshore and is usually characterized by medium to coarse, better sorted sandstones which form seaward-dipping, swash cross beds. The age of the beachrock bodies ranges from c. 7000 to 1150 cal. a BP. Although the relative sea-level record is complicated by oscillations which are probably due to minor climatic changes, it was possible to identify a general rise at c. 7000 cal. a BP which reached its maximum of +2 m c. 5000 cal. a BP and then started to fall to its present level some 300 years later. Our sea-level data are consistent with glacio-hydro-isostatic models for the region but dates of 4080 cal. a BP on shells in growth position at 4-5 m above sea level and 2780 cal. a BP on a beachrock at sea level east of the Camaubais fault point to rapid, possibly coseismic, late Holocene emergence.

Identifying and quantifying modest and local, coastal emergence or submergence caused by tectonics at passive continental margins may be hampered by eustatic and isostatic effects. There are two main problems: how to distinguish variations in sea level from local emergence and submergence of tectonic origin, and whether there is any widespread sea-level indicator which may be used to identify such variations on a tropical coast.

The Earth’s isostatic adjustment caused by deglaciation associated with ocean loading is now seen as a key explanation of sea-level changes along many coasts. Glacio-hydro-isostatic models explain such adjustments and have provided curves which have been used as trends of sea-level changes all over the world (e.g. Clark et al. 1978; Newman et al. 1980; Peltier 1988). The Brazilian coastal area is distant from the main Pleistocene glaciation centres. However, it is generally accepted (e.g. Walcott 1972; Newman et al. 1980) that such distant regions have also shown an isostatic response to deglaciation and the resulting sea-level rise: continental regions such as South America and Africa were flexed upward as a consequence of the depression of oceanic areas and the transfer of mass from regions below the oceanic lithosphere to beneath the continents. Sea-level studies in Brazil (e.g. Fairbridge 1976; Suguio et al. 1985) have shown that the Brazilian coastline has indeed undergone emergence in Holocene time and that raised beaches are a common feature. However, there is little information on how local coastal chronologies match the increasingly sophisticated glacio-hydro-isostatic models that are emerging and whether it is possible to distinguish the glacio-hydro-isostatic component from the tectonic component in relative sea-level changes.

The second problem, that of palaeo-shoreline indicators, demands a local solution. In tropical regions such as northeastern Brazil, beachrock is one of the most common coastal deposits. Beachrock may extend for several kilometres along the littoral zone and is, in places, the only available material for dating. Many studies quote beachrock as a reliable sea-level indicator in microtidal coastlines (e.g. Inden & Moore 1983; Hoppley 1986; Cooper 1991; Kindler & Bain 1993; Ramsay 1995). However, little is known about how beachrock may be used as a sea-level indicator in mesotidal coasts such as those of northeastern Brazil.

This study presents a description of the major recent coastal sedimentary deposits, mostly beachrock, along the littoral zone between Macau and
Cunhaú, northeastern Brazil (Fig. 1), and the corresponding radiocarbon chronology in an attempt to identify crustal movements in the Holocene. The occurrence of well preserved raised coastal deposits along such a littoral zone makes it an ideal site for a sea-level investigation. This paper emphasizes sedimentary structures as a basis for facies subdivision in the transitional (beach) environment where beachrock is formed. By comparing beachrock with modern beaches, we propose its use as a valid sea-level indicator in mesotidal coasts. We also compare relative sea-level changes in the area with known models of glacio-hydro-isostatic behaviour predicted for the region, to identify the local sea-level history and isolate tectonic movements. Tectonic analysis is confined to what the coastal data reveal about the local history of emergence and submergence.

**Tectonic and coastal settings**

The study area is located in the Brazilian continental margin, 5° to the south of the Equator (Fig. 1). A simplified stratigraphy for the region starts with a deformed crystalline basement of Precambrian age; Cretaceous basins formed by reactivation of Precambrian shear zones during the break-up of the South Atlantic overlie this basement, followed mainly by Cenozoic sedimentary rocks (Almeida et al. 1981; Mattos 1992). Previous evidence of late Quaternary coastal tectonics in the region was presented by Martin et al. (1986a), who recognized tectonic vertical movements of up to 3 m along the Salvador area (Fig. 1). Much of the seismicity of northeastern Brazil is less than 12 km deep. The region experiences half of Brazil’s known seismicity.
and has been subject to a series of earthquake swarms (Ferreira et al. 1987; Takeya et al. 1989). An active fault (Samambaia fault, Fig. 2), well defined by earthquake epicentres and focal mechanisms (Takeya et al. 1989), is found near João Câmara at a depth of 5–10 km but evidence that it reaches the surface or can be traced to the coast has not yet been presented.

Northeastern Brazil has a tropical climate, with average temperatures of c. 30°C. The rainfall rate is about 600–1000 mm/year (Nimer 1989). The coast has a mesotidal regime, where normal tides attain a maximum of 1.0–2.0 m and spring tides have a range of 3.2 m (Hayes 1979). Recent erosion and weathering have left the Quaternary sedimentary record relatively unscathed, although oysters and algae obscure some outcrops along the littoral zone.

Sea-level investigation and glacio-hydroisostatic models on NE Brazil

Darwin (1841) was perhaps the first to describe the coastal geomorphology of this region, in particular using the physiography of the reef off Recife (or Pernambuco, as it was then called) to infer a change in relative sea level. Since then, many of the major physiographical units such as coastal barriers, sandstone reefs, and marine terraces have been analysed by other workers (e.g. Bigarella et al. 1961). Only a few deposits have been dated, however, and the resulting chronology for this coast depends largely on relative dating based on geomorphological, sedimentary and stratigraphical attributes.

