Role of halokinesis in controlling structural styles and sediment dispersal patterns in the Santos Basin, offshore SE Brazil

by

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I hereby declare that this thesis has been composed by myself and has resulted from my own research and work, except where contributions have been cited or acknowledged otherwise, and that it has not been previously submitted for a degree at this, or any other university.
for Barbara, naturally
All is flux, nothing stays still.
Heraclitus
Abstract

The Santos Basin, offshore SE Brazil, holds the major volume of rock salt in the Brazilian margin, showing a wide variety of salt-related structures, whose precise evolution needed to be understood. Large hydrocarbon reserves have recently been discovered in the basin, both in the pre- and post-salt sequences. Salt not only acts as a very efficient seal for reservoirs, but it also plays a significant role in creating structures that can either serve as migration fairways or traps for hydrocarbons. Another relevant implication is that salt diapirs and walls can also modify the temperature distribution and maturation history in sedimentary basins. A clear definition of the processes involved in salt movement and its influence on deformation and sedimentation of this area may contribute to improve the understanding of the local petroleum system and reduce exploration risks.

This project investigated the interplay of salt movement and sediment dispersal in the deep-water central and northern Santos Basin, based on the integration of a well-calibrated interpretation of 3-D and 2-D seismic reflection data and sequential cross-section restoration. The overall aim was to increase the understanding of halokinesis and sediment interplay in the study area and to produce a holistic model of the effects of halokinesis in the development of structural styles, evolution of bathymetry and controls on sediment dispersal. This model may be of practical application to the oil industry as a guideline to deep-water exploration strategies in passive margin salt-prone sedimentary basins.

Thin-skinned gravitational gliding and spreading, driven primarily by the flow of the ductile Aptian salt in response to differential load and thermal subsidence, accounts for the main deformation within the post-salt sequence in the Santos Basin. Salt flows basinwards, giving rise to an up-slope extensional domain, characterised by listric normal faults detached on the salt layer, and to a down-slope contractional domain, dominated by folds and salt withdrawal basins. During the Late Cretaceous and early Cenozoic, the basin received a very thick sedimentation, which had greatly influenced halokinesis. As a result, a large landward-dipping listric fault formed offshore, detached on the salt layer. The Cabo Frio Fault, one of the most conspicuous structures in the Santos Basin, roughly parallels the coastline and extends for almost 200 km along strike with heaves for the Albian.
sequence that locally surpass 60 km. Since the end of the Turonian, it has been active controlling the major depocentres in the area. The precise genesis of the Cabo Frio Fault has been under debate for it is sometimes stated that it could have been controlled by pre-salt structures. Evolved cross-section restoration validates interpretation and reveals structural and stratigraphic evolution, highlighting the active structures that control depocentres at each stage. The salt layer and the overburden deformation have created structures that may act as hydrocarbon traps, controlling the depocentres and defining hydrocarbon migration fairways. The restoration demonstrates that the Cabo Frio Fault has migrated offshore throughout its evolution and consequently its present-day position, sometimes aligned with a pre-salt fault, may not correspond to its original position. Care must be taken therefore when modelling the petroleum system. Post-salt structural traps, reservoirs, migration fairways and seals position relative to the pre-salt structures and source rocks has been changing and a promising present-day correspondence does not guarantee a favourable scenario at the time of hydrocarbon expulsion.

This work demonstrates that the Cabo Frio Fault could have been produced solely in response to the differential load imposed by the presence of a thick prograding wedge, independent of any structure affecting the base salt. The genesis of the salt withdrawal basin domain is also analysed here and we suggest that, like the Cabo Frio Fault, this domain may have resulted mainly from the differential load associated with basinward progradation of sediments. The concave geometry of the continental shelf, in response to confluent directions of incoming sediments, has imposed a convergent salt flow towards the centre of the basin, which accounts for a complex interference pattern of superposed folds with intervening mini-basins in the distal domain. A gravitational cell ("Ilha Grande") is proposed for the central domain of the study area, consisting of up-dip extension and down-dip contraction, with the Cabo Frio Fault in the transition zone between these two domains and the salt withdrawal basins in the contractional domain.
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Knowledge is in the end based on acknowledgement.
Ludwig Wittgenstein

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Chapter 1

Introduction

Due to its low permeability, low density, low strength and high thermal conductivity, salt highly influences the hydrocarbon prospectivity of a sedimentary basin. The existence of a particularly thick salt layer has significant implications for the evaluation of petroleum systems (e.g. Gulf of Mexico, Brazilian and Angolan margin etc.). The clear understanding of the kinematics and timing of salt movement and its implications for deformation and sedimentation is of ultimate importance to reduce the exploration risk involved in evaporite-bearing sedimentary basins. Salt not only acts as a very efficient seal for reservoirs, but it also plays a significant role in creating structures that can either serve as migration fairways or traps for hydrocarbons. Another relevant implication is that the presence of salt diapirs and ridges can also modify the temperature distribution and maturation history in sedimentary basins.

The Santos Basin, offshore Brazil, is home to a large variety of salt-related structures. It has received increased attention during the last few years, in view of the recent important oil and gas discoveries, both in the pre-salt and in the post-salt section.
1.1 Why the Santos Basin?

Three main reasons make relevant to carry out a detailed investigation on the geological evolution of the Santos Basin.

- Highly stimulating geological aspects are associated with the post-rift sequence of the basin, which include a wide variety of salt-related structures, whose precise evolution had previously been poorly documented and understood. The Santos Basin holds the major accumulation of rock salt in the Brazilian margin that has been ascribed to the Aptian. During the Late Cretaceous, it received a thick sedimentation which greatly influenced the salt movement. As a result, a very large landward-dipping counter-regional (antithetic) listric growth fault formed offshore, essentially parallel to the coastline. This outstanding feature, termed the Cabo Frio Fault, reaches almost 200 km long, and demonstrates horizontal displacements of more than 50 km. It has been active since the end of Turonian times, controlling the major sedimentary depocentres of the extensional realm. Cenozoic rift basins onshore are approximately parallel to this offshore conspicuous normal fault.

- Practical reasons have also been taken into account. The basin has a good 2-D seismic coverage, with reasonable well-log control, and new strategic 3-D seismic reflection data are available.

- Finally, the basin was chosen for economic reasons. Significant hydrocarbon reserves have recently been discovered in the Santos Basin. A good definition of the tectono-sedimentary evolution of this
area may contribute to the better understanding of the local petroleum system and thus reduce exploration risks.

1.2 Rationale

Halokinesis is an important controlling factor for the structural evolution during the drift phase of the Santos Basin. Analogue and numerical models have previously suggested that gravitational gliding over a ductile salt layer triggered by differential loading may account for the genesis of the post-salt structures in the Santos Basin (Cobbold & Szatmari 1991; Demercian et al. 1993; Szatmari et al. 1996; Ings et al. 2004; Guerra et al. 2005a, b; Guerra & Szatmari 2008). Nevertheless, it is often stated that the post-salt structures are controlled by pre-salt faulting, either reactivated or just forming pre-existing topography on the base salt, as suggested by the eventual vertical alignment of pre-salt and post-salt structures in present-day seismic sections (Mohriak et al. 1995; Milani et al. 2005).

Based on previous analogue modelling results and on the mechanical behaviour of evaporites, the approach adopted in this thesis has been to consider that the salt and the overburden have deformed primarily by gravitational gliding of a brittle sequence over a spreading ductile layer. The main driving mechanism to trigger halokinesis was the differential load caused by a thick Upper Cretaceous to lower Cenozoic prograding wedge deposited above the Aptian salt. This would respond for the structural styles and patterns of sedimentation during most of the post-salt stage. The deformation would be entirely thin-skinned, with no need to either involve reactivation of pre-salt structures, or base salt topography. Through seismic
interpretation and cross-section restoration, the consistency of this hypothesis with the data was evaluated in this thesis, in order to propose a model for the structural evolution of the study area and creation of accommodation space for sediments.

1.3 Purpose

This project investigates the interplay of salt tectonics and sediment dispersal in the deep-water central and northern Santos Basin, in order to propose a new, unifying tectono-sedimentary model depicting the evolution of the study area, using a substantive subsurface database comprising 3-D and 2-D seismic data. The overall aim of the study was to produce a model of the effects of halokinesis in the development of structural styles, evolution of bathymetry and controls on sediment dispersal. This model may be of practical application to the oil industry as a guideline to deep-water exploration strategies, not only in the Santos Basin, but also in other divergent margin sedimentary basins involving salt tectonics.

1.4 Data Set and Methods

This project was based on a well-calibrated interpretation of 3-D and 2-D seismic reflection data integrated with sequential cross-section restoration. The seismic data set comprised 34 wells and a regional seismic reflection 2-D survey that covered an area of 64,526 km², in which 10,551 km² could be
interpreted in more detail, based on two seismic reflection 3-D surveys (Fig.1.1).

Seismic interpretation focused on the salt and post-salt sequences was carried out in a large area of the basin, comprising the central and most of the northern domains.

Fig.1.1 Location of the study area (orange outline). Black lines are 2-D seismic sections. Purple and blue outlines are the bounding boxes of the 3-D seismic surveys. Seismic interpretation was carried out in this study covering an area of 64,526 km².

The seismic interpretation was based on seismic stratigraphic methods (Mitchum et al. 1977). Depositional sequences bounded by unconformities and/or abrupt downward shift of seismofacies were the basic units used to define the stratigraphic framework and the regional correlations. Stratigraphic correlations, based on gamma-ray and sonic logs, as well as on
lithological information provided by the well data, were used to guide the interpretation. Salt and overburden patterns were identified in the seismic sections and structural maps were produced for each interpreted horizon (in two-way travel time). Bathymetry and isochore maps (in metres) were obtained by operating the TWT structural maps with mean interval velocities for each layer.

From the interpreted seismic data, seven regional cross-sections covering the study area were chosen to be restored, in order to validate the interpretation and unravel the structural evolution of the study area as well as the salt-sediment interactions, timing of fault activity, opening of salt windows and depocentres migration.

1.4.1 Software Used

Conducted by a home and away student, working with Petrobras in Brazil, the project was planned to be partly developed in Scotland, partly in Brazil. The seismic reflection data were initially interpreted within the University of Edinburgh, using Schlumberger’s GeoFrame software, and both the restoration and the rest of the seismic interpretation were carried out in Brazil. In the Brazilian season, Petrobras made available a larger data set that comprised a second 3-D seismic survey and more well logs that complemented the original data set. The seismic interpretation was continued with Landmark’s SeisWorks and Paradigm’s 3-DCanvas. Gridding and mapping, as well as operating grids that resulted in the structural, bathymetric and isochore maps were performed with the proprietary
software Sigeo. Interval velocities obtained from the seismic sections were employed to depth-convert the seismic (see Chapter 4).

During the time of this project, a proprietary 2.5-D (pseudo 3-D) restoration computer programme, ReconMS, started to be developed by a consortium between Petrobras and TecGraf (computer graphics laboratory of the university PUC-Rio). ReconMS can perform simultaneous restorations on multiple cross-sections, therefore providing a three-dimensional idea of the structural evolution of the study area. The development of the ReconMS has profited from this research project, along which the software has been intensely tested and improved.

1.5 Thesis Outline

The first chapter addresses halokinetic processes in general and their specific control on the structural styles and sediment dispersal in the Santos Basin.

Chapter 2 outlines the concepts related to evaporite composition and deposition, as well as the major controls on halokinesis and the halokinetic-related structures.

Chapter 3 describes the regional geology, especially the aspects associated with tectonics, stratigraphy and magmatism in the Santos Basin and adjacent areas, making reference to previously published work. Some aspects of the interplay between halokinesis and sediment dispersal in the Santos Basin are illustrated with insights taken from analogue models previously run in the Petrobras Research Centre and the petroleum systems found in the basin are described.
Chapter 4 shows the main results of the seismic interpretation carried out in this research, highlighting the major salt-related structures and depocentre controls in the post-salt sequence. This interpretation emphasizes the Cabo Frio Fault and the salt withdrawal basins, the most conspicuous features in the drift phase of the study area.

Chapter 5 briefly describes the concepts involved in cross-section restoration and reports the results from the sequential restoration of seven regional cross-sections, performed with the computer programme ReconMS. Since these sections are georeferenced, the visualisation of the restoration stages gives a good three-dimensional idea of the temporal and spatial structural evolution of the study area. Timing of fault activity, depocentre control and opening of salt windows could be identified in all the seven cross-sections, as well as changes in position of the post-salt structures relative to the pre-salt structures.

In Chapter 6, the results are integrated and discussed. A model for the structural evolution of the study area is proposed.

In Chapter 7, the implications for the petroleum systems are considered and the key conclusions of this research are presented in Chapter 8.
Chapter 2

Evaporites and Halokinesis

Salt is born of the purest of parents: the sun and the sea.
Pythagoras

2.1 Introduction

The term halokinesis was first proposed by Trusheim (1957, 1960) to designate the processes involved in the autonomous flow of salt influenced by gravity. It consists in the movement of salt towards regions of lower pressure and tends to lead the system from gravity instability to equilibrium. The author established the difference between halokinesis and halotectonics, the latter controlled primarily by compressive tectonic forces. Throughout this work, the term halokinesis will be used in a broad sense, meaning salt movement, either driven by gravity or by tectonics. In most of the cases salt flow is controlled by gravity forces. Halokinesis is due both to the low density of rock salt, which does not increase with burial, and its low strength. Viscosity, temperature, thickness of the salt layer and mechanical properties of the overlying rocks, surface slope and relief of the substratum are the major controllers of salt movement. As a consequence of halokinesis, a variety of salt structures may form with different sizes and shapes, most commonly as salt-rollers, salt anticlines, pillows, diapirs and walls (Jackson & Talbot 1986). Associated with them, listric growth faults, rollovers, interdome structures (such as turtle structures), crestal collapse grabens, folds and thrusts may occur, deforming the overlying sedimentary sequence. The
presence of a thick evaporite layer containing ductile rocks (halite) is the key structuring factor for the post-Aptian sequence in the Santos Basin.

2.2 Evaporite Sedimentation

Evaporite rocks are generated by precipitation of salt minerals from concentrated brines, in environments where evaporation exceeds influx, be it derived from marine, fluvial, meteoric or subsurface water. Intense evaporation requires dry weather.

The most frequent evaporitic deposits result from confined marine basins, separated from the ocean by barriers to free circulation of waters. Another evaporitic environment, not as significant, consists of salt lakes fed by rivers that carry ions dissolved in water, whose composition depends on the source area (Ojeda 1982). The barrier efficiency controls the influx of sea water into the basin and the reflux of dense brines from the basin bottom to the open sea. Three models may explain the genesis of major evaporite basins (Fig.2.1): (a) shallow water – shallow basin: demands basin subsidence to produce accommodation space for thick evaporite deposits; rates of subsidence are fast enough to keep pace with evaporite deposition; (b) shallow waters – deep basin: evaporite deposits form far below the sea level, with episodic flooding over the barrier; (c) deep water – deep basin: reflux of less dense brines occur in the basin (Kendall 1984).

Composition of evaporites that precipitate from concentrated brines depends on sea water composition, which is kept relatively constant in the oceans, with no significant changes during the Phanerozoic (Jenyon 1986). The average composition of sea water presents the major ions (in decreasing concentration order): Cl\(^-\), Na\(^+\), SO\(_4\)\(^2-\), Mg\(^++\), Ca\(^++\), K\(^+\), HCO\(_3\)\(^-\) and Br\(^-\).
Fig. 2.1 Three models to explain the genesis of major basinwide evaporite deposits (Warren 1989).

The sequence of evaporite precipitation follows the solubility sequence of the brine components, starting with deposition of the less soluble minerals. The main components of evaporites are: dolomite, gypsum, anhydrite, halite and potassium and magnesium salts such as sylvite, carnallite and tachhydrite, deposited in this sequence.

An evaporite cycle resulting from increasing salinity by evaporation may be exemplified by a model based on the North Sea Basin (Fig. 2.2), where two Upper Permian evaporite sub-basins developed (cyclical evaporites of the Zechstein Supergroup). Each Z cycle consists of a basal clastic member, product of the first marine incursion, consecutively underlain by calcite (CaCO₃), dolomite (CaCO₃MgCO₃), anhydrite (CaSO₄), halite (NaCl) and
more soluble magnesium and potassium salts (Richter-Bernburg 1986; Jenyon 1986).

The same sequential vertical distribution is displayed horizontally, in which the less soluble evaporites deposited in regions close to the boundary with the open sea and the more soluble deposited in distant regions, near the continent. In a closed basin, whose communication with the ocean is temporarily interrupted, the concentration and selective precipitation of salts may result in a concentric distribution, in which the more soluble salts are located in the centre and the less soluble on the border of the basin.

Recent and ancient evaporites are amongst the sediments with the highest deposition rates. In favourable conditions, very thick deposits can form in a few years. More than 2 km of salt in the Messinian sequence of the Mediterranean accumulated in less than 2 million years (Kendall 1984).
& Buckner (1981) estimated deposition rates of 40 m in 1,000 years for the halite sequence of the Paradox Basin.

Basinwide evaporites constitute the sedimentary fill in many ancient evaporite basins, such as the Upper Permian Zechstein Basin of Europe, the Permian Delaware Basin of USA and the Late Miocene Messinian sub-basins in the Mediterranean (Fig. 2.3). They have no modern analogue in scale and diversity and are often described as "saline giants" (Fig. 2.4). Most ancient evaporite deposits have thicknesses and horizontal extents that are two to three orders of magnitude greater than those of Quaternary evaporites.

In this study, attention is focused upon Aptian evaporite marine basins in which deposition occurred in shallow or deep waters along the South Atlantic passive continental margin of Brazil.

Fig. 2.3. Location and age of some of major evaporite deposits (Tucker 1991). The Gulf of Kara Bogaz, on the eastern side of the Caspian Sea, is the nearest modern analogue to a barred basin.
Fig.2.4 Comparative dimensions of ancient and modern evaporitic basins. The large area in red with black outline corresponds to the Aptian evaporites in the Santos and Campos basins (modified from Warren 1989).

2.3 Mechanical Properties of Evaporites

Halite (NaCl) is the prevailing salt in evaporite deposits. Its rheology differs from that presented by the great majority of sedimentary rocks, and constitutes the major controlling agent of the post-rift structuring in the basins of the Brazilian continental margin, as well as in the other evaporite-bearing passive margin basins worldwide.

Halite constitutes a polycrystalline aggregate with very low permeability that remains incompressible during burial, maintaining its density constant (approximately 2.16 g/cm$^3$), even in the presence of anhydrite or other impurities disseminated in the salt interval (Fig.2.5).
Fig. 2.5 Relation between bulk density and depth in salt and associated terrigenous clastic rocks in the US Gulf Coast (Jackson & Talbot 1986). Near the surface, salt is denser than its surrounding strata, but once it is buried beneath 1,000 m of sedimentary cover, the density contrast gets negative.

The resulting density inversion (compared to surrounding rocks that are subject to compaction) establishes a gravity instability that may compel salt to rise through the denser cover until it reaches its level of neutral buoyancy. However, the influence of density contrast in salt tectonics is limited by the strength and brittle nature of the overlying strata. Salt behaves as a pressurised fluid and differential fluid pressure drives salt flow (Vendeville & Jackson 1992a).

Due to its relatively low viscosity, halite changes from elastic to plastic stage in very low temperatures and differential stresses, rather than the other rocks and the less soluble evaporites, such as anhydrite, whose elastoplastic limit is
only reached under high temperature and pressure. These rocks have
dominantly brittle mechanical behaviour, whereas halite and the more
soluble salts have ductile behaviour and deform plastically and faster than
the majority of rocks (Fig.2.6). Rock salt constitutes a perfect detachment
surface for the faulted overlying sequence; it forms a weak layer of constant
strength under a brittle sedimentary layer whose strength increases with
depth (http://www.colorado.edu/GeolSci/courses/GEOL3120, in May 2007).

![Fig.2.6 Strength of various rock types in both tension and compression (Jackson & Vendeville 1994). Salt is much weaker than other lithologies under both tension and compression. Even overpressured shale almost always has more strength than salt. Wet salt is a viscous material and falls on the axis of zero strength.](image)

The plastic deformation of a polycrystalline aggregate of halite happens
through intra-granular movements accompanied by diffusion, pressure
solution and recrystallisation, resulting in the same deformation pattern for
all the individual crystals, for any applied stress field (Jenyon 1986). The
presence of water in the polycrystalline aggregate and the increase of
temperature and confining pressure are some of the factors that increase
ductility of materials, salt (halite) included, allowing for ductile
deformation under low differential stresses. The higher these stresses, the
faster the deformation (Fig.2.7).

Fig.2.7 Steady-state strain rate (solid lines) as a function of temperature (depth), under three
levels of differential stress. Dashed lines indicate extreme geothermal gradients for the Gulf
Coast. Strain rates depend critically on the differential stress and thermal gradient (Carter &
Hansen 1983).

Jackson & Talbot (1986) mention halite deformation rates ranging from $10^{-8}/s$
to $10^{-16}/s$, reflecting a great diversity in flow conditions. The fastest
deformation rates ($10^{-8}/s$) were recorded in bore holes closure and salt
glaciers and the slowest deformation rates ($10^{-19}$/s) were recorded in gravity-driven diapirs growth. Flow rates up to 15 m/y were related in exposed diapirs in Iran (Talbot et al. 2000).

2.4 Halokinesis: Mechanisms and Triggering Agents

Salt deforms as a viscous material having negligible ultimate strength, and it can therefore flow when it is subjected to minimal shear stresses (Urai et al. 1986; Spiers et al. 1990; Weijermars et al. 1993). The overall rheological properties of salt are effectively constant through time (Rowan et al. 1993).

The initiation of the rising movement of salt is controlled by heterogeneities in the potentially mobile layer or even in its overburden (Jackson & Galloway 1984) and may result from:

- change in the salt layer thickness;
- change in the overburden thickness;
- local change in the overburden density.

Such changes may be caused by a variety of factors that include differential sedimentation, tilting, faults on the substratum reactivated after salt deposition, unevenness of the substratum, regional tectonics etc.

Jackson and Talbot (1986) identified six mechanisms that may induce halokinesis: (i) buoyancy, (ii) differential load; (iii) gravity spreading, (iv) thermal convection, (v) extension, and (vi) compression (Fig.2.8). Gravity gliding was included by Cobbold & Szatmari (1991) as an important halokinetic mechanism (Fig.2.9).
The major mechanisms responsible for causing halokinesis are: gravity spreading, buoyancy and differential loading (Jackson & Galloway 1984; Jackson & Talbot 1986; Jenyon 1986; Rowan 1995). Also mentioned is thermal convection, which has not yet been satisfactorily proved in nature, remaining as a mathematical possibility.

Fig. 2.8 Principal mechanisms of halokinesis. All types can combine. P represents the lithostatic pressure at a point (based on overburden density and thickness) and \(\rho\) represents the mean bulk density (Jackson & Talbot 1986).
2.4.1 Buoyancy

Clastic sediments overlying salt undergo compaction with burial and salt remains incompressible. At a certain burial depth that may vary according to the lithology involved, the sedimentary cover becomes denser (due to a combination of compaction and diagenesis) than the underlying salt layer, whose density remains relatively constant. In response to temperature increase, halite actually becomes less dense with burial depth, whereas other sediments get compacted and denser (Fig.2.8A). This density inversion constitutes the gravitational instability necessary to the buoyant rise of salt.

The differential stresses generated in this process are very small and insufficient, in most cases, to trigger the rise of salt, as demonstrated by Jackson & Talbot (1986), based upon Ramberg (1981) and Carter & Hansen (1983). Those authors concluded that it is necessary an initial topography of about 150 m on top of the salt layer for the buoyancy mechanism to become effective. Moreover, the strength of the rocks overlying salt may represent a barrier to buoyancy.

2.4.2 Differential Loading

In response to a differential distribution of loads above it, a salt layer will tend to flow towards less loaded regions (Fig.2.8B).

The differential loads may be caused initially by small, random, irregularities on the top of the salt layer, as well as lateral changes in thickness, density and strength of the overburden. They are usually related to changes in
sedimentation rates or in lateral facies (alluvial fans, turbidites, volcanic cones, thrusts, etc.).

Rising salt structures can influence the depositional facies of their overburden. Depocentres established over areas of salt withdrawal increase salt supply to the relative rising structure (Jackson & Talbot 1986). This mechanism is considered more effective in forming halokinetic structures than the buoyancy driven only by density inversion. Buoyancy would act as an additional mechanism, amplifying the process.

It must be highlighted that the halokinetic process is syn-depositional, initiating concomitantly with sedimentation and not after the deposition of a thick sedimentary cover.

2.4.3 Gravity Spreading

Gravity can dissipate relief in the top of any salt body that is above its level of neutral buoyancy (Fig.2.8C). Salt on the surface becomes unstable and flows by gravity spreading down slopes as low as 3°. In Iran, salt is driven above its level of neutral buoyancy, forming extrusive domes that spread under their own weight (Jackson & Talbot 1986).

2.4.4 Thermal Convection

This mechanism is based on the high thermal conductivity and high thermal expansibility of the salt. In a thick salt layer a shallow zone is heated by the sun with daily, seasonal and secular temperature fluctuations and a deeper zone is continually warmed by the Earth's heat flow. Both sources of heat induce thermal gradients that cause density gradients, with light, hot, salt
being underlain by dense, cold, salt. This could promote thermal convection and increase the rates of the heat flow upwards through the salt (Fig. 2.8D). Halokinesis by thermal convection is still considered a hypothesis, not yet convincingly demonstrated in nature (Jackson & Talbot 1986).

2.4.5 Contraction

Salt acts as a detachment layer in many fold-and-thrust belts (Fig. 2.8E) such as the Pyrenees (Sans & Vergés 1992), Albanides-Hellenides (Velaj et al. 1999; Underhill 1988) and the Alps (Sommaruga 1999). Lithostatic pressure can retard or augment the shortening with normal (stable contraction) or inverted (unstable contraction) density stratification, respectively (Jackson & Talbot 1986).

2.4.6 Extension

Salt layer forms a decoupling zone below listric normal faults in extending overburden. The normal faults detach on the salt layer and salt tends to rise in their footwall, as asymmetric anticlines (Fig. 2.8F). Lithostatic pressures can retard or augment this flow depending on whether a density inversion is absent (stable extension) or present (unstable extension), respectively (Jackson & Talbot 1986). Extension may result from rifting or from thin-skinned gravity gliding or spreading.
2.4.7 Gravity Gliding

This mechanism consists in the gravitational sliding of salt over a tilted substratum (Fig.2.9). When the shear component of the gravity force overcomes the internal cohesion of salt and the friction between the salt layer and the detachment surface, salt flows down-dip turning the overburden rocks unstable (except when the sedimentary cover is too thick in the deeper basin, reversing the sense of salt flow).

