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Fine grain sediment transport and deposition in the Patos Lagoon–Cassino beach sedimentary system

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ABSTRACT

Extensive mud deposits superimposed on the predominantly sandy inner continental shelf adjacent to the Patos Lagoon estuary, indicates that the Lagoon is a potential source of fine sediments to the coastal sedimentary system. The lagoon is large and shallow, and the water movement is mainly controlled by wind-driven set-up and set-down. The mean river inflow is around 2000 $m^3 s^{-1}$, although peak flow rates exceeding 20,000 $m^3 s^{-1}$ have been observed during El Niño periods. Though the tidal elevations are small, tidal velocities in the lagoon's inlet can be significant due to the large extension of the backwaters. Moreover, significant exchange flows can be generated between the estuary and coastal area due to barotropic pressure gradients established as a function of wind and freshwater discharge. The predominant net flow is seawards, but opposite near-bed flows due to pronounced vertical salinity stratification can also be observed. The coastal area is characterized by small tidal effects, large scale ocean circulation, wind-induced residual flows and wave-driven currents, where the waves originate from swell or are locally generated.

Fine sediment is brought into the Patos Lagoon by the rivers and its deposits are likely to have long residence times. These fine sediment deposits can be remobilized by locally generated waves, and driven towards the channels and lagoon's shallow bays. Suspended particulate matter (SPM) concentrations within the lagoon do not exceed a few 10 $mg\,l^{-1}$, though higher values have been measured occasionally. In the southern estuarine part of the lagoon, fine sediments may accumulate due to gravitational circulation effects, yielding SPM concentrations of a few 100 $mg\,l^{-1}$. The export of fine sediment from the lagoon to the coastal area occurs predominantly during NE winds. This explains why the majority of the off shore sediment deposits, known as the Patos Facies, are more widespread towards the southern portion of the inner continental shelf. These sediments deposit in the form of fluid and more compacted mud, between the 6 and 20 m isobath, in layers with thickness varying between a few dm to 2 m causing marked lateral differences in grain size. Recent sediment core data, indicates that fluid mud occurrence increases towards the shore and that the mud depocenter remains in the same area as previously mapped two decades before. On a long-term basis, this lateral heterogeneity in sediment properties controls the geomorphology of the inner continental shelf and shoreface, and influences the shoreline accretion rate and beach morphodynamic south of the inlet. Short-term effects are associated with episodic events of mud deposition on the beach during heavy storms that often result in strong gradients in hydrodynamic processes. These gradients in turn influence the morphodynamic behavior on the sectors affected by the mud deposits and can create coastal hazards relating to beach usage.

1. Introduction

Sedimentological data demonstrate that the Patos Lagoon is a significant source of fine sediments to the adjacent shoreface and inner continental shelf of the South Atlantic region. The fine sediment deposits observed in the coastal area seem to result

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from the transport of high amounts of suspended sediment from the Patos Lagoon drainage basin (more than 2,00,000 km^2), and from anthropogenic effects associated with dredging activities and secular deforestation related to settlements and crop cultures.

Fine sediment export through the lagoon's tidal inlet is mainly observed during periods of NE winds and high freshwater discharge. These fine sediments are likely to be transported in the form of a coastal plume producing extensive mud deposits called Patos Facies (Martins, 1967). This pattern has strong influence on the inner shelf morphology, allowing the existence of a flat area between sand ridge fields which predominate over a great area of the inner continental shelf.

Historical records show that during high wave energy events, these mud deposits were remobilized and ended up depositing on the shore of Cassino beach, a high-energy dissipative beach, generating ephemeral mud banks. From 1972 to 2006, more than 30 events of mud deposition along Cassino beach have been recorded (Delaney, 1965; Villwock and Martins, 1972; Calliari et al., 2001). The oldest record dates from 1901 (Calliari et al., 2001). These events became more prominent after the 90s, when they started to cause negative impacts on tourist-related activities affecting the regular use of the beach.

Thus, the objective of this paper is to provide an overall description of the geological and physical processes relevant to the transport, deposition and erosion of fine sediments in this coastal sedimentary system, describing the resulting sedimentation patterns and the phenomenon of episodic deposition of fluidized mud within the nearshore zone and its effect on related beach and surf-zone morphodynamics.

2. Geological setting

The Rio Grande do Sul continental margin is a passive type of margin, generally displaying a low relief with the development of a wide coastal plain, where the most expressive geomorphologic

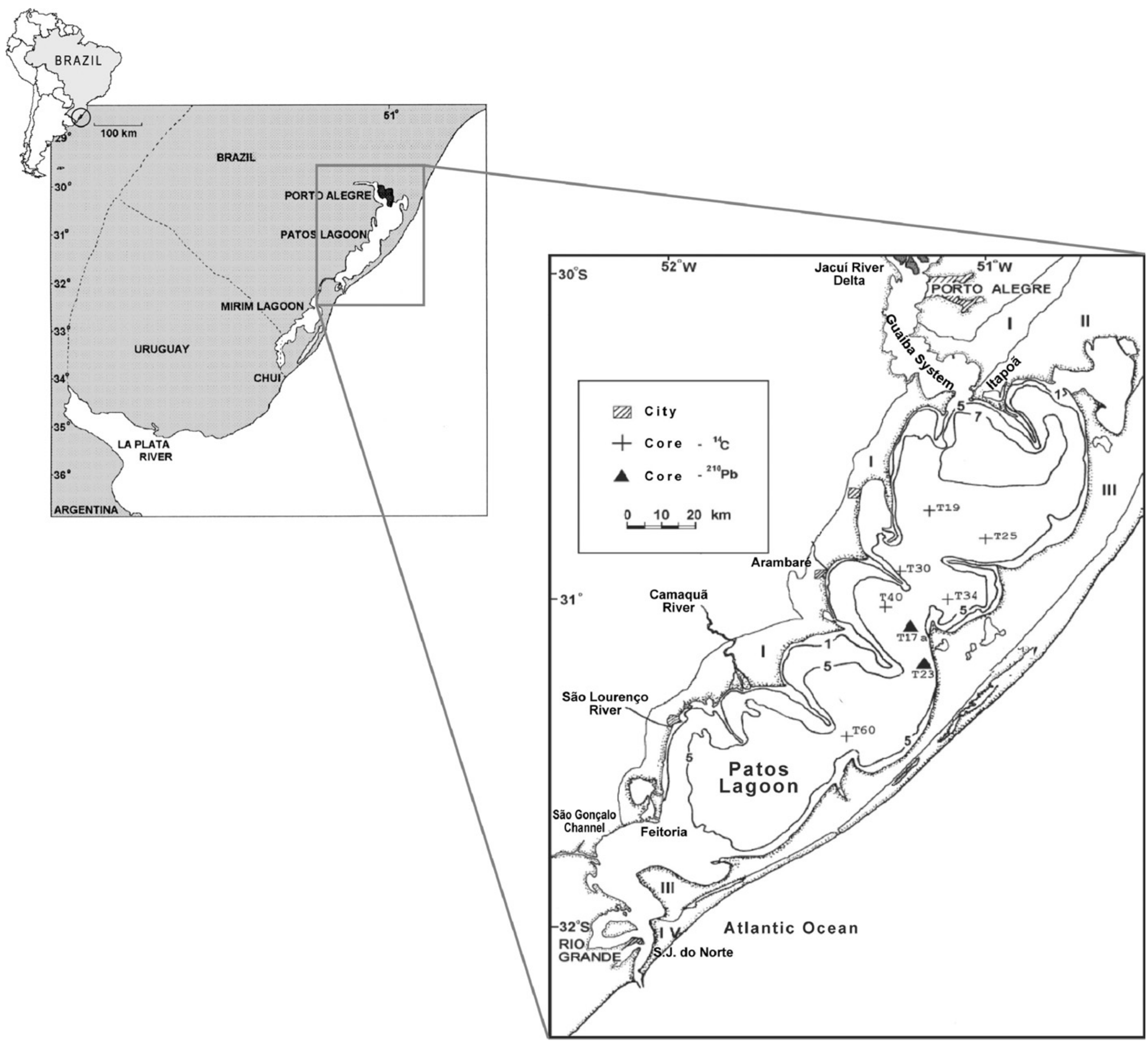

Fig. 1. Patos Lagoon, location, morphology and bathymetry of the Patos Lagoon (after Toldo et al., 2006).

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unit is the 14,000 km^2 Patos–Mirim Lagoon system (Fig. 1). This system receives runoff from a drainage basin of about 2,00,000 km^2, and is connected to the Atlantic Ocean through a single inlet which is less than 1 km wide. The evolutionary history of this system was influenced by the development of a multiple sand barrier complex as a result of eustatic sea level changes during the Quaternary. According to Villwock (1984), at least four successive transgression–regression cycles deposited presumably discontinuous barriers islands parallel to the coast (Fig. 1). The isolation of the Patos–Mirim Lagoon system was initiated by the formation of barrier III during the late Pleistocene (120,000 B.P.). However, although geomorphological events during the late Pleistocene created the original estuarine conditions, erosion and deposition processes by Holocene morphodynamics were responsible for the evolution of the modern Patos Lagoon estuary (Calliari, 1997). The last Holocene Transgression (5500 B.P.) changed barrier III deposits into a continuous scarp and deposited an external barrier (IV), composed of recent beach and foredune ridges and adjacent dunes (Villwock, 1984). Coastal waves and wave-generated longshore currents progressively closed the barrier, except for a channel through which waters from the extensive drainage basin forced their way to the ocean. The entire system can be classified as a partially closed lagoon, which was defined by Kjerfve (1986) as the worlds largest choked lagoon. This system transits to a wide continental shelf where the reworking done by oceanographic processes at the course of the Quaternary sea level oscillations imprinted differences in width, declivity, sediment bottom texture and topography.

3. The Patos Lagoon

3.1. Morphology and bathymetry

The Patos Lagoon (Fig. 1) has a surface area of 10,227 km^2 with an average depth of 5 m, and can thus be classified as a shallow lagoon. The main axis of the lagoon extends over 180 km in a NE–SW direction, while the estuarine area is 60 km long. Based on bathymetric and sedimentological elements, the submerged relief of the lagoon body is subdivided in two large morphological units (Toldo, 1991). The first is a mostly sandy unit, which encompasses the area between the margins and the 5 m isobath, while the second unit starts from there to the deeper parts of the lagoon, which is dominated by muddy sediments.