The use of radiocarbon analyses in Brazilian sea-level studies began in the 1960s. The first results were published by van Andel & Laborel (1964), who reconstructed the sea-level history of the Recife area (Fig. 1). Nevertheless, it was not until 1971 that Delibrias & Laborel studied a coast more than 3000 km long from Recife (Fig. 1) to Santo Amaro (southeastern Brazil) and made an initial attempt to understand the processes responsible for the last 7000 years of sea-level change in Brazil. Fairbridge (1976) also proposed a single sea-level curve for the Brazilian coast, which was based mainly on shell middens from the southeastern and south coasts. Five periods of transgression with amplitudes of 1–5 m were proposed.

More localized depth–time diagrams have been presented in the last two decades. The most complete investigation was published by Suguio et al. (1985), who established differentMid-
Late Holocene sea-level curves. The best example described by them was the 'Salvador curve', which is the best constrained sea-level curve for the Brazilian coast so far (diagram (a) in Fig. 1). The Salvador curve is based on 60 \(^14\)C ages of geological, biological, pre-historical (shell middens) sea-level indicators collected along 50 km of littoral zone to the north of Salvador (Fig. 1). It indicates that a relative sea-level rise began shortly before 7000 a BP, reached 5 m above the current sea level briefly before 5000 a BP, and then fell to its present position. Two regressive periods around 4000 a BP and 2700 a BP were also indicated. More recently, a study by Dominguez et al. (1990) concluded that the Salvador curve was also valid for the Recife area (Fig. 1). They dismissed samples which did not fit that curve as contaminated.

Glacio-hydro-isostatic models predict a major post-glacial sea-level rise along the Brazilian coast. The first relevant scheme to be published was by Clark et al. (1978), who divided the Earth's surface into six regions of different isostatic patterns and used a numerical model to calculate sea-level changes. Emerged beaches were predicted in four regions, including zone IV (continental shorelines). The 'eustatic' sea-level rise was assumed to be 75.6 m during the last 16 000 a BP, and ocean water volume was presumed to be constant since 5000 a BP. According to this model, most continental coasts, including that of Brazil, rose because of crustal tilting caused by loading of the ocean basins by the addition of meltwater. The sea-level curve of Clark et al. for northeastern Brazil (Recife area, diagram (b) in Fig. 1) predicted 3 m of land rise since 6000 a BP.

Later glacio-hydro-isostatic models have been refined in the light of new facts including additional sea-level data. In the first version of his model, Peltier (1982) predicted raised beaches in the Recife area up to 2 m above present sea-level elevation from about 8000 a BP and reaching their maximum elevation about 7000 a BP. In his second version, Peltier (1988) used three different melting chronologies: (a) melting chronology of ICE-2; (b) a 5000 year delay of the Antarctic component of melting; (c) a 7000 year delay of the Antarctic component of melting. These three hypotheses predicted unimodal sea-level curves with (a) 3 m (b) 1.6 m and (c) less than 0.5 m sea-level rise at about 6750 a BP for the Recife area (diagram (c) in Fig. 1). More recently, the refined model by Peltier & Jiang (1996) indicates that northeastern Brazil experienced a sea-level rise of 4.6–4.8 mm/year at 20000 cal. a BP, which increased to about 10 mm/yr at 10000 cal. a BP, and eventually decreased to 0.1–0.2 mm/year at 5000 cal. a BP. A curve by W. P. Peltier (pers. comm., 1997) for the Touros area is presented in diagram (d) in Fig. 1 and will hereafter be called the 'Touros curve'.

Holocene coastal deposits in the Macau to Cunhau littoral zone

Beachrock

Beachrock is a sedimentary rock commonly formed in the intertidal zone although it can also develop in the sublittoral zone. The mineralogy of beachrock may vary from pure silica sands to biogenic carbonate sands, whereas beachrock cement may vary from aragonite to Mg-calcite (Stoddart & Cann 1965; Alexandersson 1972). More recent reviews of beachrock cement have been given by Inden & Moore (1983), Beier (1985), Hopley (1986), Amieux et al. (1989) and Strasser et al. (1989). Beachrock is found mainly in tropical regions but occurrences up to 45°N and 30°S latitude have also been reported by Alexandersson (1972) and Ramsay (1995), respectively.

Fig. 3. Oblique aerial view of the Guaraira beachrock (foreground) and Barreta beachrock (background).
The location of the beachrock bodies investigated during this study is shown in Fig. 2. They display a great range of dimensions and shapes. In the majority of cases, they are present as elongated bodies, ranging in length from some kilometres (e.g. Perobas, Barreta, Guaraíra, Cunhau, Jacuí, Farol de Sto. Alberto, Fig. 3) to some dozens of metres (e.g. Via Costeira, Guajiru, Galinhos). Beachrock also occurs in patches (e.g. Pedra Grande, Recuado, Macau, Lagoa do Sal, Fig. 4). The beachrock bodies range in width from 50 cm to 3 m and present tabular sets whose boundaries are generally erosive. They usually show gentle seaward-dipping bedding surfaces (<10°) and are parallel to the present coastline.