![Fig.2.9 Gravity gliding.](image)

This is a very important process in passive margin basins, in which the tilting of the base salt is promoted by thermal subsidence, in the post rift stage, flexural subsidence or episodes of tectonic continental uplift (Rowan et al. 2004). Salt tends to flow basinwards, inducing extensive stresses in the overlying rocks, causing them to fracture and fault, since they are not able to deform in a ductile way. The faulted blocks translate down a gentle slope. The top of the evaporite layer acts as a detachment surface for the competent overlying rocks that will move along listric faults produced during salt migration. These faults are commonly coeval to sedimentation (growth
Salt-rollers usually form in their footwall, due to the escape of salt towards less-loaded regions. Salt movement causes intense deformation in the overlying rocks. Regions of accelerating flow (tilted substratum) favour the formation of extensional features, like listric normal faults, rollovers, turtle structures etc. Zones of decreasing gradient of the salt base, or presence of any barrier, promote flow deceleration and thickening of the salt layer, establishing a contractional regime in which folds, reverse faults, salt pillows and diapirs are common (Fig.2.10).

Fig.2.10 Differential loading and tilting are frequent mechanisms for gravity-driven halokinesis in passive margin basins (modified from Szatmari & Demercian 1993). Both mechanisms promote up-dip extension and down-dip contraction. Salt layer thins up-dip and thickens down-dip.

Gravity gliding in passive margins with shelf-slope physiography occurs along the greatest bathymetric gradient. Depending on the morphology of the margin (re-entrant, salient or straight), the gliding will be convergent, divergent or parallel, respectively (Cobbold & Szatmari 1991). In a straight
margin, salt flow is parallel. In a concave margin, salt flow is convergent and results in along-strike contraction, creating radial folds, aligned with the bathymetric gradient. In a convex margin, salt flow is divergent, resulting in along-strike extension Rouby et al. 1993) that pulls apart the overlying sediments and creates radial normal faults (Fig. 2.11).

Rowan et al. (2004) discuss the dynamics and distinctions between gravity gliding and spreading and conclude that a combination of them consists in the most common mechanism in passive margin basins (Fig. 2.12).

![Fig.2.11 Flow patterns in a passive margin basin and schematic distribution of the resultant stresses (modified from Demercian 1996). Parallel or radial gravity gliding depend on the paths followed by material particles. Black arrows indicate tangential velocity of a particle; white arrows indicate stresses due to particle flow.](Image)
Fig. 2.12 Gravity-driven deformation: (a) gravity gliding, in which a rigid block slides down a detachment; (b) gravity spreading, in which a rock mass distorts under its own weight by vertical collapse and lateral spreading; and (c) mixed-mode deformation. Shaded areas are the final stages and arrows show material movement vectors (Rowan et al. 2004).

Vendeville (2005) and Gaullier & Vendeville (2005), based on analogue modelling, discuss gravity spreading due to sediment progradation over a salt layer. Because gravity spreading does not require a seaward basal slope, it is the main process driving long-lived halokinesis after thermal subsidence has ceased. Spreading causes an overall seaward translation of the prograding wedge, causing distal shortening and proximal extension (Fig. 2.13). Even with no density inversion, diapirism can occur during spreading. The authors highlight the intricate mutual influence between sedimentation and structure at different scale. The bathymetric relief resultant from salt and overburden deformation can control the location of channels or form small intra-slope basins, hence controlling facies, thickness and dispersal pattern of syn-tectonic sediments.
As an example, the Perdido fold belt, in the deep waters of the northwestern Gulf of Mexico, marks the basinward margin of a complex, linked system of gravitational spreading above salt, comprised of extensional fault systems, salt canopies and welds, and contractional folds. Up-dip Paleogene sedimentary loading and associated extension were accommodated down-dip primarily by salt canopy extrusion (Peel et al. 1995; Trudgill et al. 1999; Rowan et al. 2000). Other toe of slope complexes, such as the Nile deep-sea fan (Gaullier et al. 2000) also comprise up-dip extension and down-dip contraction that a result from thin-skinned deformation driven by gravity gliding and spreading of a brittle layer over a ductile salt layer.

Thin-skinned gravitational flow is not only influenced by the shelf-slope physiography. It appears that the basal discontinuities, such as faults or structural highs, in the base salt may also exert influence on the flow.
2.5 Halokinetic Structures

The evaporites are originally deposited with stratiform character and horizontal top. Salt movement in response to some gravitational instability, allied with strong feedback due to buoyancy, tends to create various types of salt structures that may be concordant or discordant in relation to the overlying sequence (Fig.2.14).

![Fig.2.14 Major halokinetic structures found in the Gulf of Mexico (Jackson & Talbot 1986).](image)

The concordant structures form in the initial stage of evolution and present low amplitudes. The most common ones are:

- Salt anticlines, long, with nearly symmetric transversal section, planar base and concave top (frequently evolve to salt walls);

- Salt pillows, with dome shape, circular or slightly elliptical horizontal section and sub-planar base;
The discordant structures may result from extension or compression and buoyancy. Among the discordant extensive structures, stand out:

- Salt-rollings, small and strongly asymmetric salt anticlines, associated with listric normal faults;

- Salt walls or ridges that correspond to a more evolved stage of the salt anticlines, being more frequent in the deeper parts of the basin.

The discordant compressive halokinetic structures display a large variety of shapes and high amplitudes. The predominant ones are:

- Diapirs, stocks, columns of salt with many different shapes, whose tops may spread laterally forming overhangs and bulbs, some of them showing the upper part detached from the mother layer (rootless salt bodies). The most mature ones tend to show vertical flanks;

- Salt nappes in which great bodies of salt thrust the stratigraphic sequence along low angle faults;

- Massive salt walls, complex with deep roots and dome or pillow-shaped bulges on their crests.

Salt extrusions may occur in subaereal or subaquatic conditions and have been described in arid environments such as the Zagros Mountains, in Iran (Talbot et al. 2000; Talbot & Aftabi 2004; Letouzey & Sherkati 2004). These structures are rare and present various shapes, sometimes producing bulges and sometimes flowing as salt glaciers (namakiers) over valleys and planar surfaces. The extrusions represent the final stage of halokinesis, for they promote salt dissolution. If they occur very close to the surface, they may form a lateral salt sheet, which are identified in large areas in the northern Gulf of Mexico.
With the continuation of the halokinetic process, small structures evolve to larger ones and may turn into diapirs, that is, they may pierce the overlying rocks (Fig.2.15). Evolutive trends among these structures are common, defining halokinetic provinces. According to Jackson & Talbot (1986), strain rates for halokinesis can vary over 8 orders of magnitude from $10^4$/s to $10^{-10}$/s.

**Fig.2.15** The first salt diapir described in the geologic literature (Ville 1856). The white line indicates the southern contact of the 100-m-high "Ran el Melah" (Rocher de Sel de Djelfa), a 1.4-km-wide plug of Triassic salt in the northern fringe of the Saharan Atlas range, Algeria (after Jackson 1995).

Analogue models indicate that the thickness of the salt layer considerably influences halokinetic structuring. The higher the initial thickness of the salt layer, the fewer halokinetic structures will develop, and they will present higher amplitude and wavelength. The same effect is observed in the overburden, whose thickness is directly related to the amplitude and distance between salt structures (Ramberg 1981).

An evolutive model for diapirism under regional extension (Fig.2.16) was proposed by Vendeville & Jackson (1992a). A diapir can evolve from one stage to another, depending on the balance of regional tectonic forces and sediment progradation rates.
In a first stage ("reactive"), the extended cover thins forming grabens and half-grabens that favour the slow upward movement of the pressurised salt layer. Continued sedimentation filling the grabens thickens the layers adjacent to the diapir and tends to keep diapirism in the reactive stage.

The second stage ("active") does not depend on regional extension, for it is caused by fluid pressure at the crest of the diapir, enough to open a pathway through the thinned sedimentary cover. The thinning results from erosion or extension and is followed by fast intrusion of the diapir sourced by a pressurised underlying salt layer. Active diapirism is also known as upbuilding (Seni & Jackson 1983a, b; Jackson & Talbot 1991).

The third stage represents the "passive" diapir intrusion when it reaches or get very close to the surface. This is equivalent to the old concept of "downbuilding" (Barton 1933), where a diapir keeps its crest practically at the sea floor, as the surrounding strata subside into the source layer. The diapir at this stage is in equilibrium.

In the absence of further sedimentation, the passive diapir stops rising and may extrude or widen under regional extension. Under relative low sedimentation rates, salt may extrude, widening upwards; under high sedimentation rates, salt is onlapped by sediments and thins upwards. If the source salt layer is depleted or the flow is buttressed, the diapir may subside and the regional extensional, once responsible for its creation, will promote its extinction.
Fig. 2.16 Model of diapir initiation and growth during extension: (i) overburden faults and thins creating a reactive diapir; (ii) once the diapir is tall enough and the overburden is thin and weak enough, the diapir will actively rise through the overburden; and (iii) near the surface, the diapir will grow passively as surrounding mini-basins subside and displace the salt (Vendeville & Jackson 1992a).

Deformation inside the evaporite sequences commonly occur as folds, thrusts, normal faults and diapirs (figures 2.17 to 2.20).

Fig. 2.17 Deformation inside the evaporite sequence (sylvite and halite) in the Taquari-Vassouras Potash Mine, NE Brazil. Halite recrystallises along the fault plane. September 2006.
Fig. 2.18 Folded and thrust evaporite sequence (sylvite and halite) in the Taquari-Vassouras Potash Mine, NE Brazil (Machado & Szatmari 2008).

Fig. 2.19 Small-scale diapirs of sylvite rising through laminated halite (deformation inside the evaporite sequence), in the Taquari-Vassouras Potash Mine, NE Brazil. September 2006.
2.6 Halokinetic-related Structures

2.6.1 Faults and Folds

Faults commonly occur at the base, limbs or crest of halokinetic structures, and may constitute cause or effect for salt movement. Whilst basal faults may sometimes trigger and orientate halokinetic structuring in the basin, the others are consequence of salt movement. Basement faulting and diapirism are not necessarily coupled at all. Salt tends to decouple the deformation, so that the structural styles above and below the ductile layer can be quite different and independent. The overburden is usually draped over sub-salt faults, with salt separating and accommodating the different styles. The degree of decoupling can vary significantly over time.
Salt diapirs (or ridges) and supra-salt faults are not always located immediately above the sub-salt active faults, sometimes they form independently. However, both residual topography in the base salt (due to inactive basal faults) and reactivation of sub-salt faults may have some control on supra-salt deformation, by promoting "isostatic" unbalance between the hangingwall and the footwall blocks, as suggested by analogue modelling (Rizzo 1987; Gaullier et al. 1993; Koyi et al. 1993; Schultz-Ela & Jackson 1996, Withjack & Callaway 2000). The salt layer tends to flow more rapidly above the hangingwall block, where it is thicker. In some places, diapirs form systematically above or near the sub-salt faults, or the overburden is force-folded above the sub-salt fault into a monocline whose shape reflects the offset along the underlying basement blocks. The coupling or decoupling of supra- and sub-salt deformation are mainly influenced by the thickness of the salt, the thickness and strength of the overburden, the rate of fault slip, and the magnitude of the fault (Fig.2.21). A thick salt layer diffuses localised sub-salt slip by laterally flowing from the footwall to the hangingwall, so stresses and strains are not transmitted upwards through salt and into the overburden. Once the salt layer above the footwall has been thinned or depleted, salt can no longer flow fast enough to entirely accommodate sub-salt faulting. Hence, the fault propagates upwards as a broad forced fold whose upper hinge is affected by normal faults that accommodate further extension. Experimental and conceptual models indicate that decoupling is favoured by initially thick salt, slow sub-salt fault slip and slow or absent syntectonic sedimentation. Conversely, initially thin salt, very rapid sub-salt faulting, rapid syntectonic sedimentation, or all three promote coupled deformation, with diapirs or overburden faulting above or near the sub-salt fault (Schultz-Ela & Jackson 1996).
Fig. 2.21 Influence of active sub-salt normal faults on the growth and location of supra-salt normal faults: (a) thick salt layer and low gliding; (b) intermediate; (c) thin salt layer and rapid gliding. Close correlation lies where the salt layer is thinner. S: simple graben; D: drape graben; R: reactive graben; HP: horst and negative pop-up (Schultz-Ela & Jackson 1996).

In the case of strike-slip faults being reactivated, the salt layer cannot sustain shear stresses due to its low shear strength, hindering their propagation to the post-salt sequence. In spite of this damping effect, the halokinetic structures tend to orientate themselves along the trends of those faults, possibly by some “isostatic” instability similar to those previously discussed (Jenyon 1986).

The most frequent faults formed on the supra-salt sedimentary sequence are listric, normal, syn-depositional faults that result from salt migration. The footwall block constitutes a less loaded region, to where salt tends to move isostatically, creating a salt-roller. The main types are proximal basinward- and landward-dipping listric faults, both detaching on the salt layer. They
may be compensated by distal contraction expressed by thrusts and detachment folds (figures 2.22 and 2.23).

Fig.2.22 Illustration showing late distal contraction (left-hand side) and continuing proximal extension (right-hand side) as the wedge front reaches the distal salt pinch-out (Vendeville 2005).

Fig.2.23 Three settings for contractional salt tectonics: (a) foreland regions of collisional mountain belts; (b) deepwater regions of passive margins that fail gravitationally (gravity-induced compression); and (c) inverted rift basins (Letouzey et al. 1995).

Faults on the crest of halokinetic structures result from the extensive stresses produced during the rise of the salt layer (Fig.2.24). They usually occur in
pairs at the crest of elongated diapirs and walls, or form a radial concentric pattern over diapirs and pillows with circular basal section. If they reach the salt, these faults that commonly affect the seabed may permit the infiltration of water and consequent dissolution of the salt, causing the overburden collapse (Alsop 1996).

Fig.2.24 Schematic fault and fracture pattern produced in overburden layers above a salt diapir, based on observations of analogue models and seismic interpretation (Davison et al. 1996).

2.6.2 Raft Tectonics

Originally, the term raft tectonics was employed to the Congo and Kwanza basins (Burollet 1975; Duval et al. 1992; Jackson et al. 1998). It represents the extreme thin-skinned extension over a décollement of salt. Rafts are allochthonous fault blocks no longer in mutual contact. Burollet (1975) postulated that the Angolan margin extended by gravity gliding over a thin layer of salt. As it stretched, the overburden broke into diverging raft-like blocks separated by widening grabens or half-grabens (Fig.2.25). Sediments
rapidly filled the fault-bounded depocentres. A salt scar formed along the boundary between a gliding block and the adjoining fault-bounded depocentre was also mentioned. Raft tectonics is thin skinned and involves large displacements over the detachment layer. Complex structures result from interactions between rafts and underlying mobile salt during gravity gliding and spreading (Fig.2.26). Raft tectonics has been described not only in the Kwanza Basin, in the western divergent continental margin of Africa, but also in the South American eastern margin (Campos and Espírito Santo basins), Nordkapp Basin in Norway, eastern Mediterranean Basin, the Gulf of Mexico, the Red Sea (Gugliemo et al. 1997), and the Central North Sea (Bishop et al. 1995; Penge et al. 1993), amongst other areas.

![Fig.2.25 Illustration of extensional raft tectonics and gravity gliding (Duval et al. 1992).](image-url)
2.6.3 Salt Welds

Salt welds are salt scars in which there used to be a salt layer that was completely evacuated, hence connecting the sub-salt with the supra-salt sequences. Salt welds can also form vertically, in the place previously occupied by a salt diapir or ridge that was later on evacuated (Fig.2.27). Salt is displaced from beneath the depocentre and flows laterally into flanking areas, creating bathymetric highs. The adjacent low areas receive further sediments, which increases the differential pressure, driving further salt withdrawal and flow into the flanking highs. The process continues until the subsiding mini-basin touches down on the sub-salt strata, forming a salt weld (Rowan et al. 2004).

2.6.4 Turtle Structures

Although the centre of a mini-basin will stop subsiding once the weld forms, the flanks, which are still underlain by salt, may collapse, forming new, flanking depocentres and inverting the original depocentre into a turtle
structure (Rowan et al. 2004). Turtle structures consist of anticlines resultant from arching of the overburden due to salt withdrawal, as salt migrates towards an adjacent pillow that becomes diapiric (Trusheim 1960). These anticlines can also be formed by regional extension as diapir flanks subside (Vendeville & Jackson 1992b). Mock turtles form at the crest of a subsiding diapir, as the overburden arches and rotates against the listric faults that bound the relict salt-rollers (Fig.2.27). The intervening synclinal depocentre touches down and inverts into an anticline (mock turtle), which differs from a turtle structure by forming above a diapir (rather than between diapirs) and by missing stratigraphic section at its base (Jackson 1995).

An expulsion rollover is essentially a half-turtle. In expulsion rollovers, the initial basin touches down and welds out, just as in turtles, but flank collapse is asymmetric, so that the depocentre shifts progressively in one direction, forming a growth monocline. As the weld grows in length, salt is displaced basinwards, where it inflates and lifts a condensed overburden (http://www.colorado.edu/GeolSci/courses/GEOL3120, in May 2007).
Fig.2.27 The rise and fall of diapirs during thin-skinned regional extension creates turtle anticlines, mock turtles, crestal collapse grabens, and salt welds (Vendeville & Jackson 1992b). Welds form as the overburden subsides into the salt and comes into contact with sub-salt strata (Jackson & Cramez 1989).

2.6.5 Mini-basins

In the thin-skinned processes of gravity gliding and spreading, up-dip extension is usually compensated by down-dip contraction, accommodated by buckling that produce salt-cored folds. Salt migrates towards the less-loaded anticlines and form diapirs and ridges that bound depressed areas in which sediment accumulate. These depocentres, formed by salt withdrawal, continually subside as the adjacent diapirs or ridges rise vertically, creating more accommodation space and establishing a mutual control. The basin is thus compartmentalised into mini-basins, which are smaller sediment catchment areas (Fig.2.28). Once a basin is subsiding into salt, it will keep...
subsiding regardless of the depositional setting and the mini-basin will be filled with whatever sediment is available, including slumps off the adjacent highs (Rowan et al. 2004).

![Fig.2.28 Cross-section of analogue model showing the mini-basins formed down-dip to compensate up-dip extension, due to gravity spreading caused by a prograding wedge over a ductile layer. A landward-dipping normal fault (F) marks the approximate transition between extension and contraction (vertical exaggeration: 1.7). The dark basal layer is a polymer (GeoBR) that stands for the ductile salt layer; the overlying sequences are layers of dyed sand that stand for the brittle sedimentary overburden in the Santos Basin (Guerra et al. 2005a, b).](image)

### 2.6.6 Drag Zones

Drag zones are highly strained regions developed adjacent to the flanks of salt diapirs, resulting from folding or rotation of the overburden into steeply dipping attitudes sub-parallel to the walls of the diapir (Alsop et al. 2000). These authors have shown that the rheology and heterogeneities of the overburden at the time of diapirism control the drag profiles geometry. Narrow drag zones with high vertical relief develop preferentially in poorly lithified sediments, whereas wider drag zones are produced by competent lithologies. High competency contrast within the overburden encourages steeply outward-dipping faults (Fig.2.29).
Fig. 2.29 Schematic drag profiles adjacent to salt diapirs. Drag zone deformation within overburden that is (a) homogeneous and competent; (b) homogeneous and incompetent; (c) heterogeneous and competent; and (d) heterogeneous and incompetent with marked competency contrast (Alsop et al. 2000).

2.7 Halokinesis in the South Atlantic

In the South Atlantic passive margins, salt acts as a décollement surface for thin-skinned deformation. The evaporite layer is deposited in a restricted basin between the two rifting margins. As the plates break apart, the opposite margins gradually subside into the marine basin, in which carbonates are deposited. Gravity gliding is triggered by subsidence and sedimentary loading. The extreme extension of the margins creates rafts of the supra-salt sediments, bounded by listric normal faults detached on the salt layer and underlain by salt-rollers. Clastic sediments are deposited in the
hangingwalls that grow with continued sedimentation, as the rafts translate down-dip (Fox 1998). Salt withdrawal results in salt welds that connect sub-salt and supra-salt sequences. The up-dip extension of the margin is accommodated down-dip by formation of salt ridges associated with detachment folds, thrusts and even nappes over the oceanic crust.

In this thesis a gravity-driven mechanism of landward-dipping listric fault and mini-basin formation controlled by halokinesis will be demonstrated, based on seismic interpretation (including a strategic 3-D survey) and cross-section restoration, with insights taken from analogue modelling.
Chapter 3

Santos Basin

The widest salt basin in the South Atlantic, the Santos Basin resulted from the separation of the South American and African plates, during the opening of the South Atlantic Ocean, a process that has begun in the Mesozoic and continues at the present-day (Fig.3.1). It is located on the Brazilian continental margin, between the Cabo Frio High (23°30'S) and the Florianópolis High (28°S), and extends offshore up to the outer limit of the São Paulo Plateau, comprising an area of about 350,000 km² from the shoreline up to the 3,000 m isobath. A coastal mountain range (Serra do Mar) rims the basin from Florianópolis, in the southwest, to Rio de Janeiro, in the northeast (Fig.3.2).

Fig.3.1 Age of oceanic lithosphere on both sides of the South Atlantic (http://www.ngdc.noaa.gov/mgg, in August 2008).
Fig. 3.2 Major physiographic features of the Santos Basin and the adjacent areas on topographic and bathymetric map (http://topex.ucsd.edu, in April 2006). Universal Transverse Mercator coordinate system is centred at meridian 45° W.

3.1 Major Physiographic Features

The onshore sub-parallel mountain system comprising the coastal Serra do Mar and the inland Serra da Mantiqueira, which originated in the Late Cretaceous, roughly parallels the coastline (Fig. 3.3). Almeida & Carneiro (1998) state that during the early Cenozoic the evolution of the local relief has affected into a considerable extent the coeval clastic sedimentation of the Santos Basin. They suggest that the Serra do Mar evolved from a very different geographical position as compared to the present one and they support the hypothesis of a long-term scarp retreat from southeast to northwest due to erosive processes approximately starting at the Santos Cretaceous Hinge Line in the current continental shelf.
Between Serra do Mar and Serra da Mantiqueira, a string of Cenozoic basins occurs in an elongate narrow trough, over 900 km long (Fig. 3.3), composing the Continental Rift of Southeastern Brazil, dated from the Palaeogene (Riccomini et al. 2004), and developed within the domain of the Late Proterozoic Ribeira Belt which contains older nuclei of Archean to Middle Proterozoic rocks (Almeida 1976).

The Parafbs do Sul River, which once fed the Santos Basin with continental sediments, was diverted to the north, during the Early Palaeocene, and since then discharges into the Campos Basin, to the north, being responsible for the much thicker Cenozoic sequence of the Campos Basin when compared with that of the Santos Basin (Zalán & Oliveira 2005).

The Cabo Frio High bounds the basin to the north, marking its limit with the prolific Campos Basin. This region comprises a suite of alkaline rocks intruded and extruded during the Late Cretaceous and Palaeogene. The southern end of the Santos Basin is bounded by the Florianópolis High, which was already a basement high from the Barremian to early Aptian.

The Cretaceous deposits were confined offshore by the Santos Hinge Line, which forms the western limit of the Cretaceous sediments (Fig. 3.4). To the west of this feature, Cenozoic sediments lie directly onto the shallow basement. The subsidence during the Palaeocene and the Eocene created space to accommodate the whole sedimentation from the alluvial plain to the deep marine domain (Moreira & Carminatti 2004).
Fig. 3.3 NASA satellite image showing the mountain ranges, the intervening Cenozoic continental basins (e.g., Taubaté and Resende basins), and the Paraiba do Sul river current course, discharging in the Campos Basin, to the north of the Santos Basin.

Fig. 3.4 Seismic section displaying the Santos Hinge Line (arrow), the western limit of the Cretaceous sediments. To the west of this feature, the Cenozoic sediments occur directly onto the shallow basement (Zalán & Oliveira 2005).
In the deepwater region, from the shelf edge through the transition to oceanic crust, high sedimentary thicknesses occur in the São Paulo Plateau. This vast physiographic feature is a bathymetric high, in which the rifted crust is much wider than the rest of the Brazilian southeastern margin, reaching 500 km. Located between the 2,000 m and 3,000 m isobaths, it comprises a great variety of salt structures, usually very thick, and constitutes the southern limit of the Brazilian marginal salt basins. In this region, intense halokinesis has produced salt diapirs and ridges that affect the seabed creating pseudo-craters. The plateau is considered to be underlain by highly stretched continental crust, intruded by magmatic rocks (Kowsmann et al. 1982; Macedo 1990; Chang et al. 1992).

The Jean Charcot Seamounts are volcanic plugs located basinwards from the outer limit of the salt diapir province. The submarine chain that forms the São Paulo Ridge limits the São Paulo Plateau to the south and is considered to be genetically related to the Rio Grande Rise, an oceanic discontinuity transversal to the Mid-Atlantic Ridge (Kumar & Gambôa 1979), which is crossed by the Cruzeiro do Sul lineament that extends towards the Cabo Frio region.

The limit of the continental crust is yet to be completely defined. It is tentatively positioned near the external limit of the evaporites (Chang et al. 1992). Karner (2000) has proposed a continent-ocean boundary (COB) based on the termination of fracture zone trends and on changing directions of gravity anomalies trends. Meisling et al. (2001) have developed earlier concepts by Kumar & Gambôa (1979) and Demercian (1996) and proposed that the NE-trending gravity anomaly that comprises the Avedis Chain (described by Demercian 1996), in the distal region of the São Paulo Plateau, would correspond to a failed spreading centre emplaced in thinned
continental crust. Gomes et al. (2002) suggested that this trend could be extended to the SW of the Avedis Chain, where a proto-oceanic crust would exist (Fig. 3.5). The Outer High of the Santos Basin was interpreted as a structural basement high that has remained positive since the early Aptian acting as a regional focus of hydrocarbon migration. It is onlapped on both sides by the rift sequence. The Upper Cretaceous sequence thins over this basement high and only the Cenozoic sequence keeps its thickness roughly constant in the area (Fig. 3.6).

Fig. 3.5 Bouguer gravity anomaly map of the southern São Paulo Plateau with the Outer High and the proposed proto-oceanic crust (Gomes et al. 2002). Dashed lines are transfer zones (TZ) and dotted lines are failed spreading ridges (FS) from Meisling et al. (2001). The Avedis Volcanic Chain was defined by Demercian (1996).
Fig. 3.6 Regional strike seismic section in the southern São Paulo Plateau (location in Fig. 3.5). The Outer High of the Santos Basin is an outstanding basement high, onlapped on both sides by the rift sequence. It has acted as a regional focus for oil migration. The impressive basement structural low in the left edge of the seismic section is related to the steep gradient observed in the Bouguer gravity anomaly profile, and is interpreted as the very beginning of the proto-oceanic crust region indicated in Fig. 3.5 (Gomes et al. 2002).
Over the last two decades, the geology of the Santos Basin has been investigated in many aspects, including stratigraphy, tectonics and magmatism and a good general handle now exists for its tectono-stratigraphic development and evolution (Williams & Hubbard 1984; Pereira et al. 1986; Macedo 1989; Pereira & Macedo 1990; Mohriak & Macedo 1993; Demercian et al. 1993; Pereira 1994; Moraes Jr. et al. 1994; Mohriak et al. 1995; Cobbold et al. 1995; Moreira 2000; Demercian 1996; Cobbold et al. 2001; Meisling et al. 2001; Modica & Brush 2004; Moreira & Carminatti 2004; Milani et al. 2005; Zalán & Oliveira 2005; Freitas 2006; Moreira et al. 2006; Oreiro 2006; Caldas 2007; Gambôa et al. 2008).

