Chapter 2. Geological, oceanographical and climatic setting of the study area

2.1. Geological setting of the Azores archipelago

The Azores archipelago is located in the middle of the North Atlantic Ocean between the latitudes 37° N and 40° N and the longitudes 25° W and 31° W (Figure 17). It is the result of the volcanic activity associated with the triple junction where the American, Eurasian and African lithospheric plates meet. The Mid-Atlantic ridge (MAR) separates the American from the Eurasian and Africa plates. To the east of the ridge, the Azores-Gibraltar fracture zone marks the boundary between Eurasia and Africa (Krause and Watkins, 1970; Laughton and Whitmarsh, 1974).

The archipelago comprises nine islands, the Formigas islets and some submarine volcanoes. The islands emerge from the Azores volcanic plateau (Figure 18), which is a first-order morphological feature in the Atlantic basin. It has an overall triangular shape corresponding to a surface area of approximately 400 000 km² of elevated oceanic crust, roughly underlined by the 2000 m isobath (Lourenço et al., 1998). The plateau is mainly constructed of alkaline basalt volcanism. Geochemistry and petrology suggest a hotspot origin for this volcanism (White et al., 1976).

The plateau crosses to the west the Mid-Atlantic Ridge (MAR) and is limited to the south by the East Azores fracture Zone (EAFZ). The western group of the islands (Flores and Corvo) lie on the American plate, while the islands of the central (Terceira, Graciosa, S. Jorge, Pico and Faial) and eastern group (S. Miguel, Santa Maria and Formigas) follow a complex lineation that trends WNW–ESE from the MAR to the western limit of the Gloria Fault (Argus et al., 1989) The precise location of the western segment of the Eurasia-Africa plate boundary is still controversial. During the last three decades, several geodynamic models have
been proposed for this area relying on different approaches. Some authors have interpreted the Azores domain as a normal spreading centre (Krause and Watkins, 1970), whereas others believe it is an oblique spreading centre (McKenzie, 1972; Searle, 1980), and others support a leaky transform model (Madeira and Ribeiro, 1990).

Figure 17 – Geographical and tectonic setting of the Eurasia-Africa-North America plate boundary (modified from Argus et al., 1989). AM=American plate; AF= African plate; AGFZ= Azores Gibraltar Fracture Zone; EU= Eurasian plate; MAR= Mid-Atlantic ridge. Bathymetry of the area Azores-Gibraltar from GEBCO (IOC IHO and BODC, 2003).

Recently, various works, based on bathymetric (Lourenço et al., 1998), gravity (Luis et al., 1998), and seismic (Miranda et al., 1998) data have proposed a new model for the Azores Plateau region (Figure 19). These authors state that, presently, this region is a narrow diffuse plate boundary consisting of several tectonic blocks limited by two sets of conjugated faults striking 120° and 150°. These faults established the framework for the onset of volcanism, expressing as linear volcanic ridges or as point source volcanism. This area acts simultaneously
Chapter 2. Geological, oceanographical and climatic setting of the study area

Figure 18 – Tectonic setting of the Azores archipelago (MAR modified from Luis et al., 1994; SJLT modified from Vogt and Jung, 2004). AFZ= Açor Fracture Zone; EAFZ= East Azores Fracture Zone; FFZ= Faial Fracture Zone; GF= Gloria Fault; MAR= Mid-Atlantic ridge; NAFZ= North Azores Fracture Zone; PAFZ= Princesa Alice Fracture Zone; TR= Terceira Rift. Bathymetry of the Azores archipelago from GEBCO (IOC IHO and BODC, 2003).

as an oblique ultra slow spreading center and as a transfer zone that accommodates the differential shear movement between the Eurasian and African plates from the MAR until the western tip of the Gloria fault.

Recent geodetic measurements in the central and eastern groups of the Azores Islands (Bastos et al., 1998; Pagarete et al., 1998) confirm this strain regime in the Azores triple junction. More recently, Fernandes et al. (2004; 2006) using new geodetic data revealed that Faial, Pico, S. Jorge, Terceira and S. Miguel Islands are clearly in the deformation zone whilst Graciosa Island belongs to the stable Eurasian plate and Santa Maria Island to the African plate.

2.1.1. Seismicity in the Azores area

Due to its location on an active plate boundary, the Azores archipelago is subjected to frequent seismic activity. Although most activity consists of low to
Figure 19 - Schematic stress pattern of the Azores plateau as inferred from the morphological features. Dots represent the boundaries between the different Linear Volcanic Ridge domains. Thicker lines represent maximum compressive stress orientation ($\sigma_1$), thinner lines represent the minimum compressive stress orientation ($\sigma_3$). The intermediate compressive stress ($\sigma_2$) is vertical. Insert top are the T axis calculated from events displayed in Figure 20 in Lourenço et al. (1998) along with calculated standard deviations. The kinematic orientation of the spreading axis calculated from Nuvel-1 model (DeMets et al., 1990) in the three individualized areas is also shown for reference. Insert bottom is the schematic regional tectonic model for the Azores domain. (Lourenço et al., 1998).

moderate magnitude seismic isolated events or swarms, the islands are occasionally struck by high magnitude (M~7) earthquakes (Madeira and Brum da Silveira, 2003). The distribution of instrumental epicenters (Figure 20) shows that the present plate boundary is oriented WNW-ESE along the S. Jorge-S. Miguel alignment. Published focal mechanisms indicate: (a) dextral strike-slip with variable components of normal dip-slip in WNW-ESE oriented faults planes: (b) sinistral strike-slip with variable components of normal dip-slip in NNW-SSE oriented fault planes and (c) pure dip-slip normal events in both directions. This strain regime suggests decoupling in a transtensional regime. According to the position in time during the seismic cycle we can have pure dominant normal, dextral and sinistral strike-slip solutions on the WNW-ESE and NNW-SSE structures (Ribeiro, 2002).
2.1.2. The origin of volcanism in the Azores archipelago

The formation of the Azores Platform may have started 36 Ma ago (Campan et al., 1993), related to the northward migration of the junction, from the latitude of the Pico Fracture Zone and East Azores Fracture Zone (see Figure 18) to its present position at the vicinity of the North Azores Fracture Zone (Luis et al., 1994). 10 Ma ago, a melting anomaly originated within the Azores hotspot provoked enhanced magmatism and played a significant role in the construction of the Azores platform (Cannat et al., 1999), consequently originating a complex spreading history (Luis et al., 1994).

The oldest isotopic ages collected in sub aerial formations for each Azorean island are given in Figure 21. The data indicate emergence of Santa Maria during the Miocene (8.12 Ma), of Formigas, S. Miguel, Terceira, Graciosa and Flores Islands in the Pliocene (respectively 4.65, 4.01, 3.52, 2.5 and 2.15 Ma) and Faial, Corvo, S. Jorge and Pico Islands in the Quaternary (respectively 0.73, 0.70, 0.55
and 0.25 Ma).

Madeira and Ribeiro (1990) and Madeira (1998) suggest that the isotopic ages' determinations in Santa Maria may be higher than those determined by Abdel-Monem et al. (1975). Some samples may not have been collected in the oldest formations or may have been assigned to incorrect units. For the same reason, Madeira (1998) insists that in Faial, Pico and S. Jorge Islands, the isotopic ages published by Feraud et al. (1980) could be higher.

On the contrary, Johnson et al. (1998) presented new $^{40}\text{Ar}/^{39}\text{Ar}$ ages for the Nordeste complex of S. Miguel Island (between 0.78 and 0.88 Ma) which indicate a much younger volcanism than that proposed by Abdel-Monem et al. (1975) – K/Ar ages of 1 to 4 Ma. Similarly, Calvert et al. (2006) in face of new $^{40}\text{Ar}/^{39}\text{Ar}$ ages for the Cinco Picos Volcano of the Terceira Island, argues that the subaerial eruptions of this stratovolcano may have started between 0.40 and 0.50 Ma, in contrast with the age of 3.52 Ma inferred by White et al. (1976).
From what is mentioned above, it is clear that the age of the volcanism in the Azores islands is still controversial. Therefore, the assumed ages for the different volcanic regions from which the Faial Island is composed, will be used in this thesis carefully when trying to infer other geological processes (see Chapter 5).

2.1.3. Tsunami records

The Azores archipelago is highly vulnerable to tsunami hazards due to its location, active volcanism and plate-tectonics setting. First, its location on the northern mid-Atlantic makes it very vulnerable to far-sourced events namely those generated along the Iberian margin (Bryant, 2001). Furthermore, the compilations of Nunes et al. (2004) on the seismicity of the Azores show, that on average, every 70 years an earthquake with Ms exceeding 6.5 occurs. Finally, the volcanic islands are very prone to instability, with optimal conditions leading up to or contributing as potential triggers for tsunamis. De Lange et al. (2001) summarized the mechanisms associated to volcanism capable of generating large tsunamis which included the subaerial and submarine mass movements, submarine eruptions, caldera collapse, basal surges and shock waves as major sources.