The sedimentary structure, set geometry, petrographic characteristics and fossil content of the deposits were compared with modern beaches to identify their palaeogeography. In the study area, beaches are mainly characterized by swash cross-stratification in the middle to lower foreshore and trough cross-stratification in the lower foreshore to upper shoreface (Fig. 5). Some of the beaches are backed by cliffs in the Barreiras Formation (Late Tertiary), whereas others are gentle strandline shores with recent sand dunes to the rear.

Other mesotidal coasts on which beachrock is developed present similar physical processes and structures. Inden & Moore (1983) concluded that small to medium tabular festoon cross-beds are very common in the shoreface zone, whereas the foreshore is characterized, from bottom to top, by fine-grade beds, cross-bedding and parallel bedding. Reading & Collison (1996) stated that the beach progradation sequence comprises cross-lamination interbedded with and eventually passing to cross-bedding, capped by parallel bedding in the foreshore zone.

Local descriptions of mesotidal beaches agree with these general models. Dabrio (1982) described the sedimentary structures typical of mesotidal beaches in southern Spain as cross-laminated sand and cross-bedded sands overlain by cross-bedding, and finally by parallel-bedded sands. Semeniuk & Johnson (1982) showed that in Western Australia the shoreface is characterized by trough-bedded sand and gravel, whereas the foreshore is distinguished by parallel-bedded sand and laminated-bubble sand. All these studies agree that physical processes do not change sharply at low water but extend to the shoreface zone. Despite that, the transition between the lower foreshore and the upper shoreface is marked by the accumulation of the coarsest available grain size, which is therefore a sea-level indicator (Bourgeois 1980; Inden & Moore 1983; Shipp 1984; Dupré 1984).

Two beachrock facies were identified in the present study on the basis mainly of sedimentological features and comparison with modern beaches. They are hereafter informally called facies (a) and facies (b). Beachrock facies (a) represents the lower foreshore and the upper shoreface zones. Facies (a) is a medium to coarse, sometimes conglomeratic, sandstone. Its terrigenous constituents are quartz, limonite, fragments of marine shell and fragments of the underlying rocks. It presents a great variety of textural types and mineralogical maturity indicating different formative processes and sources. The most common sedimentary structures are trough cross-stratifications 0.2–1.5 m in thickness, interpreted as the result of migration of sinuous crested bedforms (Figs 5 & 6), which points to an important traction flow mechanism during transportation. Palaeocurrents show transport predominantly to the NNW on the N-S-trending coast as far as Touros city, where the coast bends to the W. The palaeocurrents shift predominantly to the W and to the NW on the E-W-trending coast. Both palaeocurrent patterns are similar to those observed on the present coastline, which are mainly influenced by longshore currents. Although the shoreface zone can extend from the low-tide level to the fairweather wave base (Reading & Collison 1996), the low-water level can in places be identified and used as a sea-level indicator with a precision of ±0.5 m. The low-water level is characterized by the coarsest texture associated with trough cross-stratification, sometimes capped by facies (b) (Fig. 6). Facies (a) corresponds to the lower foreshore beachrock of Flexor & Martin (1979) and the upper foreshore to lower shoreface beachrock of Oliveira et al. (1990) described in the Salvador and Cunhau–Natal littoral zones, respectively (Fig. 1).
Beachrock facies (b) corresponds to sediments deposited on the foreshore. They are usually medium to coarse sandstones, which form tabular beds and sheets from 0.1 to 1.0 m in thickness. Facies (b) is chiefly composed of quartz grains, heavy minerals (ilmenite, magnetite, zircon, tourmaline, staurolite, and rutile), and fragments of marine shells. The grain size increases on the lower foreshore, where the low-water level can sometimes be identified (Fig. 6). The most common sedimentary structure of this facies is seaward dipping, swash cross-stratification (parallel bedding of previous studies), which is a sea-level indicator of the middle to lower foreshore with a precision of ±1.0 m. Other common structures on the foreshore are ripple marks, subcritically climbing translate strata, aeolian waves, and thin deflation pavements.

Important palaeoecological implications may be drawn from correlation between beachrock fossils and modern beach fauna. There is no quantitative or qualitative difference between the fossil
content of beachrock and death assemblages of marine shells which occur in modern beaches in northeastern Brazil (Maury 1934; Campos e Silva et al. 1964; Mendonça 1966). Death assemblages of shells tend to concentrate in the lower foreshore on modern beaches. In beachrock, they are similarly more abundant in the coarsest part of facies (a). The most common species in order of abundance are Donax striata, Divaricella quadrisculcata, Tivela mactroides, Anomalocardia brasiliana, Anadara ovalis, and Ostrea sp.

Our observations of beachrock diagenesis both in the E–W- and N–S-trending coasts match results presented by Oliveira et al. (1990) for the N–S-trending coast. The complete diagenetic sequence observed in facies (a) and in facies (b) indicates five different stages of coastal emergence or submergence. The primary cement consists mainly of acicular crusts of aragonite (Fig. 7a and b) formed in the marine phreatic zone (e.g. Moore 1971; Tietz & Müller 1971). It has been replaced locally by late micrite grains of calcite which formed in the meteoric phreatic environment and indicates coastal emergence (Fig. 7c and d) (e.g. Beier 1985; Amieux et al. 1989). The third diagenetic phase is characterized by porosity reduction because of the growth of aragonite and Mg-calcite cement (Fig. 7c and d), which implies submergence (e.g. Tietz & Müller 1971; Meyers 1986; Amieux et al. 1989; Strasser et al. 1989). The fourth phase is marked by the overgrowth of crystalline calcite (Fig. 7f) from Mg-calcite in the meteoric phreatic zone, indicating another phase of coastal emergence (see Meyers 1986). The last diagenetic phase is characterized by cement dissolution and porosity growth still in the meteoric phreatic zone (Fig. 7e and f).