3.2 Tectono-sedimentary Evolution

The tectono-sedimentary evolution of the eastern Brazilian basins results from the onset and full development of the South Atlantic Ocean. The Western Gondwana break-up and the separation of the African and South American plates started in the Neocomian with rifting in the southernmost part of South America (Cainelli & Mohriak 1999). The region evolved from an initial stage of asthenospheric uplift and lithospheric thinning, followed by volcanism, which gave rise to a rift phase and a subsequent drift phase.

The Santos Basin comprises four tectono-sedimentary megasequences (Ponte & Asmus 1978; Chang et al. 1992) bounded by major unconformities of regional expression: (i) continental (Late Jurassic/Early Cretaceous rift stage with fluvial and lacustrine sediments), (ii) transitional (Aptian evaporites), (iii) restricted marine (Albian carbonates) and (iv) open marine sequences. Magmatic episodes took place in the Early Cretaceous associated with the
onset of the rift phase, in the Late Cretaceous (80-90 Ma) and in the Eocene (50 Ma).

The basin initiated with a Cretaceous rift phase (Hauterivian/Barremian) caused by increasing lithosphere stretching that created NE-SW-trending normal faults in half-grabens segmented by NW-SE transfer zones (Fig.3.7), which accommodate different stretching rates (Meisling et al. 2001). Fluvial and lacustrine sediments and conglomerates deposited in a continental environment. The rifting was preceded by crustal thinning and voluminous extrusion of tholeiitic basalts (Chang et al. 1992).

Fig.3.7 Map of main rift-related structural provinces, Campos and Santos Basins. East-west extension direction (red arrows) is from plate tectonic reconstructions of the early (Hauterivian) stage of Atlantic opening. Inferred rift transfer zones (dashed red lines) are not parallel with this extension direction (Meisling et al. 2001).

At the end of the rift phase, the topography was levelled by erosion that created a regional unconformity (the “breakup unconformity”), separating
continental from transition to marine environments of deposition (Cainelli & Mohriak 1998). The rift phase was followed by an Aptian transitional phase. The Florianópolis High and the São Paulo Ridge acted as the south barrier to oceanic water circulation (Fig.3.8), thus forming a hypersaline sea that favoured the deposition of a thick evaporite sequence (Demercian 1996).

Concerning the symmetry of the evaporite basins, the separation was uneven and asymmetric. Wide basins on one margin match with opposing narrow basins. The Santos Basin, which comprises the largest volumes of evaporites in the Brazilian margin, has no corresponding salt basin in the conjugate African margin.

During the Albian, a shallow marine environment was established with the onset of the Atlantic Ocean, leading to the creation of a large carbonate platform. Halokinesis was triggered by the overburden deposition and by basinward tilting due to thermal subsidence.

With the progressive opening of the South Atlantic Ocean, associated with production of oceanic crust, the shallow marine Albian platform was rapidly flooded in the Cenomanian (Dias-Brito 1982, 1987).

From the Late Cretaceous to the early Cenozoic, the basin was filled by a sequence of shelf and slope sandstones and shales of marine passive margin.

The Late Cretaceous sequence, which is characterised by deepening of the environment in most of the Eastern Brazilian margin, consists, in the Santos Basin, of thick successions of shallow water deposits. The regression is associated with episodes of massive clastic progradation. The Cenozoic sequence is dominated by siliciclastic rocks that prograde basinwards (Cainelli & Mohriak 1998).
Fig. 3.8 With the opening of the South Atlantic, a large Aptian salt basin developed along the Brazilian continental margin, from the Santos Basin to the Sergipe-Alagoas Basin, with counterparts in the African margin (i: after Fainstein 1999; ii: after Davison 2005).
Halokinesis accounts for the major deformation of the post-salt sequence of the Santos Basin, producing up-dip extension and down-dip contraction. It has been influenced by sediment progradation episodes, overburden extension, gravity gliding and gravity spreading.

High salt diapirs and ridges occur in the deep-water domain. Even the shelf contains salt pillows and diapirs up to a few kilometres high. Halokinesis caused raft tectonics that resulted in many carbonate turtleback structures, which form one of the main exploratory objectives in the southern part of the basin (Cainelli & Mohriak 1998). Seaward- and landward-dipping listric normal faults detached on the salt layer created many anticlinal structures both in the Albian carbonates and in the Upper Cretaceous siliciclastic sequence, also constituting good hydrocarbon plays.

Seismic interpretation, analogue and numerical modelling have been applied to investigate halokinesis in the Santos Basin, resulting in different models to explain the genesis of the Cabo Frio Fault and the salt-bounded mini-basins and proposals to divide the basin into halokinetic provinces (Pereira et al. 1986; Cobbold & Szatmari 1991; Demercian et al. 1993; Cobbold et al. 1995; Mohriak et al. 1995; Szatmari et al. 1996; Demercian 1996; Ge et al. 1997; Ings et al. 2004; Gemmer et al. 2004; Guerra et al. 2005a, b, c; Davison 2007; Guerra & Szatmari 2006, 2008).

Pereira et al. (1986) grouped the halokinetic structures in five provinces, from the shelf to the deep basin. Province I, close to the hinge line, is a narrow steep homoclinc with small amounts of salt left (anhydrite and remnants of halite that has been mobilised and partly dissolved). Province II is a belt of small-amplitude pillows in the footwall of seaward-dipping listric normal faults, and locally tall diapirs in the shelf (related to important exploratory plays). Province III is a narrow belt with few structures. Province IV is a belt
of large salt pillows in the footwall of large seaward- and landward-dipping normal faults, and salt-cored folds (Cobbold et al. 1995). Province V comprises huge salt domes and diapirs, as well as sinuous salt ridges that almost outcrop at the seabed. Folds and thrusts occur in the overburden, towards the oceanic crust limit.

Cobbold & Szatmari (1991), based on analogue modelling, proposed radial gravitational gliding to explain the Campos (divergent) and Santos (convergent) halokinetic styles. Demercian et al. (1993) divided the basin into halokinetic provinces with different patterns of gravity gliding, laterally separated by thin-skinned transfer faults.

3.2.1 The Cabo Frio Fault

Differential sedimentation caused by massive progradation drove halokinesis and created the Cabo Frio Fault (CFF), an outstanding landward-dipping listric normal fault detached on the salt layer that controlled the major depocentre from the Upper Cretaceous to the lower Cenozoic sequences (Fig.3.9). Along the Cabo Frio Fault, the Albian sequence is displaced by tens of kilometres, constituting the “Albian Gap”, which displays heaves that reach 50 km (Demercian et al. 1993). Some models have been proposed to explain the genesis of the Cabo Frio Fault, interpreted as (i) a megadetachment caused by antithetic basal shear associated with basinward salt flow and massive clastic progradation, possibly triggered by reactivation of antithetic basement-involved normal faults (Mohriak et al. 1995), (ii) a listric fault caused by differential load related to progradation and extension over a spreading ductile layer whose base is roughly flat.
(Szatmari et al. 1996; Guerra et al. 2005c), (iii) salt withdrawal and expulsion rollover associated with progradation with negligible extension of the overburden (Ge et al. 1997) and (iv) progradation over a ductile layer whose base dips landwards (Ings et al. 2004; Davison 2007).

Of all these models, only one does not consider the Cabo Frio structure as a fault. Ge et al. (1997) postulate that no significant extension is associated with the genesis of the Cabo Frio structure, and interpret it as an expulsion rollover, which is practically a half-turtle (see Chapter 2, section 2.6.4). The advance of a prograding wedge causes expulsion of the underlying salt and inflation of the salt layer on the front of the wedge. The initial basin touches down and welds out as the depocentre progressively shifts in one direction, creating a growth monocline (Fig. 3.10). According to these authors, the contact between the diapir and the proximal rollover would be "diapirical rather than faulted" (Ge et al. 1997). It is unlikely that this model applies to the Santos Basin because deformation of the post-salt sequence by thin-skin extension and contraction over a basal décollement of salt can be observed in the seismic data of the basin. The regional patterns of extension and contraction in the Santos Basin indicate that the Cabo Frio structure is
actually a landward-dipping normal fault, whose footwall moved basinwards to accommodate the massive progradation (http://www.colorado.edu/GeolSci/courses/GEO1320, in May 2007).

Fig.3.10 Model evolution proposed by Ge et al. (1997) to the Cabo Frio structure: progressive evacuation inflates distal salt, which eventually evolves into a diapir that gets buried. Note that pre-Maastrichtian sediments are absent beyond the inflated salt.
Moreover, in the model proposed by Ge et al. (1997), the first sediments deposited beyond the inflated salt dated at the Maastrichtian, which does not correspond to the Santos Basin, in which the salt layer is overlain by sediments since the early Albian, as shown by the seismic interpretation and cross-section restoration performed in this thesis.

More recently, Davison (2007) attributed the genesis of the Cabo Frio Fault to the geometry of the base salt. According to Davison (2007), based on numerical models performed by Ings et al. (2004), the base salt would have subsided under the thick sedimentary load and became a landward-dipping substratum for the halokinetic processes. The seaward flow of the ductile layer would be reverted towards the continent. However, the seismic data set used in this thesis show that many lines that cross the Cabo Frio Fault do not show a landward tilting base salt.

Previous models of the Cabo Frio Fault were based on 2-D data. Now a level of detail and accuracy can be obtained by using 3-D seismic data, which was not possible in the past works.

The interpretation of the 3-D seismic data set performed in this thesis made it possible to observe that the CFF does not always occur above a basement high or a pre-salt fault, and that the base salt does not dip landwards throughout the domain of the main fault.

During the Late Cretaceous and Cenozoic, the Santos Basin was affected by magmatism and basement uplift, probably associated with the westward movement of the South American Plate over the Trindade Plume (Zalán & Oliveira 2005). These authors state that from the Santonian to the Maastrichtian (85-65 Ma), the continental crust was uplifted in response to a mantle thermal anomaly. A large amount of sediments dumped into the basin formed a continental shelf (Juréia Formation) and sand was carried to
lower bathymetries as turbidites. The isostatic imbalance between the high mountains onshore and the adjacent progressive offshore subsiding basin established a gravity potential that promoted gravitational deformation. Collapse of the highlands started at the end of the Palaeocene and became more intense during the Middle Eocene, creating a regional unconformity (Zalán & Oliveira 2005). The intense magmatic episodes that took place during the Late Cretaceous and Eocene (Oreiro 2006; Moreira et al. 2006) were characterised by sills, dykes and volcanic cones. Due to this tectonic and magmatic activity, Zalán & Oliveira (2005) do not consider the Santos basin a typical passive margin basin.

3.3 Stratigraphy

The Santos Basin’s stratigraphic record reveals processes of crustal stretching, continental rifting, inception of oceanic crust and thermal subsidence (Fig.3.11).

The rift and thermal subsidence phases of the Santos Basin are documented by thick evaporite, carbonate and siliciclastic sediments that date from the Hauterivian to the Holocene and reach about 12 km thick (Pereira et al. 1986; Pereira & Macedo 1990; Pereira 1990, 1994).

The rift sequence started with the Neocomian basalt flows of the Camboriú Formation (120-130 Ma) that have acted as an economic basement to the subsequent sedimentary infill and shows correspondence with the lava flows of the neighbouring Paraná, Campos and Pelotas basins, to the West, North and South of the Santos Basins, respectively (Amaral et al. 1967; Fodor et al. 1983; Fodor & Vetter 1984; Zalán et al. 1990).
Fig. 3.11 Santos Basin stratigraphic chart (modified from Pereira & Feijó 1994).
As volcanism intensity decreased, the space produced in the rifting event was filled with Barremian lacustrine sediments of the Guaratiba Formation (shale, marl, limestone, coquina, and sandstone) that overlie the Lower Cretaceous tholeiitic basalts (Dias et al. 1987; Mizusaki et al. 1988). Siliciclastic sediments were deposited as prograding alluvial fans on shallow lakes, in which coquinas and anoxic shales occur. The presence of non-marine ostracods suggests Barremian to early Aptian age to this formation that can be correlated with the prolific Lagoa Feia Formation in the Campos Basin.

The syn-rift, continental, sediments of the Guaratiba Formation were deposited over the basalts, aside with sediments related to eventual marine incursions. These deposits consist of conglomerates and red sandstones, possibly from alluvial fans, with some lacustrine deposits of coquinas intercalated with sandstones (Demercian 1996). This early Aptian sequence was deposited in a rift setting (Dias 2005). The extreme anoxic conditions in the bottom of the lake, with saline to hypersaline waters of alkaline affinities, gave rise to deposition of fine, well laminated organic-rich calcareous black shales, which constitute the main hydrocarbon source rock in the basin (Mohriak et al. 1990).

The end of the rift phase is marked by a regional unconformity ("breakup unconformity") that has largely smoothed the rift topography. This erosional event provided the coarse-grained sandstones and conglomerates deposited above the "break-up unconformity".

During the transitional phase, the Florianópolis High and the São Paulo Ridge formed a barrier that hindered marine water free circulation and promoted precipitation of evaporitic minerals from a hypersaline brine, favoured by the arid conditions. A thick Aptian evaporite sequence (Ariri Formation) was deposited above siliciclastic and carbonate rocks, filling up
the elongated South Atlantic Gulf (Cainelli & Mohriak 1998). The deposition of the evaporite layer was estimated to have lasted less than 600,000 years (Dias 2005). The original evaporite thickness was variable, reaching 2,500 m in places (Macedo 1990). This evaporite sequence occurs over most of the east Brazilian margin to the north of the Pelotas Basin. Carbonates, anhydrite, halite and complex magnesium and potassium chlorides occur in several evaporitic cycles (Gambôa et al. 2003, 2008; Freitas 2006). The age of the salt in the Santos Basin is not precisely known but anhydrites and carbonates lie unconformable above volcanic rocks dated at 113.2 Ma, in the southern margin of the Santos Basin (Dias et al. 1994).

The marine sequence is associated with thermal subsidence during the drift phase, associated with cooling and contraction of the lithosphere as the plates moved away from the mid-ocean ridge.

The development of the oceanic crust effectively split up Africa and South America and enlarged the entrance of the former evaporite sea, giving rise to the formation of a passive continental margin in which deposition of a wide carbonate platform took place under shallow marine conditions. Along the ancient Albian coastline, siliciclastic fan delta systems of the Florianópolis Formation were laterally interbedded with high energy shallow carbonates of the Guarujá Formation (early to middle Albian) and later on covered by pelitic sediments of the transgressive late Albian to early Cenomanian Itanhaém Formation (Pereira et al. 1986). The Guarujá Formation consists of calcarenites and dolomites that are oil-bearing in many fields. Halokinesis was triggered both by the seaward tilting of the basin due to thermal subsidence and the deposition of the carbonate platform that increasingly loaded the underlying salt layer. The differentiation between coarser high-
energy carbonate shoals and finer low-energy carbonate depressions was synchronous with the development of salt pillows (Cainelli & Mohriak 1998).

In response to the continuing marine transgression in the Middle Albian, the basin deepened and the high-energy carbonates gave rise to a low-energy sequence formed by rhythmic intercalations of calcilutites, marls and shales of the Itanhaém Formation. The flooding event stopped the growth of the carbonate platform. Deposition of radioactive shales culminated this event. Pereira (1994), however, supports that the carbonate demise was caused by erosion and terrigenous deposition. Maximum flooding of the margin occurred during the late Cenomanian to Turonian, a time characterised by deposition of anoxic shales. Turbidite sandstones distributed in the Turonian sequence indicate relative sealevel falls in the prevailing relative sealevel rise. They subside differentially in response to intense halokinesis.

The subsidence of the basin established a transgressive marine system up to the Middle Turonian (Itajai-Ácú Formation), which was followed by regressive events during the Senonian (Santos and Juréia formations) and resulted in the seaward advance (basinward shift) of the shoreline (Pereira & Macedo 1990).

The passage from a transgressive to regressive marine setting is well-known in all the eastern Brazilian marginal basins (Chang et al. 1992). The bathyal onlapping marine sequence changes into a shallowing upward prograding wedge, whose western limit is marked by an expressive coastal plain. In the Santos Basin, this depositional style took place in the late Turonian (Pereira et al. 1986; Viviers 1986).

After the rift phase, the subsidence was relatively smooth and the basin floor topography displayed a ramp style up to the early Turonian times. A magmatic event dated at 80-90 Ma and the coastal mountain range uplift are
associated with plate reorganisation and the regional high thereafter acted as a rejuvenated relief and the major source of sediment for the subsequent sequences in the basin. The Late Cretaceous and early Cenozoic magmatism was an additional instability and triggered turbidite events that advanced basinwards (Mohriak et al. 1995; Cainelli & Mohriak 1998).

During the Late Cretaceous (Coniacian to Maastrichtian), due to the sea floor spreading combined with global eustatic high in greenhouse times, there was a general tendency of relative sea level rise in the marginal Brazilian basins. In the Santos Basin, however, the Serra do Mar uplift and the subsequent massive clastic sediment supply have compensated the sea level rise, keeping the basin in continental and shallow water environment (Macedo 1987). At that time, in the proximal domain, a platform dominated by coarse clastic sediments developed and episodes of intense sediment supply and shelf progradation were associated with the initial stage of the Serra do Mar uplift (Zalán & Oliveira 2005). Deltaic sandstones of the Juréia Formation prograded across and beyond the underlying Albian carbonate platform margin (Modica & Brush 2004). In the late Campanian, a new event uplifted the source area, maintained the progradation and promoted widespread turbidite deposition (Ilhabela Member).

During the regressive events associated with deposition of Juréia and Santos formations, sedimentation rates increased and accounted for rapid accumulation that caused a 250 km coastline advance (compared with the present one). The continental red beds and fluvial deltaic deposits advanced several tens of kilometres beyond the present-day shelf edge. Organic-rich shales of the Itajai-Açu Formation were deposited in the far shore domain. These three formations constitute a prograding and later retrograding coast-shelf-slope system.
The Paraiba do Sul drainage was modified, affecting the sediment dispersal pattern, at the latest Cretaceous (Modica & Brush 2004) or Early Palaeocene (Cobbold et al. 2001; Zalán & Oliveira 2005). Relatively immature and disorganised dip-oriented drainage systems were captured and the Paraiba do Sul River, which used to discharge directly into the Santos Basin, was deviated to the Campos Basin, in the north.

The end of the Cretaceous was characterised by a regional unconformity. The Cenozoic siliciclastic sediments are finer than the Cretaceous ones. A possible explanation is that the coarser sediments would have been captured by the Paraiba do Sul River and discharged into the Campos Basin. This could also explain the lower Cenozoic sequence being much thicker in Campos than in Santos. In its turn, the Upper Cretaceous sequence in Santos is thicker than in Campos (Macedo 1987). The strong erosion at the end of Palaeocene that created incised valleys on the shelf and slope coincides with a global sea level fall but may have been enhanced by thermal uplift around the Cabo Frio High (Cobbold et al. 2001).

During the Palaeogene, the gradual cooling of the continental crust, progressively farther from the mid-ocean ridge, promoted increasing subsidence, whilst the onshore region was uplifted. The Cenozoic sediments deposited in two major prograding sequences: from Early to Middle Eocene and from Middle Miocene to Recent (Pereira et al. 1986; Pereira & Feijó 1994).

Early Eocene transgression filled the canyons formed during the Palaeocene with deltaic sediments. In the Middle Eocene and during climax of the gravitational collapse (Zalán & Oliveira 2005), deltas prograded over the Upper Cretaceous and Palaeocene shelf in the central northern Santos Basin. Debris flows and turbidite deposits are typical of this sequence in the northern basin.
After the Late Eocene, deltaic sediments deposited on a Cretaceous and Palaeocene shelf were later on reworked and became source of debris flows and turbidites. The sediments deposited by gravity flows in the Eocene in the northern Santos Basin were studied by Moreira & Carminatti (2004), who described two different systems: a sandy (continental source) and a shaly system.

The Iguape Formation is a carbonate sequence deposited in a platformal setting created in the centre and south of the basin, influenced by alluvial fans in the proximal domain. Shales and fine-grained turbidite sandstones of the Marambaia Formation occur in the central and distal domains. The sedimentation in the Santos Basin culminates with Pleistocene siliciclastic coastal fans of the Sepetiba Formation (Fig.3.12).

3.4. Post-salt Magmatism

The drift stage of the Santos Basin has been affected by two important magmatic events: one in the Santonian-Campanian (82 ± 1 Ma), and the other
in the Eocene (48.9 Ma), dated with Ar/Ar method (Moreira et al. 2006). The Santonian-Campanian magmatism was intense in the northern Santos Basin, displaying volcanic cones, extrusive and intrusive rocks (Fig.3.13). In the area of the Cabo Frio High there is also seismic evidence for volcanic edifices formed during the Albian, Maastrichtian and Palaeocene (Oreiro 2006).

The post-salt magmatism in the study area resulted from readjustment in the South American Plate. Its genesis is attributed either to the passage of the South American Plate over a mantle plume (Szatmari et al. 2000; Thomaz Filho et al. 2005; Zalán & Oliveira 2005) or to leakage along reactivated deep faults (mainly the NW-trending transfer faults) that appear to have cut through the whole lithosphere, reaching the asthenosphere and causing partial melting of the upper mantle by pressure release (Almeida 1976; Oreiro et al. 2006; Oreiro 2006). The plume was the Trindade hot spot, responsible for the vast Abrolhos volcanic plateau and the Victoria-Trindade seamount chain, in the Espírito Santo Basin, to the north.

Fig.3.13 Seismic expression of a volcanic cone (CV) and associated lava flow deposit (DL) filling a depositional low (Moreira et al. 2006).
3.5 Halokinesis - Insights from Analogue Modelling

Analogue modelling of halokinesis can be useful for generating new concepts and for explaining the processes involved in deformation and sedimentation of salt-bearing basins, using viscous polymers for the salt layer and frictional plastic granular materials for the overburden.

A new technique for analogue modelling was created and developed at the Petrobras Research Centre, in order to improve the understanding of the interplay between halokinesis and sediment dispersal in evaporite-bearing basins. This technique was proposed as an alternative to the traditional subaerial models (Vendeville 1987; Vendeville et al. 1987; Cobbold et al. 1989). The fundamental difference from the traditional modelling is that in this method the experiments are run underwater (Guerra et al. 1998, 2000, 2005a, b, c; Guerra & Szatmari 2006, 2008). Previous analogue modelling studies of salt tectonic processes using silicone to simulate salt and dry sand to simulate the overlying sediments have led to a major advance in our understanding of salt tectonic processes. The control of salt tectonics by normal faulting of the overlying sediments has been clearly demonstrated, for this technique has enabled faulting of the overburden, which did not happen in the former experiments with immiscible fluids. Nevertheless, due to the strength of the overburden, the subaerial modelling tends to inhibit diapirism of the ductile layer, preferentially forming non-piercing structures such as pillows. A new approach, developed at the Petrobras Research Centre by the author and her colleagues, has been to submerge both the silicone and the overlying sand in water. The saturated sand layer has its shear strength reduced by increased pore pressure whilst total overburden pressure increases. As a result, buoyancy-related structures form promptly.
and the sand is more liable to produce compressive structures. The resulting structures closely resemble diapirs and canopies well known from deep-water environments. Another great advantage is that subaqueous modelling permits to combine sedimentation, including generation of turbidite channels and fans, with simultaneous salt tectonics, revealing their intricate interaction and mutual control. Exaggeration of the density contrasts thus helps to reproduce otherwise hard to achieve structures abundantly seen in nature, even though the intentionally distorted dimensioning should be treated with caution in the detailed interpretation of the results (Guerra et al. 2005c). The remarkable similarity between the resulting structures and their natural counterparts indicates that this technique is suitable to study the evolution of salt-related structures and their influence on sediment dispersal. It permits to reproduce in a dynamic way and in the same experiment the processes of erosion, transport and deposition of sediments in basins deformed by halokinesis.

A series of scaled analogue modelling was run in the Petrobras Research Centre, aiming at (i) investigating the response of the depositional space to the stresses generated by the sedimentary loading on the underlying ductile layer, in basins that display convergent physiography, and (ii) reproducing two distinct sediment sources with short-time delay, in order to analyse their control on the genesis of halokinetic-related mini-basins (Fig. 3.14). All the examples shown are from models that were downscaled to represent the lithologies found in the sedimentary basins and used a ductile polymer (silicone putty or GeoBR) as analogue for salt and quartz sand as analogue for the brittle overlying sediments (Guerra et al. 2001, 2005a, b). The continental shelf and slope were simulated by a sand wedge located at the proximal region of the model. The fundamental difference from the
traditional subaerial models is that these materials, here, were placed under a water column.

The models were designed and run by the author and colleagues at the Petrobras Geotectonics Laboratory. Although they were run prior to the onset of this project, these experiments have produced some results that shall be included here to support seismic interpretation and further discussion.

Subaqueous analogue models show how salt diapirs and walls deflect sediment pathways, and how the adjacent low areas, undergoing salt withdrawal, act as preferential depocentres. The sediments captured in these depositional troughs will represent an additional load, enhancing the rise of the neighbouring salt highs (Fig.3.15).

Fig. 3.14 Experimental rig (A) and model surface (B). The N-S segment of the shelf was the first to be deposited. The second segment, E-W, was deposited later on, after the development of a few structures (Guerra et al. 2005a, b).
The models have shown that a confluence of prograding sediments can impose convergence on salt flow. The escape of salt from under these thick sediment wedges and the related gravitational extension and contraction of the overburden have created a complex pattern of “salt”-cored ridges deformed by superposed folding establishing an interference pattern that resulted in the creation of distal salt withdrawal mini-basins (Fig.3.16). These mini-basins acted as depocentres for the sediments eroded from the shelf and carried into the deep basin in turbulent flow. Analogously, it may be expected in nature that similar structures form in response to the convergent flow of salt, in evaporite-bearing basins submitted to confluent directions of sediment supply, like in some areas of the east Brazil, west Africa and the Gulf of Mexico (Guerra et al. 2005a, b).
Fig. 3.16 In the deep water domain, “salt”-cored ridges display an interference pattern individualising mini-basins (arrow) (Guerra et al. 2005b).

Removal of the overlying sand exposed the surface of the polymer that represents salt and whose deformation recorded the principal directions of the “halokinetic” structures. In the near shore domain, positive structures correspond to the rise of the polymer in the footwall of listric faults, forming continuous “salt” rollers located under structural lows in the overburden. The polymer was almost entirely depleted in between the positive “salt” structures. A large amount of polymer rose in the footwall of the large counter-regional listric normal fault formed at the toe of slope and migrated towards the deeper basin (Fig.3.17).
Fig. 3.17 The structures observed on the top of the polymer (analogue for salt) after the slicing of the model and the removal of the top sand (Guerra et al. 2005a, b) are similar to those illustrated by Jackson & Talbot (1986). Near shore salt rollers and salt anticlines that have risen on the footwall of listric normal faults, evolve to higher amplitude salt diapirs and walls, frequently showing overhangs (near surface spreading). In the analogue model, part of the salt withdrawal basin domain is shown. This domain resulted from interference of concentric directions of progradation that induced salt to flow in a convergent fashion.