Andrade et al. (2006) have built a dataset on tsunami-induced flooding of the Azores coast based on local monographs and newspapers, web citations and technical papers that cover a period from the start of the 15th century (beginning time of colonisation of the islands) till the present day. Among its major findings is the conclusion that most of the recorded tsunamis were generated by earthquakes and that the volcanic activity apparently had not generated destructive tsunamis. Since the settlement of the archipelago at least 23 tsunamis have struck the Azorean coastal zones and the highest known run-up (11–15 m) was recorded on 1 November 1755 at Terceira Island associated to the Lisbon earthquake, corresponding to an event of intensity VII–VIII (damaging–heavily damaging) on the Papadopolous–Imamura scale (Papadopoulos and Imamura, 2001). Although Andrade et al., (2006) found no evidence of volcanogenic induced tsunamis in the historical record; they agree that it would be misleading to assume that the
volcanism in the Azores had not had the power to create tsunamis. They argue that Earthquakes, submarine explosions, pyroclastic flows (núes ardentes), caldera collapse and subsidence, avalanches of cold rock, basalt surges and shock waves, avalanches of hot rock, lahars, and atmospheric phase coupling are mechanisms associated with volcanism capable of generating tsunami waves. De Lange et al. (2001) proposed that the insular shelf of Faial had not had the power to create tsunamis. They argue that these mechanisms have the potential to generate tsunami waves. The tsunami characteristics of each mechanism are as follows:

<table>
<thead>
<tr>
<th>Mechanism</th>
<th>Source volume (km³)</th>
<th>Wave height (m)</th>
<th>Period (min)</th>
<th>Travel distance (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Earthquakes</td>
<td>1–10</td>
<td>up to 17</td>
<td>10–40</td>
<td>&lt;500</td>
</tr>
<tr>
<td>Submarine explosions</td>
<td>&lt;1</td>
<td>1–6</td>
<td>1–10</td>
<td>&lt;50</td>
</tr>
<tr>
<td>Pyroclastic flows (núes ardentes)</td>
<td>1–100</td>
<td>up to 25</td>
<td>1–40</td>
<td>&lt;250</td>
</tr>
<tr>
<td>Caldera collapse and subsidence</td>
<td>1–10</td>
<td>up to 15</td>
<td>short</td>
<td>&lt;30</td>
</tr>
<tr>
<td>Avalanches of cold rock</td>
<td>&lt;1</td>
<td>1–10</td>
<td>short</td>
<td>&lt;50</td>
</tr>
<tr>
<td>Basalt surges and shock waves</td>
<td>&lt;1</td>
<td>up to 5</td>
<td>aperiodic</td>
<td>&lt;10</td>
</tr>
<tr>
<td>Avalanches of hot rock</td>
<td>&lt;1</td>
<td>small</td>
<td>short</td>
<td>&lt;10</td>
</tr>
<tr>
<td>Lahars</td>
<td>&lt;1</td>
<td>small</td>
<td>short</td>
<td>&lt;50</td>
</tr>
<tr>
<td>Atmospheric phase coupling</td>
<td>?</td>
<td>small</td>
<td>15–40</td>
<td>&gt;1000</td>
</tr>
<tr>
<td>Lava entering the sea</td>
<td>&lt;1</td>
<td>very small</td>
<td>short</td>
<td>&lt;10</td>
</tr>
</tbody>
</table>

Figure 22 – Mechanisms associated with volcanism capable of generating tsunami waves (De Lange et al., 2001).

<table>
<thead>
<tr>
<th>Date</th>
<th>Source</th>
<th>Cause</th>
<th>Origin</th>
<th>Seismic intensity / Magnitude</th>
<th>Tsunami intensity</th>
<th>Run-up height / tsunami height (m)</th>
<th>Recorded tsunami origin</th>
</tr>
</thead>
<tbody>
<tr>
<td>24.05.1614 L</td>
<td>S→Ls</td>
<td></td>
<td></td>
<td>VIII–IX</td>
<td>III–IV, IV–V</td>
<td>3</td>
<td>Ter</td>
</tr>
<tr>
<td>10.11.1755 L</td>
<td>S</td>
<td></td>
<td></td>
<td>VI</td>
<td>III–IV</td>
<td>7</td>
<td>Ter</td>
</tr>
<tr>
<td>09.07.1757 L</td>
<td>S→Ls</td>
<td></td>
<td></td>
<td>VII–VIII</td>
<td>III–IV</td>
<td>11–15</td>
<td>All islands</td>
</tr>
<tr>
<td>31.07.1761 D</td>
<td>S</td>
<td></td>
<td></td>
<td>VI</td>
<td>III–IV</td>
<td>1</td>
<td>Ter, Gra, Fay (only)</td>
</tr>
<tr>
<td>23.01.1792 L</td>
<td>S</td>
<td></td>
<td></td>
<td>VII</td>
<td>III, IV</td>
<td>1</td>
<td>Ter</td>
</tr>
<tr>
<td>08.07.1855 L</td>
<td>S</td>
<td></td>
<td></td>
<td>VII?</td>
<td>III–IV, VI–VI</td>
<td>8</td>
<td>Sjz</td>
</tr>
<tr>
<td>03.02.1899 L</td>
<td>S</td>
<td></td>
<td></td>
<td>VI–VII</td>
<td>III–IV, VI–VI</td>
<td>5–10</td>
<td>Ter</td>
</tr>
<tr>
<td>31.08.1926 L</td>
<td>S</td>
<td></td>
<td></td>
<td>V</td>
<td>III–IV, V–VI</td>
<td>10</td>
<td>Sjz</td>
</tr>
<tr>
<td>31.08.1931 L</td>
<td>S</td>
<td></td>
<td></td>
<td>V</td>
<td>III–IV, V–VI</td>
<td>10</td>
<td>Sjz</td>
</tr>
<tr>
<td>08.05.1939 D</td>
<td>S</td>
<td></td>
<td></td>
<td>V</td>
<td>III–IV, V–VI</td>
<td>8.5</td>
<td>Fay</td>
</tr>
<tr>
<td>25.11.1941 D</td>
<td>S</td>
<td></td>
<td></td>
<td>V</td>
<td>III–II</td>
<td>1</td>
<td>Fay</td>
</tr>
<tr>
<td>28.02.1969 D</td>
<td>S</td>
<td></td>
<td></td>
<td>V</td>
<td>III–II</td>
<td>0.5</td>
<td>Ter</td>
</tr>
<tr>
<td>26.05.1975 D</td>
<td>S</td>
<td></td>
<td></td>
<td>V</td>
<td>III–II</td>
<td>1</td>
<td>Sng, Ter, Fay</td>
</tr>
<tr>
<td>01.01.1980 L</td>
<td>S</td>
<td></td>
<td></td>
<td>IV–V</td>
<td>III–IV</td>
<td>0.5</td>
<td>Ter</td>
</tr>
</tbody>
</table>

Notes: Sng — S. Migué; Ter — Terceira; Gra — Graciosa; Sjz — S. Jorge; Piz — Pico; Fay — Faial; L — local; D — distal; S — seamount; Ls — landslide; Ss — submarine slide. Tsunami run-up/height — see text for explanation. Run-up height calculated by the authors is marked bold and underlined; estimates of run-up height taken from the literature or data bases and lacking confirmation by the authors are marked with an asterisk; italics refer to estimated or reported tsunami height; the remnant data on run-up was derived by comparing the available descriptions with systems that correlate tsunami impacts on natural and human environment and structures with run-up.

Figure 23 – Historical tsunami records in the Azores archipelago (Andrade et al., 2006).
that the limited instrumental and documentary data and also the small volumes released during the historical eruptions can account for that deceive. In fact, the geological record supports their assumption showing the emission of massive pyroclastic flows and caldera explosions with potential to generate high magnitude tsunamis.

2.2. Oceanographic setting of the Azores archipelago

2.2.1. Surface currents

The basic average ocean circulation pattern of the North Atlantic is an asymmetric, large scale gyre that flows to the north, on the western side, with an intense thin jet (the Gulf Stream) and to the south, on the centre/eastern side with a multibranched current system. The Gulf Stream very efficiently transports warm water of equatorial and tropical origin into the colder northern waters. The current patterns result in the high salinity, high temperature and low nutrient regime which typify the Azores. During Winter a deep mixed layer is present around 150 m and in Summer a seasonal thermocline develops around 40 to 100 m (Santos et al., 1995). The Gulf Stream is also the source of many instability processes, meanders and eddies. This picture becomes particularly complicated when this current leaves the North American coast, at about 40º to 45º N, towards the central zone of the North Atlantic where the Azores are located (Gould, 1985; Klein and Siedler, 1989).

The Gulf Stream (Figure 24) splits into two main branches at 40º N: the North Atlantic Current (NAC) and the Azores Current (AC). Each of these also divides into two further branches: NAC1 and NAC2 and AC1 and AC2 respectively. The northern part of Azores is influenced by the NAC2 and the southern one is influenced by the AC1. The general regime is from west to east (with a mean intensity of 10 cm/s at 100m water depth) but there is a clear seasonal and half seasonal oscillation of the mean direction, with periods where NAC2 dominates (current coming from northwest) and periods where AC1 is
The insular shelf of Faial: Morphological and sedimentary evolution

The AC is a quasi-permanent feature throughout the year, centred between 33° and 35° N, with a variable eastward transport (Klein and Siedler, 1989). Surface mean speeds reach 30-40 cm/s decreasing to some 5 cm/s at about 700m water depth (Ollitrault, 1995). Local winds can alter significantly the AC, as well as the localization of Azores anticyclone which can have a major influence in the current direction (Instituto Hidrográfico, 2000).

Figure 24 – Ocean currents on the area surrounding the Azores (circle locates the Azores archipelago). (A) Summer currents. (B) Winter currents. GS – Gulf Stream; NAC – North Atlantic

Figure 24 – Ocean currents on the area surrounding the Azores (circle locates the Azores archipelago). (A) Summer currents. (B) Winter currents. GS – Gulf Stream; NAC – North Atlantic
Current: AC – Azores Current; MC – Madeira Current; CC – Canaries Current; SWEC – Southwest European Current (after Santos et al., 1995).

