Thesis presented to the University of Oxford for the degree of Doctor of Philosophy.

THE SEDIMENTATION, TECTONICS AND STRATIGRAPHY OF THE CRETACEOUS/TERTIARY SEQUENCE OF NORTHWEST SANTANDER, NORTHERN SPAIN.

BY FREDERICK MICHAEL HINES (Worcester College)

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TRINITY TERM 1986.

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VOLUME 1 : PREFACE AND TEXT.

A general view of northwest Santander, northern Spain looking southwestward from the Rio Pas over the green foothills of Mesozoic strata towards the Picos de Europa Mountains.

DEDICATION.

This thesis is dedicated to Louis Mengaud, the father of modern Cantabrian geology.

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SHORT ABSTRACT.

The facies evolution of the Cretaceous/Tertiary sequence of NW Santander is considered in relation to the Cretaceous rifting and drifting, and Tertiary partial closure of the Bay of Biscay. Overlying the Palaeozoic basement are the fluvial Lower Triassic Buntersandstone and Upper Triassic Keuper evaporitic mudstone, deposited in a failed rift, extensional basin. Overlying Lower Jurassic carbonates are the syn-rift, continental clastics of the Wealden deposited in half-grabens cut by transfer faults. The Wealden consists of two formations:- the lower, arenaceous-rich Barcena Mayor Fm. (braided stream environment) and the upper, argillaceous-rich Vega de Pas Fm. (meandering river). Overlying it is the Aptian Umbrera Fm. (calcarenite sheet), the Patrocinio Fm. (shoaling-upward sandstone/marl alternation), the San Esteban Fm. (requienid/foraminiferal biomicrite of the internal platform) and the marls of the Rodezas Fm.. The Upper Aptian Reocin Fm. is a requienid/foraminiferal biomicrite with thinned calcarenites deposited over active, diapiric palaeohighs. After initial marine and then equant calcite (meteoric phreatic) cementation, invasion of meteoric-derived groundwater over palaeohighs generated lenses of sucrosic dolomite in the Reocin Fm.. Local mixing of further groundwater and Keuper-derived, sulphate-rich waters in karstic caverns precipitated sparry, baroque dolomite and Pb/Zn sulphides (by bacterial sulphate reduction). The clastic Lower Albian is a transgressive fluvial/estuarine/inner shelf sequence with tidal estuarine channels and sandwaves. The Middle/Upper Albian (syn-drift) has basal calcarenitic tidal sandwaves and is followed by storm/wave-reworked carbonates deposited on a homoclinal ramp. The clastic Lower Cenomanian is an estuarine/inner shelf deposit with tidal sandwaves and sandbars. The Middle/Upper Cenomanian is a storm/tide-dominated calcarenite. Outer shelf marls occur in the Turonian to Middle Campanian and the Upper Campanian to Middle Eocene is a sandy, foraminiferal inner shelf limestone. The Upper Eocene/Oligocene (syn-compression) is a carbonate slope-apron-reefal flysch deposit. It includes hemipelagic marl, neritic-derived calcarenitic turbidites and rudaceous mass flow deposits with highly polymict conglomerates. These were deposited coevally with Keuper piercement and thrust reactivation and date the Pyrenean compressional deformation here.

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LONG ABSTRACT.

The facies evolution of the Cretaceous/Tertiary sequence in NW Santander is considered in this thesis in relation to the tectonic evolution of northern Spain. The northern Iberian continental margin continues onshore in the Basco-Cantabrian basin (Chapter 1), and the rifting, drifting and partial closure of the Bay of Biscay are directly linked with the facies evolution of this, the northwestern part of the Basco-Cantabrian basin.

Struturally, NW Santander can be subdivided into three zones (Chapter 2). To the west of San Vicente de la Barquera (Figs. 2-3 & 2-4) there is a zone of E-W trending Hercynian thrusts that were reactivated in the Tertiary, Pyrenean orogeny. There is a complex zone around San Vicente itself. The main part of NW Santander consists of ENE-WSW trending folds with Keuper piercements. The existing research on the area is mainly (bio)stratigraphic, and much of the sedimentology presented here is completely new. The approach utilises vertical sequence facies analysis of logged sections from the excellent coastal exposures in the area. General comparisons are made throughout the thesis with the tectonic, stratigraphic and sedimentological evolution of the rest of the Basco-Cantabrian basin.

The Palaeozoic substratum consists of the Ordovician Barrios Quartzite and the Carboniferous Caliza de Montaña limestone (Chapter 3). The Lower Triassic

Buntersandstone consists of continental alluvial fans, braided stream and meandering river sediments deposited in an extensional basin linked to the initial, failed rifting of the Bay of Biscay. The Upper Triassic Keuper is an evaporitic mudstone extensively involved in later Mesozoic and Tertiary halokinesis. The Liassic carbonates start with an evaporitic, peritidal collapse breccia (the Carniolas) which is overlain first by inner shelf limestones and then by outer shelf limestone/marl alternations in the Upper Lias.