The experimental results highlight the control that structures resultant from diapirism exert upon sediment dispersal in the basin. Sediments eroded from shelf and slope are generally carried by turbulent flow and deposited as fans from the base of slope onwards. Deposition of a turbidite fan causes the underlying ductile material to escape towards regions of less overburden, and to rise amongst the fans. The resulting bathymetry controls the
distribution of subsequent incoming sediments, which surround the evaporite structures and deposit in the adjacent bathymetric lows. Similar flows are expected in nature, in which the basinward advance of sediments is restrained by salt structures such as diapirs or walls that constitute obstacles to the seaward transport of sediments.

In these models, the main driving mechanism to trigger halokinesis was the differential loading caused by a thick prograding wedge deposited above the salt, whose base was kept horizontal.

Although planned in a rather general basis, the analogue models have produced results that could give good insights into the whole process. It is recommended to focus future models on the geology of the Santos Basin in view of the seismic interpretation carried out in this project.

3.6 Petroleum System

Although the exploratory effort is still very small in some areas, Santos is now considered a high potential basin. The discoveries in pre-salt reservoirs, in the last two years, have opened a new exploratory frontier, with a promising giant hydrocarbon province that may add huge light oil reserves.

The first find in the Santos Basin was in the shallow water Albian carbonate reservoirs of the Merluza Field, during the eighties. After that, Tubarão, Caravela, Coral and Estrela do Mar fields were found in the same kind of play. There is an established and productive Albian carbonate play in the south and an emerging Cenozoic turbidite play in the north of the basin, where important oil discoveries have been recently reported. In 2003, the large Mexilhão gas field was found in the Santonian sandstones in the
hangingwall of the Cabo Frio Fault. Since 2004, the pre-salt sequence has started to be systematically investigated, leading to important discoveries of what appears to be a giant oil province extending up north towards the Campos and Espírito Santo basins. The Tupi Field, drilled in 2006, in the ultra-deep waters (over 2,000 m water depth) of the Santos Basin, opened a promising new frontier.

Four proved plays exist in the basin: Pre-Salt, Albian, Late Cretaceous and Cenozoic. The pre-salt play was only recently discovered and is expected to continue in other eastern margin basin (Fig.3.18).

Fig.3.18 The major plays in the Santos Basin are the Cenozoic, Itajai-Açu (Upper Cretaceous), Guarujá (Albian) and Guaratiba (pre-salt) plays (www.brasil-rounds.gov.br, in February 2008).

Source

The major source rocks are rift lacustrine shales of the Guaratiba Formation (Barremian). Other source rocks in the basin are the marine anoxic shales of the Itajaí-Açu Formation (Cenomanian/Turonian).
Reservoirs

The main reservoirs are: (i) turbidites of the Marambaia Formation (Eocene/Oligocene); (ii) platform sandstones of the Juréia Formation (Upper Cretaceous); (iii) turbidites of the Ilhabela Member, Itajai-Açu Formation (Turonian to Maastrichtian); (iv) sandstones of the Florianópolis Formation (Albian); (v) calcarenites of the Guarujá Formation (Albian) and (vi) carbonates of the Guaratiba Formation (Barremian/lower Aptian).

Entrapment and Seal

The traps show various origins, displaying strong structural, stratigraphic and combined control. In the pre-salt plays, the traps resulted from basement-involved tectonics, whilst thin-skinned halokinesis, associated or not with stratigraphic features, accounts for accumulation in the post-salt carbonates and turbidite sandstones. Turbidite sandstones are sealed by deepwater marine shales, whereas calcarenites are sealed by calcilutites, marls and shales. Pre-salt carbonates are capped by shales and salt.

Generation and Migration

For the pre-salt sequence, the oil generation window started in the Albian and the generation peak was in the Cenomanian. For the drift sequence, the oil generation window started in the Maastrichtian and the generation peak was reached in the Oligocene, with generation and maturation controlled by structural lows produced by halokinesis.

The thermal evolution of the Mesozoic/Cenozoic sediments is controlled by the interplay of burial and heat flow histories. Due to the changes in the overburden thickness and to the presence of thick salt ridges, it is possible that in some places the source rocks have not reached the necessary thermal maturity to generate hydrocarbons at the same time (Mello et al. 1995).
Migration occurs mainly through rift faults, permo-porous rocks, salt welds and halokinesis-related listric faults. In addition, unconformity surfaces, canyons walls, and borders of salt diapirs and ridges may represent alternative hydrocarbon fairways. These features connect the source rocks with the many reservoir rocks in the pre- and post-salt sequences.

The information concerning the Petroleum Systems was obtained from the National Petroleum Agency homepage (http://www.brasil-rounds.gov.br, in February 2008).
Seismic interpretation was carried out in the central and northern Santos Basin, using data sets that covered an area of 64,526 km² (2-D survey), in which 10,551 km² could be interpreted in more detail based on two 3-D surveys (see Fig.1.1). The seismic interpretation was calibrated by well-log data from all available wells but with two important constraints. Firstly, the well-data did not cover the entire study area and secondly, the whole post-salt sequence was not drilled by most of the wells. Using the seismic stratigraphic methods (sensu Mitchum et al. 1977), depositional sequences bounded by unconformities and/or abrupt downward shift of seismofacies were defined and used to build the stratigraphic framework and regional correlations. Stratigraphic correlations¹, based on gamma-ray and sonic logs, alongside lithology information from the well data, were used to guide the interpretation in the salt and post-salt sequences. Top salt and sedimentary/structural patterns in the post-salt sequence were identified using the seismic sections and structural maps were produced (in two-way travel time) for each interpreted horizon, as well as bathymetric and isochore maps (in metres). All the seismic interpretation and mapping presented in this chapter were performed by the author; none of the interpretation or maps provided relies on other interpreters' work.

¹ They were kindly provided by Petrobras stratigrapher Francisco A. Lima Martins and cannot be presented here due to proprietary reasons.
The software 3-DCanvas was adopted, to give a better visualisation of the 3-D seismic cube and to allow the fault enhancement applications, such as *Semblance* and *AFE*, conjugated with amplitude, which favoured the identification of faults.

To enhance seismic imaging of certain structures, the seismic data was processed using TecVa (volume of amplitudes technique), developed by Bulhões (1999) and improved by Bulhões & Amorim (2005). TecVa highlights amplitude contrasts by enhancing coherent seismic attributes (continuous reflectors), thereby increasing the imaging of structural and stratigraphic features. Continuous reflectors constitute true time lines and appear in light colours, whilst seismic discontinuities (faults, fractures, unconformities) appear in dark colours.

### 4.1 Method and Assumptions

In total, ten well-calibrated horizons were interpreted in the seismic data set, namely: Basement, Base Salt, Top Salt, Albian, Santonian, Campanian, Maastrichtian, Middle Eocene, Miocene and Seabed. Some of them had to be inferred in some places, for their seismic signature was not always clear.

In the post-salt interval, the sequences bounded by these horizons have internal and external geometries that reveal the presence or absence of halokinetic activity during their deposition. In addition to the horizons interpreted as sequence boundaries, some igneous features were also interpreted in the study area.

Stratigraphic correlation based on 34 wells provided input for the interpretation of the post-salt horizons that have seismic expression all over
the study area and bound sedimentary sequences that reveal the tectono-sedimentary history of the basin.

The nine interpreted sequences are described below, along with their lithology and corresponding stratigraphic units (see the stratigraphic chart in Fig. 3.10).

I. from Basement to Base Salt – corresponds to the rift sequence and may include the initial stages of the post-rift sag (thermal) sequence, in some places. This sequence comprises pre-Aptian fluvial and lacustrine clastic sediments of the Guaratiba Formation and contains petroleum source rocks.

II. from Base to Top Salt – corresponds to the Aptian evaporite layer of the Ariri Formation. Halite is the dominant component of this layer, but anhydrite and salts of potassium and magnesium are also present, sometimes showing a multilayered pattern suggestive of more than one cycle of evaporite deposition.

III. from Top Salt to Albian – corresponds to the carbonate platform deposited over the evaporites, in the beginning of the marine stage. Associated with the Guarujá and Itanhaém formations, it comprises the carbonate reservoirs of the first finds in the Santos Basin. The Turonian marine shales that correspond to the final drowning of this platform are sometimes included in the next sequence.

IV. from Albian to Santonian – corresponds to the clastic sediments that entered the basin in the open marine stage, mainly sourced by erosion of the onshore uplifted area.
V. from Santonian to Campanian – evidence from well-logs suggests that the sediments in this sequence were generally regressive whilst neighbouring basins were undergoing transgression.

VI. from Campanian to Maastrichtian – bounded on top by the K/T unconformity, and hence, are commonly eroded. Like the two preceding sequences, it corresponds to the Santos, Juréia and Itajai-Açu formations, comprising the turbidite rocks of the Ilhabela Member, which are excellent reservoirs.

VII. from Maastrichtian to Middle Eocene - bounded by a well-defined erosive surface that can be mapped all over the basin and possibly results from regional relative sea level fall. Corresponds to the Iguape and Marambaia formations, like the next two sequences.

VIII. from Middle Eocene to Miocene – corresponds to a quiet stage, not intensely deformed by halokinesis.

IX. from Miocene to Recent Seabed – this sequence is characterised by quiet deposition but shallow faults that reach the seabed can be observed in the vicinity of the large listric faults flanked by salt walls, demonstrating halokinetic activity in the basin.

Concerning the names of the interpreted reflectors, *Santonian* refers to the top of the Santonian, more precisely the base of the Campanian sequence. Accordingly, *Campanian, Maastrichtian, Middle Eocene* and *Miocene* refer to the base of the Maastrichtian, base of the Palaeocene, base of the upper Eocene and base of the Pliocene sequences. *Albian* refers to the top of the Albian sequence, sometimes also comprising Cenomanian sediments.
Although a key marker through which to understand the evolution of the post-salt structures, the top of the Turonian sequence was interpreted only on the seismic sections chosen to be restored, for it was not possible to identify this reflector in much of the study area.

Less importance was given to the interpretation of the pre-salt sequence, and the top of the basement was loosely inferred in the regional 2-D data set, due to poor seismic image and lack of well data. The base of the salt layer, however, is a very clear reflector that could be easily interpreted all over the basin. Its geometry usually mimics the one of the top salt, due to the “pull up” effect caused by the high sound velocity in salt body in contrast to much lower velocities in the adjacent sediments. Even in the depth-converted sections chosen to be restored, this artifact is still present, indicating that the velocity cube used was not perfect. The depth conversion, nevertheless, was considered a good approach, in view of the scarce well-log data that did not cover the whole basin.

Because the seismic data set used for interpretation was in time, estimated sonic velocities for the interpreted sequences (interval velocities) were used to convert the isochron maps produced into isochore maps. Accordingly, the sound velocity in water was used to convert the interpreted seabed into a bathymetric map. The faults are included in the maps, represented by gaps whose width reflects the fault heaves.

The magmatic rocks that are known to occur in the northeastern region of the study area were also identifiable through seismic interpretation. Two major magmatic pulses, one in the Late Cretaceous and the other in the Eocene (dated at ~83 and 50 Ma, respectively) are reported, but in the study area all the interpreted magmatic rocks appeared to be of Santonian age. Seismic criteria to identify and classify magmatic events in any context where they...
are associated with sedimentary sequences have been proposed by Moreira et al. (2006) and Oreiro (2006). The seismic signatures of some non-magmatic events, which comprise a range of geologic features, may be erroneously interpreted as signatures of magmatic rocks. Amongst these features are evaporites, turbidite mounds, mud volcanoes originated by gas escape, slump seismofacies and carbonate deposits (Oreiro & Guerra 2005). The purpose of this project does not include a detailed study of the magmatism in the basin. It is important to note, however, that the presence of magmatic rocks in the sedimentary cover temporarily heats the surrounding area, accelerating halokinesis in those areas affected by igneous intrusion. Moreover, where higher density magmatic rocks are present, they may represent a further weight that may add to the differential loading mechanism tending to expel the underlying salt and thus enhancing local salt withdrawal.

4.2 Results

From the interpretation of the 3-D (detail) and the 2-D (regional) seismic reflection data set, the following maps were produced:

i. Bathymetry (in metres) of the area covered by the 2-D and 3-D surveys

ii. Structural maps (TWT, in ms) of the 2-D and 3-D surveys for the following reflectors:

- Seabed
- Miocene
- Middle Eocene
iii. Isochore maps of the 2-D and 3-D surveys for the following layers:

- Seabed-Miocene
- Miocene-Middle Eocene
- Middle Eocene-Maastrichtian
- Maastrichtian-Campanian
- Campanian-Santonian
- Santonian-Albian
- Albian-Top Salt
- Top Salt-Base Salt (salt layer)
- Base Salt-Basement (pre-salt sediments)*

The maps coordinates are from the Brazilian National Grid (UTM projection).

* Only for the area of the 3-D surveys.
4.2.1 Salt and Overburden Patterns Identified in Seismic Sections

The main faults interpreted in the post-salt sequence of the study area, in the 3-D seismic surveys, are landward-dipping listric normal growth faults (figures 4.1 to 4.3). Crestal and flank collapse associated with salt diapirs and walls are frequent in the distal domain.

Most of the seismic sections shown here are located on the Top Salt structural map (cold colours are deep, hot colours are shallow).

Fig.4.1 Seismic volume depicting the post-salt faults interpreted in the area of the 3-D surveys.
Fig. 4.2 Landward-dipping listric normal growth faults dominate in the post-salt sequence (3-D surveys). The Cabo Frio Fault (CFF) is indicated by the arrow. Seismic section location is shown in the adjacent Top Salt structural map.
The Cabo Frio Fault has been the subject of previous studies (see section 3.2.1), which were wholly dependent on regional 2-D seismic data. These studies conclude that the Cabo Frio structure (i) may be controlled by a pre-existing high or reactivated fault in the substratum (Mohriak et al. 1995; Milani et al. 2005); (ii) is not a fault (Ge et al. 1997); or (iii) is controlled by reverse of salt flow due to a landward-tilting base salt (Davison 2007). Part of the objective of this chapter is to use previously unavailable, high resolution, 3-D seismic data to develop structural understanding of the Cabo Frio Fault and to evaluate the previous models.

The Cabo Frio Fault (CFF) is the most conspicuous structure in the study area. It is a large landward-dipping listric normal growth fault that soles out on the top of the salt layer and generally has a salt ridge on its footwall (figures 4.2 and 4.4). The Cabo Frio Fault controls the major depocentre of the post-salt sediments in the study area. The layers on its hangingwall are rotated against the fault plane and show large heaves increasing with depth.
The salt layer has thinned to a minimum under the hangingwall block. It may have been squeezed out by the thick prograding wedge formed by Upper Cretaceous to lower Cenozoic sediments that were supplied to the Santos Basin whilst the onshore area was uplifted. The footwall comprises a salt structural high, probably formed by salt withdrawn from under the prograding wedge. The fault activity appears to be reduced after the Middle Eocene. Seabed disturbance seen just above this structure, however, indicates that it is still active to date.

![Seismic Section Diagram]

**Fig. 4.4** The regional seismic section (in depth) depicts the extensive and compressive halokinetic domains, separated by the Cabo Frio Fault (CFF) and the adjacent (also) landward-dipping normal fault (indicated by arrows). Salt has moved basinwards, expelled from beneath the thick sedimentary wedge, and left a salt weld behind, onto which the tilted layers of the hangingwall are grounded. Those salt welds connect the pre-salt and the post-salt sequences. See map for location (Sec III).
Small-scale shallow listric normal faults were identified in the neighbourhood of the Cabo Frio Fault and other important adjacent listric faults. They seem to be very young seaward- and landward-dipping growth faults that affect only the shallow, poorly lithified strata and disturb the seabed. (figures 4.5 to 4.7). Their preferred concentration neighbouring elder and deeper, large, predominantly landward-dipping listric faults detached on the salt layer suggests that halokinesis is still active in the basin, affecting the current bathymetry and creating accommodation space.

The study area can be divided into two major halokinetic domains: a near shore extensional domain and a far shore contractional domain. The Cabo Frio Fault and its neighbouring landward-dipping faults are in the transition zone between those domains. NW- and SE-dipping secondary faults form on the rollover associated with the Cabo Frio Fault. Those faults accommodate the tilting and drag of the thick overburden against the main fault that controls the major depocentre in the area (Fig.4.8).

Salt structures grade from small-amplitude salt rollers up-dip into higher amplitude salt pillows and diapirs (or walls) down-dip. Seawards of the Cabo Frio Fault, the salt layer thickens and forms high-amplitude pillows, diapirs and ridges, flanking salt withdrawal basins, in the deeper-water region (see Fig.4.4). The salt withdrawal basins are the second most outstanding feature in the study area. They appear to result from superposed folds formed as confluent directions of sediment supply imprinted a convergent salt flow. They are important sites for deposition of sediments that overrode the troughs of the large landward-dipping listric normal faults or for sediments that have entered the basin laterally.
Fig. 4.5 Faults formed in the shallow sequences, reaching up to the Seabed (arrows). The Cabo Frio Fault is interpreted in blue. The VA technique (TecVa) was applied to enhance fault imaging. See map (Top Salt) for location.
Fig. 4.6 *Semblance* cube depicting the newly formed depocentres for young sediments associated with shallow faults in the neighbourhood of deeper faults such as the Cabo Frio Fault (interpreted).

Fig. 4.7. *Semblance-Amplitude* combination highlighting the normal faults that still occurs in the basin. Halokinesis is still active creating accommodation space for young sediments.
In the northeastern region (relative to the regional study area), a seaward-dipping listric fault has a strong expression and controls important depocentres of the post-salt sequence (Fig.4.9). In the extreme NE, where the Cabo Frio Fault is not present, it is possible to identify a series of small scale and high frequency seaward-dipping (mainly) listric growth faults that deform the Albian sequence in a raft tectonic style (Fig.4.10). This pattern is common in other salt-prone basins in the Brazilian eastern margin and in the
conjugate African western margin, but in the Santos Basin it is not identified in the Cabo Frio Fault domain.

Many seismic lines illustrate salt diapirs (or walls) associated with troughs on the seafloor above them that result from crestal collapse over a rising salt diapir or wall (Fig. 4.11). Collapse also occurs on the flanks of the salt structures. These extensional faults are found in the contractional domain, associated with buckling. Salt dissolution is expected along the crestal collapse faults.

![Image](image.png)

Fig. 4.9 A seaward-dipping listric normal fault (arrow) controls important depocentres in this area, outside the Cabo Frio Fault domain.
Fig. 4.10 In the NE region, the lower Albian sequence is deformed along a series of small-scale seaward-dipping listric normal faults, in a raft tectonics style (arrow). Volcanic cones, sills and dykes occur in the Santonian sequence in this part of the basin.
Fig. 4.11 Collapse above and on the flanks of a salt wall are common structures in the basin. The rising salt controls the deformation of the overburden. Section located on the adjacent map (grey line).

The salt layer does not always seem to be homogeneous in the Santos Basin. Instead, distinct patterns can be identified inside a salt ridge. The upper layered domain possibly consists of intercalated evaporites of different compositions, including anhydrite, whereas the lower blind domain reflects the presence of a more homogeneous rock, possibly pure halite (Demercian et al. 1993; Freitas 2006; Gambôa et al. 2008). In many salt ridges of the distal mini-basin domain, it is possible to identify deformation inside the salt layer (Fig. 4.12). Those ridges were formed with contribution of down-dip and lateral salt flow, adding to the original salt layer.
In some places, a series of salt-cored short-wavelength, low-amplitude folding develops between salt diapirs or ridges. The shortening that caused those folds may result from reduction in space whilst they sank adjacent to the rising diapirs that confine them (Fig.4.13). Another explanation could be that the early, short-wavelength folds gave way to larger-wavelength folds as the overburden thickened.

Intense halokinetic deformation has in some places promoted the tilt of the older sequences that were eventually exposed at the surface and eroded (Fig.4.14).

Depocentre shifts are identified both in the extensional domain and in withdrawal basins, where the bounding faults and the flanks of the opposite diapirs (or ridges) show different times of activity (figures 4.15 and 4.16), which suggests that the faults and diapir growth had different timing, or even that some fault has changed vergence through time. In some places depocentres are inverted. Many evidences of previously formed troughs that were later on inverted into a structural high are found in the study area (Fig.4.17).
Fig. 4.13 Salt diapirs with intervening short-wavelength, small amplitude salt domes.

Fig. 4.14 Tilted sequences (arrows) due to intense halokinesis (TecVa processing).
Fig.4.15 Shift in depocentres through time, associated with different timing of salt structures growth. The depocentres for the Santonian and Campanian sequences (limited at the top by the orange and blue horizons) were controlled by seaward-dipping faults whose activity stopped at the end of the Campanian. The Maastrichtian and Middle Eocene (green and red horizons) depocentres were controlled by landward-dipping faults. Whether these faults are new or they resulted from a change in vergence of previously seaward-dipping faults is not clear. Note that in the centre of the section the Maastrichtian to Middle Eocene sequence was strongly eroded.
Fig. 4.16 Seismic evidence for stratigraphic shift in depocentres through time from the flank of the diapir on the left to the centre of the mini-basin indicated by the black arrow. The growth strata against the diapir on the left indicate that a seaward-dipping normal fault was active during the Albian and part of the Santonian. The geometry of the upper part of the Santonian sequence and the Campanian sequence indicates that at that time this region was under shortening and the mini-basins were already established. The depocentre shift to the right from the Campanian and Maastrichtian to the Cenozoic sequences suggests that the salt diapir (or ridge) on the left started to rise faster in the Cenozoic than the opposite diapir.

Fig. 4.17 Inversion of depocentre. The area presently deformed as an anticline (arrow) was once a depocentre, as indicated by the layer growth.
In the proximal extensional domain, the salt layer has thinned so intensely that it formed a salt weld that connects the pre-salt sequence with the post-salt sediments (Fig. 4.18). These windows opened in the impermeable salt layer represent favourable fairways for hydrocarbon upward migration from the pre-salt source rocks to the post-salt reservoirs. In some places, salt welds connect the pre-salt with the Albian carbonates; in other places, the pre-salt is in direct contact with younger sequences (Santonian, Campanian and even Maastrichtian). Conversely, in the distal contractional domain, the thick salt ridges constitute a highly efficient seal for hydrocarbons accumulated in pre-salt reservoirs.

![Fig. 4.18 Salt weld (indicated by dots), resulting from the escape of salt towards less loaded regions, establishes a connection between pre-salt and post-salt sequences.](image)

The position of the halokinetic structures does not show a strict preference regarding the pre-salt structures. Salt diapirs and ridges occur either above pre-salt highs or lows (Fig. 4.19).
Fig.4.19 Salt diapirs form either above a pre-salt low or a pre-salt high. The latter appears to be more frequent, though. Care must be taken regarding the pull-up effect associated with high impedance contrasts that may cause the pre-salt features to be imaged in a false higher position.

Magmatic rocks, both intrusive and extrusive, were identified in the NE part of the study area. They occur mostly in the Santonian sequence, as volcanic cones (sometimes with identifiable feeder dyke) and saucer-shaped sills (figures 4.20 to 4.22). These rocks may have contributed to accelerate the salt flow by locally loading and heating the area.
Fig. 4.20 Magmatic rocks interpreted in the NE domain of the study area. Volcanic cones are coloured in their TWT values; sills and dykes are shown in red colours.
Fig. 4.21 Volcanic cone with feeder dyke and saucer-shaped sills.
Fig. 4.22 The Santonian magmatism is expressed in the NE area by volcanic cones, sills and dykes (in red). Feeder dykes are sometimes identified beneath volcanic cones.

The distal domain consists of a series of salt ridges formed by buckling, which evidences the shortening caused by the down-dip accumulation and deceleration of salt that has moved from under the up-dip loaded region. Sediments that managed to surpass the domain of the listric normal growth faults, or that travelled along a lateral way, accumulate in the synclines of salt-cored folds. In turn, these sediments act as a local load, forcing the salt
underneath to move towards the adjacent anticlines (Fig.4.23). The salt withdrawal thus creates more space for further sediments that reach this domain. The growth of strata within the salt withdrawal basins indicates that they are currently active and the crestal faults that cut up to the seabed attest that halokinesis is still playing a role in controlling deformation and sedimentation in the area.

Fig.4.23 Strike section across salt withdrawal basins. The shapes of the cover layers indicate that the salt withdrawal basins control depocentres at least since the deposition of the Santonian sequence. Crestal faults that reach the seabed indicate that halokinesis is still active. The arrows indicate the salt movement from beneath the thick sediments towards the adjacent salt ridges.
4.2.2 Salt and Overburden Patterns Identified in Map View

The seismic interpretation has provided important information through which to understand the structural evolution of the salt layer and the post-salt sequence. The resultant structural and isochore maps depict the halokinetic-related structural styles and sediment dispersal over the stages of basin evolution.

Although these maps may give good insights into the overall distribution of salt and sediments, they do not provide absolute thicknesses. In an attempt to estimate the sedimentary thickness in depocentres and the height of the salt structures, the isochron maps were depth converted into isochore maps, assuming a sonic velocity of 4,500 m/s for the salt layer and interval velocities that range from 2,500 m/s to 5,340 m/s for the various layers of the post-salt sequence and 4,600 m/s for the pre-salt sequence above the basement. A more precise depth conversion would require a more detailed velocity analysis (certainly problematic due to the lack of well data in large areas of the basin). For our approach, however, those assumptions were considered acceptable and indicated that the salt structures can be as thick as 8,450 m in the area (4,750 m in the area of the 3-D survey) and the overlying sequence can accumulate up to 9,600 m thick (~6,500 m in the area of the 3-D survey).

The water depths in the area range from 100 m in the shelf to 2,634 m in the distal domain. The bathymetry in this part of the Santos Basin is quite regular in the near shore domain, with the seabed dipping towards SE at about 1.3° in the upper slope. The gradient decreases in the central area, which shows some irregularities. In the far shore domain, the seabed is disturbed by N35-40E, N60-70E and N-S oriented troughs, which reflect halokinetic activity (Fig.4.24).
The bathymetry of the 3-D survey shows the seabed intensely disturbed by small faults at the slope break, and crestal collapse grabens above the salt ridges in deeper water (Fig.4.25).

The following structural and isochore maps (figures 4.24 to 4.47) illustrate the structural and sedimentary evolution of the study area. The isochore maps demonstrate the control of the halokinetic-related structures on the sediment dispersal along the geological history of the central and northern Santos Basin, highlighted by the thickening of the sedimentary sequences in salt-cored mini-basins and in faults hangingwall.

**Fig.4.24** Bathymetry (2-D survey), covering an area of 64,526 km². VE (vertical exaggeration):15. Sea water depths in the study area range from 100 m in the shelf to 2,634 m in the distal domain. The shelf and slope deposition is tranquil. At the base of slope, however, the high bathymetries and the disturbed seabed reflect active halokinesis. The black outline locates the 3-D survey shown in Fig.4.25.
Fig. 4.25 Bathymetry (3-D survey), covering an area of 10,551 km². VE:10. Colour bar shows water depth values in metres. Aside from the gently dipping slope, in the north, seabed is intensely disturbed by fault-bounded circular features, segmented faults at the slope break, and crestal collapse grabens above salt ridges in deeper water.