2.2.2. Tides

The Azores archipelago is subjected to semidiurnal regular tides. In the Faial Island the annual mean tidal range is about 0.9 m and the annual mean tidal elevation is about 40 cm above the local mean sea level (MSL) (Instituto Hidrográfico, 2000). Consequently, this coast is subjected to a microtidal regime according to Hayes’s classification (1979). In the Azores archipelago, tidal currents have not been studied yet. Nevertheless, empirical observations show that these exist, but with moderate intensity both in the flood tide and the ebb tide. In general these currents are stronger during spring tides, between the flood and ebb tides, reaching the highest velocities near the headlands of the coast. During the neap tides, the currents are weak or even do not exist (Instituto Hidrográfico, 2000).

2.2.3. Waves

Carvalho (2002; 2003) characterized the sea conditions in the Azores central archipelago during the period 1989-2002, based on the open sea-wave model (MAR3G). The model is a third generation wind-wave model that uses a square grid over a stereographic polar projection, with a node distance of 300 km at 60ºN. The grid has 20 x 22 points and for each point of the model calculates a directional spectrum of 12 directions and 13 frequencies and several wave parameters are calculated. From all the parameters supplied by Carvalho (2003) only the Significant Wave Height ($H_s$) and the Significant Wave Period ($T_s$) are required for the calculations in Chapter 6. Unfortunately, the model only provides the Significant Wave Height from spectral moments ($H_s(m_0)$) and the Peak Period ($T_p$) – see Tables 1 and 2. Therefore, a relationship between $H_s(m_0)$ and $H_s$ and $T_p$ and $T_s$ must first be found to obtain the proper wave parameters. The significant wave height ($H_s(m_0)$) provided by Carvalho (2003) is calculated from the spectral moments (Equation 1):

$$H_s(m_0) = 4m_0^{1/2}$$

(1)
\[ m_n = \int_0^{\infty} f^n S(f) df \quad (2) \]

Where \( S(f) \) is the omnidirectional wave spectrum in terms of the variance of surface elevation and \( f \) the frequency of each wave in the spectra.

According to Tucker (1991) the use of \( H_s(m_0) \) has been generally accepted since presently the majority of the records are analyzed as spectra. For a wave system whose spectrum contains only a narrow band of frequencies; \( H_s = H_s(m_0) \), but for more typical sea states, \( 0.9 H_s(m_0) < H_s < H_s(m_0) \). Therefore, the \( H_s \) can be substituted by the \( H_s(m_0) \) in the calculations (Equation 3) without introducing significant errors.

\[ H_s = H_s(m_0) \quad (3) \]

Concerning the period given from Carvalho (2003), this is the Peak period \( (T_p) \) which is the period at which the wave spectra \( S(f) \) has its highest value. Fortunately, Tucker (1991) provides also the relationships for the Peak period \( (T_p) \) with the Zero crossing period \( (T_z) \) and for \( T_z \) with the Mean period \( (\bar{T}) \) (Equations 4 and 5). These are based on Carter’s (1982) formulae derivation using the JONSWAP spectrum for the case of constant wind velocity with either limited fetch or limited duration. The equations 6 and 7 are respectively the definitions of \( T_z \) and \( \bar{T} \) based on the spectral moments:

\[ T_p = 1.286 \, T_z \quad (4) \]

\[ \bar{T} = 1.073 \, T_z \quad (5) \]

\[ T_z = \frac{m_0}{\sqrt{m_1}} \quad (6) \]
The Significant Wave Period \(T_s\) is the period that corresponds to the mean wave period of one third of the highest waves in the wave spectrum. Considering the definition of the wave period parameters based on the spectra (Equations 6 and 7), \(T_s\) will lie between the values of \(T_p\) and \(T_z\), and is most probably very similar to the value of \(\bar{T}\). Therefore, the value of \(\bar{T}\) will be used in Chapter 6 to substitute \(T_s\) in the calculations (Equation 8).

\[
T_s = \frac{m_0}{m_1} \quad \text{(7)}
\]

\[
T_s = \bar{T} \quad \text{(8)}
\]

Tables 1 and 2 are taken from Carvalho (2003) and represent annual averages for a period of 14 years. They are expressed into eight directional sectors (each one representing 45º) and for each sector it is represented the respective yearly percentages of the wave heights (Table 1) or periods (Table 2) (e.g. in Table 1, wave heights of 1.5 m that come from NE occur 3.71% of the year). The last two columns were calculated by the author, because their values were needed for this work. The calculations in Chapter 6 required the use of the mean annual significant wave height per quadrant \(H_{sav}\) and the mean annual significant period per quadrant \(T_{sav}\).

The \(H_{sav}\) can be determined for each quadrant in Table 1 using the following formula:

\[
H_{sav} = \frac{\sum_{i=1}^{n} f_i \times H_s(m_0)_i}{\sum f_i} \quad \text{(9)}
\]

Where \(H_s(m_0)_i\) is the significant wave height and \(f_i\) its relative frequency (e.g. in Table 1, for quadrant NE, \(H_s(m_0)_i\) is 0.50, 1, ......,16 m and \(f_i\) is 0.420, 2.968,......,0%). Because \(H_s = H_s(m_0)\), then \(H_{sav} = H_s(m_0)_{sav}\).

Similarly, to calculate the mean annual peak period per quadrant \(T_{pav}\), the following formula was used:
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\[ T_{pav} = \sum_{i=1}^{n} f_i \times T_{pi} \sum_{i=1}^{n} f_i \]  \hspace{1cm} (10)

Where \( T_{pi} \) is the peak period and \( f_i \) its relative frequency (e.g. in Table 2, \( T_{pi} \) is 5, 6,......,19s and \( f_i \) is 0.044, 0.685,......0%). After, Equations 4, 5 and 6 are used to determine \( T_{pav} \) after \( T_{pav} \).

The annual wave frequency of a certain directional sector \( (f_{ws}) \) is the percentage of time in a year that waves from that sector occur and was calculated in Table 1 summing all the frequencies \( (f_i) \) of the wave heights of the respective directional sector. (e.g. in Table 1, for waves that come from NE, \( f_i \) are 0.42, 2.97,.....0.00% and the yearly frequency is 12.33%):

\[ f_{ws} = \sum_{i=1}^{n} f_i \]  \hspace{1cm} (11)

The same procedure was used to calculate in Table 2 the yearly period frequency of the directional sectors \( (f_{ps}) \).

The report of Carvalho (2002) provides also the minimum and maximum annual significant wave heights. According to the model (Figure 25), the annual prevailing wave directions come from NW and W (29% and 24% respectively). Less frequent, but also relevant are the waves from N and NE (16% and 12% respectively). The analysis of the significant wave heights along the year (Carvalho, 2002) shows a clear division of the data in four seasonal wave regimes: Autumn (September to November), Winter (December to February), Spring (March to May) and Summer (June to August). Although the waves from SW are often in fifth position in terms of decreasing frequencies, they are quite high which ranks them in the third position in terms of decreasing heights.

The Autumn (September to November) prevailing wave directions (Figure 26) comes from the NW (29%). Less frequent, but also relevant are the waves from the W, N and NE (20%, 19% and 12% respectively). During this period, the maximum significant wave heights can reach 8m, with average significant wave heights between 2 and 3 m.
Table 1 – Average annual significant wave height ($H_s(m_0)$) for the quadrant directions and respective frequencies, concerning the period 1989-2002 (Carvalho, 2003)

| Significant wave Height ($H_s(m_0)$) | 0.5 (m) | 1 (m) | 1.5 (m) | 2.0 (m) | 2.5 (m) | 3.0 (m) | 3.5 (m) | 4.0 (m) | 4.5 (m) | 5.0 (m) | 6.0 (m) | 7.0 (m) | 8.0 (m) | 9.0 (m) | 10 (m) | 12 (m) | 14 (m) | 16 (m) | $f_{Rv}$ (%) | $H_s(m_0)_{av}$ (m) |
|-------------------------------------|---------|-------|---------|---------|---------|---------|---------|---------|---------|---------|---------|---------|---------|---------|---------|---------|---------|---------|---------|---------|---------|
| NE                                 | 0.420   | 2.968 | 3.706   | 2.225   | 1.090   | 0.797   | 0.469   | 0.210   | 0.137   | 0.215   | 0.073   | 0.020   | 0       | 0       | 0       | 0       | 0       | 0       | 12.330  | 1.870   |
| E                                  | 0.176   | 0.953 | 1.232   | 1.085   | 0.665   | 0.284   | 0.279   | 0.108   | 0.015   | 0.034   | 0.005   | 0       | 0       | 0       | 0       | 0       | 0       | 4.836   | 1.913   |
| SE                                 | 0.029   | 0.191 | 0.753   | 0.768   | 0.386   | 0.288   | 0.152   | 0.142   | 0.064   | 0.015   | 0       | 0       | 0       | 0       | 0       | 0       | 0       | 2.788   | 2.211   |
| S                                  | 0       | 0.161 | 0.797   | 0.807   | 0.509   | 0.376   | 0.259   | 0.220   | 0.068   | 0.137   | 0.005   | 0       | 0       | 0       | 0       | 0       | 0       | 3.344   | 2.456   |
| SW                                 | 0.005   | 0.347 | 1.305   | 1.203   | 1.115   | 0.905   | 0.792   | 0.562   | 0.401   | 0.631   | 0.083   | 0.024   | 0.020   | 0.015   | 0.005   | 0       | 0       | 7.687   | 2.999   |
| NW                                 | 0.303   | 2.464 | 3.511   | 3.550   | 3.129   | 2.660   | 2.019   | 1.658   | 1.071   | 1.570   | 0.714   | 0.523   | 0.318   | 0.152   | 0.093   | 0.039   | 0.005   | 0       | 23.779  | 2.964   |
| N                                  | 1.002   | 3.545 | 3.496   | 2.851   | 1.858   | 1.217   | 0.777   | 0.460   | 0.328   | 0.342   | 0.132   | 0.073   | 0.020   | 0       | 0       | 0       | 0       | 16.101  | 2.018   |
| Total                              |         |       |         |         |         |         |         |         |         |         |         |         |         |         |         |         |         |         | 100.00  |