The most striking features along both margins of the lagoon are the presence of cuspate spits, which are 1 m above the water level on average (Toldo, 1991). Such features commonly form as a result of bidirectional wind blowing parallel to the main axis of the water body in microtidal areas (Zenkovich, 1967; Rosen, 1975), which determine the dominant wave-approach direction in elongated coastal lagoons. This is the case of the Patos Lagoon, where two major wind components of the regional climate, the NE and the SW wind blow along the main axis. The result of the erosional–depositional system in the cuspate spit process is the segmentation of the lagoon body by bars accreting from the end of the spit.

3.2. The hydrodynamics of the Patos Lagoon

3.2.1. Tides, wind and freshwater discharge

The Patos Lagoon has microtidal characteristics, and the tides are mixed, mainly of the diurnal type, having a mean amplitude of 0.47 m near the entrance area (Garcia, 1997). This is due to an amphidromic point for the M2 constituent situated off the Rio Grande do Sul coast. Like most choked coastal lagoon systems, the Patos Lagoon is wind forced rather than tidal driven, and experiences little short-term variability (Kjerfve, 1986).

The region is influenced by a predominantly NE–SW wind regime. Winds from the NE quarter (mean velocity 3.6–5.1 m s^{-1}) occur during 22% of the year, while winds from the SW quarter (mean velocity 5.7–8.2 m s^{-1}) during 12% of the year (Garcia, 1997). Seasonal wind patterns are represented by frequent and strong NE quarter winds between September and April, and by SW quarter winds from May to October (Delaney, 1965). Garcia (1997) comments that these winds blow along the NE–SW main axis of the lagoon body, and these were early identified as the principal forcing factor in the Patos Lagoon system (Malaval, 1922).

The wind action in the Patos Lagoon can be observed through the difference in water level generated inside the lagoon and between the lagoon and the ocean. Data analysis of water level, wind and freshwater discharge carried out by Moller et al. (1996, 2001) confirmed that the wind acting on time scales associated with the passage of frontal systems (6–15 days) tends to be the most important factor promoting circulation. Local wind effect is responsible for most of the observed water level variance in the central and inner parts of the lagoon. However, the remote wind effects, generating large scale set-up and set-down of the ocean's water levels tend to drive the exchanges between the estuarine area and the adjacent continental shelf, being the main mechanism responsible for the inflow and outflow of oceanic water (Fernandes et al., 2002).

The orientation of the lagoon coincides with both the local and non-local wind direction. However, their effects can be separated because they force water level variations of opposite phases in the estuarine area (Fig. 2). Furthermore, their relative importance on the lagoon-continental shelf exchange processes can be distinguished because of the morphology of the lagoon. The entrance area not only filters the astronomical tides, but also attenuates the long period oscillations generated offshore (Fernandes et al., 2004). This result is an important role for the local wind on the dynamics of the inner lagoon and also on the exchanges with the continental shelf.

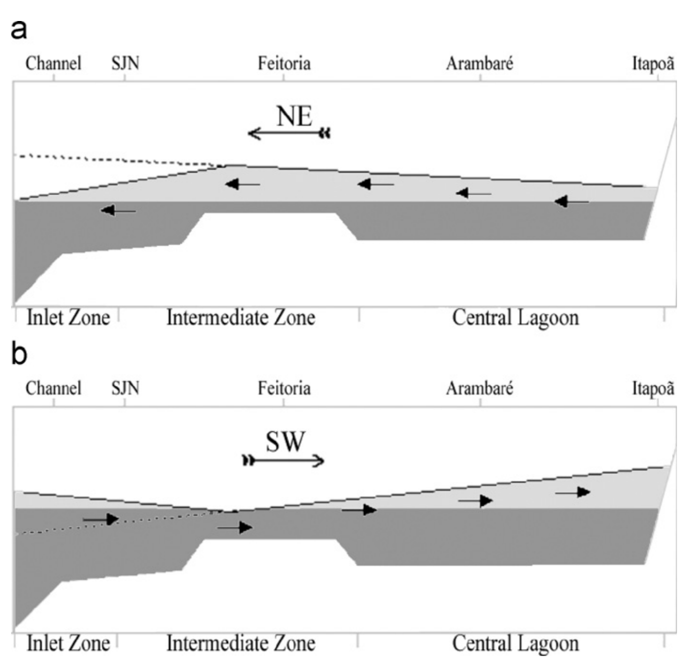

Fig. 2. Schematic representation of the water level oscillations in the lagoon induced by NE (A) and SW (B) wind (after Moller et al., 2001).

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According to the wind dominant effect, Moller et al. (2001) distinguished three different areas in the lagoon (Fig. 2): (1) the inlet zone, from the mouth to São José do Norte (SJN), where the circulation is mostly driven by the remote wind action; (2) the central region situated between SJN and Feitoria, where the local wind dominates the circulation but the remote wind has still some influence; (3) the central and inner lagoon which are exclusively driven by the local wind through set-up/set-down effects. Maximum or minimum water level values associated with the wind regime have always been found at the transition between the central and inner lagoon area. It may be considered as a nodal point where pressure gradients conditioned by the wind have their signals changed.

Fig. 2 presents a schematic representation of the wind effects on the water levels in the lagoon during NE and SW wind events. When the wind comes from the NE (Fig. 2A), its local effect decreases the water level in the north of the lagoon and piles up water at Feitoria. The non-local effect, driven by the Ekman transport acting 90° to the left of the wind (southern hemisphere), reduces the water level close to the mouth. The combination of these two forcing effects generates a barotropic pressure gradient towards the ocean, which favours flushing of the lagoon water. When the wind comes from the SW (Fig. 2B), its local effect increases the water level in the north of the lagoon and decreases at Feitoria, and the non-local effect driven by the Ekman transport is now piling up water close to the mouth. The combination of both effects generates a barotropic pressure gradient towards Feitoria, which favours water penetration into the lagoon. This mechanism, which combines the non-local and local wind action, is the most important factor that forces exchanges between the lagoon and the continental shelf. The stronger the wind, the more important the pressure gradient established between the coast and Feitoria region, and the magnitude of water advection into or out of the lagoon (Fernandes et al., 2002).

The second way by which the winds influence the circulation in the Patos Lagoon is through the cross-shore sea breeze wind component, which is common during summer and spring, with a well-marked preference of easterly (E) winds. Moller et al. (2007) concluded that in the inner parts of the lagoon, 24 h oscillations are mainly forced by the combined effect of diurnal tides and sea breeze action. These oscillations are tied with a natural period of oscillation of 24 h.

Studies showed that although the winds decisively control the Patos Lagoon circulation, salinity distribution and water levels, the freshwater discharge may also become an important forcing factor during periods of high discharge ($>4000\,m^3\,s^{-1}$). Each part of the lagoon has its own tributaries (Fig. 1). In the northern region, the main tributary is the Guaiba River, which receives freshwater from the Jacuí–Taquarí river system, is responsible for 85% of the average total freshwater input in the lagoon. The mean annual freshwater contribution in the north of the Patos Lagoon is $2000\,m^3\,s^{-1}$. Seasonal variations range from $700\,m^3\,s^{-1}$ during summer (late December–March) up to $3000\,m^3\,s^{-1}$ during spring (September–early December). In the central region, the main tributary is the Camaquã River, responsible for the remaining 15% of the freshwater discharge.

In the middle of the estuary, the 70-km-long São Gonçalo Channel connects the Patos Lagoon to the Mirim Lagoon (surface area of $3749\,km^2$), forming the Patos–Mirim system (Fig. 1). Although information about the freshwater contribution of this channel to the Patos Lagoon estuary is restricted, estimates indicate a mean fluvial discharge of $700\,m^3\,s^{-1}$ (Moller and Castaing, 1999). The rivers that flow into the lagoon have a total catchment area of $200{,}000\,km^2$. They exhibit a typical mid-latitude pattern of high discharge in late winter and early spring, followed by low to moderate discharge through summer and autumn. Significant rainfall variation, however, occurs in the Patos Lagoon drainage basin due to large scale events as the El Niño Southern Oscillation (ENSO), where wet and drought periods prevail under El Niño and La Niña events, respectively. Moller et al. (1991) observed river discharge peaks of 12,000 and $25{,}000\,m^3\,s^{-1}$ associated with the meteorological phenomenon El Niño. During high floods, the lagoon can remain fresh for several months and the estuarine mixing moves into coastal waters (Moller et al., 1991). The residence time of river water in the Patos Lagoon was calculated by Moller et al. (1996) as 1.5 yr for a mean annual freshwater discharge of $1000\,m^3\,s^{-1}$. Fernandes et al. (2002) applied a two-dimensional (2D) numerical model of the lagoon to study the dynamics of the system under the El Niño influence, and calculated residence times of 135, 85 and 68 days when considering river discharges of 5000, 8000 and $10{,}000\,m^3\,s^{-1}$ at the head of the lagoon, suggesting a significant increase in the lagoon's flushing as the discharge increases.

3.2.2. Waves

Information on the wave contribution to the dynamics of the lagoon is limited. Toldo et al. (1996) estimated the wave behavior inside the Patos Lagoon based on hourly wind speed and direction data collected between 1986 and 1989. Two stations were installed in the lagoon's west margin and two other in the east margin, providing information to statistically calculate wave parameters such as wave height, period and direction. They concluded that the east margin of the lagoon receives less wave energy than the lagoon's west margin, although the east margin is subject to high-energy waves resulting from storms. They also present the predictions of waves in the lagoon based on 1 yr wind observations in 1988. They concluded that the dominant wave direction on the west side of the lagoon is from the NE and ENE, with mean annual significant wave heights of 0.6 m and periods of 2.9 s due to the prevailing northeastern winds in summer and spring. On the east side of the lagoon, however, the mean annual significant wave height is 1.6 m and the period is 4.8 s due to prevailing winds from WNW and WSW occurring during autumn and winter. The authors also comment that the high frequency of winds with velocities between $0–2\,m\,s^{-1}$ and their short duration are the principal limitation for wave generation inside the lagoon. The small depth of the lagoon is also a limiting factor. Wind velocities of $14\,m\,s^{-1}$ or more, on the other hand, produce storms in the lagoon with modal significant wave height of 1.0 m and period of 4.0 s.