The Miocene structural map shows a series of small faults roughly aligned with the troughs mapped on the Seabed (Fig. 4.26). Most of them are associated with crestal collapse above salt walls, especially in the distal domain, as recognised in seismic sections. At the Miocene, the main listric faults have become dormant. The major depocentres occur in the minibasins, as indicated by the thickening of sequences in the isochore maps (figures 4.27 and 4.28).
Fig. 4.26 Miocene structural map (TWT). VE:10. Regional 2-D (A) and 3-D (B) surveys. Deformation occurs along small listric normal faults and on salt withdrawal basins. At this stage, many faults are associated with crestal collapse above salt walls.
Fig. 4.27 Seabed-Miocene isochore map. VE:5. Regional 2-D (A) and 3-D (B) surveys. The highest thickness (up to 2,400 m) for this sequence is concentrated in the very proximal and in the western area (A). Also important depocentres are controlled by N60-70E-trending faults and by salt withdrawal basins that are still active at this time.
Fig. 4.28 Miocene-Middle Eocene isochore map. VE:5. Regional 2-D (A) and 3-D (B) surveys. The major depocentres (up to 1,240 m) are controlled by salt withdrawal basins, in the eastern region.
Siliciclastic sediments are transported beyond the shelf edge into deep water by sediment gravity flow processes and are deposited on the continental slope and in the basin.

Fig. 4.29 Middle Eocene structural map (TWT). VE:10. Regional 2-D (A) and 3-D (B) surveys. Deformation occurs along small listric normal faults and on salt withdrawal basins; some faults are associated with crestal collapse above salt walls.
Fig. 4.30 Middle Eocene-Maastrichtian isochore map. VE:5. Regional 2-D (A) and 3-D (B) surveys. Major depocentres (up to 3,800 m) are controlled by the Cabo Frio Fault (arrow). Secondary depocentres are controlled by the other important NE-trending listric faults and by the salt withdrawal basins. Significant depocentres are associated with the canyons (i) that affect the slope in the NE domain (B).
Fig.4.31 Maastrichtian structural map (TWT). VE:10. Regional 2-D (A) and 3-D (B) surveys. Deformation occurs along large landward-dipping listric normal faults (CFF being the major one) and on salt withdrawal basins. Many faults are associated with crestal collapse above salt walls. Canyons desiccating the slope (i) are important sediment pathways for transporting clastics into the basin, in the NE area.
Fig. 4.32 Maastrichtian isochore map. VE:5. Regional 2-D (A) and 3-D (B) surveys. Major depocentres (up to 2,578 m) on the hangingwall to the CFF, secondary depocentres on the hangingwall to the other landward-dipping NE faults and on the salt withdrawal basins.
Fig. 4.33 Campanian structural map (TWT). VE:10. Regional 2-D (A) and 3-D (B) surveys. The Cabo Frio Fault is the dominant fault. Other landward-dipping faults occur in the deeper waters. Faults in the distal active salt withdrawal basin domain are related with crestal collapse. Relay ramps appear to connect the faults, representing favourable pathways for sediment transport into the deeper waters.
Fig. 4.34 Campanian isochore map. VE:5. Regional 2-D (A) and 3-D (B) surveys. Major depocentres (up to 3,800 m) are in the hanging wall to some listric landward-dipping normal faults, especially the Cabo Frio Fault. Secondary depocentres occur in the salt withdrawal basins.
Fig. 4.35 Santonian structural map (TWT). VE:10. Regional 2-D (A) and 3-D (B) surveys. Deformation occurs along large listric normal faults (CFF being the major one) and on salt withdrawal basins. Relay ramps form between faulted blocks; they represent sediment pathway to the deeper water domain.
Fig. 4.36 Santonian-Albian isochore map. VE:5. Regional 2-D (A) and 3-D (B) surveys. Major depocentres (up to up to 5,287 m) are in the hangingwall of the CFF and neighbouring landward- and seaward-dipping listric normal faults. Secondary depocentres form between salt ridges in the northeastern domain.
Fig. 4.37 Albian structural map (TWT). VE: 10. Regional 2-D (A) and 3-D (B) surveys. The Albian sequence is offset by large faults; it has accumulated deformation during the whole post-salt history of the basin.
Fig. 4.38 Albian isochore map. VE:5. Regional 2-D (A) and 3-D (B) surveys. The Albian sequence presents large gaps, especially in the western domain. Major depocentres (up to 4,740 m) are in the northwestern area and in the hangingwall to some listric normal faults. Also important depocentres occur in the northeastern domain.
The structural map of the Top Salt shows salt walls aligned in the NE-SW, ENE-WSW and NW-SE directions (Fig.4.39). These structures are bordered by large troughs in which sediments have accumulated through time. In the near shore domain, these structures occur on the footwall of large landward-dipping listric normal faults. In contrast, those troughs located in the far shore domain are completely surrounded by salt walls, forming perfect mini-basins. This configuration suggests that the salt has escaped from the loaded areas and progressively migrated towards the neighbouring highs. The salt-withdrawal basins represent preferential sites for deposition. Further north, some local diapirs are present.

When compared with the Seabed structural map (Fig.4.25), the Top Salt map (Fig.4.39) shows that the salt highs coincide with bathymetric lows. This is well illustrated by the seismic lines, in which crestal collapse grabens commonly occur above salt diapirs and walls, deforming the entire cover sequence up to the seabed. This indicates that the halokinetic processes are still active.

The salt diapirs and walls rose in the footwall of the main N65E and N30-40E-trending faults and on the anticlines of the salt withdrawal basins. Isolated diapirs or very narrow ridges occur aligned with N25-30W. Short wavelength and low amplitude salt domes are frequent within the salt withdrawal basin or in confined places bordered by salt walls. They may have resulted from the reduction in space within the confined basin, as the bounding salt walls were rising and their configuration may indicate variations in the local stress field. The area of the 3-D survey is divided into two domains (Fig.4.39B): the western domain, in which the Cabo Frio Fault plays a key role, and the eastern domain, outside the influence of the Cabo Frio Fault. The western domain is dominated by N30-40E and N65E-trending
Fig. 4.39 Top Salt structural map (TWT). VE:5. Regional 2-D (A) and 3-D (B) surveys. A) salt ridges show an intricate pattern in the mini-basin domain. In the distal SE area, the salt ridges display curved shape. B) the CFF plays a key role in the western domain, which is dominated by N30-40E and N65E-trending salt ridges, whilst the eastern domain, not deformed by the CFF, is dominated by N25-30W-trending salt ridges or aligned diapirs, and by a N65E-trending salt ridge (tip of the main fault indicated by black circle). Note the short-wavelength, low-amplitude salt domes within confined areas bordered by salt walls.
salt ridges, whilst the eastern domain is dominated by N25-30W-trending salt ridges or diapirs, and by a N65E-trending salt ridge. In the distal salt withdrawal basin domain, the salt ridges display an intricate pattern denoting interference. The salt ridges assume curved shapes in the distal SE area.

The isochore maps constructed for the salt layer and for the overlying sequence (figures 4.40, 4.41 and 4.44) demonstrate that the major depocentres for the post-salt sediments are located where the salt layer is very thin. On the whole, the salt layer thickens from shelf to basin and is structured mainly as elongated walls. The salt highs are in many places surrounded by regions with almost no salt left, filled with sediments, suggesting that salt has migrated from those regions to form the neighbouring high structures.

The salt layer thickens from shelf to basin. Salt escapes from under the thicker overburden, rises on the footwall of listric normal faults and forms ridges bordering salt withdrawal basins. This movement promotes up-dip extension and down-dip contraction (figures 4.40 and 4.41). The major depocentre of post-salt sequence occurs on the northern area (thinned salt layer), in the hangingwall of major listric faults and in synclines adjacent to salt diapirs and walls.
Fig 4.40 Salt isochore map (2-D), showing the extensional (A) and the contractional (B) halokinetic provinces. VE:5. Salt thickness reaches ~4.5 km and, locally, 8.5 km (high diapir in the NW). Apart from a few remnant diapirs and domes, the salt layer has thinned to a minimum in the near shore domain. The concentration of thick salt walls and diapirs in the central and distal domains indicates that salt has flown to these areas, coming from different directions. The low areas adjacent to the salt highs are favourable sites for sediment accumulation. The sediment trapped in those areas, in turn, constitute an extra load that will force the ductile underlying layer to move towards less loaded areas and also to rise through the overlying brittle sequence.
Fig. 4.41 Salt isochore map (3-D). VE:5. The western domain shows a N40E-trending salt wall (i) that rose on the footwall of the CFF, limiting the salt weld domain in the north, and approximately N45E-trending salt walls (ii), further down-dip, that rose in the footwall of listric normal faults; the eastern domain shows a N70E-trending salt wall (iii) that rose on the footwall to a landward-dipping listric fault and marks the transition to the deeper-water contractional mini-basins domain. Up-dip, a series of N30W-trending salt diapirs or thin walls (iv) formed, possibly associated with the Maastrichtian to Eocene submarine canyons (see figures 4.30 and 4.31).

The base of the salt gently dips towards the southeast. In the Base Salt structural map, the highest regions in the far shore domain coincide with the highest points on the Top Salt map and derive from pull-up effects caused by the high velocity contrast between salt and adjacent sediments (Fig.4.42). Pull-up effects may obscure the structures that affect the pre-salt session.
Fig. 4.42 Base Salt structural map (TWT). VE:5. Regional 2-D (A) and 3-D (B) surveys. Influenced both by basement and salt geometries, the Base Salt mimics the Top Salt structural map where the salt layer is thick. In the proximal area, where salt has thinned dramatically, the Base Salt shows NE-trending structures that follow the orientation of the rift faults in this domain.
Fig. 4.43 Top Basement structural map (TWT). VE:5. Deformation is controlled by NE-trending rift normal faults, in the west and northeast, and by NW-trending normal faults in the south. Again, this reflector may be influenced by the high-velocity salt layer that may cause pull-up effects in the reflectors below. This influence can be noted in the areas (indicated by arrows) that lie right beneath thick salt ridges.

The post-salt sequence comprises up to 9,600 m in the study area (more than 6,500 m in the area of the 3-D survey). The major depocentre for the post-salt sediments is located on the northern part of the study area, where the salt layer is very thin. Other significant depocentres occur in the rim synclines of diapirs and in mini-basins bordered by salt ridges (Fig. 4.44).
Fig. 4.44 Isochore map of the entire post-salt sequence (from Top Salt to Seabed), that reaches more than 9,000 m. Regional 2-D (A) and 3-D (B) surveys. VE:1. The thinnest regions (blue) correspond to high salt ridges underneath (or to a basement high, on the SW corner of the regional map).
The pre-salt sequence (from Top Basement to Base Salt) comprises up to 3,730 m of sediments (in the area covered by the 3-D survey). The major depocentres display a NE-trending direction deflecting to NW in the south (Fig.4.45), highlighting the control of the rift faults on the distribution of the pre-salt sediments.

Fig.4.45 Isochore map of the pre-salt sequence of the 3-D survey (from Basement to Base Salt). VE:1. The NE-trending normal faults in the northern domain deflect to the NW in the south. The major depocentres occur on their hangingwalls.

The entire sedimentary sequence comprises up to 9,780 m of sediments (in the area covered by the 3-D survey). The major depocentres display a NE-trending direction (Fig.4.46).
Fig. 4.46 Isochore map of the whole sedimentary infill, including the salt layer (from Basement to Seabed). VE:1. Major depocentres (up to 9,780 m) are aligned in a NE direction.

4.3 Analogy with salt-prone basins located along other divergent margins

The main driving mechanisms for the salt movement in the study area are the basin tilt caused by thermal subsidence and the differential load caused by the thick Upper Cretaceous to lower Cenozoic prograding sedimentary wedge deposited above the Aptian salt, associated with the uplift of coastal mountains. Salt escapes from the loaded near shore domain and moves towards the deeper basin, where it accumulates and creates a contractional domain, characterised by salt-cored detachment folds and intervening mini-basins. A later source of sediment supply was established in the early Cenozoic, with continental sediments entering the basin from the NNE. This was related to a change in the drainage pattern, after the Paraiba do Sul River was captured and deviated to the north, starting to discharge in the Campos
The sediment income created a loaded area in the northeastern domain, which forced the salt layer to move to the SW, also creating up-dip extension and down-dip contraction.

The Cabo Frio Fault resulted from the differential load triggered by the thick Late Cretaceous progradation of continental sediments supplied from the NW. The extensional domain is marked by seaward and landward-dipping faults and the compressional domain is marked by salt-cored detachment folds. The confluent directions of the continental sediment supply (initially from NW and in the Cenozoic shifting to NNE) have promoted a convergent flow towards the deeper basin, superposing the younger salt ridges to previously formed ones. This interference pattern resulted in the geometry of the salt withdrawal basins in the contractional domain of the study area.

An analogous scenario occurs in the Gulf of Mexico, where sedimentary differential loading is pointed as the driving mechanism for the allochthonous Sigsbee salt nappe (Humphris 1978). The Cenozoic structural evolution of the northern Gulf of Mexico is controlled by progradation of the massive Plio-Pleistocene sediment dump of the Mississippi River over deforming, largely allochthonous salt structures derived from an underlying autochthonous Jurassic salt (Lawrence & Anderson 1993; Diegel et al. 1995; Peel et al. 1995). The progradation of the Cenozoic clastic margin upon the Jurassic salt has resulted in salt evacuation producing distinct domains of extension and contraction in the basin (Fig. 4.47). A large variety of structural styles can be found in both domains, especially in the latter, where during progradation, progressive salt withdrawal from tabular salt bodies on the slope formed salt-bounded mini-basins. It is worth noticing that, like the Santos Basin, the Gulf of Mexico displays an overall concave coastal
geometry that could similarly have imprinted convergence to the salt flow (Fig.4.48).

Fig.4.47 Linked system of up-dip extension and down-dip contraction. The allochtonous Sigsbee Escarpment results from salt advance driven by differential loading due to the Cenozoic progradation (Peel et al. 1995).

Fig.4.48 The Sigsbee Escarpment, Louisiana Slope, Gulf of Mexico. The slope bathymetry is pockmarked by salt ridges and salt withdrawal mini-basins (http://www.3d-geo.com/tips_pics/04052003/hex4d.jpg on http://worldwind.arc.nasa.gov).
Comparison with the conjugate West African margin is more obvious. Geological sections from the Angolan margin show the same structural style for the salt and post-salt sequences that can be interpreted in the Santos Basin (Fig.4.49), with the important exception of raft tectonics, which is more expressive. The halokinetic-related up-dip extension and down-dip contraction with associated structures such as seaward- and landward-dipping listric faults that control the major depocentres and salt withdrawal mini-basins that influence secondary (but also important) depocentres closely resemble the structures that deform the post-salt sequence in the Santos, Campos and Espírito Santo basins, in the eastern Brazilian margin. They resulted from similar structural evolution experienced by both margins as they separated during the opening of the South Atlantic. In the Kwanza basin, thick continental sediment was carried into the basin in response to the Cenozoic reactivation and uplift of the Congo Craton with minor uplifts in the shelf (Jackson et al. 2005).

![Fig.4.49 Halokinetic-related deformation in the Kwanza Basin (Jackson & Hudec 2000).](image)

**4.4 Post-Aptian Structural and Sedimentary Evolution of the Basin**

The seismic interpretation has provided important new information by which it is possible to understand and interpret the structural evolution of the salt layer and the post-salt sequence, as well as salt-sediment interplay.
The produced structural and isochore maps depict the halokinetic-related structural styles and sediment dispersal over the stages of basin evolution.

In the study area, the pre-salt sequence is characterised by extension along essentially planar normal faults, defining a pattern of grabens and half-grabens. The depocentres of the rift sediments were limited by defined structural highs. Deformation within the post-salt sequence results from thin-skinned gravitational gliding and spreading, controlled primarily by the presence of a ductile salt layer at depth. In response to thermal subsidence and to the differential load caused by a significant prograding sedimentary wedge, salt flows basinwards, giving rise to an up-slope extensional domain, marked by landward- and seaward-dipping listric normal faults detaching on the salt layer, and to a down-slope contractional domain, dominated by salt-cored buckle detachment folds.

One of the most conspicuous structures observed in the basin is the Cabo Frio Fault, a large landward-dipping (NW-dipping) listric normal growth fault that detaches on the top of the salt layer and forms the seaward limit of a thick Upper Cretaceous prograding wedge. It has acted as a growth fault since the end of Turonian times, controlling the major depocentres in the study area, especially from the Santonian to the Middle Eocene. Sediments in the hangingwall are rotated against the fault and show a large displacement that increases with depth. The salt layer has thinned to a minimum under the hangingwall block as it has been squeezed out by the thick overlying sequence, moving down-dip and upwards. The salt ridge in the footwall of the Cabo Frio Fault resulted to a great extent from upward migration of the salt that withdrew from under the prograding wedge. The fault activity appears to be reduced after the Middle Eocene. Evidence for a disturbed
seabed just above this structure, however, indicates that it is still active to date.

During the Albian, the post-salt sequence was structured by raft tectonics, with the lower Albian sediments segmented by small, mainly seaward-dipping, listric normal growth faults, detached on the salt layer. Salt rollers rose in the footwall of the listric faults. This structural style is preserved in the northeastern area, outside the Cabo Frio Fault domain.

After the end of the Turonian, thick deposits of continental sediments prograded into the basin from a NW continental source to create a clastic wedge that squeezed out the underlying salt towards the less-loaded deeper water region. Large landward-dipping listric normal growth faults (including the Cabo Frio Fault) appeared in the transition zone between the extensional and the contractional domains. The extensional domain was characterised by listric normal faults (dominantly landward-dipping) with salt rollers, diapirs and walls in the footwall block, and by rotation of the overlying sequence in the hangingwall block, forming roll-overs, alongside with many salt welds that directly connected the pre- and the post-salt sequences. In the distal contractional domain, buckle detachment folds appeared to accommodate down-dip shortening. The up-dip listric normal faults strongly controlled the major depocentres from the Santonian to the Middle Eocene, whereas the down-dip salt-bounded mini-basins controlled the secondary depocentres, trapping sediments that managed to travel further offshore or that came by a lateral pathway.

In the early Cenozoic, continental sediment supply came from a second source area, located in the NNE and established another configuration of differential loading that interfered with the previous system. The salt-cored
ridges displayed an intricate pattern and bounded intervening polygonal salt withdrawal basins.

Two phases of folding were identified in the central and distal domains of the study area. A first generation of NE folds involved horizontal shortening accommodated by buckle folding, which compensated the up-dip extension. A second generation of folds formed with ESE direction and was superposed with the previously formed folds, creating a complex interference pattern that may explain the geometry of the salt withdrawal basins domain. Both folding phases resulted from gravity-driven thin-skinned tectonics, strongly controlled by halokinesis. This conclusion is consistent with analogue models results (Guerra et al. 2005c), which showed that convergent prograding wedges over a ductile layer (analogue for salt) causes interference of “salt”-cored folds and results in the creation of polygonal mini-basins (see figures 3.14 to 3.16).

The isochron maps constructed for the salt layer and for the overlying sequence complement the seismic sections in unravelling the salt-sediment interplay. Salt structures grade from small-amplitude salt rollers up-dip to high-amplitude salt pillows and diapirs (or ridges) down-dip. The main depocentre for the post-salt sediments lies in the northern part of the study area, where the salt layer is very thin. The hangingwalls of the listric normal faults, especially the Cabo Frio Fault, represent preferential sites for deposition. As a whole, the salt layer thickens towards the deeper water region. In this domain, the post-salt sequence is deposited into salt withdrawal mini-basins bordered by salt diapirs or walls. Many interpreted sections illustrate both the salt wall and a trough on the seabed above it, resulting from crestal collapse, which indicates active halokinesis.
Syn-kinematic growth of strata is less evident in the Neogene than it used to be from the Late Cretaceous to the Middle Eocene. Since the Late Eocene the sedimentation is roughly calm and does not show much influence from halokinesis, except in areas affected by shallow faults such as crestal collapse grabens above salt ridges and shallow listric faults that deform the upper sedimentary sequence including the seabed. Although restricted to the shallower sequences, these faults lie in the neighbourhood of important deeper landward-dipping listric faults detached on the salt layer, hence suggesting that halokinesis is still active.

Different patterns of deposition are recorded in the post-salt sequence, from growth syntectonic rollovers, to prograding and aggrading sequences and submarine fans. Halokinesis created space for sediment deposition, and sediment load, which in turn enhanced salt movement, in a feedback process. Magmatic rocks in the northeastern area, dated at the Late Cretaceous and early Cenozoic, may have locally accelerated the halokinetic process by temporarily heating the area and by adding an extra load to the overburden.

Seismic interpretation provided powerful information to improve the understanding of salt-sediment interplay in a large area of the Santos Basin. It is important to note, however, that the sections and maps based on seismic interpretation illustrate the present configuration of each interpreted horizon and isochores for individual sequences, which when combined with restoration allow their true position, geometry and geological evolution of the basin to be determined.
Chapter 5

Cross-section Restoration

Cross-section restoration is an important tool to help to define the geometry of the geological layers and the structures that dissect them through time, and validate a geological model by restricting alternative interpretations. This technique is adopted in the oil industry for its results contribute to improve seismic interpretation and modelling of petroleum systems, increasing the confidence in the geological model and thus reducing the risks involved in exploration. The method deployed consists in sequentially removing the effects of compaction and deformation towards the past in order to take the section to its undeformed state.

The sequence of restored sections provides insights on the structural and stratigraphic evolution of the area and indicates which structures have been active controlling depocentres in the Santos Basin at each stage.

5.1 Brief History

Cross-section restoration techniques were initially developed to estimate and quantify the shortening in fold-and-thrust belts, study the orogenic evolution and determine the depth to the décollement level. Bally et al. (1966) and Dahlstrom (1969) were pioneers in discussing the methods and principles to produce balanced (palinspatically-restored) sections. Later on, during the
80's, the restoration methods were adapted to extensional settings (Gibbs 1983; Elliot 1983), and were widely used in petroleum exploration. Worral & Snelson (1989) extended the use of restoration to areas affected by halokinesis in an attempt to understand and balance observable up-dip extension with down-dip contraction.

5.2 Concepts

A balanced geological section has no change in the layer length or area between undeformed and deformed states. The principle of area conservation requests that no rock material enter or leave the plane of the geological section. During the restoration process, plane strain is assumed, so the section is chosen in the direction of the tectonic transport.

According to Elliot (1983), if a geological section can be restored to its undeformed state, it represents a viable model. The following grades of confidence were defined:

1. Non-balanced section - preliminary investigation showing inferred structures.

2. Non-restorable section - section chosen in an inadequate orientation. This could be a viable section but its interpretation cannot be validated by restoration.

3. Viable (restorable) section - oriented in the direction of tectonic movement.

4. Admissible section - comprises structures that can be found in nature.
5. Valid balanced section - viable and admissible. Although it does not constitute the unique solution, the integration of data from various sources (wells, surface geology, geophysics, etc.) may reduce alternative interpretations.

Only after being balanced, a geological section can be considered to represent a viable and admissible model.

5.3 Deformation and Restoration Mechanisms

Different models were proposed to determine the geometry of the faults in the restored sections, in cases where deep structures are not well imaged. Verral (1981), Gibbs (1983), Suppe (1983), White et al. (1986), Williams & Vann (1987), Rowan & Kligfield (1989), Xiao & Suppe (1989, 1992) and Dula (1991) discussed the mechanisms of deformation and techniques that could be used to restore cross-sections deformed by listric normal faults. Suppe (1983), Suppe & Medwedeff (1990), Shaw et al. (1999, 2004) and Corredor et al. (2005) discussed the deformation mechanisms involved in thrusts and fault-related folds.

The restoration mechanisms try to reproduce the kinematics of deformation. In fold belts, flexural slip tends to be the major deforming mechanism. Thrust belts can display fault-bend folds and fault-propagation folds that may be restored by shear along faults. Extensional settings are mainly structured along planar or listric normal faults. Planar rotational faults (domino style) cause both the hangingwall and the footwall blocks to rotate, with no internal deformation in the faulted blocks. These are typical geometries of rift systems. Blocks deformed by planar rotational faults may be restored by a
combination of rigid translation and rotation. Listric normal faults cause rotation of the hangingwall layers. These structures are commonly found in basins deformed by halokinesis, where the listric faults detach on the top of the ductile salt layer. Simple shear mechanism is proposed for extensional rollovers. Shear angles and mechanisms are suggested according to different authors: (i) flexural slip (Gibbs 1983; Davison 1986; Moretti et al. 1988; Rowan & Kligfield 1989), (ii) simple shear along vertical lines (Verrall 1981), (iii) antithetic simple shear (White et al. 1986; Crespi 1988; Dula 1991), and (iv) synthetic and antithetic shear (Rowan 1987; Schultz-Ela 1992), and (v) fault-bend folding (Suppe 1983; Groshong 1989).

5.4 Restoration and Halokinesis

Due to the high mobility of the salt layer, the amount of salt present in section varies considerably through time. Salt flows towards the deeper basin and laterally (into and out of the plane of any section) and also rises through the overlying sediments to form domes and diapirs. Therefore, salt does not obey the basic balancing premises (the movement of the ductile layer does not always follow the transport direction of the brittle sequences and the area is not preserved). Moreover, Hossak & McGuinness (1990) addressed important questions concerning the use of restoration in halokinetic areas, for it is not easy to differentiate extension from salt withdrawal. Cross-section restoration in a basin deformed by salt tectonics is thus controversial, and this must be taken into account when analysing the results. Nevertheless, this method has been increasingly adopted for halokinetic-controlled basins. Rowan (1993) proposed a systematic method for sequential restoration of salt
structures, which consists of calculating and removing the effects of processes such as sedimentation, compaction, isostatic adjustment (Airy), thermal subsidence, faulting, and salt withdrawal/diapirism. It makes no assumptions about salt kinematics and generally results in the area of the salt layer changing through time. This method was partly adopted to restore cross-sections in areas of salt tectonics where deformation is confined to the salt and upper layers (Rowan 1993), and thus its basic principles were adopted for the restorations in this work. We use a slightly different method that includes contractional domains and considers flexural isostasy rather than the Airy’s model.

5.5 ReconMS

The proprietary ReconMS computer program has been developed by Petrobras and TecGraf (PUC-Rio). It permits to restore multiple geological sections simultaneously and to verify the consistency of their structural interpretation. Its development has benefited from the restorations carried out during this project.

The restoration involves layer-by-layer removal of the stratigraphic sequence, accompanied by flexural backstripping and sediment decompaction, as the section is unstrained towards the past.