Table 2 – Average annual Peak period ($T_p$) for the quadrant directions and respective frequencies, concerning the period 1989-2002 (Carvalho, 2003)

| Peak Period ($T_p$) | 5 (s) | 6 (s) | 7 (s) | 8 (s) | 9 (s) | 10 (s) | 11 (s) | 12 (s) | 13 (s) | 14 (s) | 15 (s) | 16 (s) | 17 (s) | 18 (s) | 19 (s) | $f_{Rv}$ (%) | $T_{Pav}$ (s) |
|---------------------|-------|-------|-------|-------|-------|--------|--------|--------|--------|--------|--------|--------|--------|--------|--------|-------|--------|-------|
| NE                  | 0.044 | 0.685 | 2.303 | 3.124 | 2.943 | 1.843  | 0.826  | 0.411  | 0.137  | 0.015  | 0      | 0      | 0      | 0      | 0      | 0      | 12.331 | 8.626  |
| E                   | 0.020 | 0.440 | 1.306 | 1.501 | 1.037 | 0.465  | 0.054  | 0.015  | 0      | 0      | 0      | 0      | 0      | 0      | 0      | 0      | 4.837  | 7.986  |
| SE                  | 0.005 | 0.225 | 0.817 | 0.861 | 0.449 | 0.328  | 0.054  | 0      | 0      | 0      | 0      | 0      | 0      | 0      | 0      | 0      | 2.789  | 8.013  |
| S                   | 0.005 | 0.108 | 0.724 | 1.012 | 0.758 | 0.577  | 0.152  | 0      | 0      | 0      | 0      | 0      | 0      | 0      | 0      | 0      | 3.346  | 8.434  |
| SW                  | 0     | 0.166 | 1.169 | 1.359 | 1.545 | 1.584  | 0.933  | 0.543  | 0.225  | 0.078  | 0.015  | 0.005  | 0      | 0      | 0      | 0      | 7.687  | 9.321  |
| W                   | 0     | 0.484 | 2.132 | 3.208 | 3.833 | 3.574  | 3.418  | 2.826  | 2.039  | 1.081  | 0.689  | 0.318  | 0.132  | 0.044  | 0      | 0      | 23.778 | 10.318 |
| NW                  | 0.005 | 0.528 | 3.589 | 4.425 | 4.430 | 4.190  | 3.916  | 3.119  | 2.235  | 1.213  | 0.714  | 0.469  | 0.215  | 0.064  | 0.034  | 0      | 29.137 | 10.143 |
| N                   | 0.015 | 1.027 | 2.474 | 3.237 | 3.027 | 2.670  | 2.019  | 0.963  | 0.406  | 0.152  | 0.078  | 0.029  | 0.005  | 0      | 0      | 0      | 16.102 | 9.085  |
| Total               |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       |       | 100.00 |
The insular shelf of Faial: Morphological and sedimentary evolution

Figure 25 – Average, minimum and maximum annual significant wave heights and respective frequency (after Carvalho, 2002).

Figure 26 – Average, minimum and maximum significant wave heights and respective frequency in the Autumn (after Carvalho, 2002).

The Winter (December to February) prevailing wave directions (Figure 27) come from the W and NW (28% and 27% respectively). Less frequent, but also
relevant are the waves from the N, SW and NE (13%, 10% and 9%). During this period the maximum significant wave heights can reach 11m, with average significant wave heights between 3 and 4 m.

The Spring (March to May) prevailing wave directions (Figure 28) come from the NW and W (26% and 23% respectively). Less frequent, but also relevant are the waves from the N and NE (18%, and 13% respectively). During this period the maximum significant wave heights don’t go beyond 8m, with average significant wave heights between 2 and 3 m.

The Summer (June to August) prevailing wave directions (Figure 29) come from the NW (31%). Less frequent, but also relevant are the waves from the W, N and NE (20%, 19% and 15% respectively). During this period the maximum significant wave heights don’t go beyond 5m, with average significant wave heights between 1 and 2 m.

Borges (2003) characterized the wave climate of the central group of the Azores Islands using ocean wave data available from the publication nº 700 of the U.S. Naval Oceanographic Office. He used the swell and sea observations
(Figures 30 and 31) of the mensal mean wave heights to construct the trimester graphs of the mean wave heights and frequency.
The wave climate from Borges (2003) is presented in a different way than that produced by Carvalho (2002; 2003) making difficult the comparison between them. Nevertheless, they show almost the same results for the swell observations: the prevailing waves are those from the NW and W. The main difference is that in the MAR3G model the next prevailing waves in order of importance comes from the N whilst in the publication n° 700 they come from the SW. The sea observations are quite different, as it should be expected since these are generated by local winds. They show a predominance of the SW sector, followed by the NW, W and S in decreasing order of importance.

Figure 30 – Average wave heights and respective frequency for swell conditions. A – Spring; B – Summer; C – Autumn; D – Winter; E - Annual (Borges, 2003).
In spite of that, the wave data in all of them is consistent with the NW and W directions. The SW direction also appears in the three, with the frequency having more significance in publication nº 700 and the height more significance in the MAR3G model. The N and NE sector in the publication nº 700 has less importance in terms of frequency and height when compared against the MAR3G model.

Taking into account that the annual significant wave heights are in all the
quadrant directions between 1.9 and 3 m and that the mean tidal range is 0.9 m, the coast can be classified as a wave dominated coast according to the classification of Davis and Hayes (1984). Nevertheless, it is should be reminded that the tide in the channel Faial-Pico has probably a remarkable importance in the sedimentary dynamics. Although there is a lack of measurements to support this supposition, it is known that the flood currents are directed to NNE and the ebb currents to SSW, with mean velocities between 1 and 2 knots (Instituto Hidrográfico, 2000).

![Figure 32 – General relationships between tidal range and wave height (Davis and Hayes, 1984).](image)

2.2.4. Storminess records

Borges (2003) built a database of the historic storminess of the Azores archipelago based on the records of two Azorean newspapers. He selected the period from 1835 to 1988 and used the information from the Açoriano Oriental and the Açores newspapers. Using several statistical approaches he managed to characterize the Azores storminess for that period (Table 3) and grouped the storms into a four point intensity scale based on a classification previously made.
by Andrade et al. (1996). He reached the conclusion that the low intensity storms (classes I and II) correspond respectively to 30% and 56% of the total records, whilst the more extreme events (Class IV) accounted only for 5%. Consequently a low intensity event occurs, in average, four times every five years while a violent storm occurs once every seven years. The more struck season (October to March) accounts for 76% of the entire year. During this season, December and February are the months more prone to be hit by storms (20% each), followed by November (15%), October (12%) and March (11%). The average annual frequency is 3.1 storms per year.

Table 3 - Statistical resume of the historic storminess (Borges, 2003)

<p>| | |</p>
<table>
<thead>
<tr>
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<tbody>
<tr>
<td>Average storm duration (days)</td>
<td>2.3</td>
</tr>
<tr>
<td>Maximum storm duration (days)</td>
<td>9.0</td>
</tr>
<tr>
<td>Average frequency (Storms/year)</td>
<td>3.1</td>
</tr>
<tr>
<td>Extreme events frequency (Classes III and IV)</td>
<td>14%</td>
</tr>
<tr>
<td>Average duration of the extreme events (Class IV) [days]</td>
<td>1.1</td>
</tr>
<tr>
<td>Maximum duration of the extreme events (Class IV) [days]</td>
<td>2</td>
</tr>
<tr>
<td>Average of the extreme events (Class IV)/year</td>
<td>0.15</td>
</tr>
<tr>
<td>Average duration of the Class I events (days)</td>
<td>1.8</td>
</tr>
<tr>
<td>Average duration of the Class II events (days)</td>
<td>2.1</td>
</tr>
<tr>
<td>Average duration of the Class III events (days)</td>
<td>3.0</td>
</tr>
<tr>
<td>Number of storms</td>
<td>509</td>
</tr>
<tr>
<td>Number of stormy days</td>
<td>1166</td>
</tr>
<tr>
<td>Number of analyzed days</td>
<td>59536</td>
</tr>
</tbody>
</table>