**Other coastal deposits**

Other coastal deposits were used as complementary evidence of sea-level changes. Coral reefs overlie the Barreiras Formation bedrock. The Pirangi and Jacumã coral-reef banks (Fig. 2) form lines parallel to the current shoreline. They rise from the sea floor to a height of 5 m, 1 m of which is exposed at low tide. Exposure to sunlight and dry conditions during low tides may have killed the coral reefs. Most of the reef mass exposed in both of the exposures that were sampled is made up of dead organisms, so that original growth position may account for only a small part of the whole bank and represent the minimum low-water level at the time of their death. A detailed investigation of the coral-reef fauna of the study area was carried out by Kempf & Laborel (1967). They concluded that the most common species are *Mussismilia* sp., *Siderastrea stellata*, *Millepora alcicornis*, *Dendropoma* sp. and the sessile foraminifer *Homo-tree*, which is very common in open spaces.

Raised tidal flats are composed of shell-rich layers 10–30 cm in width and are associated with sand and mud. These tidal flats are located to the south of the Galinhos beach and are partly covered by vegetation. The area is protected from direct wave action by a large spit to the north. These tidal flats represent mean sea level with a precision of ±1.0 m. Their fossil content is very similar to that described for the Holocene beachrocks. Peats of mangrove-swamp origin
Fig. 7. Beachrock cement phases: (a) acicular crusts of aragonite (plane-polarized light); (b) acicular crusts of (cross-polarized light); (c) micrite calcite cement (cross-polarized light); (d) micrite calcite cement (plane-polarized light); (e) cement dissolution and secondary porosity growth (plane-polarized light); (f) secondary porosity growth over crystalline calcite and a cicular aragonite (cross-polarized light). Q, quartz; Ac, acicular calcite; M, micrite calcite; Cc, crystalline calcite; Sp, secondary porosity.

display a minimum thickness of 1.5 m and are composed mainly of wood fragments (tree branches) near the top and mud near the bottom of the deposit (Fig. 8). The upper part of the peats was deposited in the middle to upper foreshore, and the bottom part in the middle to lower foreshore (Martin et al. 1996b). Shells in living position are found mostly in the latter (Fig. 9), which can be used as sea-level indicator with a precision of ±1.0 m. A marine fauna in growth position occurs to the east of São Bento (Fig. 2). Its position indicates the minimum height of low-water level before its death.

Sample collection, pre-treatment and radiocarbon dating

A group of 25 samples of fossils were selected for dating: 17 samples of shells in beachrock, mostly in facies (a); four samples of shells in peat and tidal flat deposits; three whole-rock samples in coral reef; and one sample of live shell (Table 1). Shells in living conditions were found in a few sites (Catanduba, Rio-do-Fogo, and Recuado). Other shell samples correspond to death assemblages but, as Richards (1982) has shown, wave-deposited fauna can be a better indicator of
former waterlines than molluscs in growth position, especially if these can tolerate a wide range of depths. Whole-rock samples in beachrock were avoided because they are usually composed of fragments of other rocks, contain fossils of various ages, and may have passed through several stages of recrystallization, overgrowth, neomorphism, and dissolution.

Altitude measurements of the various coastal features, including samples, were determined by levelling and were combined with horizontal positioning by global positioning system (GPS). Correction to local standard ports (Macau and Natal) followed procedures recommended by the Admiralty (1996) using tide-table predictions by the Brazilian Navy (1996). The zero level to which all measurements and corrections were made was the Brazilian ‘Corrego Alegre’ National datum.

Radiocarbon dating by first-order assay was carried out at University College London according to the method developed by Vita-Finzi (1983, 1991). Careful pre-treatment procedures were carried out to avoid or reduce contamination. Mechanical cleaning and acid leaching were the main operations used to remove contamination. These techniques were monitored by X-ray diffraction and inspection of acetate peels by light microscopy (Table 1). Samples showing signs of contamination were also analysed by scanning electron microscopy to identify primary and secondary calcite and aragonite (Fig. 10). Samples with clear signs of diagenetic alteration were rejected or subjected to further mechanical cleaning and acid leaching.

The errors cited for the ages determined at University College London are based on sample activity, background and modern standard; the
Table 1. List of $^{14}C$ ages used in this study