The input data are interpreted depth-converted seismic sections, lithology, density, porosity curves and palaeobathymetry for each geological layer. ReconMS has recently been implemented to consider lateral facies variations, and this improvement allowed for more realistic decompaction along the regional sections.
To correctly perform decompaction, it is necessary to remove the effects of compaction. The change in porosity with depth is given by the equation:

$$\phi = \phi_0 e^{-cz},$$

where $\phi$ is the porosity at depth $z$, $\phi_0$ is the initial porosity of the sediment, and $c$ is the constant of decay, which describes the rate of porosity decrease with depth (Athy 1930).

By decompacting, closing the faults displacement and unfolding, the section is restored to its undeformed state.

**5.6 Cross-section Restoration in the Santos Basin**

The seven regional selected seismic sections are NW-SE oriented and cover the study area (Fig.5.1). They were depth converted applying Landmark's SeisWorks software and were loaded into ReconMS, with their UTM coordinates specified. Although transfer faults might exist in the basin, the regional cross-sections were chosen in the best possible direction, and they showed good results.

**5.6.1 Method**

- The post-salt sequence was restored to the age of the following horizons:
  
  i - Miocene
  
  ii - Middle Eocene
  
  iii - Maastrichtian
iv - Campanian

v - Santonian

vi - Turonian

vii – Albian

Although the Turonian reflector could not be identified over the entire study area, it was tentatively interpreted in the seven cross-sections selected for restoration, for it is considered a key horizon in the structural evolution of the basin.

- Image files of the geological sections, always in depth, were imported by the program (with vertical and horizontal scales equal). The sections were georeferenced with their UTM coordinates used as primary locators.

- The key horizons and faults were digitised. Properties such as porosity, constant of porosity decay with burial, density and lithology were attributed to each geological layer. Important lithology-dependent differences occur in compaction behaviour; therefore the individual porosity-depth relationships were determined for each lithology\(^1\). Palaeobathymetries used in the restorations were based on Cretaceous palaeobathymetric maps (Viviers et al. 2005).

- Before each stage of restoration, the sections were decompacted by removal of the upper (younger) layer, considering porosity decay with depth (based on Athy’s law) and flexural backstripping (as a function of overburden thickness, elastic thickness of lithosphere, and densities of mantle, overburden and water).

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\(^1\) Data kindly provided by Laury Medeiros (Petrobras geologist), taken from the basin modelling software Petromod.
• The section was subdivided into modules that encompass rock bodies submitted to the same type of deformation, so that they could be restored by the same deformation mechanism.

• Each module was restored to its undeformed state, being rotated, translated and unstrained (by applying deformation mechanisms such as simple shear and flexural slip), trying to attain the best fit to the adjacent modules in the result section. Except for the salt layer, the module area was kept constant during the restoration process.

• The salt layer was removed before the restoration mechanisms were applied to the overlying sequence. Afterwards, the salt layer occupied the remaining space between the pre- and post-salt sequences.

• Different scenarios were tested for each restoration.

• The process was repeated for the older stages.

• The main faults active at each stage were identified; their names refer to the same fault at all the restoration stages of a certain section. Faults with the same name in neighbouring sections are not necessarily correlated.

• The position of the Cabo Frio Fault (CFF) - or the major landward-dipping fault (F) in the adjacent sections (I, VI and VII) - was measured at each section for every restoration stage.

• Elongation (e) was calculated for each restoration stage in each section as

\[
e = \frac{l - l_0}{l_0},
\]

where \(l\) is the final length and \(l_0\) the original length of the section.
5.6.2 Restoration Results

Sequential restoration of seven regional cross-sections (Fig.5.1) along the Santos Basin reveals the structural and sedimentary evolution of the study area since the deposition of the Albian carbonates over the Aptian salt layer and throughout the late Cretaceous and Cenozoic, indicating fault activity and depocentre controls at each stage.

Timing of faults activity, depocentre migration and opening of "salt windows" is of ultimate importance to the understanding of the petroleum system of the basin and, therefore, may contribute to reduce the risks involved in exploration.

The simultaneous observation of several regional restored cross-sections through time permits a good three-dimensional analysis of the post-salt tectonic evolution of the study area (Fig.5.2). The multi-section restoration produces scenarios that can be regarded as a series of vertical sections cutting through palaeostructural and palaeobathymetric maps. They give the true isopachs at each restoration stage, since the sections have been decompacted and backstripped as the deformation was being removed towards the past.

Due to the loss of rock material along fault planes through continued deformation over the last 100 my, and also because at the beginning of the drift phase the direction of movement could have been different, the Albian sequence may be highly incomplete in the restored sections.
Fig. 5.1 Location of the seven restored cross-sections on the Top Salt structural map.
Fig. 5.2 Sequence of restoration along seven regional cross-sections in the Santos Basin. Georeferenced interpreted depth-converted seismic sections (a). Sections restored to the Albian (b), Turonian (c), Santonian (d), Campanian (e), Maastrichtian (f), Middle Eocene (g) and Miocene (h). The grey area (b to h) indicates the current geometry of the pre-salt sequence, over which salt has spread and the post-salt sequence has gravitationally glided since the Albian until reaching the present-day configuration (i). Note that in sections II to V, where the Cabo Frio Fault is well characterised, with the largest heaves, the elongation is higher than in the adjacent sections.
Chapter 5 - Cross-section restoration
Chapter 5 - Cross-section restoration
Chapter 5 - Cross-section restoration

Maastrichtian
65.5 Ma
Chapter 5 - Cross-section restoration
Chapter 5 - Cross-section restoration

Miocene
5.3 Ma
Chapter 5 - Cross-section restoration
5.6.2.1 Section I

5.6.2.1.1 Albian

Depocentres for this sequence are located in the NW portion (left hand side) of the section, with thicker regions controlled by F2, F3, F5 and F7 (Fig.5.3).

![Fig.5.3 Sec-I restored to the Albian. Depocentres are in the central and NW domains, associated with seaward- and landward-dipping listric normal faults.](image)

5.6.2.1.2 Turonian

Major depocentres developed in the central and NW domains, during the Turonian, either on the hangingwall of seaward and landward-dipping listric normal faults or in the rim synclines of salt diapir S1 (Fig.5.4). Faults F1, F2 and F4 ceased their activity before the end of Turonian. Nevertheless, F1 has exerted a local control during part of the Turonian deposition. Comparing the sections restored to the Albian and to the Turonian, it appears that F7 has changed vergence, as suggested by the inversion of depocentres from one side of diapir S4 to the other. No significant activity was recorded in the SE
domain. Elongation for this section is 2.1% in relation to the Albian section (Fig.5.11).

![Diagram](image)

Fig.5.4 Sec-I restored to the Turonian. Major depocentres are in the central and NW domains: in the hangingwall of seaward- and landward-dipping listric normal faults and in the rim synclines of salt diapir S1.

5.6.2.1.3 Santonian

At this stage, the mini-basins domain has already been established, in the SE area (Fig.5.5). At the same time, a large landward-dipping listric normal fault (F11) has developed controlling the major depocentre of this sequence. F8 appears to have changed vergence since the Turonian stage, for the depocentre has shifted to the other side of the salt diapir S5. F5 activity seems to have ceased during the Santonian. Elongation for this section is 16.4% in relation to the Albian section (cumulative) and 14% in relation to the Turonian section (Fig.5.11).
Fig.5.5 Sec-I restored to the Santonian. Depocentres are in the central and NW domains, associated with seaward- and landward-dipping listric normal faults, and in the mini-basins domain. Major depocentres in the hangingwall of F11; significant depocentres in the newly formed syncline between S2 and S3; and in the west flank of S1.

5.6.2.1.4 Campanian

Depocentres are controlled by seaward- (in the central domain) and landward-dipping faults (in the NW domain), and by the salt withdrawal mini-basins (in the SE domain). The major depocentres are controlled by F11, F8 and F15 (Fig.5.6). Elongation for this section is 35.4% in relation to the Albian section (cumulative) and 16.4% in relation to the Santonian section (Fig.5.11).

Fig.5.6 Sec-I restored to the Campanian. Depocentres are in the central and NW domains, associated with seaward- and landward-dipping listric normal faults, and in the mini-basins domain.
5.6.2.1.5 Maastrichtian

Aside from F6, which controls one of the major depocentres, landward-dipping faults dominate at this stage of the restorations (Fig.5.7). F13, F14 and F also control important depocentres. Secondary depocentres are controlled by F9 and by the syncline between S5 and S6. The folded salt withdrawal basins domain continues to capture sediments on the fold synclines. The large landward-dipping listric normal fault (F) has developed during this stage and may be correlated with the Cabo Frio Fault (though considerably younger), maybe a late propagation of the CFF to the west. The diapir S2 continued to rise but F11 ceased its activity. F16 is incipient at this stage. Fault F7 appears to have changed vergence since the Campanian to turn into F (or another SE-dipping fault existed at this stage controlling depocentres that later on shifted to the opposite flank of the diapir). Crestal collapse faults are common in the distal SE domain. Elongation for this section is 46.4% in relation to the Albian section (cumulative) and 8.1% in relation to the Campanian section (Fig.5.11).

Fig.5.7 Sec-I restored to the Maastrichtian. Aside from F6, which controls the major depocentre, landward-dipping faults dominate at this stage. Braces indicate regions faulted by collapse on crest and flanks of diapirs or ridges.
5.5.2.1.6 Middle Eocene

The major depocentre is controlled by fault F. Important depocentres are located in the rim synclines of S2, S4 and S5 (Fig.5.8). Buckling is very intense in the mini-basins domain. Crestal collapse faults are abundant on top of S2, S5 and in ridges of the SE region. F16 displays clear activity during this stage whilst activity along F13, F6 and F14 greatly decreases. A salt weld tends to form in the proximal domain, connecting the pre-salt and the post-salt sequences. Elongation for this section is 54.9% in relation to the Albian section (cumulative) and 5.8% in relation to the Maastrichtian section (Fig.5.11).

![Fig.5.8 Sec-I restored to the Middle Eocene. Depocentres are in the hangingwall of F and in the rim synclines of S2, S4 and S5. Braces indicate regions faulted by collapse on crest and flanks of diapirs and ridges. The pair of dots indicates a salt weld.](image)

5.6.2.1.7 Miocene

The major depocentre is in the proximal region, to the west of S2 (Fig.5.9). Although not controlling depocentre, F is still active. Crestal collapse faults are abundant above S2, which is still rising, and in ridges of the SE region. Activity of F17 starts and F13 is no longer active at this stage. Elongation for
this section is 56.5% in relation to the Albian section (cumulative) and 1% in relation to the Middle Eocene section (Fig.5.11).

Fig.5.9 Sec-I restored to the Miocene. The major depocentre lies in the proximal domain, to the west of S2.

5.6.2.1.8 Recent

Some faults are still active disturbing the seabed, and the salt withdrawal mini-basins still control local depocentres. Nevertheless, sediment accumulates mostly by aggradation in the NW domain. In this section, no landward-dipping fault would clearly correlate with the Cabo Frio Fault (Fig.5.10). A candidate to be the lateral continuation of CFF would be fault F, although it is much younger and has no significant heaves for the faulted horizons. Although not controlling expressive depocentres, F10 has remained active since the Santonian (and maybe even Turonian). F3 has been active since the Turonian; F6 and F9 are active since the Albian. F7 was active in the Albian and appears to have changed vergence in Maastrichtian becoming F. Of all these long-lived faults, F6 and F7 are the only ones that played a major role in controlling depocentres along the structural evolution recorded in this cross-section. Other faults and diapirs that controlled depocentres had shorter activity.
Halokinesis has decreased in intensity but still exerts control on sedimentation. F is still active controlling depocentres, as well as F10 and F17, which deforms the present-day seabed. Salt continues to move upwards and laterally creating troughs in which sediment accumulates. A new salt weld formed in the proximal domain, immediately above a pre-salt fault, connecting the pre-salt and the post-salt sequences. It is located directly above a pre-salt fault and a basement high, constituting a favourable scenario for hydrocarbon migration to the post-salt reservoirs.

Fig. 5.10 Sec-I - Recent. Halokinesis has decreased in intensity but still exerts control on sedimentation. A new salt weld formed in the proximal domain, immediately above a pre-salt fault, connecting the pre-salt and the post-salt sequences.

Elongation for this section is 57.6% in relation to the Albian section (cumulative) and 0.8% in relation to the Miocene section, indicating that gravitational gliding and spreading over salt is still occurring (Fig. 5.11). Although having decreased in intensity when compared with the late

Chapter 5 - Cross-section restoration
Cretaceous to early Cenozoic strongly halokinetic stages, lateral flow of the salt layer is a process that evidently remains active in the Santos Basin.

Fig.5.11 Sec-I: Graphs illustrate the elongation for the restored sections relative to the previous restoration stage or relative to the Albian section (cumulative).
Fig. 5.12 Sec-I: Fault F migrated to the deeper SE basin. The change in position was more intense in the early stages and decreased significantly after the Maastrichtian (K/T), tending to cease towards the Middle Eocene (ME) and then to recommence after the Miocene (M).
5.6.2.2 Section II

5.6.2.2.1 Albion

Depocentres for this sequence are located in the NW half of the section, with thicker sedimentation controlled by the listric normal faults F1, F2 and F3 (Fig.5.13).

![Fig.5.13 Sec-II restored to the Albian. Active faults at this stage are shown. F1, F2 and F3 control major depocentres.]

5.6.2.2.2 Turonian

Depocentres were mostly located in the NW half of the section during the Turonian, with thicker regions controlled by ‘CFF’ (relatively incipient at this stage), F3 and possibly F1, whose activity has ceased before the end of the Turonian. It seems that the mini-basins activity has also started at this stage, with salt structures S6, S7 and S8 locally controlling depocentres on its rim synclines (Fig.5.14). Elongation for this section was 1.9% in relation to the Albian section (Fig.5.21).
5.6.2.2.3 Santonian

The major depocentre lies in the hangingwall of the Cabo Frio Fault (CFF). Also important depocentres are controlled by F3 and F10. Secondary depocentres are controlled by F5, F6, F8 and F9. Synclines between salt diapirs or ridges also control depocentre at this stage (Fig.5.15). Elongation for this section is 22.9% in relation to the Albian section (cumulative) and 20.6% in relation to the Turonian section (Fig.5.21).
5.6.2.2.4 Campanian

The major depocentre is controlled by the CFF. Important depocentres are controlled by the synclines formed between S1 and S2 and, in the distal domain, between S5 and S6. The activity along F11, F12 and F6 have ceased during this stage (Fig. 5.16). Elongation for this section was 54.8% in relation to the Albian section (cumulative) and 26% in relation to the Santonian section (Fig. 5.21).

![Diagram of Campanian section](image)

*Fig. 5.16 Sec-II restored to the Campanian. Major depocentres are controlled by the CFF. F5, F6, F11 and F12 are active, controlling minor depocentres. Important depocentres are controlled by salt structures S1, S2, S5 and S6.*

5.6.2.2.5 Maastrichtian

The major depocentre for the Maastrichtian sediments is strongly controlled by the CFF and secondary depocentres are controlled by faults F6 and F13. Rim synclines of the salt ridges S1, S2, S3, S4, S5, S6, S7 and S8 also control depocentres. Crestal collapse faults are common above S2 (Fig. 5.17). Elongation for this section was 70% in relation to the Albian section (cumulative) and 9.8% in relation to the Campanian section (Fig. 5.21).
5.6.2.2.6 Middle Eocene

The major depocentre is controlled by CFF, whose activity appears to have ceased during this stage or shifted eastwards to the position indicated by F (Fig.5.18). S1 has deflated since the Maastrichtian and the ancient syncline between S1 and S2 that has controlled important depocentres for the Campanian and Maastrichtian sediments is now tilted along the NW flank of S2. The mini-basin domain is in full activity during the Middle Eocene, capturing sediment in the fold synclines. Crestal collapse is frequent in this domain. Activity of F6 seems to have ceased before the end of this stage. The salt layer thickness is reduced to a minimum under the hangingwall of the CFF and salt welds tend to appear, locally connecting the pre-salt with the post-salt sequences. At least the westernmost salt weld is aligned with a pre-salt fault, implying that a promising migration fairway formed at this time. Elongation for this section is 74.6% in relation to the Albian section (cumulative) and 2.8% in relation to the Maastrichtian section (Fig.5.21).
Fig. 5.18 Sec-II restored to the Middle Eocene. The major depocentre is controlled by the CFF that may have ceased its activity during this stage or shifted to the site indicated by F. Salt withdrawal mini-basins are in full activity controlling depocentres in the SE domain. Activity along F6 appears to have stopped before the end of the Middle Eocene. Salt tends to weld at locations indicated by a pair of dots. Crestal collapse faults (indicated by braces) are frequent above salt ridges, especially S2 and those on the distal domain.

5.6.2.2.7 Miocene

The major depocentres at this stage are located in the salt withdrawal mini-basins domain and in the through related with the CFF (Fig. 5.19). Crestal collapse grabens are frequent in the distal domain, as well as above S2. The westernmost salt weld has remained approximately at the same position since the Middle Eocene, aligned with a pre-salt fault, so the migration fairway still exists. Elongation for this section is 75.2% in relation to the Albian section (cumulative) and 0.4% in relation to the Middle Eocene section (Fig. 5.21).
Fig. 5.19 Sec-II restored to the Miocene. Depocentres controlled by CFF and salt withdrawal basins. Salt tends to weld at locations indicated by pairs of dots.

5.6.2.2.8 Recent

The major depocentres are still controlled by the CFF (or F) and by the salt withdrawal basins, depicting the long-lived halokinetic activity (Fig. 5.20). Apart from the salt weld that appears to have formed between S5 and S6, the salt welds have not significantly changed since the Miocene. That salt weld formed in the distal domain is aligned with a basement high and a fault in the pre-salt sequence. This is a promising scenario for hydrocarbon migration from pre-salt source rocks to post-salt reservoirs. It must be taken into account that it formed after the Miocene so it is necessary to know if this matches the timing of generation and expulsion. On the other hand, the salt weld implies that at this stage a salt seal is no longer effective for pre-salt reservoirs. Elongation for this section is 76% in relation to the Albian section (cumulative) and 0.5% in relation to the Miocene section (Fig. 5.21).
Fig. 5.20 Sec-II - Recent. Disturbed seabed is suggestive of recent fault activity. Depocentres are controlled by F and salt withdrawal basins. Apart from the salt weld that appears to have formed between S5 and S6, the salt windows are the same as in the Miocene. Note that this distal salt weld is aligned with a basement high and a fault in the pre-salt sequence.
Fig. 5.21 Sec-II: Graphs illustrate the elongation for the restored sections relative to the previous restoration stage or relative to the Albian section (cumulative).
The Cabo Frio Fault has migrated to the deeper water basin (Fig.5.22) imposing gaps in the faulted horizons that are registered in Fig.5.23. The overall activity of this fault reached a maximum from the Campanian to the Middle Eocene and then was dramatically reduced, but never ceased. Since the Campanian, the gaps in the Albian sequence along the Cabo Frio Fault have been very large: more than 40 km in the Campanian, reaching 60 km in the Maastrichtian and over 60 km at present. The Turonian follows the same pattern, only with slightly smaller heaves. The Santonian also shows large heaves, ranging from about 32 km in the Campanian to 51 km at present. Since the Maastrichtian, the displacement for the Campanian horizon has remained constant (~20 km). No significant gaps were found for the Maastrichtian.

Fig.5.22 Sec-II: The Cabo Frio Fault migrated to the deeper SE basin. The change in position was intense in the early stages and decreased significantly after the K/T boundary, tending to cease after the Middle Eocene (ME) and then to recommence after the Miocene (M).
Fig. 5.23 Sec-II: Horizontal component of displacement along the Cabo Frio Fault at each restoration stage. The heaves for the Albian, Turonian, Santonian and Campanian horizons increase up to the Middle Eocene and then remain stable. Gaps in the Campanian horizon were kept constant at ~20 km. No significant gap was identified in the Maastrichtian horizon.
5.6.2.3 Section III

5.6.2.3.1 Albian

This sequence was deformed in blocks segmented by listric normal faults that detach on the salt layer (Fig.5.24). A considerable part of the Albian may have been lost due to continued deformation throughout the drift phase.

Fig.5.24 Sec-III restored to the Albian. Depocentres are controlled by seaward- and landward-dipping listric normal faults in the proximal (NW) domain.

5.6.2.3.2 Turonian

Turonian depocentres are mainly controlled by F1, F3, F7 and F8. The Cabo Frio Fault ('CFF') was incipient at this time (Fig.5.25). Elongation for this section was 5.2% in relation to the Albian section (Fig.5.32).
Fig.5.25 Sec-III restored to the Turonian. Active faults are shown. Incipient CFF appear in the site of former F4.

5.6.2.3.3 Santonian

At this time, the Cabo Frio Fault is very active and controls the major depocentre. Important depocentres are also controlled by F5, F6, F7 and F8, which detach on the flanks of salt ridges S2, S3, S4 and S5. Activity in the salt withdrawal basins is intense (Fig.5.26). Elongation for this section was 39.1% in relation to the Albian section (cumulative) and 32.2% in relation to the Turonian section (Fig.5.32).

Fig.5.26 Sec-III restored to the Santonian. The CFF and the mini-basins are intensely active and control important depocentres. Faults F3, F5, F6, F7 and F8 also control depocentres.
5.6.2.3.4 Campanian

The major depocentre is controlled by the CFF (Fig.5.27). Important depocentres are controlled by F3, F5 and the syncline between S2 and S3. F8 appears to have changed vergence since the Santonian stage, as suggested by the depocentre that shifted to the opposite side of S5. Elongation for this section is 58.5% in relation to the Albian section (cumulative) and 13.9% in relation to the Santonian section (Fig.5.32).

![Fig.5.27 Sec-III restored to the Campanian. The major depocentre is located in the hangingwall of the CFF. Secondary depocentres are in the hangingwall of F3 and F5 and in the synclines of the salt withdrawal domain, especially between S2 and S3.]

5.6.2.3.5 Maastrichtian

The major depocentre is in the hangingwall of the CFF (Fig.5.28). Important depocentres are in the hangingwall of F3 and F5 and also in the salt withdrawal basins. Again, the F8 vergence appears to have changed. Salt ridge S2 has become much wider than it used to be in the Campanian.
Collapse faults at the crest and flanks of the salt ridges are frequent. Elongation for this section was 97.9% in relation to the Albian section (cumulative) and 24.9% in relation to the Campanian section (Fig.5.32).

![Fig.5.28 Sec-III restored to the Maastrichtian. CFF controls the major depocentre. Other important depocentres are controlled by the salt withdrawal basins and by F3 and F5. Collapse on the crest and flanks of the salt structures are frequent (indicated by braces).]

### 5.6.2.3.6 Middle Eocene

The CFF controls the major depocentre (Fig.5.29). Secondary depocentres are controlled by F5, F8 and by the salt withdrawal basins. At this time, the thickness of the salt layer is intensely reduced in the proximal domain, tending to form a long salt weld that creates a large area connecting the pre-salt and the post-salt sequences. Pre-salt faults and basement highs occur in this area in which the salt may no longer act as a seal, therefore opening excellent hydrocarbon fairways. Crestal and flank collapse faults are frequent in the mini-basins domain. Elongation for this section was 115.2% in relation to the Albian section (cumulative) and 8.7% in relation to the Maastrichtian section (Fig.5.32).
Fig.5.29 Sec-III restored to the Middle Eocene. The major depocentre is on the hangingwall of the CFF and other important depocentres are on the hangingwall of F5 and F8, as well as in the synclines of the mini-basins domain. Salt welds were formed at this stage in the proximal domain.

5.6.2.3.7 Miocene

At this stage, the CFF still has some activity, but very subtle (Fig.5.30). Although the mini-basins are still active controlling depocentres, the thickest deposits are those in proximal domain, accumulated by aggradation, with no evident structural control; they. Newly formed salt welds appear under the hangingwall of the CFF. Elongation for this section was 114.9% in relation to the Albian section (cumulative) and -0.1% in relation to the Middle Eocene section, which indicates shortening but may be due to lack of rock material caused by erosion (Fig.5.32).
Fig. 5.30 Sec-III restored to the Miocene. CFF and mini-basins domain are still active, although not controlling very thick deposits. Braces depict sites of faults associated with collapse on top and flanks of the salt diapirs (or ridges).

5.6.2.3.8 Recent

Disturbed seabed in the vicinity of the CFF indicates the present activity of this fault (Fig. 5.31). The seabed is also deformed by crestal collapse grabens formed above all the salt ridges in this section. The existence of faults connecting the salt bodies with the ocean strongly favours dissolution that would account for significant salt losses to the sea. Salt welds in the proximal domain are favourable scenarios for hydrocarbon migration towards the post-salt reservoirs. The thinning of the salt layer between S1 and S2 and between S2 and S3 will eventually lead to the opening of salt windows at the base of these salt withdrawal basins. Elongation for this section was 115.8% in relation to the Albian section (cumulative) and 0.4% in relation to the Miocene section (Fig. 5.32).
Fig. 5.31 Sec-III: Recent. Disturbed seabed in the vicinity of the CFF indicates the present-day activity of this fault. The seabed is also deformed by widespread crestal collapse grabens. Salt welds are present at the proximal domain and eventually two salt welds will form at the base of the mini-basins flanked by S1, S2 and S3.
Fig. 5.32 Sec-III: Graphs illustrate the elongation for the restored sections relative to the previous restoration stage or relative to the Albian section (cumulative).
The Cabo Frio Fault has migrated to the deeper water basin (Fig. 5.33) imposing gaps in the faulted horizons that are registered in Fig. 5.34. The overall activity of this fault reached a maximum from the Campanian to the Middle Eocene and then was dramatically reduced, but it never ceased.

The gaps in the Albian and Turonian sequences along the CFF have increased up to the Middle Eocene, when they reached about 42 km and then remained stable for the Albian and slightly decreased for the Turonian. The Santonian follows the same pattern as the Albian, but with smaller heaves (ranging from 10 to 32 km). The heaves for the Campanian increased from ~12 km in the Maastrichtian, to ~23 km in the Middle Eocene, after which it remained constant. The gaps in the Maastrichtian horizon remained constant at about 12 km.

![Diagram of the Cabo Frio Fault](image)

**Fig. 5.33 Sec-III:** The Cabo Frio Fault migrated to the deeper SE basin. The change in position was intense in the early stages, decreased significantly after the Maastrichtian (K/T), tending to cease towards the Middle Eocene (ME) and then to recommence after the Miocene (M).
Fig. 5.34 Sec-III: Horizontal component of displacement along the Cabo Frio Fault at each restoration stage. The heaves for the Albian, Turonian, Santonian and Campanian tended to increase until the Middle Eocene, and then remained stable. The heaves for the Maastrichtian horizon was constant.
5.6.2.4 Section IV

5.6.2.4.1 Albian

The major depocentre is in the proximal NW domain. The Albian sequence is deformed by faults F1, F2 and F3 (Fig.5.35).

![Diagram showing faults F1, F2, and F3](image)

**Fig.5.35** Sec-IV restored to the Albian. Depocentres for this sequence is mostly in the proximal NW domain. Faults F1, F2 and F3 are active at this stage.