2.2.5. Sea level variations in the Azores archipelago

In Chapter 5, a sea level oscillation curve will be used to estimate the shelf widening erosion rates. For that purpose there is the need to use a curve that covers 800 Ka, which is the oldest know subaerial age of Faial Island. There are many sea level published curves, but the problem is that many of them are based on the ratio of oxygen isotopes contained in benthic or deep sea foraminifera (Siddall, 2005). Although it is true that the ratio of $^{18}$O/$^{16}$O increases in the water during glacial periods due to the incorporation of $^{16}$O into the ice, there are
complications when it comes to extract the effect of ice-cycles from this ratio. The first complication is the preferential evaporation and subsequent incorporation of the lighter oxygen isotope in the ice-sheets during glacial conditions, which varies where and when the shell was formed, affecting the ocean's $^{18}$O/$^{16}$O value (Bintanja et al., 2005). Secondly, differences in the isotopic ratio between water masses, coupled with changes in deep-ocean circulation, in turn affect the $\delta^{18}$O value incorporated into the shells of benthic foraminifera (Siddall, 2005). The best option is to use sea-level curves based on fossil coral terraces, since they can provide a precise indicator of sea level when the colony lives up to the mean low sea level, as is the case of the micro-atoll formations (Chappell, 1983; Woodroffe and McLean, 1990). In other situations, the range of coral habitat may be as large as 5 to 10 m (Lighty et al., 1982; Montaggioni et al., 1997), being nevertheless more precise than using the ratio of oxygen isotopes (Chappell and Shackleton, 1986). The sea level oscillation curve from Thompson & Goldstein (2006) is the more recent and accurate curve for the last 240 Ka based on corals (Figure 33). It also uses a new approach to correct many of the problems associated with imprecise conventional U-Th ages, producing sea-level reconstructions with improved accuracy and resolution (millennium step) that confidently correlate with SPECMAP, the benchmark of $\delta^{18}$O records (Thompson and Goldstein, 2006). This curve shows a continuous record for the first 129.3 Ka, but since there is the need to go beyond that age, another solution must be found. One hypothesis is to use another curve, for instance one based on $\delta^{18}$O records and use a corrected proxy between concentration and sea level. The Bintanja et al. (2005) approach takes that into account, using the benthic $\delta^{18}$O records (LR04 stack) from Lisiecki & Raymo (2005), a 5.3 Ma stack from 57 globally distributed sites and with a reliable age model calibration (Figure 34). The central assumption from Bintanja et al. (2005) is that deep ocean temperatures are related to the surface-air temperatures at high latitudes in the Northern Hemisphere by a simple, linear transformation. This enables these authors to link algorithms that describe temperatures in the deep ocean to those that calculate the ratios of oxygen isotopes in the ice sheets of the Northern Hemisphere. The authors then adjust the linked benthic and high-latitude temperatures of their coupled model iteratively, to generate the right
Figure 33 – Eustatic sea level curve from Thompson & Goldstein (2006). Error bars correspond to two times the standard deviation and those not visible are smaller than plotted symbols.

Figure 34 – Eustatic sea level curve for the last 800 Ka (after Bintanja et al., 2005).
amount of ice growth and benthic temperature to best fit the record of benthic oxygen isotope ratios, and thus arrive at mutually consistent records of temperature and ice volume at high latitudes in the Northern Hemisphere (Siddall, 2005). A similar approach had already been tried by Siddall et al (2003) in the Red Sea, reaching also good results when compared with other independent sea-level reconstructions. The Red Sea is strongly affected when lowered sea-level restricts circulation with the Indian Ocean. Therefore, Siddall et al (2003) used a conversion of a salinity record from the Red Sea, via a hydraulic model, to infer the sea-level oscillations.

The uncertainties reported so far are only a small fraction of the processes that can affect the prediction of the sea-level oscillations. There are processes mainly related with the rheological properties of the Earth that may introduce considerable variations in the sea-level depending on the geographical position of the area under study (Farrell and Clark, 1976). The magnitude and spatial extent of the ice-water mass exchange during a glacial cycle is sufficient to produce a significant deformation of the solid earth that proceeds until a state of isostatic equilibrium is reached. The surface mass flux and the associated solid earth response are commonly described by the general term Glacial Isostatic Adjustment (GIA). GIA induced sea-level changes are driven by a number of mechanisms. The ice-water mass redistribution is the most important GIA forcing. It contributes to the direct mass exchange between the ice sheets and the oceans and at the same time induces load deformations of the geoid and the solid surface. Recent improvements of the traditional sea-level theory of Farrell & Clark (1976) have had an impact on the predictions of sea-level changes. These include the change in the Earth’s rotation vector which acts to deform the geoid and the solid surface (Milne and Mitrovica, 1998b), the water influx into regions vacated by a retreating marine-based ice margin (Milne et al., 1999) and the impact of a time-varying shoreline, which depends on the detailed ocean-continent geometry and bathymetry (Milne and Mitrovica, 1998a). In the near-field regions, the uplift of the solid surface during and following ice load removal is the dominant process resulting in a sea level fall of up to hundreds to several hundreds of meters (Lambeck and Chappell, 2001; Milne et al., 2002). In contrast, the addition of
glacial meltwater to the global oceans, either by isostatic effect of water loading or simple water excess, is the dominant contribution to the observed sea-level rise in far-field regions (Lambeck and Chappell, 2001; Milne et al., 2002). These examples are, of course, a simplified view of the end-members of the global sea level response. It is likely to expect that at mid-latitude places like the Azores, these two components of the GIA sea-level contribute with different magnitudes creating a more complex signal. It would be difficult to make previsions for mid-latitude places since operational models for glacial rebound with lateral variation do not yet exist and different parameters may need to be used according to the representative mantle conditions (Lambeck and Chappell, 2001). Furthermore, significant and systematic discrepancies between previous predictions of sea-level change subsequent to the last glacial maximum and observations are recorded even in far-field regions (up to 14 m), suggesting that also here the several components that control the sea-level play a significant role (Bassett et al., 2005; Milne et al., 2002). In conclusion, the relative sea level change therefore exhibits complex spatial patterns, even when the localities lie relatively near to each other (Lambeck and Chappell, 2001). Pflaumann et al. (2003) present a comprehensive new Sea Surface Temperature reconstruction of the Atlantic for the last glacial period. It shows that sea ice was restricted to the north western margin of the Nordic seas during glacial northern summer, while the central and eastern parts were ice-free. During northern glacial winter, sea ice advanced to the south of Iceland and Faeroes (between 50º and 60º N). This does not mean that this had to be the case during more extreme glacial cycles like MIS 12 (474–427 Ka). However, it is pretty likely that during most of the glacial times the polar front did not reach the Azores islands. According to Dennielou et al. (1999) the Azores Plateau was located south of the maximum extension of the polar front which was at 42ºN.

Presently, there are also other processes that affect the sea level surface and although they are now constant, they may have varied in the past. The marine geoid is the sea undulating surface related to the gravity field that reflects differences below the surface of the Earth (e.g., density). This surface (Figure 35) can account for sea level variations of almost 200 meters between two ocean
regions thousands of kilometers apart (Cullen and Moore, 2001). Superimposed over this surface is the sea surface height, also called the dynamic topography. It is defined as the height of the sea surface relative to the geoid and is directly caused by ocean currents (Figure 36). Differences of one meter are recorded around the North Atlantic due to wind and buoyancy (heat and freshwater) forcing, resulting in a nonzero dynamic topography and a corresponding circulation of the ocean (Bingham and Haines, 2006).

Figure 35 – Global mean sea surface from ERS-1 geodetic phase data (Cullen & Moore, 2001).

Figure 36 – Composite MDT (mean dynamic topography) for the period 1993-2001 over the North Atlantic (Bingham & Haines, 2006).
In the Azores region a high resolution mean sea surface, named AZOMSS99, has been derived using altimeter data from ERS-1 and ERS-2 35-day cycles, spanning a period of about four years, and also data from an ERS-1 geodetic mission (Fernandes et al., 2000). It shows that the mean sea surface near Faial Island is 58 meters above the zero reference level of the Earth’s ellipsoid (Figure 37). This data also shows that the mean sea surface in the Azores may vary by 10m (see Figure 37) and that using sea-level variations extracted from cores in this area may introduce this amount of error and we are only considering the error transmitted by the mean sea surface variation. The sum of all the errors (if possible) resulting from using local cores in the Azores would be considerable. Therefore, it is preferable to use a compilation of results from different ocean basins and latitudes that would average out local effects. Taking this into account, the sea-level curves from Thompson and Goldstein (2006) and from Bintanja et al. (2005) were chosen as the ones that would introduce fewer errors in the modeling. The resulting sea-level curve is a compilation which includes the first 129.3 Ka of the curve of Thompson and Goldstein (2006) and from 129.3 Ka to 800 Ka the curve of Bintanja et al. (2005).

![Figure 37 - Regional geoid of the Azores archipelago (Fernandes et al., 2000).](image)
However, the author is aware that these sea-level curves do not include most of the corrections for the uncertainties discussed above. Nevertheless, the emphasis in Chapter 5 is not to reproduce the details of an Azorean sea-level curve, but to have a rough estimation of the amount of time that sea has been at the different levels (with 10 m interval).

2.3. Climatic Setting of the Azores archipelago

The Azores archipelago is located in the Northern Hemisphere subtropical anticyclones zone, under the direct influence of the Azores anticyclone. During most of the year (September to March), the polar-front jet often migrates to south and the Azores region is affected by low-pressure systems causing stormy weather and intermittent periods of high winds. It is during this season that three quarters of the total annual precipitation occurs. During late Spring and Summer, the Azores climate is influenced by the Azores anticyclone and there is less rain (Ferreira, 1981; Santos et al., 2004).

The islands are characterized by an oceanic temperate climate, with mild temperatures all year round, at low altitudes, and a rather wet climate. The distribution of rain is highly controlled by topography (the precipitation is 20 to 25% higher in the northern slopes than the southern), rainy at high altitudes and drier in coastal areas. In fact, one of the main processes responsible for the production of precipitation in these islands is the ascent of moist air over the islands terrain. The climate can be considered as oceanic cold in the higher areas where precipitation is higher and temperature decreases, with temperature decreasing 0.6 ºC and the precipitation increasing 250 mm every 100 m. The annual precipitation ranges between 800 mm and 2200 mm (Bettencourt, 1979).