3.3. Sediments

The major volume of clastic sediments in the lagoon originates from the southeast basin (Martins et al., 1981), which drain the highlands to the north (500–1000 m elevations), the lowlands to the west (100–500 m elevations) and the coastal plain, with elevations lower than 100 m (Toldo, 1994). According to these authors, the main tributaries from the highlands loose their transport capacity when they flow into the wide depositional basin ($500\,km^2$) of the Guaíba complex, forming the Jacuí fluvial delta complex in the northern part of this basin (Fig. 3), where the major part of the coarse, sandy sediments are deposited.

Additional sediments to the main lagoon came from the lowlands to the west, drained by the Camaquã River (Fig. 3), which forms a delta in the lagoon, and from two other small rivers and several small tributaries. Coastal erosion of the lagoon margins are also a source of sediments, since several sectors of the west margins are under this process. The fine sediments composed of silt and clay enter the lagoon, and are deposited in the deeper parts and in shallow bays protected by long sand spits.

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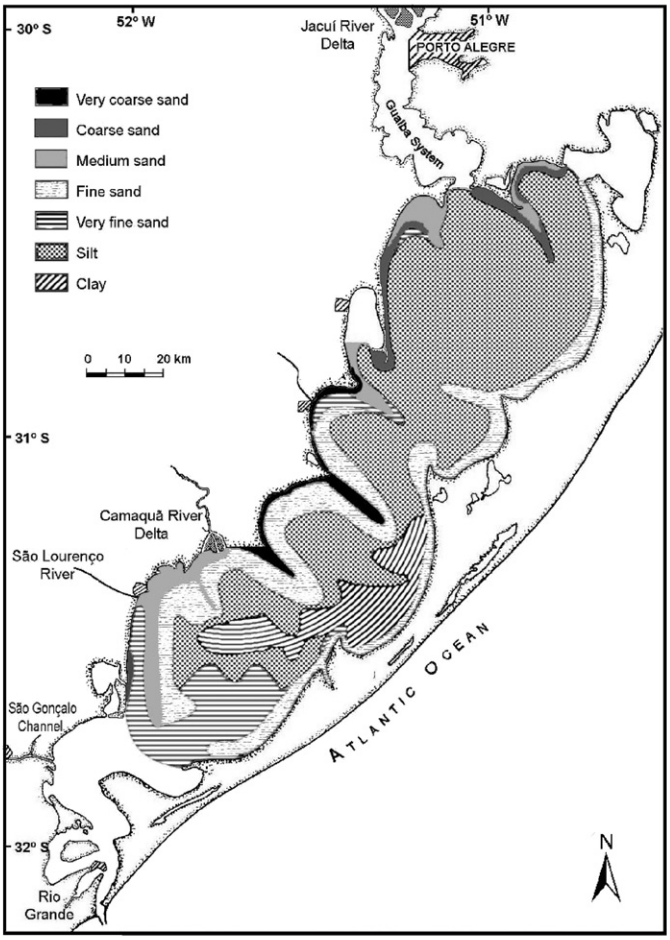

Fig. 3. Bed composition within the main lagoon body (after Toldo, 1994).

Toldo (1994) identified seven bottom types along to the main body of the Patos Lagoon (Fig. 3). Very coarse, coarse and medium fine sand is found in the northern part of the lagoon and in the surface of the sand spits located at the central part of the lagoon. Fine sand bordering these bottom types characterizes the lateral variation along the west margin. Fine sand is also bordering the east margin, where it connects laterally with very fine sand, silt and clay bottom types. There is a remarkable difference in sediment maturity between the west, east and north margin of the lagoon. The west margin consists of arkosean medium to coarse sands (Martins and Vilwock, 1987), which reflects the source area characterized by the internal coastal depositional system and the actual contribution of the drainage of the lowlands and coastal plain throughout the fluvial streams mentioned above. The east and north margins are composed of mature fine sands from a coastal-marine depositional system. The silty facies dominate the major part of the lagoon bed, while the clay facies are restricted to the southern sector of the main lagoon body. In the transition zone between the main lagoon and the estuarine system, both margins and the lagoon bed are covered by a wide deposit of fine sand, which probably indicates the existence of an ancient flood tide delta (Toldo, 1994).

Baisch and Wasserman (1998), analyzing the granulometric distribution of the fine fraction of the superficial bottom sediments, found a visible decrease in the silt content from the northern to the southern part of the lagoon. Their result also showed that in the same direction there is an increase of the very fine-grained fraction represented by fine and coarse-grained clay. According to their interpretation, the very high silt content in sediments at the mouth of the Guaíba system (Fig. 1) is the first stage of a sedimentary process at a local level of Patos Lagoon. At a distance of 130 km from the Guaíba system, there is a considerable increase in the silt fraction due to the contribution of the Camaquã River forming an intra-lagoon delta. In this southern cell, the bottom fine sediments consist of a mixture of fine-grained sediments originated from the Guaiba system and coarse-grained sediments from the Camaquã River delta. The increase in clay content in the southern cell and in the estuary coincides with the northern limit of the salt water intrusion during periods of low discharge associated with southern quadrant winds.

Three years of sampling along the lagoon main axis and at the mouth of the main tributaries of the Patos Lagoon indicates that the average amount of suspended matter can vary between 43 and 196 mg/l (Hartmann et al., 1986), being extremely dependent on salinity. Generally, increase in salinity along the lagoon causes abrupt decrease in suspended matter in the water column. Textural analysis of the suspended matter indicates that it is mainly composed of silt ($>80\%$), followed by clay (15%) and sand (5%), which is almost absent. This data indicates that the lagoon is a source of silt (range of 5–8 phi) and also clay for the inner continental shelf.

Based on Holocene sediment thickness, stratigraphic correlations and local sea level curves, Toldo et al., 2006) estimated the long-term sedimentation rate for the lagoon at 0.75 mm yr^{-1}. Such rate is inside the range of 0.5–3.8 mm yr^{-1} found for coastal lagoons in the US (e.g. Nichols, 1989). Short-term sedimentation rates based on ^{210}Pb data from two surface deposits give values of 3.5 and 8.3 mm yr^{-1}, which is fairly high and similar to the values found in lagoons with active deltaic sedimentation (Nichols, 1989). The authors attributed this high rate to the deforestation of the watershed and/or other human-related activities in the last 150 years.

4. The Patos Lagoon estuary

4.1. Estuarine morphology and bathymetry

The southern part of the Patos Lagoon, an area of 900 km^2, is characterized by the presence of sedimentary islands, large shoals areas with depths varying between 1 and 5 m and a large number of shoals and spits, forming shallow bays. The deeper parts are represented by natural and artificially dredged channels with average depths of 10 m, where most of the navigation routes are located. The main navigation channel, called Canal do Norte, is 14 m deep and connects the lagoon with the Atlantic Ocean. The channel is fixed by two 4-km-long jetties with widths varying between 700 m at the entrance up to 3 km at its northern end. This 15-km-long channel needs periodic dredging, since large volumes of fine sediments deposit in the deeper parts of the channels.

The high level of precipitation causes considerable fluvial input of fine sediments (silt and clay) from the drainage basin, and thus the estuary receives sediments from a variety of sources. The reducing flushing current velocities in the Patos–Mirim Lagoon complex lead to a large sediment deposition. Apart from fluvial sediment input, the hydrodynamic erosion of estuarine margins, especially those comprised of Pleistocene and aeolian Holocene formations, lagoon terraces, marshes and benthic estuarine and flora represent important source of organic sediments.

4.2. Hydrodynamics

Closs (1963) has suggested that the limit for the Patos Lagoon estuary can be defined around Feitoria region (Fig. 1), located 60 km away from the entrance. For very dry seasons and SW wind conditions, this limit can migrate northward, and during flood periods it can be more restricted to the entrance of the channel.

Möller et al. (1996) applied the numerical model of Wang and Connor (1975) to analyze the propagation of a diurnal tidal wave into the estuary. Their results showed that 70% of the original O1 wave, the main tidal constituent, can be attenuated 22 km landwards. However, no data was available to validate this study. More recently, Fernandes et al. (2004) studied the tidal attenuation throughout the estuarine access channel using a 2D finite element model. Their main objective was to propose an alternative method for studying the attenuation of tidal and subtidal oscillations in an area where obtaining simultaneous time series recorded by several tide gauges throughout the estuary is difficult. Their results suggest that the low-frequency (subtidal) oscillations are less attenuated and propagate further than the high-frequency (tidal) oscillations, and that semi-diurnal and quarter-diurnal signals are more attenuated than the diurnal ones in the Patos Lagoon estuary.

The mean annual salinity in the estuarine area is 13 ppt (Castello, 1985), with instantaneous values ranging from 0 to 34 ppt. Although coastal lagoons are considered as well as mixed types of estuaries, the longitudinal salinity distribution presented by Moller and Castaing (1999) indicates that this is not always the case in the Patos Lagoon estuary. In this area, stratification and destratification cycles depend on the relative strength of the barotropic circulation induced by the wind and the river flow. Surface salinity distribution data for the estuarine area presented by authors for different seasons (summer, autumn and spring), suggest that the strongest part of the longitudinal salinity gradient is observed north of the entrance channel. The authors concluded that the advection related to the freshwater discharge and the low-frequency oscillations related to the non-local wind forcing are the main processes controlling this longitudinal salt transport in the Patos Lagoon estuary entrance channel. Lateral salinity gradients are also observed in the area where the cross-sectional area increases (Fernandes et al., 2005).

Moller and Castaing (1999) also comment that the vertical salinity distribution varies from the salt wedge type to a well-mixed structure, which was confirmed by the three-dimensional (3D) numerical experiments carried out by Fernandes et al. (2005). The morphology of the estuary may also play an important role in establishing the traditional estuarine circulation, as the funneling at the main access channel tends to enhance the effect of river flow. Furthermore, the deep inlet (14 m) is also favorable to induce a two-layered vertical profile.