5.6.2.4.2 Turonian

The major depocentres are located in the hangingwall of F1 and in the extreme NW, possibly controlled by some out-of-section structure (Fig.5.36). F3 controls a secondary depocentre. The CFF is still very incipient at this stage. Elongation for this section was 8.7% in relation to the Albian section (Fig.5.43).
Fig. 5.36 Sec-IV restored to the Turonian. Depocentres are in the proximal region and in the hangingwall of F1. Other active faults at this stage are F2, F3 and an incipient Cabo Frio Fault ('CFF').

5.6.2.4.3 Santonian

The major depocentres are controlled by F1, F2, F3 and CFF, with secondary depocentres controlled by the mini-basins. Salt diapirs and ridges S1, S2, S3 and S4 rise at the footwall of those faults. The CFF and the mini-basins appear to have started their activity during this stage, although an incipient CFF may have existed at the end of the Turonian (Fig. 5.37). Elongation at this stage is 35.6% in relation to the Albian section (cumulative) and 24.7% in relation to the Turonian section (Fig. 5.43).
Fig. 5.37 Sec-IV restored to the Santonian. Depocentres are controlled by seaward- and landward-dipping listric normal faults and by the salt withdrawal mini-basins.

5.6.2.4.4 Campanian

Campanian depocentres are highly controlled by the CFF and F3 (Fig. 5.38). Also important depocentres are controlled by the salt withdrawal basins. F1 and F5 control local depocentres. Crestal collapse grabens form above S1 and ridges of the distal domain. Elongation for this section is 60.4% in relation to the Albian section (cumulative) and 18.3% in relation to the Santonian section (Fig. 5.43).

Fig. 5.38 Sec-IV restored to the Campanian. The CFF controls the major depocentre. F1, F3, F4, F5 and the mini-basins control other important depocentres.
5.6.2.4.5 Maastrichtian

The Cabo Frio Fault strongly controls the major depocentre (Fig.5.39). Other important depocentres are controlled by F4 and F5. It appears that F4 has changed vergence since the Campanian. Crestal collapse grabens form above S1 and S6. Elongation for this section is 95% in relation to the Albian section (cumulative) and 21.5% in relation to the Campanian section (Fig.5.43).

Fig.5.39 Sec-IV restored to the Maastrichtian. The CFF controls the major depocentre. F4 and F5 are also active controlling smaller depocentres. Crestal collapse grabens occur above S1 and S6.

5.6.2.4.6 Middle Eocene

The major depocentre is in the hangingwall of the CFF (Fig.5.40). Important depocentres are controlled by F4, F5 and the salt withdrawal basins. Crestal collapse grabens occur mainly above S1 and S6. A salt weld appears under the hangingwall of the CFF, roughly aligned with a fault in the pre-salt sequence and a basement high. Elongation for this section was 105.5% in relation to the Albian section (cumulative) and 5.4% in relation to the Maastrichtian section (Fig.5.43).
Fig. 5.40 Sec-IV restored to the Middle Eocene. The major depocentre is controlled by CFF. Other important depocentres are controlled by F4 and F5. Braces indicate regions of intense crestal collapse grabens.

5.6.2.4.7 Miocene

The CFF is still active, as well as the salt withdrawal mini-basins and the crestal collapse grabens (Fig. 5.41). Although the salt withdrawal basins are still active controlling depocentres, especially between S5 and S6, the thickest deposits are those aggrading in the proximal domain, which do not evidence a clear structural control.

A new salt weld appears to the NW of the first one. It is not aligned with any fault in the pre-salt sequence. Whilst the first salt weld connected the pre-salt with the Maastrichtian sediments, this one connects the pre-salt with the Albian sediments. The first would favour the charge of Maastrichtian turbidites and the latter would favour the accumulation on Albian fractured carbonates. Crestal collapse is intense above S1, S5 and S6. Elongation for this section was 106% in relation to the Albian section (cumulative) and 0.2% in relation to the Middle Eocene section (Fig. 5.43).
Fig. 5.41 Sec-IV restored to the Miocene. Although with reduced activity, the CFF is still active, as well as the mini-basins and the crestal collapse faults above S1, S5 and S6. Two salt welds are present at this stage.

5.6.2.4.8 Recent

The major depocentres are still controlled by the CFF and by the salt withdrawal mini-basins, depicting the long-lived halokinetic activity (Fig. 5.42). Two salt welds are present in this section. The first one appeared in the Middle Eocene and connects the pre-salt with the Maastrichtian layer, highly tilted against the CFF. The last one appeared in the Miocene and connects the pre-salt with the Albian sequence. Both are still open and, if they matched the timing of hydrocarbon generation and expulsion, they may have acted as hydrocarbon migration fairways for the pre-salt source rocks to the post-salt Maastrichtian turbidites and Albian fractured carbonates. Elongation for this section was 106% in relation to the Albian section (cumulative) and only 0.1% in relation to the Miocene section (Fig. 5.43).
Fig. 5.42 Sec-IV: Recent. The CFF and the mini-basins are still active, demonstrating that halokinesis has not ceased. Two salt welds connect the pre-salt and the post-salt sequences, under the hangingwall of the CFF.
Fig. 5.43 Sec-IV: Graphs illustrate the elongation for the restored sections relative to the previous restoration stage or relative to the Albian section (cumulative).
The Cabo Frio Fault has migrated to the deeper water basin (Fig. 5.4.4) imposing gaps in the faulted horizons that are registered in figure 5.4.5. The overall activity of this fault reached a maximum from the Campanian to the Middle Eocene and then was dramatically reduced, but it never ceased.

The gaps in the Albian and Turonian sequences along the CFF were roughly the same and have increased from 6 km in the Santonian to about 33 km in the Middle Eocene, and then remained stable. The Santonian gaps reached ~27 km in the Middle Eocene. The heaves for the Campanian increased from ~14 km in the Maastrichtian, to ~21 km in the Middle Eocene, after which it remained constant. The gaps in the Maastrichtian horizon remained constant at about 7 km.

![Diagram of the Cabo Frio Fault](image)

**Fig. 5.4.4** Sec-IV: The Cabo Frio Fault migrated to the deeper basin. The change in position was intense in the early stages and decreased significantly from the Maastrichtian (K/T) to the Middle Eocene (ME), tending to cease towards the Miocene (M) and then to recommence.
Fig. 5.45 Sec-IV: Horizontal component of displacement along the Cabo Frio Fault at each restoration stage. The heaves for the Albian, Turonian, Santonian and Campanian tended to increase until the Middle Eocene, and then remained stable. Heaves for the Maastrichtian horizon remained constant.
5.6.2.5 Section V

5.6.2.5.1 Albian

The Albian sequence is deformed mostly by seaward-dipping listric normal faults (Fig. 5.46). The depocentre is in the NW domain and the main active faults are F1, F2, F3, F4 and F5.

![Fig. 5.46 Sec-V restored to the Albian. Depocentres are controlled mostly by seaward-dipping listric normal faults in the NW domain, especially F1, F2 and F4.]

5.6.2.5.2 Turonian

The major depocentre is located in the NW domain, controlled by F1 (Fig. 5.47). At this time the CFF does not exist. In the Turonian, at this place there is no evidence of a fault. Elongation for this section was 3.1% in relation to the Albian section (Fig. 5.54).
Fig. 5.47 Sec-V restored to the Turonian. Depocentres are concentrated in the NW, mostly controlled by seaward-dipping listric normal faults (mainly F1).

5.6.2.5.3 Santonian

The major depocentres for the Santonian are located in the proximal domain, either controlled by an out-of-section seaward-dipping fault or by a newly formed landward-dipping listric normal fault (F9). Important depocentres are controlled by the CFF and secondary depocentres are located down-dip, controlled by the seaward-dipping listric normal faults F6 and F7 (Fig. 5.48). Minor depocentres occur in the SE, indicating activity on the salt withdrawal basins. Elongation for this section was 28% in relation to the Albian section (cumulative) and 24.2% in relation to the Turonian section (Fig. 5.54).
Fig. 5.48 Sec-V restored to the Santonian. Major depocentres are concentrated in the NW, controlled by an out-of-section structure (an inferred seaward-dipping fault) and by the landward-dipping listric normal fault F9. The CFF control an also important but not as thick depocentre at this stage.

5.6.2.5.4 Campanian

The major depocentres are controlled by the CFF. Important depocentres are in the hangingwall of F9 and in the salt withdrawal basins. The activity of F7 has ceased and the depocentre it controlled during the Santonian stage has now shifted to the hangingwall of F8. Considering the extensive character of this domain, it is unlikely that a thrust fault would have formed at the place indicated by F6; possibly this fault, located at the top of S1 has changed vergence after the Campanian (Fig. 5.49). Elongation for this section was 48% in relation to the Albian section (cumulative) and 15.6% in relation to the Santonian section (Fig. 5.54).
Fig. 5.49 Sec-V restored to the Campanian. The major depocentres are concentrated on the NW, controlled by two landward-dipping listric normal faults CFF and F9. Secondary depocentres are located down-dip, controlled by seaward- and landward-dipping listric normal faults and by the salt withdrawal basins.

5.6.2.5.5 Maastrichtian

The Cabo Frio Fault controls the major depocentre at this stage (Fig. 5.50). Secondary depocentres are found in the salt withdrawal basins and in the hangingwall of F7. F6 and F9 are active and control minor depocentres. F6 and F7 may have changed vergence since the end of the Campanian. F8 is no longer active. Elongation for this section was 62.2% in relation to the Albian section (cumulative) and 9.6% in relation to the Campanian section (Fig. 5.54).
Fig.5.50 Sec-V restored to the Maastrichtian. The major depocentre is strongly controlled by CFF, in the centre of the section. Secondary depocentres are controlled by landward-dipping listric normal faults (especially F7) and salt withdrawal basins.

5.6.2.5.6 Middle Eocene

The major depocentre is controlled by the CFF (Fig.5.51). Also important depocentres are controlled by F6, F7 and F10. Salt structure S3 has risen and widened since the Maastrichtian, associated with the creation of F10. This wide salt ridge S3 here marks the transition from the extensional to the contractional domain. Elongation for this section was 78.5% in relation to the Albian section (cumulative) and 10% in relation to the Maastrichtian section (Fig.5.54).
Fig. 5.51 Sec-V restored to the Middle Eocene. The Cabo Frio Fault (CFF) remains in full activity, controlling the major depocentre of the section. Another important depocentre is controlled by a landward-dipping fault F10, flanked by a high salt wall (S3). Secondary depocentres are controlled by landward-dipping listric normal faults and salt withdrawal basins.

5.6.2.5.7 Miocene

The Cabo Frio Fault is still active, although with decreased activity, controlling the major depocentre of the section. Another important depocentre is controlled by F10, flanked by the salt ridge S3. Secondary depocentres are controlled by landward-dipping listric normal fault F6. The salt withdrawal basins seem to have reduced significantly their activity. Only the westernmost mini basin, adjacent to the salt high S, appears to have controlled a significant depocentre. Salt welds form in the proximal domain, connecting the pre-salt sequence with the Albian carbonates. The eastern salt weld is located immediately above a pre-salt fault (Fig. 5.52). Elongation for this section was 78.7% in relation to the Albian section (cumulative) and 0.1% in relation to the Middle Eocene section (Fig. 5.54).
Fig. 5.52 Sec-V restored to the Miocene. Depocentres at this stage are controlled by the CFF, F6 and F10 and by the salt ridges S1, S2, S3 and S4. Salt welds are indicated in the proximal domain, under the hangingwall of the CFF.

5.6.2.5.8 Recent

Depocentres are still controlled by the CFF, F6, F10 and the salt withdrawal basins, indicating that halokinesis is still active (Fig.5.53). The salt welds found in the present-day section may indicate promising scenarios for hydrocarbon upward migration from the pre-salt source rocks to the post-salt reservoirs. Attention should be paid to the timing of these welds, for the restoration shows that they have existed since the Middle Eocene and it is important to know whether this matches or not the timing of salt generation in this region. Elongation for this section was 80% in relation to the Albian section (cumulative) and 0.7% in relation to the Miocene section (Fig.5.54).
Fig. 5.53 Sec-V: Recent. Halokinesis is still active and depocentres are controlled by the CFF, F6, F10 and the salt withdrawal basins. Salt welds in the proximal NW domain may represent promising hydrocarbon fairways.
Fig. 5.54 Sec-V: Graphs illustrate the elongation for the restored sections relative to the previous restoration stage or relative to the Albian section (cumulative).
The Cabo Frio Fault has migrated to the deeper water basin (Fig.5.55) imposing gaps in the faulted horizons that are registered in Fig.5.56. The overall activity of this fault reached a maximum from the Campanian to the Middle Eocene and then was dramatically reduced, but it never ceased. The gaps in the Albian and Turonian sequences along the CFF were roughly the same and have increased from 3-4 km in the Santonian to ~30 km in the Middle Eocene, and then remained roughly stable. The Santonian gaps reached ~25 km in the Middle Eocene. The heaves for the Campanian increased from ~11 km in the Maastrichtian, to ~17 km in the Middle Eocene, after which it remained constant. The gaps in the Maastrichtian horizon remained constant at about 7 km.

Fig.5.55 Sec-V: The Cabo Frio Fault migrated to the deeper basin. The change in position was intense in the early stages and decreased significantly from the Maastrichtian (K/T) to the Middle Eocene (ME), tending to cease towards the Miocene (M) and then to recommence.
Fig.5.56 Sec-V: Horizontal component of displacement along the Cabo Frio Fault at each restoration stage. The heaves for the Albian, Turonian, Santonian and Campanian tended to increase until the Middle Eocene, and then remained roughly stable. Heaves for the Maastrichtian horizon remained constant.
5.6.2.6 Section VI

In this section, a seaward-dipping listric normal fault, rather than the landward-dipping fault, plays a major role in controlling depocentres.

5.6.2.6.1 Albian

Normal listric faults, mainly seaward-dipping, control the deformation of the Albian sequence (Fig.5.57). Depocentres are concentrated in the NW proximal domain.

![Diagram of listric normal faults](image)

Fig.5.57 Sec-VI restored to the Albian. Depocentres are located towards the continent (NW), locally controlled by seaward-dipping listric normal faults.

5.6.2.6.2 Turonian

As in the Albian, the major depocentres for the Turonian sediments are located towards the continent (NW), controlled by seaward-dipping listric normal faults (F1 and F2). Other listric faults (F3, F4, F5 and F6) control minor distal depocentres (Fig.5.58). Elongation for this section was 3.8% in relation to the Albian section (Fig.5.65).
5.6.2.6.3 Santonian

The major depocentre for the Santonian is controlled by F1 (Fig.5.59). Other important depocentres are controlled by (landward- and seaward-dipping) listric normal faults (F, F5, F6, F7 and F8). Secondary depocentres are located in the hangingwalls of F9, F10, F11 and F12, as well as in the mini-basins. Elongation for this section was 44.2% in relation to the Albian section (cumulative) and 39% in relation to the Turonian section (Fig.5.65).
5.6.2.6.4 Campanian

The major Campanian depocentre is controlled by F7, a seaward-dipping normal fault (Fig.5.60). F is a landward-dipping fault that would be correlated with the CFF but it apparently does not have the same role in controlling depocentres. F5, F6, F8, F9, F10, F11 and F12 control secondary depocentres. The activity in the salt withdrawal basins is intense. F1 seems to have ceased its activity at this stage. Elongation for this section is 67.4% in relation to the Albian section (cumulative) and 16% in relation to the Santonian section (Fig.5.65).

![Fig.5.60 Sec-VI restored to the Campanian. Major depocentre is controlled by the seaward-dipping F7. Other listric faults, seaward and landward-dipping, as well as the salt withdrawal basins, control secondary depocentres.](image)

5.6.2.6.5 Maastrichtian

The major Maastrichtian depocentres are controlled by F and F8, and also important depocentres are controlled by F6 and F7 (Fig.5.61). Although having controlled important depocentres, F7 and F8 cease their activity...
before the end of this stage. F9 and F13 control secondary depocentres. Minor faults are also active at this time, especially associated with crestal collapse grabens above salt walls, but they did not play an important role in depocentres control. The salt-withdrawal basins were active and controlled deposition in the distal domain. Elongation for this section is 73.4% in relation to the Albian section (cumulative) and 3.6% in relation to the Campanian section (Fig.5.65).

![Diagram of Sec-VI restored to the Maastrichtian](image)

**Fig.5.61 Sec-VI restored to the Maastrichtian.** Important depocentres are controlled by F, F6, F7 and F8. F7 and F8 seem to have ceased their activity during this stage. Secondary depocentres are controlled by F9, F11 and the salt withdrawal basins.

### 5.6.2.6.6 Middle Eocene

The Middle Eocene reflector is a highly eroded surface and even though erosion was taken into account during restoration, there may still be some rock material lacking. The major depocentres are controlled by F6 and F (Fig.5.62). Also important depocentres are controlled by F9 and F11. The salt withdrawal basins are intensely active, with fold synclines controlling depocentres in the distal domain. Salt welds form at this stage, in the
proximal domain, connecting the pre-salt sequence with Albian, Santonian and Campanian sediments, favouring hydrocarbon upward migration. Elongation for this section was 75.1% in relation to the Albian section (cumulative) and 1% in relation to the Maastrichtian section (Fig. 5.65).

![Diagram of cross-section restoration](image)

Fig. 5.62 Sec-VI restored to the Middle Eocene. Major depocentres are controlled by F6 and F. Secondary depocentres are controlled by F9, F11 and synclines associated with salt withdrawal basins. Salt welds form in the proximal domain, directly connecting the pre-salt with Albian, Santonian and Campanian sediments.

5.6.2.6.7 Miocene

The major depocentres are controlled by F, F9, F11 and the salt withdrawal basins. F6 has reduced its activity during this stage. Salt welds in the proximal domain are promising hydrocarbon fairways, connecting pre-salt rocks with Albian, Santonian and Campanian sediments in the post-salt sequence. Crestal collapse faults are frequent in the distal domain (Fig. 5.63). Elongation for this section was 74.6% in relation to the Albian section (cumulative) and -0.3% in relation to the Middle Eocene section (Fig. 5.65).
Fig. 5.63 Sec-VI restored to the Miocene. Depocentres are controlled by F, F9, F11 and salt withdrawal basins. Salt welds in the proximal domain connect the pre-salt sequence with Albian, Santonian and Campanian sediments.

5.6.2.6.8 Recent

This section has no significant landward-dipping fault that may correlate with the CFF. A candidate to be the lateral continuation of CFF would be fault F, but it is younger, does not display significant heaves and its evolution differs from that of the Cabo Frio Fault. The fault F position has not significantly changed since the Miocene. However, disturbed seabed close to this fault suggests that it remains active. Depocentres are controlled by listric faults and salt withdrawal basins, demonstrating the halokinetic control (Fig. 5.64). Elongation for this section was 75% in relation to the Albian section (cumulative) and only 0.2% in relation to the Miocene section (Fig. 5.65).
Fig. 5.64 Sec-VI at Recent. Interpreted depth-converted seismic section and geological section used for restoration. The fault F position has not significantly changed since the Miocene. However, disturbed seabed close to this fault suggests that it remains active. Depocentres are controlled by listric faults and salt withdrawal basins, demonstrating halokinetic activity.
Fig. 5.65 Sec-VI: Graphs illustrate the elongation for the restored sections relative to the previous restoration stage or relative to the Albian section (cumulative).
Fig. 5.66 Sec-VI: Fault F migrated to the deeper basin. Its position changed from the Santonian to the Maastrichtian (K/T) and remained roughly stable afterwards.
5.6.2.7 Section VII

5.6.2.7.1 Albian

The Albian sequence is deformed mainly by seaward-dipping listric faults (F1 to F6) and by the landward-dipping fault F (Fig.5.67).

Fig.5.67 Sec-VII restored to the Albian. Seaward-dipping and some landward-dipping listric normal faults are active at this stage.

5.6.2.7.2 Turonian

The major Turonian depocentres are controlled by F and F1 (Fig.5.68). Elongation for this section was 2.8% in relation to the Albian section (Fig.5.75).
The major depocentres are controlled by F, F1, F4 and F7 (Fig.5.69). Although having controlled important depocentres, F2 and F3 cease their activities during this stage. The salt withdrawal basins are active at this stage and control secondary depocentres. Elongation for this section was 21.5% in relation to the Albian section (cumulative) and 18.1% in relation to the Turonian section (Fig.5.75).

Fig.5.69 Sec-VII restored to the Santonian. Major depocentres are controlled by seaward- and landward-dipping listric normal faults. Secondary depocentres are controlled by salt withdrawal basins.
5.6.2.7.4 Campanian

The major Campanian depocentre is controlled by F and secondary depocentres are controlled by the salt withdrawal basins and by F1 (Fig.5.70). Elongation at this stage was 36.9% in relation to the Albian section (cumulative) and 16.7% in relation to the Santonian section (Fig.5.75).

Fig.5.70 Sec-VII restored to the Campanian. F controls the major Campanian depocentre. Secondary depocentres are controlled by F1 and the salt withdrawal basins.

5.6.2.7.5 Maastrichtian

The major depocentre for the Maastrichtian is controlled by F (Fig.5.71). Another important depocentre is located in a syncline (Sy) formed to the east of F, possibly associated with escape of salt towards the adjacent salt ridge. F1 is still active, although it does not control a significant depocentre. The salt withdrawal basins are very active controlling secondary depocentres. Crestal collapse grabens are common above salt ridges in the distal domain. Elongation for this section was 44.5% in relation to the Albian section (cumulative) and 5.6% in relation to the Campanian section (Fig.5.75).
5.6.2.7.6 Middle Eocene

The major depocentres are controlled by F and Fc (a set of listric faults formed on the crest of the salt ridge S1) (Fig.5.72). The salt withdrawal basins are intensely active and control important depocentres. In this domain, crestal collapse grabens are frequent. F1 is still active, though with reduced activity. Elongation for this section was 53.8% in relation to the Albian section (cumulative) and 6.4% in relation to the Maastrichtian section (Fig.5.75).
5.6.2.7.7 Miocene

Depocentres at this stage are controlled by a set of faults formed on the top of S1 (Fig.5.73). F has reduced its activity and only controls a minor depocentre. The mini-basins distal domain still controls important depocentres. A salt weld forms in the centre of the section, connecting the pre-salt sequence with the Albian sediments. Elongation for this section was 53.7% in relation to the Albian section (cumulative) and -0.1% in relation to the Middle Eocene section (Fig.5.75).

![Fig.5.73 Sec-VII restored to the Miocene. Depocentres are controlled by Fc, and the salt withdrawal basins. Braces indicate sets of faults associated with crestal collapse on top salt ridges and diapirs. Salt weld occurs in the centre of the section, connecting the pre-salt sequence with the Albian carbonates.](image)

5.6.2.7.8 Recent

The major depocentre is controlled by F8 and minor depocentres are controlled by the salt withdrawal basins (Fig.5.74). The fault F recent position has not changed significantly since the Campanian. This landward-dipping fault evolution was very different from the Cabo Frio Fault. Three more salt
welds appeared since the Miocene. They connect the pre-salt sequence with Albian and Santonian sediments and may represent promising hydrocarbon fairways. Elongation for this section is 54.4% in relation to the Albian section (cumulative) and 0.5% in relation to the Miocene section (Fig.5.75).

Fig.5.74 Sec-VII: Recent. Interpreted depth-converted seismic section (top); geological section used for restoration (bottom). The fault F recent position (55km) has not changed significantly since the Campanian. F8 controls the major depocentre. Important depocentres are controlled by the salt-withdrawal basins. Salt welds connect the pre-salt sequence with Albian and Santonian sediments.
Fig. 5.75 Sec-VII: Graphs illustrate the elongation for the restored sections relative to the previous restoration stage or relative to the Albian section (cumulative).
Fig. 5.76 Sec-VI: This fault has not changed position over the past 70 My, i.e., since the Campanian. It is the only important landward-dipping fault in this section and does not seem to correlate with the Cabo Frio Fault.

5.7 Sequential Evolution of the Margin

The restoration of seven regional cross-sections has provided important information on the structural and sedimentary evolution of the study area. They constitute viable models of the kinematic evolution of the structures and register different stages of salt structures development and depocentre migration. The structures that have been active at each stage controlling the depocentres could be clearly defined. The multi-section restoration has also given insights on spatial distribution of the structures and depocentres through time. The timing and position of the key structures and the timing of the opening of salt windows are crucial information for a confident evaluation of the petroleum system. In most of the study area (covered by Sections I to VI), salt welds started to form during the deposition of the Middle Eocene-Maastrichtian sequence, except in the northeastern area (covered by Section VII), in which salt welds started to form later, during the deposition of the Miocene- Middle Eocene sequence.
The restoration has highlighted the role of halokinesis in controlling the structural styles of the post-Aptian sequence and its interplay with sediment deposition.

The area occupied by salt in the restored sections was much higher at the Albian than it is at present. This implies that salt has moved basinwards throughout the drift stage of the basin and the salt reduced from the area of the basin covered by the sections may have contributed to form the salt ridges found in the distal domains. Much of the salt present in the ultra-deep water ridges constitutes salt that has been expelled from below the sediments in the near shore and central Santos Basin since its deposition, driven mainly by differential load caused by sedimentation.

The sections restored to the Albian suggest that many salt structures started to grow during the deposition of the Albian sequence. Conversely, in the restored sections some of the areas filled by salt may in fact be gaps in the Albian carbonate sequence. These gaps possibly resulted from out-of-section movement of the Albian rafts, in the early stages of faulting, and also due to accumulated deformation through time. Continuous deformation since Albian times may have produced small blocks of the Albian sequence that are not easily identified in seismic sections and thus may lack on the restored sequence.

Extension prevailed over contraction (figures 5.77 and 5.78) and this is much more evident in the central area covered by regional cross sections II to V, which show elongation higher than 100% (meaning that the length of the section has more that doubled from Albian to the present day). In this central domain, the large landward-dipping Cabo Frio Fault has played a major role in controlling depocentres, especially from the Santonian to the Middle Eocene.
Extension has prevailed over contraction, except for the Miocene stage, when a small amount of contraction (up to 1%) took place (observed in sections II, VI and VII).

Along-strike variations between the seven restored cross-sections can be summarised in the diagrams below. The highest elongation was observed in the central area (sections III, IV and V).

The landward-dipping listric faults interpreted in the lateral sections (Section I, VI and VII) had a different style of evolution and do not seem to correlate with the Cabo Frio Fault, which dominates in the central domain, occupying most of the study area, from WSW to ENE.

![Diagram showing total elongation (since the Albian) obtained for the seven restored cross-sections. The diagram illustrate the along-strike variability in extension rates.]

Fig.5.77 Total elongation (since the Albian) obtained for the seven restored cross-sections. The diagram illustrate the along-strike variability in extension rates.
Fig.5.78 Elongation through time along the seven restored cross-sections. The sections show maximum sequential elongation during the Santonian, except for Section II, whose highest partial elongation was reached in the Campanian.