2.3.1. Wind

The highest wind velocities occur from January to March and are related with the bigger horizontal gradient of the atmospheric pressure in the North Atlantic during this period.
In Horta Observatory (Figure 38A) the highest frequency is from SW (21.9%) being the highest velocities from the SW and S, with annual mean values of about 30 km/h. The annual mean days with velocities equal or above 36 km/h (force 5 in the Beaufort’s scale) and equal or above 55 km/h (force 7 in the Beaufort’s scale) are respectively 119 and 28 (Instituto Hidrográfico, 2000). In Madalena Observatory (Figure 38B) the highest frequency is from SW (24.3%) being the highest velocities from SW, with annual mean value of about 24 km/h.

The SW and S prevailing winds are consistent with the generation of local sea waves coming from the SW and S (see Figure 31 to compare) shown in Borges (2003). However, the other prevailing directions of local sea waves shown in Figure 31 are not consistent with those of Horta and Madalena observatories, which may indicate that the data presented by Borges (2003) may not apply for all the islands of the central archipelago.

**Figure 38 - Average annual wind speeds and respective frequency. A) Horta Observatory (Faial Island). B) Madalena Observatory (Pico Island). Data provided from the Horta and Madalena observatories of the Instituto de Meteorologia.**

### 2.3.2. Soil erosion rates in the Azores archipelago

The Azorean soils are mainly Andisols, i.e. soils that have formed in volcanic ash or other volcanic ejecta (Madruga et al., 2001).

Fontes et al. (2004) used modelling and direct measurement to obtain
values of runoff and erosion on pasture lands, the predominant cover of the Azorean cultivated soils. Results show that under pasture, runoff is < 1% of total rainfall but increases sharply to near 20% of the rainfall when the soil is tilled and not protected by vegetation. The sediment produced by erosion corresponded to less than 50 kg/km$^2$/year under pasture cover, but increased to almost 150 ton/km$^2$ during the period of 8 months when the soil was disturbed by tillage and less protected by vegetation.

Louvat & Allègre (1998) estimated erosion rates for three drainage basins in S. Miguel Island. They used the chemical composition of the river waters sampled to infer chemical denudation rates of 26-50 ton/km$^2$/year. Using a steady-state model of erosion that assumes a relationship between chemical and mechanical erosion rates, they reached a rate of mechanical erosion of 70-500 ton/km$^2$/year.

### 2.4. The Faial Island

The Faial and Pico Islands define the subaerial section of a major WNW-ESE topographic lineament that extends over 100 km, from 10 km west of Capelinhos in Faial Island to 10 km east of Ponta da ilha in Pico Island (Figure 39). To the east, this lineament changes direction to NW-SE as the Pico submarine volcanic ridge, with approximately 83 km long and 9.6 km wide (Stretch et al., 2006). This ridge seems to be predominantly formed from fissure eruptions along the ridge axis, with lava flowing down its flanks (Stretch et al., 2006).

Both islands rise from a depth of 1200-1600 m and extend over an area of about 2000 km$^2$, whereas the subaerial parts are a mere 660 km$^2$. (Pico’s surface is about 480 km$^2$ and Faial’s is 180 km$^2$).

In contrast to the fairly well known geological evolution of the subaerial part of the island, the submarine portion has remained almost unknown till the present time. The only data source available to understand its submarine development was the hydrographic map of the Faial Island (Instituto Hidrográfico, 1999). The approximate shelf extent of the island is shown by the 100 m contour (Figure 40), where an obvious increase in slope (higher than five degrees), reflects the shelf edge. One exception is the northeast part of the shelf where the shelf limit
changes abruptly from the 100 m to the 200 m contour. Another exception is the southern part of Faial where the E-W shelf edge occurs at the 50 m contour. The third exception is the eastern part of the Faial Island where the shelf extends beneath the channel Faial-Pico and merges with the western shelf of the neighboring Pico Island.

Figure 39 – Faial-Pico topographic lineament and Pico submarine volcanic ridge. 1. Capelinhos; 2. Ponta da Ilha. Bathymetric curves are spaced 200 m. Bathymetry of the Azores archipelago from GEBCO (IOC IHO and BODC, 2003).

The channel Faial-Pico (Figure 41) is the submarine extension of the two islands. It constitutes an 8 km-wide shelf between the islands and the adjacent coasts whose depths vary from 200 meters at both entrances to an average of 80 meters in the middle. There is a clear bathymetric separation into two sectors by a WNW-ESE lineament between the Ponta da Espalamaca and the Madalena village. The northern sector has an average depth of 80 meters in its axis, whilst the southern is deeper, falling immediately to 130 m south of the lineament. Apparently this lineament is composed of submarine cones of the surtseyan type.
Chapter 2. Geological, oceanographical and climatic setting of the study area

(Nunes, 1999). The more prominent cone is the Baixa do Norte with 50 m height and its peak only 16.2 m below sea level. Westerly of Baixa do Norte there are also other prominent features in the same alignment with peaks just 41 and 46 m below sea level. Easterly of Baixa do Norte are the subaerial features of the WNW-ESE lineament, the Ilhéus da Madalena which are also surtseyan cones. In the southern sector there is a remarkable bathymetric feature which also corresponds to a submarine cone (Nunes, 1999), the Baixa do Sul, with 90 m height and its peak only 7.1 m below sea level.

![Slope of the submarine area of Faial Island](image)

Figure 40 – Slope of the submarine area of Faial Island obtained from the hydrographic map of Instituto Hidrográfico (1999), using ArcGis 9.0 GIS running the 3D Analyst extension. In red are the bathymetric curves.

2.4.1. Geomorphology of the Faial Island

The island of Faial has approximately a pentagonal shape, elongated along the WNW-ESE direction with a maximum length of 21 Km and width of 14 km. The sub-aerial part of the island can be divided into four different physiographic regions (Figure 42): **Caldeira Volcano, Pedro Miguel Graben, Horta-Flamengos-Feteira Region and Capejo Peninsula** (Madeira, 1998). Although Madeira (1998) has
only divided the island into four regions, the author decided to differentiate the coastline into eight different sectors plus a subdivision of one of them. This division is based on the geological mapping of the shelf (see Figure 57 in Chapter 3) and it is the reason why there is a detailed description in this Chapter of the defined coastal sectors that composed the island. The subaerial description of these sectors will be useful to understand some of the differences found on the shelf, because there is, as it will be discussed in Chapters 3, 4 and 5, a high degree of correspondence between the shelf geology and the emerged geomorphology.

Figure 41 – Channel Faial-Pico (after Instituto Hidrográfico, 1999), showing the WNW-ESE lineament composed of submarine elevated features which are thought to be surtseyan cones.
The **Caldeira Volcano** is a stratovolcano which dominates the central part of the island and constitutes the main relief, reaching 1043 m at Cabeço Gordo. It has 15 Km in diameter at sea level and the summit is truncated by a 2 km wide caldera. It is composed by a range of volcanic rocks that includes basalts to trachytes and can be divided into two volcanostatigraphic units (Madeira, 1998). The lower unit, the Cedros Volcanic Complex (Cd in Figure 43), is mainly composed of sub aerial basalt to benmoreite lava flows and some scoria cones and trachytic domes. According to Féraud et al. (1980) and Chovelon (1982) this unit ranges from 470 Ka to 11 Ka. The upper unit, the Caldeira Formation (p and i in Figure 43), corresponds to a Holocene sequence of trachyte pumice fall deposits, phreatic and phreatomagmatic breccias, surges, ignimbrites and lahars, emplaced by explosive eruptions (sub-plinian style from the formation of the summit caldera). The radiocarbon dating made by Madeira et al. (1995) revealed ages that range from 10 to 1 Ka. Pacheco (2001) presents a finer stratigraphy of
the Caldeira Formation, dating it from 16 Ka to the present and assigning it to the Grupo Superior do Complexo Vulcânico dos Cedros (Figure 44) and also assigning the Cedros Volcanic Complex to Grupo Inferior do Complexo Vulcânico dos Cedros.

Although, Pacheco’s (2001) thesis has a more up to date stratigraphy and
nomenclature for the recent products of the Caldeira Volcano, the adopted stratigraphy in this work is from Madeira (1998). The reason is because in Madeira and Brum da Silveira (2003) mapping (p and i in Figure 43), the recent products of the Caldeira Volcano show a geographical distribution that is useful to understand their influence on the shelf sedimentation (see Chapter 4).

![Stratigraphic correlation between the Caldeira Formation (Madeira, 1998) and the finer stratigraphy of the Grupo superior do Complexo Vulcânico dos Cedros of Pacheco (2001).](image)

At its surface, the **Caldeira Volcano** is almost completely covered by the Caldeira Formation (Figure 43) whose thickness can reach 100 m in the northern part of the caldera, diminishing towards the littoral zone. On the top of this Formation, stream drainage has developed eroding the superficial pyroclast deposits and reaching the lava flows of Cedros Volcanic Complex beneath. The drainage has a radial pattern (Figure 42), although in some places the tectonic control reveals its influence, with drainage showing considerable angles to the steepest slopes. These streams are ephemeral over much of the year and end...
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over suspended valleys on the littoral cliffs. The littoral zone is manly constituted by rocky cliffs and often shows accumulation of rock debris at the base of the cliffs and can be divided into four coastal segments.