Garcia (1997) commented that as a consequence of the reduced tidal influence, salinity distribution does not show tidal variability but correlates with wind forcing and variations in freshwater input on scales of hours to weeks. During periods of low discharge (summer and autumn), onshore SE and SW winds force seawater through the inlet into the lower estuary and occasionally as far as 150 km into the lagoon. In contrast, NE winds together with high fluvial discharge significantly decrease estuarine salinity. Fluvial discharge in excess of 3000 $m^3 s^{-1}$ causes pronounced salinity stratification in the inlet, and higher values extend the estuarine mixing zone into the coastal waters (Moller et al., 1991). Thus, although our understanding of the system's hydrodynamics is still limited, it is evident that both the regional climate and the hydrological cycles are the principal forcing factors controlling the lagoon and estuarine circulation patterns and salinity variations.

4.3. Sediments

Six different bottom types can be related to the depth and hydrodynamic energy levels in the estuary (Fig. 4). Silty clay sediments dominate the deep channels and shallow protected bay environments (Calliari, 1997). Transitional environments change with decreasing depth from silty clay, clayey silt, mixed sand-silt-clay and clayey sand to silty sand bottoms. Sandy bottoms, the predominant sedimentary cover in the estuary, are found on the shallow marginal part of large shoals but occasionally occur in channels with high hydrodynamic energy.

Detailed studies based on 179 sediment bed samples (Antiqueira and Calliari, 2006) allowed to define six bottom types along the lagoon entrance (Fig. 5). Sandy sediments are found in shallow areas adjacent to a sand spit attached to the root of the east jetty. This sand spit is generated by longshore sand transport induced by flood currents during periods of storm waves from the southeastern quadrant.

Behind the spit, calm conditions develop and fine sediments consisting of clayey silt, clayey sand and silty clay are deposited. As the depth increases, the sediments become finer, but mixed bottoms predominate. An increase in the amount of fine sediments transverse to the inlet as described for the estuarine channels located in the medium and upper part of the estuary missing. The non-occurrence of a silty sand bottom is remarkable, and probably indicates that the sedimentation patterns at the inlet are driven by bidirectional currents. In the northern part of the inlet a clayey silt area occurs, most likely as the result of trapping the sediment by the salt wedge. The sediment mapping did not indicate the presence of a sandy bottom near the end of the inlet. This is probably due to the fact that during the sampling

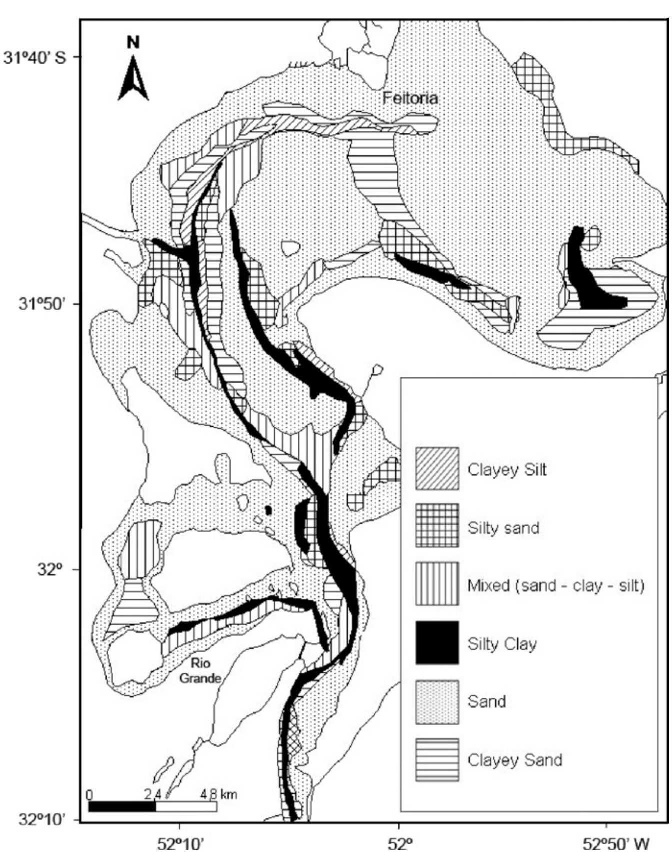

Fig. 4. Bottom types in the estuary based on the proportions of sand silt and sand and clay composition of estuarine bed (after Calliari, 1997).

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year the ebb flow predominated reaching in some situations 10,000 $m^3 s^{-1}$, carrying large loads of fine sediments towards the ocean. Thus, the bottom mapped at the very end of the inlet can represent an ephemeral situation, since previous mapping identified lateral segregation of sand and mud deposits.

In order to maintain the design depth of the navigation channels, seven major sectors inside the estuary are subject to periodic dredging. In the last 20 years, approximately $30 \times 10^6 m^3$ have been dredged with an annual average of $1.5 \times 10^6 m^3$. Although the major part of the dredged sediments is silty sand, large quantities of finer sediments (clayey silt and silty clay) deposited on the deep channel bed and in the harbor basins are also dredged. According to the port authorities, 83% of the total dredged volume came from the main channel entrance. The remaining 17% of the total volume which was dredged from the basins mainly consisted of fine sediments (clayey silt and silty clay). The historical disposal site (more than 50 years) for the clayey sediments from the Rio Grande New Harbor Basin and other muddy places is a large sandy shoal, close to the foot of the west jetty (Fig. 5).

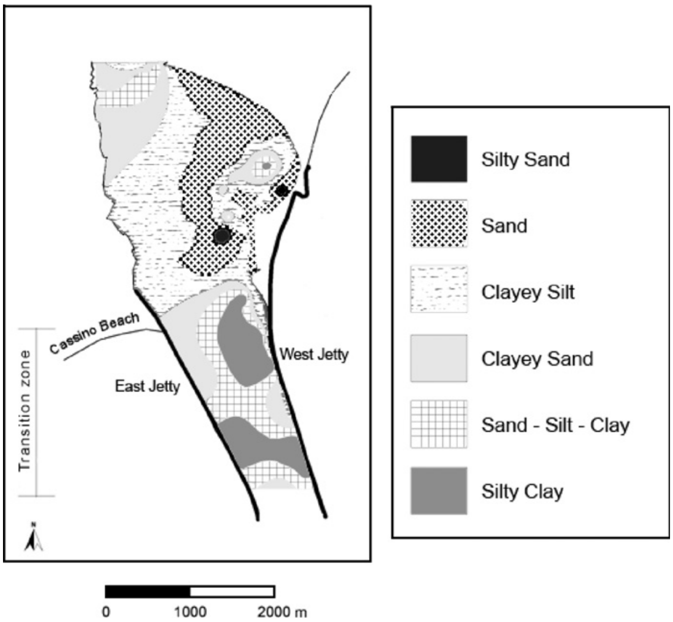

Fig. 5. Bed composition at the inlet (after Antiqueira and Calliari, 2006).

5. The continental shelf

5.1. Morphology and bathymetry

According to Castro and Miranda (1998), the limits of the Southern Brazilian Shelf (SBS) are established between 28.5° and 34°S (Fig. 6), and can be considered relatively smooth. The continental shelf is narrow in the northern part (110 km) and widens up to 170 km in the south, being the shelf break located around the 180 m isobath. According to Correa and Ponzi (1978),

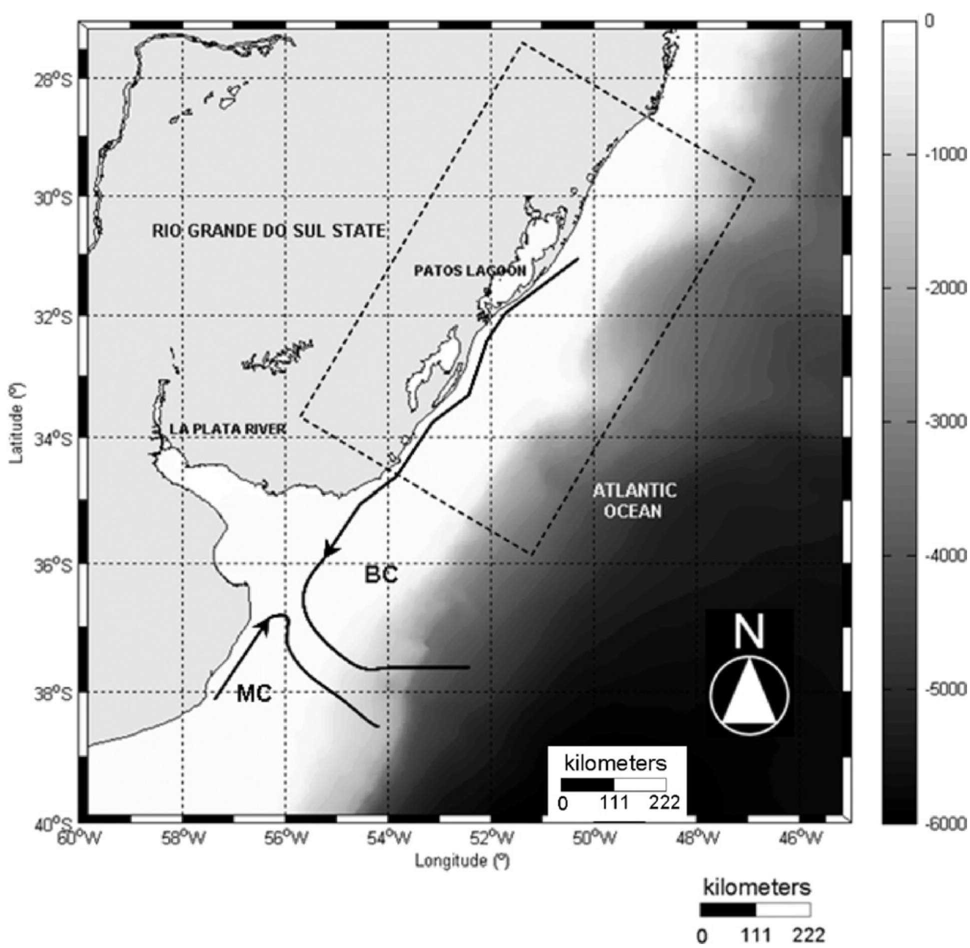

Fig. 6. The SBS and the main physical forcing in the outer and inner shelf.

the continental shelf displays several drowned river valleys, long sand ridge fields and beach rock outcrops. The most notable features are the sand ridges, which can extend 70 km and have heights varying from 4 to 10 m, with a preferential northeast orientation forming a 30° angle with the shoreline. Such features occur at depths between 8 and 30 m and concentrate along the inner shelf, except for the central part of the coast adjacent to the mouth of Patos Lagoon. During storms, the sand ridges migrate and shelly gravel is deposited in the troughs.