<table>
<thead>
<tr>
<th>Sample source</th>
<th>Height (m asl)</th>
<th>Lab number</th>
<th>Deposit</th>
<th>Species</th>
<th>X-ray</th>
<th>$^{14}C$ age (a BP)</th>
<th>Calibrated age (a BP at 2$\sigma$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>PG/(1)</td>
<td>+0.60</td>
<td>UCL-423</td>
<td>b</td>
<td>*</td>
<td>A</td>
<td>2700 ± 80</td>
<td>2350$^{+380}_{-160}$</td>
</tr>
<tr>
<td>MC1/(1)</td>
<td>+1.80</td>
<td>UCL-354</td>
<td>b</td>
<td>*</td>
<td>A</td>
<td>1600 ± 40</td>
<td>1151$^{+150}_{-100}$</td>
</tr>
<tr>
<td>MC2/(1)</td>
<td>+1.80</td>
<td>UCL-418</td>
<td>(live)</td>
<td>*</td>
<td>-</td>
<td>Modern</td>
<td>-</td>
</tr>
<tr>
<td>CDB1/(1)</td>
<td>+0.30</td>
<td>UCL-433</td>
<td>tf</td>
<td>*</td>
<td>A</td>
<td>3950 ± 110</td>
<td>3930$^{+320}_{-290}$</td>
</tr>
<tr>
<td>CDB2/(1)</td>
<td>+0.30</td>
<td>UCL-434</td>
<td>tf</td>
<td>*</td>
<td>C</td>
<td>4500 ± 130</td>
<td>4710$^{+280}_{-340}$</td>
</tr>
<tr>
<td>GA/(1)</td>
<td>+1.10</td>
<td>UCL-416</td>
<td>b</td>
<td>*</td>
<td>A</td>
<td>3550 ± 100</td>
<td>3430$^{+240}_{-220}$</td>
</tr>
<tr>
<td>FSA1/(1)</td>
<td>-0.50</td>
<td>UCL-410</td>
<td>b</td>
<td>*</td>
<td>A</td>
<td>6050 ± 190</td>
<td>6460$^{+390}_{-290}$</td>
</tr>
<tr>
<td>REC1/(1)</td>
<td>+3.90</td>
<td>UCL-397</td>
<td>b</td>
<td>*</td>
<td>A</td>
<td>5100 ± 140</td>
<td>5450$^{+390}_{-390}$</td>
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<tr>
<td>REC2/(1)</td>
<td>+5.40</td>
<td>UCL-393</td>
<td>mt</td>
<td>*</td>
<td>A/tr sC</td>
<td>4050 ± 110</td>
<td>4070$^{+280}_{-280}$</td>
</tr>
<tr>
<td>GU/(1)</td>
<td>+0.60</td>
<td>UCL-431</td>
<td>b</td>
<td>*</td>
<td>A</td>
<td>3050 ± 90</td>
<td>2759$^{+230}_{-110}$</td>
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<tr>
<td>SAL/(1)</td>
<td>+1.20</td>
<td>UCL-417</td>
<td>b</td>
<td>*</td>
<td>A</td>
<td>3950 ± 110</td>
<td>3930$^{+320}_{-290}$</td>
</tr>
<tr>
<td>PB/(1)</td>
<td>+0.10</td>
<td>UCL-420</td>
<td>b</td>
<td>*</td>
<td>A</td>
<td>4500 ± 130</td>
<td>4710$^{+280}_{-350}$</td>
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<tr>
<td>RF/(1)</td>
<td>-0.20</td>
<td>UCL-409</td>
<td>pd</td>
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<td>A</td>
<td>3750 ± 110</td>
<td>3670$^{+270}_{-270}$</td>
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<tr>
<td>PJC/(1)</td>
<td>-0.40</td>
<td>UCL-424</td>
<td>cr</td>
<td>†</td>
<td>-</td>
<td>1450 ± 40</td>
<td>970$^{+60}_{-60}$</td>
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<tr>
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<td>UCL-413</td>
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<td>*</td>
<td>A</td>
<td>4950 ± 150</td>
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<td>VC/(1)</td>
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<td>UCL-430</td>
<td>b</td>
<td>*</td>
<td>A</td>
<td>6300 ± 200</td>
<td>6750$^{+450}_{-450}$</td>
</tr>
<tr>
<td>PR2/(1)</td>
<td>+0.10</td>
<td>UCL-425</td>
<td>cr</td>
<td>†</td>
<td>-</td>
<td>1150 ± 30</td>
<td>680$^{+40}_{-40}$</td>
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<tr>
<td>PR3/(1)</td>
<td>-0.50</td>
<td>UCL-361</td>
<td>cr</td>
<td>†</td>
<td>-</td>
<td>950 ± 30</td>
<td>530$^{+30}_{-30}$</td>
</tr>
<tr>
<td>BR1/(1)</td>
<td>+2.20</td>
<td>UCL-403</td>
<td>b</td>
<td>*</td>
<td>A</td>
<td>4700 ± 140</td>
<td>4880$^{+420}_{-320}$</td>
</tr>
<tr>
<td>BR2/(1)</td>
<td>+1.80</td>
<td>UCL-404</td>
<td>b</td>
<td>*</td>
<td>A/tr sC</td>
<td>4500 ± 120</td>
<td>4710$^{+260}_{-260}$</td>
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<tr>
<td>GR1/(1)</td>
<td>+0.20</td>
<td>UCL-419</td>
<td>b</td>
<td>*</td>
<td>A</td>
<td>5600 ± 170</td>
<td>5970$^{+380}_{-350}$</td>
</tr>
<tr>
<td>GR2/(1)</td>
<td>0.00</td>
<td>UCL-421</td>
<td>b</td>
<td>*</td>
<td>A</td>
<td>6550 ± 210</td>
<td>7070$^{+360}_{-360}$</td>
</tr>
<tr>
<td>GR3/(1)</td>
<td>+0.70</td>
<td>UCL-405</td>
<td>b</td>
<td>*</td>
<td>A</td>
<td>5950 ± 170</td>
<td>6370$^{+370}_{-390}$</td>
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<tr>
<td>CH2/(1)</td>
<td>+1.70</td>
<td>UCL-432</td>
<td>b</td>
<td>*</td>
<td>A/sC</td>
<td>5400 ± 170</td>
<td>5750$^{+420}_{-330}$</td>
</tr>
<tr>
<td>CH1/(1)</td>
<td>+1.50</td>
<td>UCL-414</td>
<td>b</td>
<td>*</td>
<td>A</td>
<td>6550 ± 210</td>
<td>7070$^{+360}_{-360}$</td>
</tr>
<tr>
<td>P20/(2)</td>
<td>+1.0-2.0</td>
<td>-</td>
<td>b</td>
<td>*</td>
<td>-</td>
<td>6067 ± 80</td>
<td>6490$^{+320}_{-190}$</td>
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<tr>
<td>P21/(2)</td>
<td>+1.0-2.0</td>
<td>-</td>
<td>b</td>
<td>*</td>
<td>-</td>
<td>6067 ± 100</td>
<td>6370$^{+230}_{-200}$</td>
</tr>
<tr>
<td>P14/(2)</td>
<td>+3.00</td>
<td>-</td>
<td>b</td>
<td>*</td>
<td>-</td>
<td>4737 ± 130</td>
<td>4970$^{+340}_{-330}$</td>
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<tr>
<td>P29/(2)</td>
<td>+3.00</td>
<td>-</td>
<td>b</td>
<td>*</td>
<td>-</td>
<td>4609 ± 100</td>
<td>4830$^{+290}_{-280}$</td>
</tr>
<tr>
<td>BL/(3)</td>
<td>+1.0(?)</td>
<td>-</td>
<td>mfs</td>
<td>cement</td>
<td>-</td>
<td>6060 ± 80</td>
<td>6480$^{+200}_{-180}$</td>
</tr>
<tr>
<td>S-6/(3)</td>
<td>-3.0 to +4.0</td>
<td>-</td>
<td>mfs</td>
<td>shell</td>
<td>-</td>
<td>2340 ± 60</td>
<td>2350$^{+140}_{-140}$</td>
</tr>
</tbody>
</table>