The restored sections clearly demonstrate that the Cabo Frio Fault and many structures of the post-salt sequence were originally formed in positions entirely different from the ones they have at present and they have migrated to the deeper SE domain (Fig.5.79). This has significant impact on the study of petroleum systems, for some migration pathways that exist today possibly did not exist in the past, matching the time of hydrocarbon generation and expulsion. Also, the structural traps and seals, as well as the location of reservoirs may have changed position through time. The precise definition of the temporal and spatial evolution of these key elements can be achieved by applying restoration techniques that will provide the necessary information to improve the petroleum systems assessment.
Chapter 5 - Cross-section restoration
Evolved section restoration validates interpretation and reveals structural and stratigraphic evolution, highlighting the active structures that control depocentres at each stage. The salt layer and the overburden have deformed primarily by gravitational gliding of a brittle sequence over a spreading ductile layer, creating structures that may act as hydrocarbon traps, controlling the depocentres and defining hydrocarbon migration fairways. The restoration demonstrates that the Cabo Frio Fault has migrated offshore throughout its evolution and therefore its present-day position sometimes aligned with a pre-salt fault does not correspond to its original position. Care must be taken when modelling the petroleum system. Post-salt structural traps, reservoirs, migration fairways and seals position relative to the pre-salt structures and source rocks have been changing and a promising present-day correspondence does not guarantee a favourable scenario at the time of hydrocarbon expulsion.

*Fig. 5.79* The spatial and temporal evolution of the Cabo Frio Fault is illustrated from the end of Turonian, when it was very incipient (if present), to the Recent. The dashed line indicates that the lateral correlation is inferred and may not exist. Note the change in position through time of the post-salt structures in relation to the pre-salt structures, especially from the Santonian to the Middle Eocene (the arrow points to a reference position in the pre-salt sequence).
Halokinesis is a key factor in controlling the structural evolution of the post-salt sequence in the Santos basin. The analysis of seismic interpretation and restored cross-sections led to the proposal of a halokinetic model for the post-salt evolution of the study area.

The main driving mechanisms for the salt movement in the study area are the basin tilt due to thermal subsidence and the differential loading caused by the thick Upper Cretaceous and lower Cenozoic sedimentary wedge deposited above the Aptian salt. Salt escaped from the loaded near shore domain and moved towards the deeper basin, where it accumulated and created a contractional domain, characterised by salt-cored folds.

The up-dip extensional domain is marked by seaward- and mainly landward-dipping listric normal growth faults, such as the Cabo Frio Fault, which resulted from the flow of the ductile salt layer from beneath a thick prograding wedge towards less-loaded regions. It was originally formed up-dip and has migrated basinwards controlling the major depocentre in the study area (Fig.6.1). Down-dip contraction is accommodated by a combination of detachment folds, partly squeezed diapirs, and shortened and inflated salt massif with a thin, folded overburden comprising the post-salt stratigraphic section.

A later source of sediment supply was established in the Cenozoic, with continental sediments entering the basin from the NNE. The sediment input
created a new loaded area in the northeastern region, which forced the salt layer to move to the SSW, again creating up-dip extension and down-dip contraction.

Fig. 6.1 Simplified diagram illustrating the gravity-driven evolution of the Cabo Frio Fault (CFF). The salt layer acted as a detachment zone for the extended blocks of the brittle post-salt sediments. A Late Cretaceous to early Cenozoic thick prograding wedge caused by massive continental sediment supply promoted salt withdrawal and a landward-dipping listric growth fault formed at the toe-of-slope region. The CFF resulted from the basinward flow of the ductile salt layer that escaped from beneath the prograding wedge towards less-loaded regions. It remained active controlling the major depocentre of the study area and the sediment captured in its hangingwall added to the differential load and maintained its activity.
The linked system of extension and contraction, comprising the Cabo Frio Fault, appears to constitute a gravitational cell hereinafter named Ilha Grande Gravitational Cell (IGGC), as a reference to the large and beautiful island of Ilha Grande that lies just north of this area. This proposed feature is interpreted in the salt structural map and in the salt isochore map (figures 6.2 and 6.3) and appears to have migrated to the SE deeper basin.

The eastern limit of the IGGC, probably a transfer zone, is oriented N-S and its projection towards the continent coincides with the entrance of the Guanabara Bay (Fig.6.2). Although a preliminary analysis did not show any significant structure onshore aligned with this trend, it is recommended to give continuity to this work by thoroughly investigating the lineaments on the continent.

![Proposed outline of the Ilha Grande Gravitational Cell](image)

Fig.6.2 Proposed outline of the Ilha Grande Gravitational Cell (dashed grey line), with up-dip extension and down-dip contraction, on the Top Salt structural map. The eastern and western borders of the IGGC are indicative of lateral gradients of slippage. Sense of sediment supply shown by grey arrows.
The area limited by the dashed line (Fig.6.3) seems to have moved towards the SE as a gravitational cell. A prominent structural high on the SW may have partly buttressed the salt movement. The IGGC possibly deviates from this basement high and moves towards the SE region, imposing further compression on the salt withdrawal basins previously formed by interference of confluent (concentric) directions of continental sediment supply (source areas at NW and N). The curved shapes of the salt walls, on the west, are suggestive of further deformation of the salt structures in response to the advance of the IGGC.

Fig.6.3 Salt isochore map with the approximate outline (dashed) of the proposed IGGC, which comprises linked up-dip extension and down-dip contraction. The Cabo Frio Fault lies in between (black line). Grey arrows indicate the overall sense of salt flow in the IGGC and adjacent domains. Black arrows indicate relative movement along transfer zones that limit the IGGC. Curved-shaped salt ridges in the SE result from reworked salt walls.
The IGGC eastern limit is N-S oriented and probably consists of a thin-skinned transfer zone. This direction does not match the NW-SE oriented transfer zones recognised by Demercian (1996) and Demercian & Szatmari (1999) and interpreted as thin-skinned halokinetic structures, neither matches the NW-SE transfer zones that segment the rift (figures 6.4 and 6.5), described by Meisling et al. (2001). The Bouguer gravity map of residual anomalies (Fig.6.4) shows no NS-trending structure that could correlate with the eastern limit of the Ilha Grande Gravitational Cell. This limit coincides with a shift in the interpreted near shore Moho uplift, interpreted by Meisling et al. (2001) as the effect of a NW-SE transfer zone (TZ).

Moreover, the N-S orientation of the transfer zone that borders the IGGC does not coincide with the basement faults. Due to the poor seismic definition for the basement in the 2-D seismic sections, in this work the basement structures have been interpreted only in the area of the 3-D seismic survey. They are mainly oriented in a NE direction, shifting to a NW direction in the south of the area (see Fig.4.43). Likewise, the structural map for the Base Salt (see Fig.4.42) does not show any NS-trending structure that would correlate with the eastern border of the IGGC. This suggests that the eastern limit of the IGGC is a thin-skinned structure. The whole IGGC appears to be a thin-skinned gravitational cell.

A schematic diagram illustrating the proposed model for the halokinetic evolution of the study area is displayed in Fig.6.6.
Fig. 6.4 Map of gravity anomalies, Campos and Santos basins (Lamont-Doherty Geological Observatory). Interpreted anomalies: MU (near shore Moho uplift) and FS (failed spreading ridge), by Meisling et al. (2001). No NS-trending structure correlates with the eastern limit of the Ilha Grande Gravitational Cell (dashed purple line).

Fig. 6.5 Map of main rift-related structural provinces, Campos and Santos basins (Meisling et al. 2001).
Stage I:

Sediment supply from NW

Sediment coming from NW creates a thick prograding wedge that pushes salt to the deeper waters, resulting in gravitational linked system of updip extension and downdip contraction.

Stage II:

Interference pattern associated with superposed salt ridges due to convergent directions of prograding sediments.

Sediment supply from NW

Sediment supply from NNE

Salt withdrawal basin surrounded by salt ridges

Stage III:

Reworking of the salt ridges due to continued advance of the IGGC to the SE.

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Fig. 6.6 Schematic diagram illustrating the proposed model for the evolution of the halokinetic-related structures and creation of depocentres in the study area. In response to differential loading, salt moved basinwards and laterally, creating up-dip listric normal faults, dominantly landward-dipping such as the Cabo Frio fault that control the major depocentres of the area, and down-dip shortening that created salt-cored ridges. The IGGC moved towards the SE, driven by the thick prograding sedimentary wedge formed by a large amount of sediment dumped into the basin during the Late Cretaceous. The mini-basins domain resulted from the interference of convergent flow of salt escaping from beneath sediments that were supplied from NW and, more recently, from NNE. The confluent directions of the sediment income have promoted a convergent flow towards the deeper basin, which resulted in a complex pattern of superposed salt ridges bounding depressed areas in which sediments were caught. The continued migration of the IGGC over the mini-basin domain may have imposed further compression on the already deformed salt ridges in the distal SE area. The curved shapes of these salt ridges indicate reworking.
Two phases of folding are identified in the central area. A first generation created NE-trending folds as a result of horizontal shortening accommodated by buckling. A second generation of folds formed with ESE direction and was superposed with the previously formed folds, creating a complex interference pattern that accounts for the geometry of the mini-basins domain. Both folding phases are related to gravity-driven thin-skinned tectonics, strongly controlled by halokinesis (figures 6.6 and 6.7).

Fig.6.7 Schematic diagrams illustrating the formation of mini-basins as a result of two different dominant directions of sediment supply, first from the NW (grey arrows) and then from the NNE (blue arrows). The black arrow indicates the Cabo Frio Fault (CFF).
This conclusion is supported by analogue models, which demonstrate the role of the interference pattern of superposed folds in the configuration of the salt withdrawal mini-basins (Fig.6.8).

![Underwater analogue modelling](image)

**Fig.6.8** Underwater analogue modelling. A) experimental rig; B) surface view. Thin-skinned extension in the shelf and slope; contraction in the deep-water domain. The confluent directions of the sedimentary wedges promote convergent salt flow to the centre of the basin. Salt-cored ridges formed in the contractional domain to accommodate shelf and slope extension due to the first (NW-SE) sedimentary input interact with the newly formed ridges associated with the NNE supply and form an intricate pattern of superposed folds that closely resembles the Top Salt and the Bathymetric maps in the Santos Basin and may explain the polygonal shape of the mini-basins (modified from Guerra et al. 2005a, b).

The continued migration of the IGGC over the mini-basin domain may have imposed further compression on the already deformed salt ridges in the distal SE area. The curved shapes of these salt ridges indicate that they have been reworked.
A significant proportion of the world's hydrocarbon reserves is found in halokinetic-related structures, including many in the Middle East, the South Atlantic passive margins (Brazil, Gabon and Angola) and the Gulf of Mexico. Halokinesis plays a major role in the development of petroleum systems, for salt not only acts as a highly efficient seal, but also its movement can create structural traps and migration fairways from the pre-salt source rocks to the post-salt reservoirs. Halokinetic-induced bathymetry determines sites for turbidite deposition. Moreover, salt influences thermal gradient and the presence of thick salt layers in the distal basin may alter the timing of hydrocarbon generation and expulsion in the pre-salt source rocks. Conversely, the post-salt source-rocks would receive an extra heat.

Evolved restoration permitted to determine the timing of creation of a series of elements that influence the petroleum system. The multi-section restoration has given insights on spatial distribution of the faults, folds and depocentres through time. The timing and position of the key structures and the timing of the opening of salt welds are crucial information for a confident evaluation of the petroleum system. The restoration of seven regional cross-sections in the Santos Basin has shown that a linked system of up-dip extension and down-dip contraction has migrated basinwards since the end of the Turonian, therefore indicating that the position of the key elements for hydrocarbon migration, such as salt welds and faults has changed through
time. A present-day connection between pre-salt and post-salt faults may not guarantee a correspondence in the past, matching the timing of hydrocarbon generation and expulsion.

**Maturation:** Due to their high thermal conductivity, the presence of salt rock influences geothermal gradient by reducing the temperature below and increasing the temperature above (Mello et al. 1995). Thick salt layers or salt ridges cool down the underlying source rock, thus delaying hydrocarbon generation. The restored cross-sections revealed the timing of creation of halokinetic positive structures and thick salt layers in the study area, providing information to satisfactorily assess the petroleum generation and expulsion in the pre-salt source rocks.

**Migration:** Halokinesis has controlled some key factors that influence hydrocarbon migration. Hydrocarbon hosted in post-salt reservoirs has risen through salt welds, faults and permeable pathways. Salt has thinned to a minimum under the hangingwall of landward-dipping listric normal faults such as the Cabo Frio Fault, intimately associated with halokinesis. These faults may be aligned with pre-salt structures, favouring hydrocarbon migration to the post-salt reservoirs. Nevertheless, these halokinetic-driven faults have changed position through time. A new position may no longer represent a favourable scenario for hydrocarbon migration. Restoration techniques help determine the timing of salt welding and the position of the post-salt structures relative to the pre-salt structures at each stage.

**Reservoirs:** Halokinesis has strongly influenced bathymetry and controlled sediment dispersal in the basin. Sand-rich reservoirs are expected to occur in the salt-induced bathymetric lows adjacent to positive salt structures. The sediments that entered the basin were trapped in the hangingwall to the evolving landward-dipping faults and the sediments that managed to

*Chapter 7 - Exploratory Implications*
surpass this extensional domain or that have travelled laterally were deposited into the bathymetric lows adjacent to salt ridges, in the far shore mini-basin domain. Faults and fractures created by halokinesis in the Albian carbonate rocks may have also resulted in excellent reservoirs.

**Trap:** Tilted blocks against salt-related listric faults and inverted depocentres are preferred sites for hydrocarbon accumulation. The rollovers and turtle structures associated with listric faults detached on the salt layer represent important hydrocarbon plays in the Santos Basin. Restoration helps determine the timing of formation of such features and how they evolved through time.

**Seal:** Rock salt is impermeable and therefore constitutes an excellent seal. The presence of thick massive salt layer in the distal part of the Santos Basin has acted as an efficient seal that favoured large hydrocarbon accumulation recently discovered in pre-salt plays.
Chapter 8

Conclusions

The Santos Basin resulted from Mesozoic tectonic events associated with the Gondwana break-up that led to the opening of the South Atlantic and consequent separation of African and South American plates. A large and thick Aptian evaporite sequence deposited in the basin turned out to be essential to petroleum accumulation. The understanding of the salt layer kinematics is a key factor for hydrocarbon exploration, because salt mobility is responsible for the major structural traps, reservoir distribution and migration fairways for the hydrocarbon sourced in the rift sequence. Based on seismic interpretation and cross-section restoration, this project investigated the halokinetic-related structural evolution and sedimentation of the drift sequence in the central and northern Santos Basin.

A new seismic-based structural and stratigraphic interpretation undertaken on well-calibrated, strategic 3-D seismic surveys and on regional 2-D seismic lines permitted to study the role of salt mobility in controlling the sediment dispersal and structural evolution of the Santos Basin, a component part of the Brazilian passive continental margin.

Structural and isochore maps obtained from seismic interpretation illustrate the faults, salt ridges and depocentre distribution along the study area for all the interpreted sequences and indicate the structural styles, the dominant sediment fairways, the active faults at the time of deposition, and the regions of salt highs and lows, demonstrating the control exerted by the halokinetic-
related structures on sediment dispersal through time. However, this is relative to the present-day salt and sediment geometries, which may not represent the past stages of halokinetic evolution. A complete analysis of the post-salt history of the basin requires the use of restoration techniques in order to unravel the past configuration of the post-salt sequence and more accurately determine the geometry and position of the halokinetic-related structures and depocentres at each stage. Evolved restoration of seven regional geological cross-sections that cover the central and northern Santos Basin was performed using a proprietary restoration software (ReconMS) that was tested and improved with this project. The kinematic restoration validated the interpretation and revealed the post-Aptian evolution of the study area, depicting the structures that have been active at each stage, controlling depocentres and the timing of formation of salt welds. The simultaneous observation of several regional restored cross-sections through time permitted a good spatial investigation of the post-salt tectonic evolution of the study area. The multi-section restoration produced scenarios that give the true isopachs at each stage, since the sections were decompacted and backstripped as the deformation was removed towards the past.

In the study area, the pre-salt sequence is characterised by extension along landward- and seaward-dipping, essentially planar normal faults, defining a pattern of grabens and half-grabens. The depocentres of the rift lacustrine source rocks are limited by structural highs. Deformation within the post-salt sequence results from gravitational gliding and spreading, controlled primarily by the presence, at depth, of a ductile salt layer, whose movement has affected the entire sequence.

Two major halokinetic domains are observed in the post-salt sequence: a near shore extensional domain characterised by salt welds, small-scale salt pillows
and asymmetric salt-rollers in the footwall of listric normal growth faults, and a far shore contractional domain dominated by mini-basins. The landward- and seaward-dipping listric normal faults are detached on the salt layer and deform the sedimentary cover in a pattern of rollovers and turtle anticlines. Salt-withdrawal basins and long salt ridges are widespread in the deep-water domain, where the sedimentary cover is deformed by salt-cored detachment folds. Normal crestal collapse grabens are frequent above the top of the salt diapirs and ridges.

From the integrated analysis of the structural and isochore maps obtained by interpretation of well-calibrated 2-D and 3-D seismic reflection data with sequential restoration of seven regional cross-sections, and insights from analogue modelling, an evolutive model could be delineated for the drift phase of the study area:

1. A thick Aptian evaporite layer, possibly comprising multiple evaporite cycles, was deposited during the transition phase.

2. Albian (to Cenomanian) carbonates deposited over the salt layer promoted salt movement. Thermal subsidence promoted the seaward tilt of the margin, producing small tilted rafts. Gravity gliding of a brittle layer over a spreading ductile layer created listric normal growth faults associated with small-scale rollovers and salt-rollers. This raft tectonic style is still recorded in the northeastern part of the study area, in the lower Albian sequence. These structures do not appear in the centre of the study area; they were possibly obscured by subsequent massive sedimentation and recurrent faulting that affected this domain.

3. The carbonate platform was progressively drowned and transgressive Turonian shales deposited as open marine conditions were established, with the opening of the South Atlantic Ocean.
4. During the Late Cretaceous, the coastal uplift favoured erosion onshore and accumulation of continental sediments into the basin. A thick prograding wedge was formed by clastic sediments that fell abruptly into the basin. Whilst the neighbouring basins were under transgression, in the Santos Basin the transgressive stage ended during the Turonian. Massive sediments supplied from northwest compensated the regional sea level rise and loaded the underlying salt differentially, accelerating its movement towards the deeper basin. Basin tilt and differential loading over the ductile salt layer enhanced gravity gliding and spreading that gave rise to up-dip extension and down-dip contraction, creating accommodation space for sand-rich sediments that entered the basin. Pillowing and withdrawal occurred in response to salt mobility.

5. In the extensional province, the post-salt sequence is deformed by syn-depositional listric faults, mainly landward-dipping, detached on the salt layer. The most outstanding of these faults is the Cabo Frio Fault, a large listric landward-dipping normal growth fault that extends for 185 km along strike and displaces the Albian sequence with heaves that reach 60 km. It was originally formed up-dip and has migrated basinwards acting as a growth fault and controlling the major depocentre in this part of the basin, especially from the Santonian to the Middle Eocene. Its rate of migration has decreased since the end of the Cretaceous. A set of large landward-dipping faults is observed in the vicinity of the Cabo Frio Fault, generally flanked by long and high-amplitude salt walls that have risen in their footwall. The contractional domain starts just after this set of large listric landward-dipping growth faults. Mini-basins flanked by salt walls formed down-dip, as salt escaped from beneath the massive prograding wedge and accumulated in the deeper basin. Major depocentres are controlled by the Cabo Frio Fault and
secondary depocentres are controlled by the salt withdrawal mini-basins. In general, the salt layer has thinned in the extensive domain and thickened in the compressive domain.

6. With the individualisation of the NE-SW-trending Serra do Mar and Serra da Mantiqueira mountains, in the Early Palaeocene, the ancestral drainage was captured and deviated to the northeast, forming the Paraíba do Sul River that started to discharge further northeast, in the Campos Basin. The Santos Basin, whose major continental sediment supply during the Late Cretaceous came from the northwest, started to receive large volumes of sediment from the north-northeast in the Cenozoic.

The massive sedimentary supply from the NW caused by the onshore uplift has created a peculiar halokinetic style in the centre of the study area. The resulting large landward-dipping faults, such as the Cabo Frio Fault, associated with significant salt withdrawal possibly obscured previous deformation. The northeastern region, in its turn, not directly affected by that thick wedge, could maintain the initial halokinetic style, hence preserving early Albian rafts. Later, in the Cenozoic, when the strong sediment supply shifted to NNE, creating a new prograding wedge, the salt layer there had already considerably evacuated from the near shore domain and maybe it was not thick enough to give rise to a landward-dipping fault as impressive as the Cabo Frio Fault, with such large displacements. The differential loading over a much thinner ductile layer resulted in a different halokinetic style, not as expressive as the one that dominated during the Late Cretaceous in the centre of the study area.

7. Igneous activity during the Late Cretaceous and early Cenozoic may have affected salt movement by locally accelerating halokinetic processes, due to additional load and temporary heating. Dykes, volcanic cones and
lava flows represent an additional weight that may have contributed to salt withdrawal.

8. Based on the results of seismic interpretation and cross-section restoration, a model was elaborated to explain the halokinetic evolution of the post-salt sequence of the study area, as well as the genesis of the salt withdrawal basins and the superposed folds. This model proposed the existence of a gravitational cell, designated “Ilha Grande Gravitational Cell”, a NW-SE linked system of up-dip extension and down-dip contraction comprising the Cabo Frio Fault, in the centre of the study area.

9. The Ilha Grande Gravitational Cell (IGGC) was proposed based on the conjugate analysis of seismic interpretation, especially salt isochore and Top Salt structural maps, and cross-section restoration. Driven by the thick prograding sedimentary wedge formed by massive sediment supply during the Late Cretaceous, the IGGC has moved towards the southeast, with flow rates different from the lateral domains. Although not easily identifiable on seismic data, transfer faults are likely to occur at the lateral borders of the IGGC and, if present, they may favour hydrocarbon migration from source to reservoir.

10. Massive clastic sediments were initially supplied from NW and, more recently, also from NNE. The confluent directions of the sediment income have promoted the movement of the salt layer towards the deeper basin in a convergent flow, resulting in a complex pattern of superposed salt ridges that bound depressed areas in which sediments were caught. The interference pattern of superposed folding in the distal contractional domain resulted in polygonal mini-basins.
11. The continuing advance of the IGCC towards the distal southeast area may have imposed further compression on the already deformed distal salt walls, whose curved shapes are suggestive of refolding.

12. The salt layer has thinned to a minimum in most of the up-slope domain, eventually cutting the salt supply, as salt welds form, establishing a fairway for hydrocarbon migration from the pre-salt source rocks to the post-salt reservoirs. Beneath the Cabo Frio Fault hangingwall, salt welds connect the pre-salt rocks with Albian, Turonian, Santonian and younger sequences. The closer we get to the fault, the younger the sequence that touches down onto the pre-salt rocks. Restoration showed that in most of the study area (Sections I to VI), salt welds started to form during the deposition of the Maastrichtian-Middle Eocene sequence, except in the northeastern area (Section VII), in which salt welds started to form later, during the deposition of the Middle Eocene-Miocene sequence.

13. The restored sections showed that in the extensional domain the salt layer was much thicker in the past. This implies that salt has moved down-dip and laterally and contributed to form the salt ridges of the contractional domain. Much of the salt present in the ultra deep-water ridges consists of salt that has been expelled from beneath the depocentres towards relatively unloaded regions. Deformation inside the massive salt ridges includes thrusts and recumbent folds.

14. Intense regional erosion affected the basin during the Middle Eocene. After the Middle Eocene, the relatively quiet sedimentation suggests that halokinesis intensity was significantly reduced. However, crestal collapse grabens and shallow normal faults that deform up to the seabed attest that halokinesis is still active in the basin.

*Chapter 8 - Conclusions*
Both the seismic interpretation and cross-section restoration performed in this project have provided important information on the evolution of the study area, highlighting the role of halokinesis in controlling sedimentation and deformation of the overlying sequence. Throughout the post-Aptian evolution of the basin, salt has been flowing and imposing stresses in the overlying sediments, giving rise to transient structural geometries in the Santos Basin and creating accommodation space for sand-rich sediments deposition that, in its turn, has modified salt flow patterns, in a feedback process.

Halokinesis has been the major controlling factor for the structural evolution during the drift phase of the Santos Basin. It has been crucial to the creation of listric normal growth faults and salt withdrawal basins that have segmented the basin and created important fairways for the sand-rich sediments, whose deposition was controlled by salt-induced bathymetry. Differential loading and basin tilt were the main driving mechanisms to trigger halokinesis. This accounted for the structural styles and sedimentation patterns during most of the drift phase.

The multi-section visualisation provided by ReconMS gave a "2.5-D" insight on the structural and sedimentary evolution of the basin, permitting to identify different structural domains, depocentre migration, timing of creation of salt welds and the structures that have been active controlling depocentres at each stage.

The genesis of the structures interpreted in the post-salt sequence can be explained by gravitational gliding and spreading of brittle rocks over a detachment layer that thinned up-dip and thickened down-dip. The Cabo Frio Fault was formed and evolved as a listric fault detached on the salt layer. It was nucleated further up-dip and has migrated towards the deeper
basin. It does not show any evidence of being triggered nor controlled by presence or reactivation of pre-salt structures. Likewise, all the post-salt deformation in the study area appears to be entirely thin-skinned, as suggested by the basinward migration of the post-salt sequence. The ductile salt layer tends to decouple the structures of the overlying sequence from any deformation that might have affected the pre-salt sequence. It must be taken into account that the seismic section just shows the present-day configuration. Vertical alignment of pre-salt and post-salt structures identified in seismic sections may sometimes erroneously suggest that the post-salt deformation is controlled by the pre-salt structure, either reactivated or just forming pre-existing topography at the base salt.

The post-salt structures do not occupy any longer the position in which they nucleated. The basinward migration of the post-salt sequence suggests that the post-salt structures are independent of pre-salt structures vertically aligned with them in a seismic section. The gravity-driven movement of this sequence over a mobile salt has created faults, salt welds, structural traps and depocentres.

This has a considerable impact on the petroleum system assessment, since the positions of post-salt faults, salt welds and salt ridges have changed throughout the drift phase and thus their present-day configuration on a geological section does not represent the past stages of the basin. Also, there were times when the connection between pre- and post-salt sequences were favoured either by aligned faults or by salt welds, establishing promising migration fairways for hydrocarbons sourced in the pre-salt rocks to rise towards the post-salt reservoirs. However, the restoration evidenced that these scenarios have changed with the continued thin-skinned basinward migration of the salt and post-salt sequence. Therefore, it is strongly
recommended to restore all the geological sections used as input for the petroleum systems modelling, in order to obtain confident scenarios of the structural evolution of the basin and make sure that the key elements are in their true position at each stage.

The analysis of the structural and isochore maps produced after the interpretation of well-calibrated 2-D and 3-D seismic reflection data, conjugated with sequential restoration of seven regional cross-sections and supported by analogue models provided holistic understanding on the role of halokinesis in controlling structural styles, bathymetry evolution and sediment dispersal throughout the post-salt history of the northern and central Santos Basin. The results of this research may contribute to enhance the understanding of the interactions between halokinetic processes and sediment dispersal likely to occur not only in the Santos Basin but also in other passive margin sedimentary basins involving salt tectonics.
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