Fig. 7 (Fachin, 1998) shows the relief of the continental shelf up to the 33 m isobath, which includes the inner shelf and the shoreface. The presence of three distinct morphological regions can be distinguished in this figure: from the lagoon mouth up to 50 km southwards, the shoreface and inner shelf are extremely smooth, and significant expressive topographic features are absent. This area is limited to the north and to the south by two irregular areas characterized by the presence of linear sand ridges with heights up to 7 m. The figure also shows that the northern and southern regions display two different patterns of sand ridge orientation in relation to the shoreline, the latter being more oblique and shorter than the sand ridges to the north, which are longer and more distant to the coastline. The existence of this smooth area between two regions of irregular bathymetry indicates different bottom types composed of mixtures of sand and mud. It is well known (Calliari and Fachin, 1993) that the muddy deposits of the Patos Facies (Martins et al., 1967) can extend beyond the 20 m isobath in this area.

5.2. Hydrodynamics of the SBS

5.2.1. Currents and hydrography

The outer SBS is under the influence of the convergence of two western boundary currents, the Brazil Current, which flows over the slope transporting warm water southward, and the cold northward flowing Malvinas Current, forming the Brazil–Malvinas Confluence (Fig. 6). Campos et al. (1996) mapped this tongue of low temperature and low salinity up to approximately 24°S, well inside the SBS, during the austral winter of 1993. Seasonal latitudinal migrations of the Brazil–Malvinas Confluence result in a stronger influence of the Brazil Current on the SBS during the austral summer, when this current reaches its maximum southward displacement (Olson et al., 1988).

The inner shelf is dominated by the coastal currents originating along the Patagônia coast (Piola and Rivas, 1997) and by the La Plata River discharge along the Uruguayan coast (Framiñan and Brown, 1996). The cold northward flowing Malvinas Current receives the large continental freshwater runoff from the La Plata River drainage basin (3,170,000 km^2). The La Plata River, the second main hydrographic basin of South America (Fig. 1), does not show significant seasonal variations. The advection of the river plume towards the SBS depends on the coastal winds (Guerrero et al., 1997), which are favorable for northeast transport during austral autumn and winter, and unfavorable in the austral spring and summer.

The study of the low-frequency currents and water mass distribution on the SBS conducted by Soares and Moller (2001) concluded that riverine, low salinity water inflow to the SBS is high in the autumn, winter and spring austral seasons due to: (1) the Patos Lagoon runoff, which is high in the austral winter and spring and (2) to the coastal winds over the Argentina shelf, which are favorable to northward transport of the La Plata River plume in the austral autumn and winter periods. Soares and Moller (2001) also comment that local winds in the SBS are very energetic in few-days period bands due to the atmospheric frontal systems and the transient high-pressure centers that reach the region with periods ranging from 3 to 10 days (Stech and Lorenzzetti, 1992).

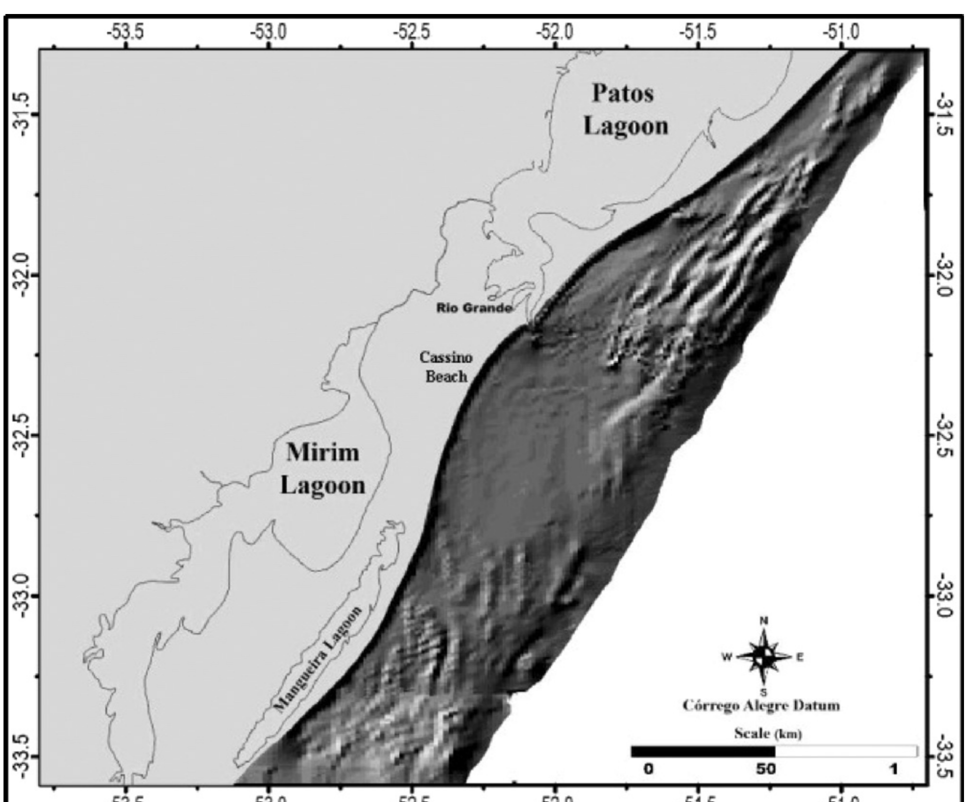

Fig. 7. Relief map of the upper part of the inner shelf and shoreface of the area between Mostardas and Albardão lighthouse. The offshore limit corresponds to the 35 m isobath (after Fachin, 1998).

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Therefore, the SBS low-frequency current variations should be expected to contain combined influences of the seasonal and short-term wind variations, the seasonal Brazil–Malvinas Confluence migrations, the meso-scale processes of the Brazil Current (meanders and eddies), the cycles of the La Plata and Patos Lagoon runoffs, and the short-term variations of the latter.

Marques et al. (2008) comment that studies on the dynamics of the coastal plumes on the SBS are sparse. At the southern boundary, the coastal plume resulting from the La Plata River discharges 22,000 $m^3 s^{-1}$ freshwater in the coastal zone (Framiñan and Brown, 1996). Towards the north, the Patos Lagoon discharges on average ten times less water over the shelf with a seasonal and inter-annual variability. Hartmann and Silva (1989) observed the wind contribution for the Patos Lagoon coastal plume formation under low to moderate discharge conditions and, more recently, Zavialov et al. (2003) monitored the river plume behavior under high discharge conditions using salinity and temperature observations. Marques et al. (2008) studied the barotropic mode of the response of the Patos Lagoon coastal plume to the main physical forcing (rotation, tides, discharge and wind) based on 2D numerical modeling experiments and EOF analysis. Their modeling results showed that rotational effects, tidal currents formed by the interaction between tides and topography, and southerly and southwesterly winds contribute to the freshwater northward transportation. The tides intensify the horizontal mixing process spreading the freshwater over the shelf. The interaction of the rotational and tidal effects, the southerly and southwesterly winds contribute to the coastal current formation. On the other hand, the dominance of the northerly and northeasterly winds enhances the spreading of the coastal plume off the coast. The main spatial variability mode of the EOF analysis indicate that 70% of the plume variability is associated to the plume head oscillation in response to the freshwater intensity and as function of the prevailing winds direction. The second mode explains 24% of the variability and was associated to the seasonal set-up and set-down mechanism of the coastal plume.

5.2.2. *The wave climate in the SBS*

Data on the SBS wave climate are sparse and lack resolution in space and time. The main field data in shallow/intermediary waters for the coastal area of Rio Grande do Sul State (RS) resulted from two studies accomplished in the surroundings of Tramandaí beach (Motta, 1963) and Rio Grande city (Strauch, 1998), which define the northern and southern limits of the RS coastal area.

Motta (1963) determined the maximum wave height in Tramandaí beach based on historical data obtained between October 1962 and September 1963 (12 months of data), using a non-directional wave rider, anchored at 17.5 m depth. The observed maximum wave height was 7 m in April 1963, the maximum value of the significant wave height was 4.8 m. The direction of wave incidence was acquired visually, indicating that crests were almost parallel to the coast.

Strauch (1998) carried out measurements with a directional waverider, located close to the mouth of the Patos Lagoon estuary at 15 m depth. The data showed mean significant wave heights of 1.0 m in the 100 × direction (SE) due to sea conditions, and mean significant height of 1.5 m in the 160° direction, due to swell conditions.

Coli (1994) combined the analysis of data of wave height and direction obtained through opportunity ships (between 1946 and 1979), with data of the satellite Topex/Poseidon (1993), and determined the spatial and temporal variation in wave height and direction in the coastal and oceanic area of the RS State. His results showed that the winter waves were larger than summer waves. The northeast, east and southeast waves were more frequent in spring time, while the north, west and southwest waves were amplified during autumn and winter time. Waves of medium height mainly came from the southwest direction and, secondarily from the south, west and north directions. Regarding the annual mean for each of the wave propagation direction quadrants, the data demonstrated significant variability, with waves coming mainly from the northeast and from the south. Through the analysis of the Topex/Poseidon data, it was possible to determine that in the southern oceanic limit of the study area the largest waves prevail in relation to the north. Below 33°S of latitude, historical maps demonstrated the occurrence of centers of larger wave heights, intensifying during autumn, and reached its maximum during winter.

5.3. *Sediments*

In general, the inner and outer continental shelf are respectively dominated by quartz sand and muddy deposits. The sediment patterns of the continental shelf and upper slope are predominantly terrigenous. Based on the proportion of sand, silt and clay they can be classified in eight distinct bottom types (Martins et al., 1967; Corrêa, 1987), Fig. 8. The inner and outer relict facies composed of medium and fine sand were deposited during the Pleistocene Regressive and reworked at the course of the Holocene Transgression. They are similar to the recent beach and dune sands of the actual beaches and cover practically the entire inner shelf (Martins et al., 1967) being interrupted by a carbonate facies in three areas: north of the Albardão lighthouse; approximately 30 km to the north of Rio Grande city and close to Mostardas. These superficial calcareous deposits, represented by shell fragments, cover approximately an area of 1000 km^2 and are located between depths of 15 and 30 m (Martins et al., 1967; Côrrea and Ponzi, 1978). They are mostly coupled with linear sand ridges and at specific locations (Calliari and Abreu, 1984) are associated to beach rocks outcrops and coarse fluvial relict sand. Relict facies of silty clay and clayey silt dominates the medium and outer shelf and probably were deposited during the Holocene Transgression under the effect of the continental drainage. Transitional facies constituted by a mixture of sand, silt and clay in equal amounts and mixed with other calcareous sediments characterize the other facies.