Sample source: (1) this study; (2) Oliveira et al. (1990); (3) Silva (1991). Coastal deposit code: b, beachrock; cr, coral reef; pd, peat deposit; mt, marine terrace; td, tidal flat; mfs, marine and fluvial sediments. Species: * fossil assemblage on text; † coral-reef content on text. X-ray code: A, primary aragonite; C, primary calcite; sC, secondary calcite; (tr) mineral in trace quantity.

$^{14}C$ ages were calibrated using the curve of Stuiver et al. (1986), and rounded up or down to the nearest multiple of ten. The activity determined for a specimen collected live (MC2), 1 cpm above the pre-bomb background level of 7.7 cpm, shows that the residual effect of the atmospheric bomb tests swamps any apparent reservoir effect within the resolution of the method (C. Vita-Finzi, pers. comm.). Further ages determined by conventional radiometric dating in previous studies (Oliveira et al. 1990; Silva 1991), are also presented in Fig. 2 and Table 1 but were not included in the time-depth diagram because of the lack of information on sample quality.
Field and chronological correlation of beachrock

Some of the coastal deposits which give overlapping ages can also be identified with a single palaeo-shorelines on the basis of field evidence. In the Cunhãu to Natal littoral zone, the Cunhãu, Guaraira and Via-Costeira beachrock bodies range in height from 0.5 m below mean sea level to 2.0 m above mean sea level and present sedimentary sections similar to the one depicted in Fig. 6b. There is good agreement between their ages (VC, GR1–GR3 and CH1) and those obtained by Oliveira et al. (1990) for the Guaraira beachrock (samples P20 and P21). If the error bounds are extended to 2σ, ages VC, GR1 and GR3 overlap at 6300–6350 cal. a BP, which is likely to be the real age of this group. Samples GR2 and CH1 are 200 years older than the others and are perhaps redeposited death assemblages. It is possible to extend the unit to include the Farol de Santo Alberto beachrock on the E–W-trending coast (Fig. 2) on the basis of age and field character. Chronological correlation is also possible between the Recuado, Barreta and Jacumã beachrock bodies, whose ages overlap at 5060–5300 cal. a BP at 2σ. Another possible chronological association is found between the Galinhos and Lagoa-do-Sal beachrock bodies and the Rio-do-Fogo peat, whose ages overlap at 3640–3670 cal. a BP.

Age correlation between other samples is less evident. A few coastal deposits present discordant 14C ages, which might reflect several contributing factors. The age difference between CDB1 and CDB2 in the Catanduba tidal flat is likely to reflect the varied provenance of the detrital material. The age of the upper part of the Guaraira beachrock, where sample GR1 was collected, suggests that cementation processes continued for thousands of years. Despite these anomalies, it was possible to establish a local sea-level chronology and to compare it with predicted changes.

Comparisons between coastal chronology and glacio-hydroisostatic predictions: implications for tectonics

The sea levels indicated by our findings, with the exception of beachrock facies (b) and tidal flats, do not represent the whole intertidal zone but reflect shells in the lower foreshore and upper shoreface transition zone during Holocene time. From Fig. 11 it can be seen that the great majority of Holocene coastal deposits in the study area are elevated relative to the present sea level. Sea-level
behaviour, as indicated by beachrock data and other coastal deposits, points to two main phases of sea-level change. The first was a transgression which started at c. 7000 cal. a BP and reached its maximum at c. 5500–5000 cal. a BP. It was followed by a regressive phase which lasted until the present day. Samples of the same age and subenvironment but at different locations (e.g. BR1 and PB) plot at different heights, indicating sea-level oscillations. These sea-level oscillations are confirmed by the relationship between facies (a) and (b) and by different beachrock-cement phases.