The segment between Baía da Ribeira das Cabras and Ponta dos Cedros has a NE-SW direction and is slightly concave (Figure 45). The cliffs are developed in the lava flows of Cedros Volcanic Complex and can reach 60 m at Ponta dos Cedros, increasing to Baía da Ribeira das Cabras where they can reach 300 m. There is a remarkable incision (gullying) of Ribeira Funda and also Ribeira das Cabras on the cliff. Lava flows from a recent unit (the Capelo Volcanic Complex) have prograded into the sea (where is now the Fajã village) and acted as a buffer zone to marine erosion (Figures 43 and 45), protecting the old cliff of Baía da Ribeira das Cabras. This protection against marine erosion was enough to build a small beach in front of Ribeira das Cabras, composed of sand near the Fajã village and boulders towards its northeastern sector.

The segment between Ponta dos Cedros and Salão fishing Port has a WNW-ESE direction and is slightly convex (Figure 46). The cliffs are also developed in the lava flows of Cedros Volcanic Complex and can reach 30 m at Salão Port, increasing to Ponta dos Cedros where they can reach 60 m.

The segment between Varadouro and Morro do Castelo Branco has a NNW-SSE direction and is slightly concave (Figure 47). The cliffs are also developed in the lava flows of Cedros Volcanic Complex and can reach 230 m at the northern sector, decreasing to Morro do Castelo where they can reach 60 m.

The streams make small incisions (gulling) on the cliffs. Similarly to Baía da Ribeira das Cabras (Figure 45), the old cliffs of Varadouro are protected by the progradation into the sea of lava flows (represented in the Figure 47 by the low-lying and irregular coast just next to Varadouro) of the Capelo Volcanic Complex. This allowed also the building of a small beach (dot in the Figure 47) in front of the Varadouro thermal Spa, composed of sand in its western tip and pebbles towards its eastern tip. The Morro do Castelo Branco is a small peninsula linked to the land by a narrow isthmus. It is an erosion relief more resistant to wave attack due to its nature – a trachytic dome.

The segment between Morro do Castelo Branco and the geodetic pillar of
Chapter 2. Geological, oceanographical and climatic setting of the study area

Figure 45 – Coastal segment between Baía da Ribeira das Cabras and Ponta dos Cedros (after Instituto Geográfico do Exército, 2001d). Black lines represent contours spaced every 10 m below 50 m height and spaced 50 m above 50 m height.

Rocha Alta has a W-E direction and is slightly convex (Figure 48). The cliffs in this segment are less developed, with heights from 10 to 50 m, sculpted in the lava flows of the Cedros Volcanic Complex. One coastal segment, south of the Horta airport has no cliffs, suggesting that also here there was recent progradation of the volcanism into the sea (see Chapters 4 and 5).

The old cliffs from the Caldeira Volcano used to extend from the Fajã village (Figures 43 and 45) to the eastern part of Cabeço do Capelo (black square in
in Figure 52) and from this place to Varadouro (Figures 43 and 52), before being covered by the more recent flows of the Capelo Volcanic Complex.

Although in Figure 43 the Caldeira Formation covers almost the Cedros Volcanic Complex, near the cliffs of the Caldeira Volcano region the thickness of the cover has almost no expression.

The Pedr o Miguel Graben region comprises the eastern area of Faial, from the Salão village to Ponta da Espalamaca (Figure 42). It is a rather complex region since the explosive and effusive materials from the Caldeira Volcano lay on top of the lava flows of the oldest volcanic structure of the island, the Ribeirinha shield volcano (800-580 Ka old; Féraud et al., 1980). It is predominantly composed of sub aerial hawaiitic lava flows mapped under the designation of Ribeirinha Volcanic Complex (Rb in Figure 43). This region is characterized by a WNW-ESE trending graben structure composed of seven normal dextral faults. The graben is partially fossilized by Holocene pyroclastic fall and flow deposits from the Caldeira Formation in the central area of the island.
The drainage pattern is controlled by the normal faults that constitute the graben, following the base of the fault scarps (Figure 42). The stream drainage north of Lomba Grande is well developed (Figures 42 and 43), eroding the superficial pyroclast deposits and reaching the lava flows beneath (Madeira, 1998). It is less developed between the Lomba Grande and Lomba da Espalamaca (Figures 42 and 43) because it cuts very recent materials from the Caldeira with 1200 years (Madeira, 1998).
Figure 48 – Coastal segment between Morro do Castelo Branco and the geodetic pillar of Rocha Alta (after Instituto Geográfico do Exército, 2001a). Black lines represent contours spaced every 10 m below 50 m height and spaced 50 m above 50 m height.

Figure 49 – Coastal segment between Salão fishing port and Ponta da Ribeirinha (after Instituto Geográfico do Exército, 2001c). Black lines represent contours spaced every 10 m below 50 m height and spaced 50 m above 50 m height.
The littoral zone is mainly constituted by rocky cliffs and often shows accumulation of rock debris at the base of the cliffs. The littoral can be divided into two coastal segments.

The segment between Salão fishing port and Ponta da Ribeirinha is more irregular than the previous ones, nevertheless revealing a NNW-SSE trend (Figure 49). The height of the cliffs is quite variable, ranging from 30 m at the Porto do Salão to 120 m in Ponta da Ribeirinha. The more concave segment can reach up to 150 m.

The segment between Ponta da Ribeirinha and Ponta da Espalamaca is very irregular with heights ranging from 0 to 120 m (Figure 50). The cliffs normally have accumulation of rock debris at the cliff toe and in some areas several pocket beaches have developed. There are, from north to south, two boulder beaches, the Porto da Boca da Ribeira and the Praia da Rocha Vermelha and four sandy beaches, Porto Pedro Miguel, Foz da Grotta da Relvinha, Praia dos Inglesinhos and Praia do Almoxarife. The Praia do Almoxarife is fed by the Ribeira da Praia which has a well developed alluvial plain that extends 500 m landward. The southern tip of Almoxarife beach is composed of boulders that result from the erosion of Ponta da Espalamaca headland.

The **Horta-Flamengos-Feteira** region (Figure 42) is characterized by a set of scoria cones surrounded by low slope areas formed by the accumulation of pyroclasts deposits and lava flows from a basaltic fissural unit, the Almoxarife Formation (Al, in Figure 43). This Formation comprises Hawaiian/Strombolian volcanism that overlies the southeastern slope of the Caldeira volcano and the southern part of Ribeirinha Volcano. This region is covered in the western part by a thin blanket of pyroclasts from the Caldeira Formation (Figure 43). The only available age for this formation is a low quality K/Ar date of 30 ±20 Ka (Feraud et al., 1980). The Almoxarife Formation also appears (Figure 43) in the tectonic block of Praia de Almoxarife (between Espalamaca fault and Rocha Vermelha fault which is not seen at the scale of the map of Figure 43) and in the central part of the Pedro Miguel Graben (Madeira, 1998).

The Horta Bay (Figure 51) is protected at the northern part by Ponta da Espalamaca headland and at the southern by Monte da Guia. This entire bay used
to have a sandy beach before the urbanization of this area (Madeira, 1998). Now, the only remaining of this beach is the sandy beach of Conceição fed by the Ribeira dos Flamengos in the northern part of bay.

The Monte da Guia is a small peninsula (a surtseyan cone) linked to the land by a narrow isthmus. This isthmus is composed of sand in its western part where it forms a bay and the famous sandy beach of Porto Pim. The eastern part of the isthmus has also a small unnamed pocket beach composed of pebbles.

Figure 50 – Coastal segment between Ponta da Ribeirinha and Ponta da Espalamaca (after Instituto Geográfico do Exército, 2001b; after Instituto Geográfico do Exército, 2001c). Black dots represent beach locations. Black lines represent contours spaced every 10 m below 50 m height and spaced 50 m above 50 m height.
The segment between Monte da Guia and Feiteira (Figure 51) has very low cliffs less than 30 m) cut into the lava flows of the Almoxarife Formation. These are plunging cliffs punctuated with several small pocket beaches on this coastal segment. The biggest and the only sandy beach is the Feiteira, whose sediments are fed by the Ribeira da Granja.

The hydrographic network in this region is almost absent; the few exceptions are the Ribeira dos Flamengos and the Ribeira da Granja.

The Capelo Peninsula is located on the western region of Faial and is formed by a dextral en échelon alignment (WNW-ESE direction) of basaltic Hawaiian/Strombolian and surtseyan cones that form the central spine of this peninsula (Figures 42, 43 and 52). The activity from these cones spread lava flows both to the north and south, enlarging the area of the primitive island (which before 10 ka only reached the black dot in Figure 52). The eastern half of the peninsula is built on the west slope of the Caldeira volcano, which means that the other
western half has started as submarine eruptions. These materials belong to the Capelo Volcanic Complex (Cp in Figure 43), whose age is less than 10000 years (as it overlies the oldest pyroclasts of Caldeira Formation), and includes two historical eruptions (1672 and 1957 in Figure 43). There is alternation between the later deposits of the Caldeira Formation and these from the Capelo Volcanic Complex in the eastern sector of the Capelo Peninsula. Capelinhos lies on the most westerly tip of the island and started as a classic surtseyan event in which rising basaltic magma came into contact with sea water to produce violent eruptions of ash, lapilli and steam (Cole et al., 1996; Cole et al., 2001). The Surtseyan activity changed to magmatic Strombolian activity when seawater was prevented from accessing the vent, by the piling of the submarine pyroclastic deposits. At least three tuff cones were formed, with each new cone being built to the east of the previous one. The first two cones were destroyed by collapse events and the third tuff cone still remains today.