A close look at the Patos Facies (Fig. 9) adjacent to the estuarine part of the Patos Lagoon, mapped on the basis of 150 bottom samples and hundreds of miles of side scan sonar records indicates that it consists of mud, muddy sand and sandy mud sediments (Calliari and Fachin, 1993), derived from the lagoon complex and mixed with the transgressive sand sheet. These facies are more widespread in front of the lagoon's mouth, and until 1984 reached the 22 m isobath with a length and width of the order of 40 and 14 km, respectively (Calliari and Abreu, 1984; Corrêa, 1987).

Silty clay and clayey silt bottoms (Calliari and Fachin, 1993) have been mapped towards the south of the lagoon's mouth between 15 and 17 m isobaths. It is noteworthy that fine sediments were found up to the 22 m water depth. Although not mapped, we speculate that fine sediments are present at larger depths towards the south as well.

More than 30 cores taken from these facies during several marine geologic expeditions under the Marine National Program of Geology and Geophysics (PGGM) throughout the 80ths, displayed significant amounts of muddy sediments. Inter-beded sand and mud with sharp, abrupt transitions were common both to the north and south of the lagoon's mouth. The isopach map of total mud (sum of all the mud layers) constructed in 1984 (Fig. 10), indicated an increase in thickness toward the beach being more

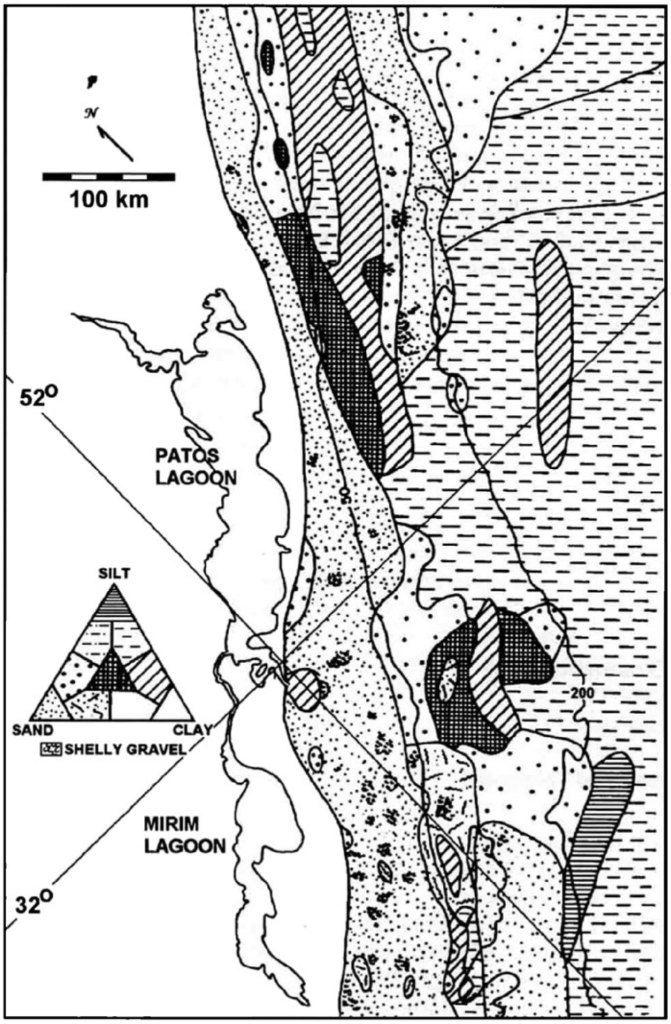

Fig. 8. Superficial sediments of the continental shelf and upper continental slope of the SBS (after Côrrea, 1987).

widespread than the superficial distribution, demonstrating the strong influence of the lagoon as a source of fine sediments to the inner continental shelf.

Lenses of fluid mud were also evident during the 80^{ths}; however, their extension was better mapped after a fluid mud depositional event occurred in 1998, when stormy conditions during cold front passages between March and May, caused extensive deposits of fluidized mud on the shoreface, surf-zone and foreshore of Cassino beach. A total of 21 cores and 12 grab samples (Calliari et al., 2001) allowed mapping of the fluid mud over an area of 30 km^2. Although the isopach map indicated that the total amount of mud increased offshore, fluid mud thickness increased landward (Fig. 11). Cores located in areas with less than 7 m deep displayed mud contents of the order of 93% and an average density of 1230 kg m^{-3}. Sandy laminations (average thickness of 10 cm) with abrupt transitions were common along the cores located at water depths greater than 10 m.

Recent data obtained between 2004 and 2005 (Sperle et al., 2005; Calliari et al., 2007) using echobathymetry, seismic reflection and grab and core sampling allowed a detailed mapping of the fluid and more compacted mud between the isobaths of 5 and 14 m, an area where the deposits were not mapped yet (Fig. 12). The shallower cores showed superficial fluid mud interbeded with sand laminations. Cores located at higher depths displayed more compacted mud interbeded with sand with variable thickness (Fig. 13). During 2007, two bottom sampling and seismic campaigns, designed to map the shoreface at both, north and south of the inlet, allowed to detect a continuous fluid mud belt between the 10 and 12 m isobath, as far as 80 km north of the lagoon mouth. Fluid mud as shallow as 6 m, however, was only found immediately south of the inlet. Such fact suggests that fluid mud only is transported towards the beach when it occurs at depths less than 10 m, otherwise episodes of fluid mud deposition would occur all along the sampled shoreface and not only in the vicinities of the depocenter, as has been the case for the last hundred years.

Sedimentation rates based on the ^{210}Pb method from a core at water depth of 14 m produced an accumulation rate of 25 ± 0.09 mm yr^{-1}. This short-term accumulation rate spanning decades indicates a value almost three times of that found by Toldo et al., 2006 for the main lagoon, and is high compared with the accumulation rate of 2 mm yr^{-1} found by Tessler (2001) when analyzing cores from the inner and middle shelf of São Paulo state located approximately 1500 km to the north.

5.4. The influence of mud on Cassino beach

Since the Patos Lagoon Facies is more widespread southward of the inlet, fluid mud is found at shallow depths in front of Cassino beach. Stormy conditions associated with periodic cold front passages, which occur frequently on the area, can rework and transport the fluid mud from the shoreface to the surf-zone and foreshore of Cassino beach (Fig. 14). Such events have strong influence on the surf-zone processes, beach morphodynamics, shoreline changes and coastal hazards.

More than 15 years of beach profiling and monitoring indicate that based on the morphodynamic classification proposed by Wright and Short (1984), beaches along Cassino area are dissipative (Calliari and Klein, 1993). Beaches at this area are flat and wide, composed of very fine quartz sand and display multiple parallel bars. Variations in sediment properties (grain size and composition) at the beach (Figueiredo and Calliari, 2006), surf-zone and shoreface imply a beach differentiation up to 25 km south of the inlet. This lateral variability is identified by three sectors with distinct morphodynamic behavior. In the first 6 km (north sector), beach, surf-zone and shoreface are composed of very fine sands. For the next 9 km, which characterizes the central part of Cassino, beach sediments are fine while the shoreface displays muddy sediments mostly composed of clayey silt, silty clay and silty sand with fluid mud at shallow depths (6 m). In the last 10 km fine sand predominates and fluid mud occur at depths between 10 and 12 m.

The lateral heterogeneity along the shoreface and surf-zone of this beach causes visible differences in beach morphodynamic behavior, which is a response to the hydrodynamic conditions. Due the absence of mud in the shallow parts of the shoreface, which can substantially reduce the wave energy, beaches in the first and third sector display more mobility and are subject to higher set-up when compared to the central sector of Cassino beach.

Another effect caused by the muddier shoreface in front of the Cassino beach is the high rate of shoreline accretion on the order of 4.10 m yr^{-1} found at 10 km south of the inlet (Lélis and Calliari, 2006). Historical records show that before the inlet stabilization and fixation by two jetties (Fig. 5), alternated phases of shoreline accretion/erosion occurred in conjunction with the changeable morphology of the ebb tide located at the lagoon mouth. After the end of the jetties construction, which begun in 1911 and ended in 1915, the shoreline started to accrete at higher rate. The shoreline

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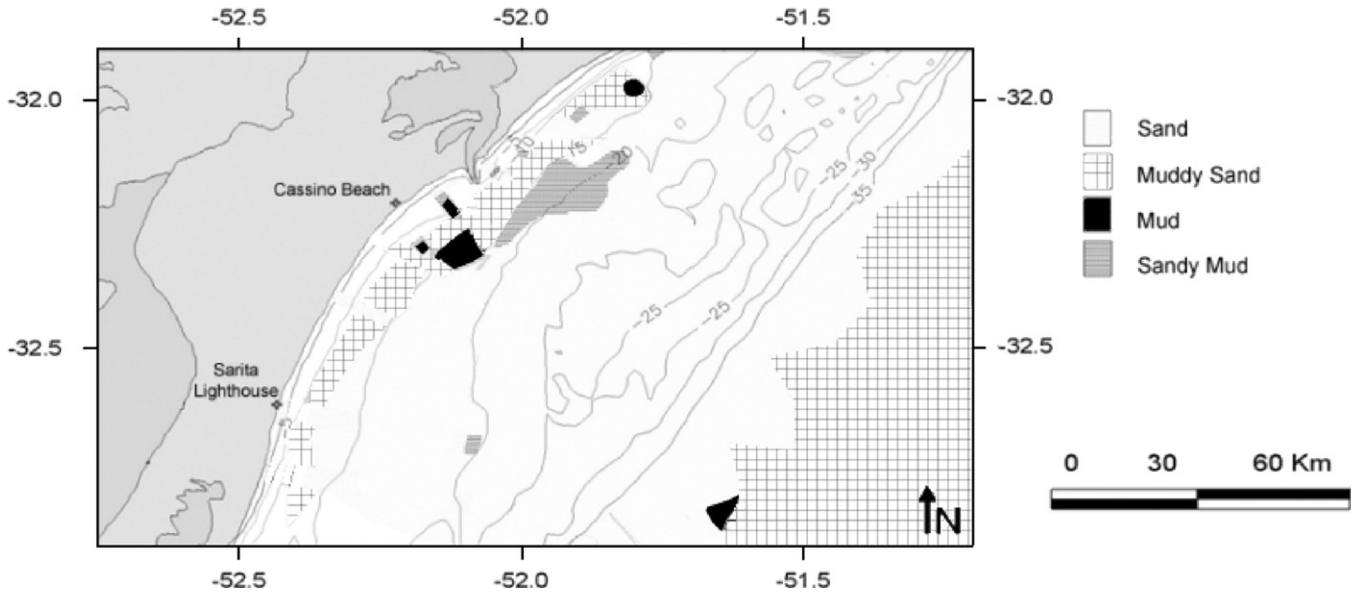

Fig. 9. Bottom types associated with the Patos Facies (after Calliari and Fachin, 1993).