Our results also vindicate the data presented by Silva (1991) for the littoral zone near Macau, where a transgressive sequence starting at 7460 (+190/−160) cal. a BP and reaching a maximum at 5330 (+290/−310) cal. a BP is overlain by a regressive sequence of sediments. The data, which predominantly bear on low-water level, plot slightly below or on the Touros curve by W. R. Peltier (pers. comm., 1997), which represents mean sea level. The Touros curve correctly predicts a Holocene highstand in the study area at c. 5500–5000 cal. a BP. Some sea-level oscillations of less than 2 m are also indicated by samples on and below the Touros curve (see Fig. 11).

Some published Holocene coastal features are consistent with the relative sea-level fall predicted by the Touros curve, which is possibly indicated by the absence of any age higher than 3 m after 4000 cal. a BP. At least three generations of an extensive Holocene aeolian sand cover (Perrin & Costa 1982) occur in the current backshore zone of the study area. It is related to a relative sea-level fall in Holocene time which caused dune building by raising intertidal sandflats above the mean high tide level. In the Macau area, Silva (1991) presented an age of 2350 (+350/−140) cal. a BP (sample S-6) for deposits formed by the migration of sands from the shallow platform to the continent, presumably during the main regression phase.

The values which do not fit the Touros curve by W. R. Peltier (pers. comm., 1997) may be the product of low-magnitude climatic oscillations, as proposed by Suguio et al. (1985) and Fairbridge (1992). Nevertheless, a recent study by Angulo & Lessa (1997) favours Holocene sea-level curves with few or no oscillations along the Brazilian coast. It may also be that minor changes in wind or currents can account for the anomalies (Damuth & Fairbridge 1970; Isla 1989; Gonzales & Weiler 1994).

In some areas, however, a tectonic explanation is more appropriate. On the littoral zone near São Bento (Fig. 2), coastal emergence to the east of the Carnaubais fault system is matched
by submergence to the west. In the Macau High, the Galinhos beachrock (facies (a)) plots slightly below the Touros curve (mean sea level), indicating no uplift (Fig. 12). Similarly, near the Umbuzeiro Graben, which is directly affected by the Carnaubais fault system, the Farol de Sto. Alberto beachrock and the Catanduba tidal flat do not display any emergence (Fig. 13), in contrast to coastal deposits younger than 4000 cal. a BP east of Sao Bento, where sudden coastal emergence is indicated by a rich marine bivalve fauna in growth position (REC2), dated to 4240–3910 cal. a BP. This unit is continuous along more than 5 km of the littoral zone to the east of the Carnaubais fault system but not to the west of it (Fig. 13). It lies 5.5 m above mean sea level, 1.5 m above the Recuado beachrock dated to 5600–5290 cal. a BP, and 5 m above the Guajiru beachrock dated to 2910–2740 cal. a BP (Fig. 6a). From Fig. 11, it can be seen that local emergence by at least 5 m indicated by sample REC2 and GU, occurred during a short period of about 1320 years. Such local emergence points to tectonic movement and is in contrast to the much smoother sea-level regression that occurred after 5500–5000 cal. a BP.

Additional field evidence for tectonic movement is found at Sao Bento, which caused coastal emergence to the east and submergence to the west of the Carnaubais fault. Tectonic emergence and submergence is in agreement with geoelectric soundings across the Carnaubais fault system to the southwest of Sao Bento (Fig. 12) by Caldas et al. (1997), which showed a sequence extending from Neocomian to Holocene time and 120 m thick in the Umbuzeiro Graben. The base of the sequence displays a vertical offset of 60 m along the Carnaubais fault system which gradually dies out upwards. Holocene activity is consistent with the coseismic faulting described by Takeya et al. (1989) in the area (Fig. 2).

Conclusion

In this attempt to analyse sea-level changes in the Holocene sedimentary record of NE Brazil, two beachrock facies associated with (a) the lower foreshore to upper shoreface and (b) the middle to lower foreshore were recognized, and beachrock was found to be a useful sea-level indicator in a mesotidal regime. Our sea-level data were found to fit the Touros curve proposed by W. R. Peltier (pers. comm., 1997) and confirmed the presence of the Holocene palaeo-shorelines predicted by him. In addition, however, sea-level oscillations superimposed on the Touros curve might be the result of minor climatic changes or wind–current changes in Holocene time. Tectonic influences can be detected locally, notably near the Carnaubais fault system, where rapid emergence of at least 5 m to the east of Sao Bento occurred at c. 4080–2780 cal. a BP. The significance of such intraplate activity for seismic hazard is clear and needs to be investigated further.

We are greatly indebted to C. Vita-Finzi, I. Stewart, R. W. Fairbridge and K. Suguio for useful comments and suggestions. We also thank W. R. Peltier for providing the Late Quaternary sea-level prediction for the study area. This research was funded mainly by University College London, and by the PADCT II-UFRN Project (No. 65.91.0366.00). We are very grateful to a CNPq Project (Vales Tectônicos do Rio Grande do Norte) co-ordinated by Allaoua Saadi, which contributed with financial support for the last field trip.