This area, due to its youth and the resistance of the geology (mainly basaltic flows), does not have a well-developed hydrographic network. Therefore, the drainage occurs along the morphology structures of the flows or can be criptorreic through subterranean lava tubes (Madeira, 1998). The littoral zone is manly constituted by rocky cliffs and often shows accumulation of rock debris at the base of the cliffs. The littoral can be divided into two coastal segments.

The segment between Baía da Ribeira das Cabras and Capelinhos has a WSW-NNE direction and has two sandy beaches, one in the northern part of the Capelinhos volcano and the other just east of the old cliff before the Capelinhos eruption (Figure 52). In the older sector of the Capelo Volcanic Complex the cliffs range from 30 to 120 m whilst in the littoral covered by the lava flows of 1672, the sea has already cut cliffs that reach 5 to 50 m height (Madeira, 1998).

The segment between Capelinhos and Varadouro has a WNW-SSE direction and one sandy beach in the southern part of the Capelinhos volcano (Figure 52). To the SSE there are two beaches, manly composed of pebbles and is also very frequent the accumulation of rock debris at the cliff toes. The littoral covered by the lava flows of 1672 has cliffs of 20 to 45 m height and the older sector of Capelo Volcanic Complex has cliffs from 5 to 90 m height.
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Figure 52 – Coastal segment of the Capelo Peninsula between Varadouro and Baía da Ribeira das Cabras (from Instituto Geográfico do Exército, 2001d). The hatched red line corresponds to the coastline of November 1958 limit and the solid red line the coastline of 1981 (from Machado and Freire, 1985). Black dots represent beach locations. Black lines represent contours spaced every 10 m below 50 m height and spaced 50 m above 50 m height.

According to Machado e Freire (1985) the littoral zone of the Capelinhos volcano has been eroding with a rate of 20 m/year. These rates were derived from five topographical surveys during 1958 and 1981 (see 1958 and 1981 coastline in Figure 52). Using the 2001 coastline from the map of Instituto Geográfico do Exército (2001d) it is possible to estimate a rate of 10 m/year for the last 20 years when comparing against the 1981 coastline.

2.4.2. Tectonic setting of the Faial Island

The eastern part of Faial Island is characterized by a WNW-ESE trending graben structure (Pedro Miguel Graben) composed of seven normal dextral faults
(Figure 43). Other WNW-ESE important faults are located south of the caldera and in the western part of the island. There is also a less important, secondary conjugate fault system, trending NNW-SSE to NW-SE, composed of normal left lateral faults. Their geomorphic expression is less developed and the fault lengths are smaller.

2.4.3. Climatic setting of the Faial Island

According to Coutinho (2000) the density of the observation stations is not adequate to the dimension of the island. There are only two meteorological stations, six udometric stations and two climatologic stations to characterize the Faial climate. Nevertheless, Coutinho (2000) has calculated the water balance based on real data and using several statistical approaches to interpolate the missing data. He estimated values for the precipitation, actual evapotranspiration and the surplus water that will generate runoff on these 10 stations. Annual precipitation ranges from 980 to 1031 mm/year in the coastal area and 2860 mm at 700 m above sea level. The precipitation recorded from September to March represents 70 to 75% of the total precipitation. The mean annual temperature is 17.5 ºC and the temperature decrease is 0.6 ºC every 100 m of altitude. Mean annual actual evapotranspiration and surplus estimated by the methods of Thornthwaite (1948), Turc (1955), and Coutagne (1954), is respectively 736 and 244 mm/year at 60 m above sea level, reaching 711 and 2049 mm/year at 700 m.

2.4.4. Oceanographic setting of the Faial Island

In section 2.2.3, the review of the wave climate has shown the direction and height of the predominant waves for the Azores central archipelago. However, the different coastal sectors of the island are subjected to more than one wave directional component and therefore to compare its vulnerability to erosion, two parameters must first be calculated, the Directional annual wave frequency (DAWF) and the Directional annual maximum wave height (DAMWH). The coastal sectors presented in Figure 53 are defined in the following Chapter (see Figure 57 and text for explanation in Chapter 3) where a letter for reference has been
assigned to each one. DAWF and DAMWH (Table 5) were calculated for all these sectors using the respective wave directional components that each sector is exposed, which can be three to four depending on the sector orientation (Figure 53). For sector E these two parameters were not calculated because this sector suffers from the sheltering effect of the nearby Pico and S. Jorge Islands.

Figure 53 – Coastal segments defined in Chapter 3 and the wave directions used to calculate the Directional annual wave frequency (DAWF) and the Directional annual maximum wave height (DAMWH).

The DAWF is calculated using Table 1 simply by summing the annual wave frequency \( f_{w} \) of each of the directional components that affect the respective coastal segments (e.g., in Table1 for coastal sector B, the \( f_{w} \) of the W, NW and N are summed which gives 69.2%). Table 4 is from Carvalho (2002) and represent the annual averages of the Maximum Wave height regarding a period of 14 years and is used to calculate the DAMWH. The calculation is simply a weighted mean of the Annual Maximum Wave height of each coastal sector (e.g., in Table 4 for coastal sector B, the frequencies of the W, NW and N are multiplied by the
respective wave height and the result divided by the sum of the frequencies of the quadrants involved).

Table 4 – Annual maximum significant wave height and respective frequency (Carvalho, 2002)

<table>
<thead>
<tr>
<th>Quadrant</th>
<th>Maximum significant wave height</th>
<th>Frequency (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>N</td>
<td>6.06</td>
<td>16.03</td>
</tr>
<tr>
<td>NE</td>
<td>5.14</td>
<td>12.19</td>
</tr>
<tr>
<td>E</td>
<td>4.05</td>
<td>5.00</td>
</tr>
<tr>
<td>SE</td>
<td>3.92</td>
<td>2.79</td>
</tr>
<tr>
<td>S</td>
<td>4.65</td>
<td>3.35</td>
</tr>
<tr>
<td>SW</td>
<td>6.89</td>
<td>7.49</td>
</tr>
<tr>
<td>W</td>
<td>7.95</td>
<td>24.02</td>
</tr>
<tr>
<td>NW</td>
<td>8.24</td>
<td>29.13</td>
</tr>
</tbody>
</table>

Table 5 - Directional annual wave frequency (DAWF) and the Directional annual maximum wave height (DAMWH) calculated for the different coastal sectors

<table>
<thead>
<tr>
<th>Sectors</th>
<th>DAWF (%)</th>
<th>DAMWH (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A1</td>
<td>37.60</td>
<td>7.26</td>
</tr>
<tr>
<td>A2</td>
<td>81.35</td>
<td>7.15</td>
</tr>
<tr>
<td>B</td>
<td>69.02</td>
<td>7.63</td>
</tr>
<tr>
<td>C</td>
<td>62.41</td>
<td>6.74</td>
</tr>
<tr>
<td>D</td>
<td>62.41</td>
<td>6.74</td>
</tr>
<tr>
<td>F</td>
<td>13.82</td>
<td>5.73</td>
</tr>
<tr>
<td>G</td>
<td>13.82</td>
<td>5.73</td>
</tr>
<tr>
<td>H</td>
<td>34.81</td>
<td>7.41</td>
</tr>
</tbody>
</table>

One of the mechanisms described in Chapter 1 for the cross-shelf transport of sediments was the downwelling offshore currents provoked by the piling of water against the coast during storm surges where intense wave agitation affects the coastline. Youssef (2005) studied the physical environment of two sites on the southern coast of Faial. He used current meters deployed at 19 and 23 meters depth to characterize the bottom currents at these depths. During 10 months (from August to June) these sensors measured the bottom current at these two sites and found several extreme bottom current events, the majority of them (80%) related with the increase of the significant wave heights and directed offshore (Figure 54). The highest bottom velocity (200 m/s) was recorded during 24th March, during a
period in which the significant wave height reached 5.12 m. It is likely to expect even higher bottom velocities since according to Carvalho (2002) the annual maximum significant wave heights can reach up to 8 m.

![Figure 54 – Relationship between the measured maximum bottom current velocity and the significant wave height (Youssef, 2005).](image)

### 2.5. Summary

The aim of this chapter was to introduce the state of the art of the scientific knowledge of the Faial Island and the surrounding physical environment. The amount of information exposed was selected concerning its relevance for the scientific problems under study.

First, the tectonic and volcanic setting of the Azores archipelago is outlined to understand the geological origin of the Faial Island. Within this general view, the historical records of earthquakes and tsunamis were mentioned because of their relevance for the evolution of the subaerial as well as the submarine part of the island.

Secondly, the Azores oceanographic environment is also described in detail. The importance of the waves, tides and surface currents in the modeling of
the coastline and the shelf environment is taken into account when facing differences in the coastline of the same volcanic terrains.

Thirdly, the climatic setting of the Azores is also addressed in this chapter because the type of weathering is directly related with the climate. The subaerial erosion rates are needed to fully understand its role in the contribution of sediments to the shelf deposits. The sea-level variations are of extremely importance because it is the oscillating level of the sea through wave erosion that is responsible for the development of an abrasional shelf. The lowest sea levels control the depth of the shelf break, while intermediate stillstands are responsible for the presence or absence of erosional terraces and the longer periods of highstand favor the widening of the shelf.

Finally, the geomorphology of the emerged part of the Faial Island as well as the age of the different volcanic regions is delineated. The subaerial geomorphology is the result of the subaerial processes that have impinged on it. In addition, the age of the terrains normally shows a linear relationship between their age and the degree of evolution of the shelf sectors. Both are good indicators of the expected shelf development because it is often found a high degree of geological correspondence between the subaerial and the submarine part of the island.