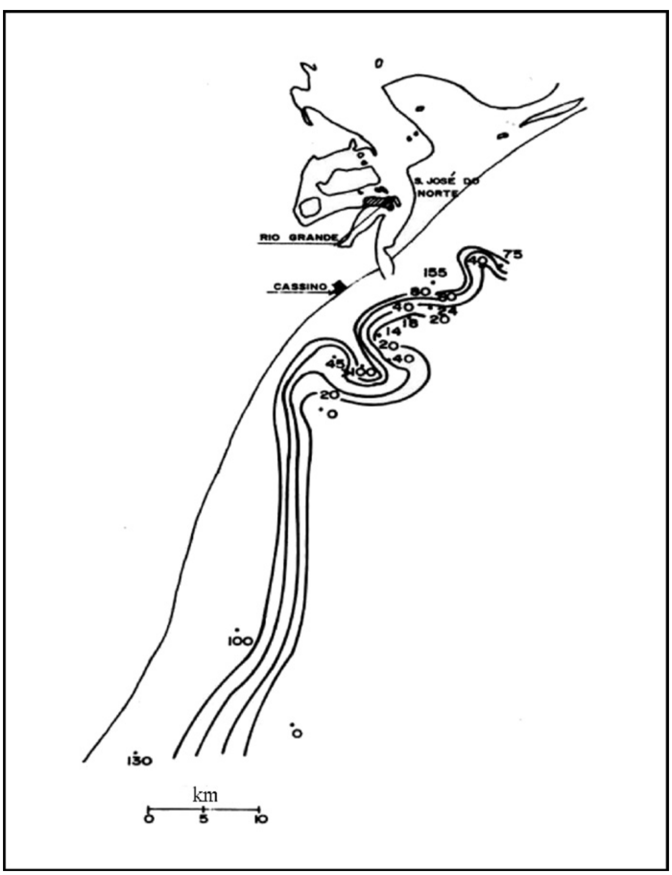

Fig. 10. Isopach map of total mud (sum of all the mud levels of the cores (after Calliari and Fachin, 1993).

accretion to the south of the inlet can be explained by the fact that the net littoral drift along the RS coastline is towards the northeast. However, the higher rate of accretion coincides with the depocenter of the deposit mapped on the basis of thickness of total mud content of the sediment cores (Fig. 10), as well as the recent fluid mud thickness map (Fig. 12). The stretch of beach adjacent to this area is also the place where mud deposits on the beach appear more frequently. These data strongly suggest that there is an association between the shoreface mud deposits and the high shoreline accretion rate found in this area. The higher frequency of mud deposition at the beach and shoreface can substantially reduce the local wave energy (Fig. 14) and act like a sink for the sediments, which arrive there by littoral drift. Shoreline accretion and stabilization due to the presence of fluid mud on the shoreface has also been reported (Wells and Roberts, 1980) for sectors of the chennier plain coast of southwestern Louisiana. Similar processes have been described by Wells and Coleman (1981) where waves transiting fluid mud bottoms along the coast of Surinam, South America were progressively attenuated nearshore without breaking. Similarly, Mathew et al. (1995) analyzed wave data from Kerala coast in India and found that wave height reduction as a result of wave propagation over fluid mud layers can reach 85%. According to these authors, the wave energy can be completely damped in a distance corresponding to a few wavelengths during the stage of mud banks formation along the southwest coast of India.

On at least two occasions the presence of fluid mud in the surf-zone during episodes of high wave energy represented a potential hazard for surfers. Due to lateral gradients in wave energy along the surf-zone alongshore currents transported surfers to areas where the surf-zone was isolated from the beach by fluid mud which had thickness of 1.5 m or even more. Such situation caused surfers to get trapped in the surf-zone under risk of life, since they suffered hypothermia due to long exposure in autumn temperatures waters. There was also some concern that they could be immobilized and buried by the fluid mud deposits. Fortunately locals used cables and body boards to come across the surf-zone to rescue them.

The presence of mud between the beach face and backshore represents an additional risk for car traffic. The fact that Cassino beach is typically a highly dissipative wide and very fine sandy beach with a high degree of compaction (close-grained) encourages a steady flow of vehicles. Any mud deposited across the beach is rapidly covered by a thin layer of sand as soon as the sea

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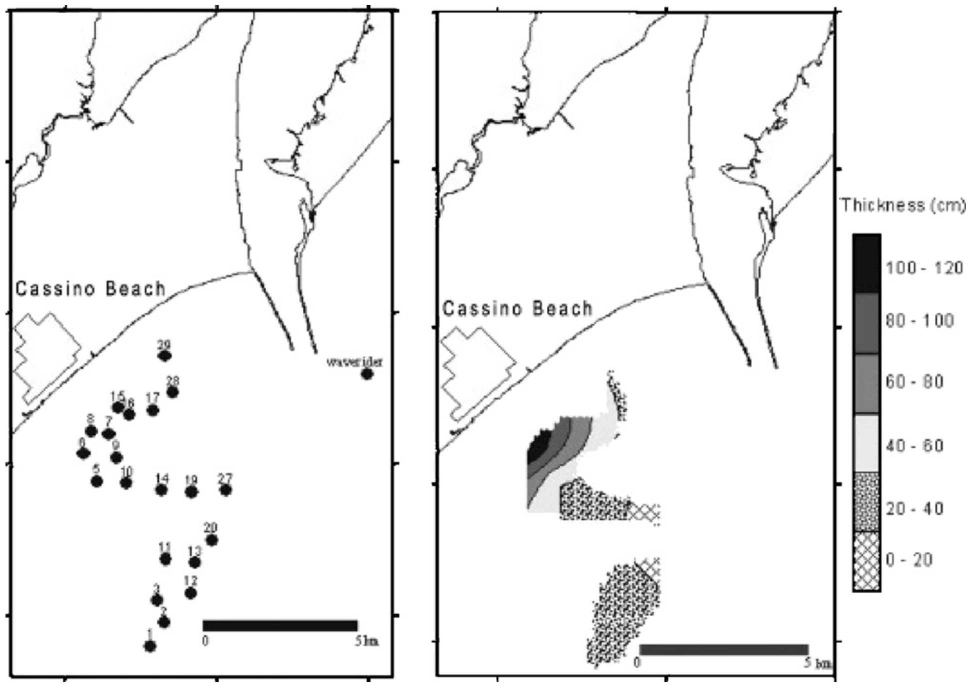

Fig. 11. Isopach map of fluid mud thickness during the 1998 mud deposition event.

level drops after the storms and acts as a trap for cars which get stuck in the mud sometimes reaching more than 0.3 m thick. It becomes impossible to remove the cars without the use of heavy machines.

6. Conclusions

The suspended matter in the Patos Lagoon is mainly composed of silt and clay, which respectively decreases and increases towards the estuarine part of the lagoon. Although the amount of suspended matter of the lagoon waters are not elevated enough to generate a hyperpicnal plume, the volumes discharged from the drainage basin, as well as the patterns of seasonal variations of rainfall and wind, make the estuary predominantly dominated by ebb flows. Thus, the amount of suspended matter that can be exported to the shoreface and inner shelf is large, and the most important advective offshore transport mechanism along the shoreface and inner shelf within the study area appears to be this ebb-dominated flow of turbid lagoon-estuarine waters. Active ebb-flow plumes of turbid water displaying both offshore and alongshore drift components have been well documented on the shelf by satellite imagery. Although the discharge of the La Plata River is high and has a marked influence on the seasonal salinity variation on the SBS, it is unlikely that its suspended load has a significant contribution to the modern Patos Lagoon Facies. The lack of fine sediments on the inner and outer continental shelf south of 32°30′S is a strong indication that the main, if not the only, source of fines has been the lagoon over a very long period of time.

The occurrence of a 4000 km^2 smooth bathymetric area between two sand ridges where the partial textural mapping indicates the predominance of fine sediments reinforce the role of the lagoon as a source of fine sediments to the inner shelf. The elongation of this area towards the south reflects the high frequency of the lagoon discharge under the influence of the northeast wind associated with the local and overall wind climate. This sedimentation pattern is corroborated by numerical models experiments and EOF analysis.

According to the coring data, obtained in 1984, the mud depocenter is located approximately 9 km southward of the estuarine mouth, 10 km offshore at 15 m water depth. Depocenter location is a reflex of the plume clockwise rotation due to the deceleration process after it leaves the lagoon entrance. Although the data from 2005 is related to fluid mud and from a restricted area, it indicates the same sedimentation pattern. The limits of the modern mud blanket southwards of the Patos Lagoon vary between 6 and 25 m. The position of the mud boundary is mainly a function of the interaction between mud supply and the erosional capacity of the prevailing flows, which in the shoreface is highly dependent of storm waves, while in the inner shelf is dominated by swell and wind-driven currents. As one of these factors prevails, the location of the fluid mud deposits can be expected to change. Cores from the depocenter and from near the outer boundary supports such variability. Layers of mud, presumably deposited at times of high sediment supply relative to the current strength, are layered with sand, silty sand and mixed layers with equal amounts of sand, silt and clay. Such pattern possibly represents periods of fast mud supply and the subsequent advance of the transgressive relict sand blanket over the recent mud deposits. The general increase in the thickness of the fluid mud layers detected in cores at shallow depths is an indication that mud is eroded from the shoreface during storms and the resulting fluid mud can be deposited nearshore and eventually reach the surf-zone and the beach. Although a continuous fluid mud belt occurs along the shoreface both to the north and south of the inlet, beach deposits only appears immediately to the south because fluid mud is only deposited at shallow depths around the depocenter. These ephemeral deposits can be eventually resuspended and pushed towards the beach by wave-generated storms.