References


COASTAL TECTONICS IN NE BRAZIL


WALCOTT, R. I. 1972. Past sea levels, eustasy and deformation of the Earth. Quaternary Research, 2, 1–14
Electrical sounding (resistivity) data from Queiroz et al. (1985) presented in Cross-section A-A’ (Fig. 5.3)

- Curve A - Second log from A to A’
- Curve C - Fourth log from A to A’
- Curve D - Third log from A to A’
- Curve E - Second log from A to A’
Cross-section B-B' (Fig. 5.3)
First well from B to B'

Argila pouco arenosa, voceaiada.
Arenito fino a gros. arg. col. crema.
Arenito fino a conglomerático argiloso col. crema.
Arenito fino com horizonte argiloso col. esbranquiçada.
Arenito fino siltoso col. esbranquiçado.
Arenito fino a médio col. esbranqu.arenito fino a médio col. esbran.
Arenito fino a médio col. esbran.
Arenito fino siltoso col. esbran.
Calcário dolomítico col. branca.

Calcário arenoso col. cinza.

Arenito fino a médio calcífero, col. esbranquiçado.
Cross-section B-B' (Fig. 5.3) - Second well from B to B'.
### Descrição e Litológica

<table>
<thead>
<tr>
<th>Localização</th>
<th>Perfil Litológico</th>
<th>Profundidade (m)</th>
<th>Perfil do Poço</th>
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</thead>
<tbody>
<tr>
<td>0,40</td>
<td>Sólo aren-argiloso, marron</td>
<td></td>
<td></td>
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<tr>
<td></td>
<td>Argila plástica, marron clara</td>
<td></td>
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<td></td>
<td>Silte arenoso, caulínico, branco</td>
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<tr>
<td></td>
<td>Argila plástica, creme amarelada</td>
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<tr>
<td></td>
<td>Arenito fino a grosseiro com cascalho e peq. seixos, matriz muito argilosa, amarelada</td>
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<td>AL</td>
<td>Argila plástica, marron</td>
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<td>BR</td>
<td>Argila plástica esv. c/ nódulos preto</td>
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<td>Arenito fino a grosso com cascalho e seixos, mal selecionado pouco argiloso, creme</td>
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<td>Arenito fino, calcífero, creme</td>
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</tbody>
</table>

**Observações:**

Escala: 1:200

Local: Sede do município
REVESTIMENTO

PERFIL LITO-GEOLÓGICO

Cross-section B-B' (Fig. 5.3)
Fourth well from B to B'

Arenito argiloso, col. avermelhada.

15m

Argila arenosa, col. amarelada.

24m

Argila arenosa, col. cinza.

26m

Arenito de fino a médio, col. esbranquiçada.

31m

Argila amarelada.

32m

Arenito de fino a médio, pouco argiloso.

36m

Arenito de fino a médio, col. esbranquiçada.

45m

Arenito calcífero, col. crema.

48m

ESCALA 1:400

[Signature]
[Geologist: GGEA 2197-8.08]
[Scale: 080009-90-0]
REVESTIMENTO

PERFIL LITO-GEOLOGICO

34m
Arenito de médio a grosseiro.

40m
Argila.

43m
Arenito de médio a grosseiro.

46m
Argila arenosa, col. avermelhada.

50m
Arenito.

57m
Argila.

62m
Arenito calcífero.

Gelma Teste Barros
Telma Tostes Borba
Geólogo CREA 2181-B-CE
CPF 091002023-04

ESCALA 1:500
**PERFIL LITOLÓGICO E CONSTRUTIVO**

**POÇO Nº: 2**  
**LOCAL:** Conjunto Soledade  
**MUNICÍPIO:** Natal

### DESCRIÇÃO LITOLÓGICA

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<th>Perfíl Litológico</th>
<th>Profundidade</th>
<th>Perfíl do Poço</th>
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<td>Silt argiloso, roxo claro a esbranquiçado</td>
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<tr>
<td>Silt argiloso crese</td>
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<tr>
<td>Arenito fino a grosso, esbranquiçado</td>
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<td>Silt argiloso, pouco arenoso acinzentado</td>
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<tr>
<td>Silt argiloso, com pouco calcário</td>
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<td></td>
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<tr>
<td>Cultura crese-argiloso, fino, poroso, cinza</td>
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</tbody>
</table>

**OBSERVAÇÕES:** Perfil litológico e construtivo

**LEGENDA**

- **CIMENTAÇÃO**
Cross-section G-G' (Fig. 5.3)
Third well from G to G'

PERFIL LITÓLÓGICO

<table>
<thead>
<tr>
<th>DESCRICAO LITOLÓGICA</th>
<th>PERFIL</th>
<th>PROFUNDIDADE (m)</th>
<th>PERFIL DO POÇO</th>
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<tbody>
<tr>
<td>EIA FINA MARRON, C/GRS MÉD.</td>
<td>0.00</td>
<td>+ 0.35</td>
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<td>EIA FINA/MÉD. CREME, PÇO ARG.</td>
<td>0.00</td>
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<td>N. FINO/GRS., AVERM. COMPAC. ARG.</td>
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<tr>
<td>N. SILTICO-ARG., FİNO/GRS. CREME</td>
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<td>- 22.00</td>
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<tr>
<td>G. COMP. VARIEGADAS</td>
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<td>- 60.00</td>
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<td>N. FİNO ESBRANQ. PÇO ARG. C/GRS MÉD.</td>
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<tr>
<td>N. FİNO/MÉD. CREME ESBRANQ. ARG. C/ARN. CARB. NA BASE</td>
<td>0.00</td>
<td>- 83.00</td>
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MÉD/GRS. CREME, PÇO ARG.
N. FİNO/ED. C/ESBRANQ. PÇO ARG.
N. FİNO/GRS. C/GRS MÉD.