The sedimentation rate in the Patos Facies may also be affected by anthropogenic effects, i.e. the dredging of fine sediments in the estuarine part of lagoon and dumping in the shoreface and inner

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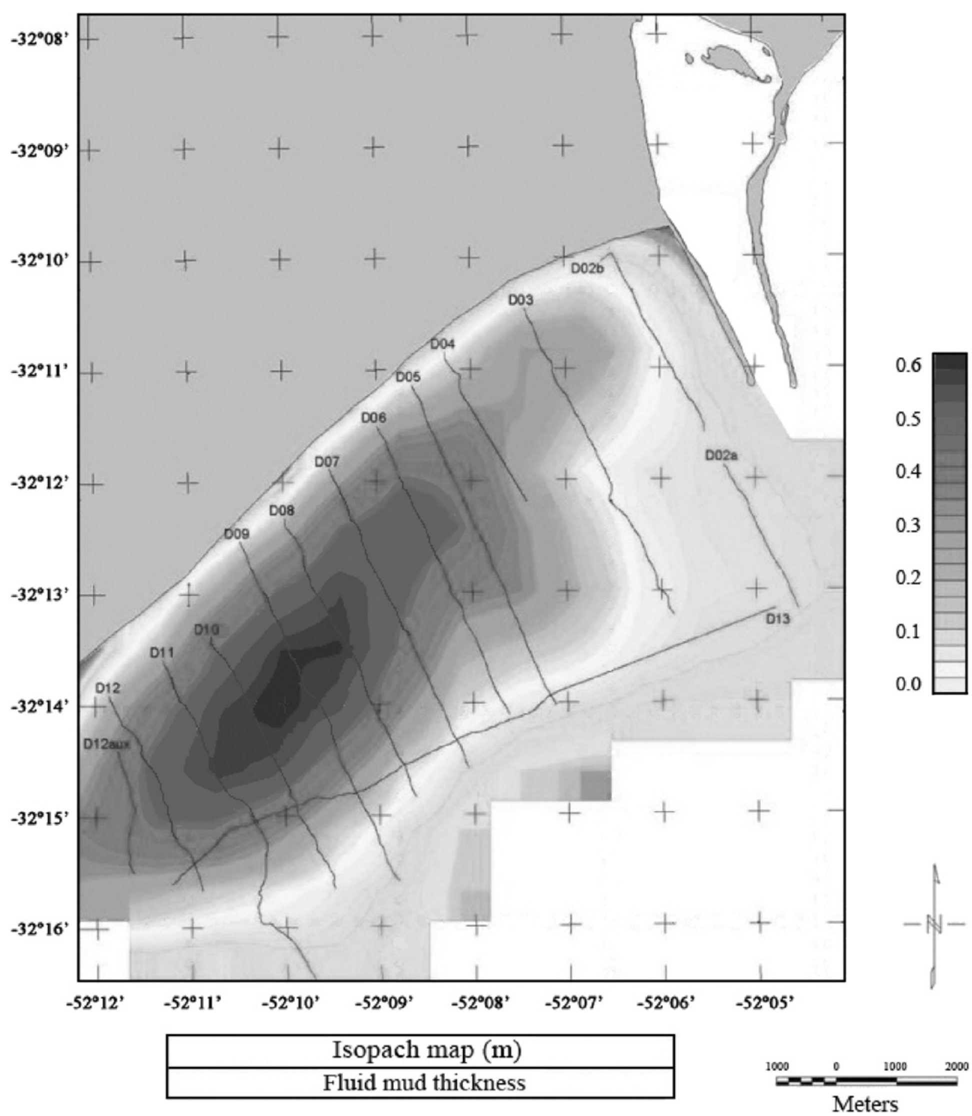

Fig. 12. Fluid mud thickness obtained from geophysical and coring data during 2004/2005 at the Cassino shoreface. (after Sperle et al., 2005; Calliari et al., 2007).

shelf. However, this contribution to the overall fine sediment balance is not known. We hypothesize, supported by the results of numerical model studies (Vinzon et al., 2008), that fine sediments exported from the Patos Lagoon during periods of northern winds accumulate in Cassino shoreface. Eventually, these sediments will contribute to the large amount of fine sediments found on the shoreface, but their amounts are small (estimated at about $4\,\mathrm{M\,ton\,yr^{-1}}$) compared to the existing amount of mud deposited since ancient times.

The well-mixed structure of many of these samples, though sometimes enriched with sandy laminations, is an indication of severe reworking, which can only be attributed to wave action. These fluid mud layers can be transported along the nearshore, and even uphill to the beach by wave-induced radiation stresses and possibly other hydrodynamic forcing. Deposited on the beach and immediate foreshore, its fluid state cannot be maintained and the mud is slowly re-eroded by waves and currents, cleaning up beach and foreshore.

We suggest few mechanisms to explain the occurrence of fluidized mud in the area. The waves can liquefy the mud, upon which the waves are damped by viscous dissipation within the fluid mud. The waves can also carry the fluid mud further onshore, even to the beach. An alternative explanation is that the historical mud deposit can be eroded during extreme storms to bring large volumes of mud into suspension, locally at the deposit. Conversely, high sediment concentrations generated by flushing of the lagoon could result in fluid mud depositing on the shoreface, perhaps over moderately long periods of time, which can be eventually moved on to the shore. However, because of the complex interaction between mud-induced wave damping and wave-induced mud transport, fluid mud attachment to the beach is highly unpredictable, as the mud transport seems to depend not only on the actual meteorological conditions, but also on previous events and mud characteristics.

Sporadic mud deposition events during storms create short-term morphodynamic differentiation. Beach sectors under the influence of mud in the sub-aerial beach, surf-zone and part of the shoreface display low mobility and low hydrodynamics when compared with sectors without mud. Mud deposition at the surf-zone and beach can represent a non-permanent potential coastal hazard for bathing, surfers and car traffic.

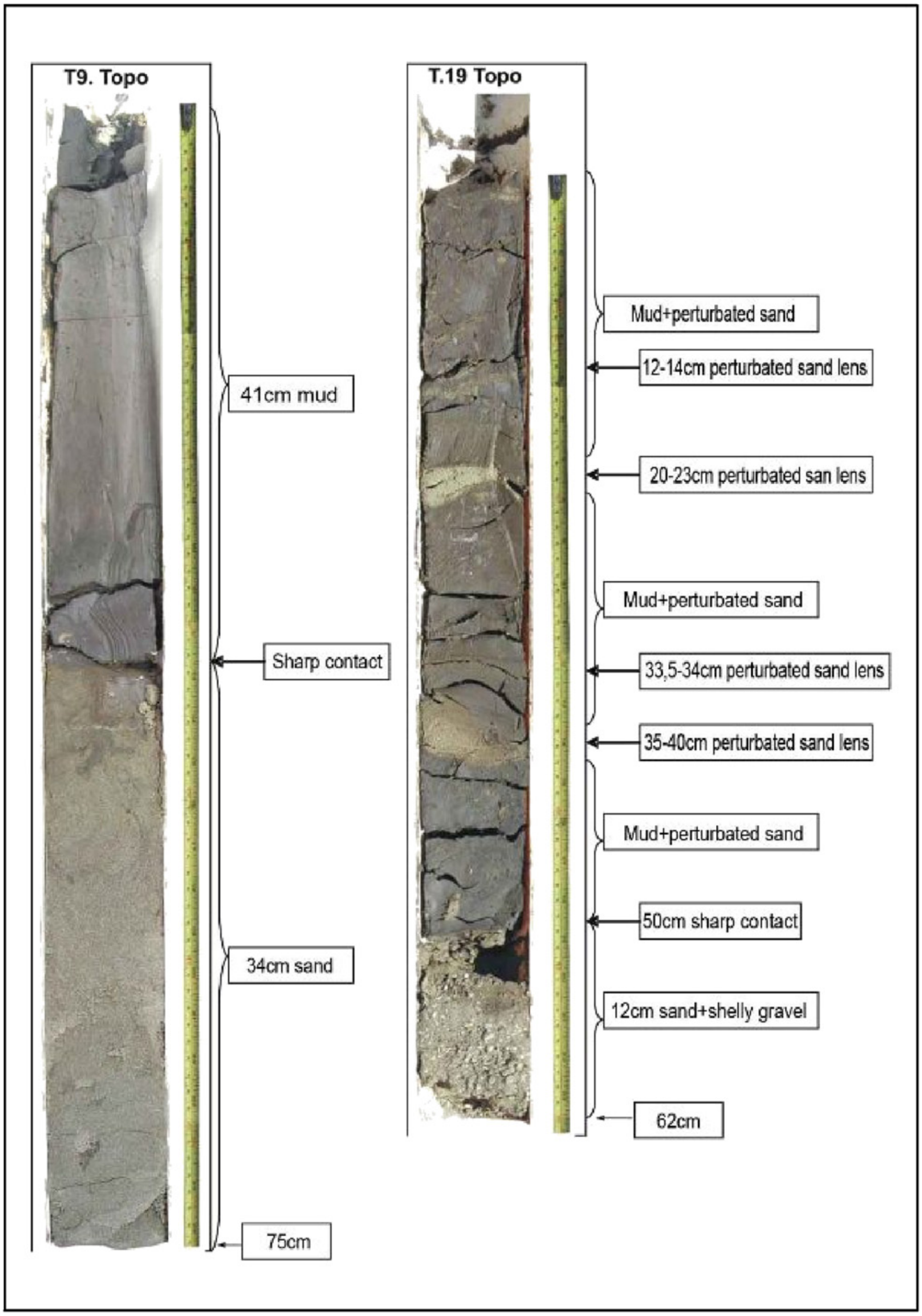

Fig. 13. Cores at 6 and 14 m of water depth.

Fig. 14. Wave attenuation and lateral variability in breaker height, surf-zone width and set up associated with presence of mud.

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