A major pre-rift unconformity separates the Jurassic carbonates from the Valanginian-Barremian, syn-rift, continental clastics of the Wealden (Chapter 4; Fig. 2-1). NW Santander was a half-graben with down-to-the-north, extensional faulting on the E/W-oriented Escudo de Cabuerniga fault and forms the northern part of the Santander basin. The NW Santander graben was compartmentalized by N/S trending transfer faults that had a strong, local effect on thickness and facies patterns. Virtually no Purbeckian exists in NW Santander as this area exists on the rotated, up-dip edge of the Santander basin extensional half-graben and was subjected to major pre-Wealden uplift and erosion. The Wealden itself consists of two formations:- the lower, arenacous Barcena Major Fm. (deposited in a braided stream environment) and the upper, argillaceous Vega de Pas Fm. (deposited in a meandering river environment). The latter has typical fluvial channels, crevasse splay sandstones, overbank mudstones (the dominant lithology) and lacustrine limestones.

The Santander Wealden basin is primarily a rift basin (not a strike-slip related pull-apart one) and can be related to corresponding syn-rift basins on the northern margin of the Bay of Biscay and elsewhere. Good correlation exists between the offshore seismic stratigraphy and the Cretaceous/Tertiary sequence of NW Santander. The Wealden corresponds to the syn-rift Formation 4.

Urgonian (Aptian) marine carbonates (Chapters 5 & 6; Fig. 2-1) overlie the Wealden clastics diachronously. Directly above the Wealden is a thin (20m), basal, transgressive, high-energy, sheet calcarenite - the Umbrera Fm.. This subtidal, orbitolinid grainstone shoal passes up-dip towards the palaeoshoreline

(S/SW) into tidal channel calcarenites and clastic-rich limestones. The Lower Aptian transgression continued on up into the basal Patrocinio Fm. (an alternation of dm-bedded sandstones and open marine marls - Fig. 2-1). The Patrocinio Fm. consists of a coarsening- and thickening-upward deposit of sandstone tempestites interpreted as a shoaling-upward sequence.

The San Esteban Fm. is the first Urgonian, rudistid, low-to-medium energy carbonate platform (50m thick). This *Requiena/Toucasia*/benthic foraminiferal biomicrite was deposited on an internal platform with a homoclinal ramp transition to deeper waters. The Rodezas Fm. above has a basal sandstone-rich Member but chiefly consists of marly limestones and marls containing ammonites, exogyras and echinoderms.

The main Urgonian, rudistid biomicrite was the Upper Aptian Reocin Fm. (Chapter 6; Fig. 2-1) which was also deposited in an internal platform environment. The Reocin Fm. was dominated by requienid/orbitolinid/miliolid lime mudstones/wackestones with local rudistid packstone banks, coral patch reefs and cross-bedded, skeletal grainstones above diapiric palaeohighs. The Upper Aptian saw the initiation of major, Keuper diapirism with thinned, higher-energy Reocin Fm. carbonates over the diapiric palaeohighs while thickened, micrite-rich sequences formed in the associated rim synclines.

The key feature of the diagenesis of the Reocin Fm. was its early exposure to meteoric-derived waters from the overlying subaerial exposure horizon at its top. After synsedimentary marine phreatic micritization and initial isopachous, fringing cementation, an early generation of non-ferroan, equant cement was precipitated associated with the leaching of high-magnesium calcite (HMC)/aragonite or neomorphic transformation of HMC to LMC under active meteoric phreatic conditions. Invasion of slightly reduced, Fe-rich, meteoric-derived groundwater over palaeohighs generated extensive lenses of sucrosic dolomite (the Novales Mb.) in the seawater/groundwater mixing zone. These surface-derived groundwaters became progressively more reduced and enriched in Fe and Mn and precipitated coarser-grained dolomites richer in Fe/Mn and lighter in $\delta^{18}O$. A

second, more local, phase of karstification in the basal Cenomanian brought in further estuarine-derived groundwaters and precipitated sparry, baroque dolomite. At Reocin, these hyposaline groundwaters mixed with hypersaline, sulphate-rich, Keuper-derived waters in karstic caverns at the base of the Novales Mb. and precipitated Pb/Zn sulphides by bacterial sulphate reduction.

The thin, Albian sequence (140m) of NW Santander (Chapter 7) commences with the siliciclastic Lower Albian Las Penosas Fm.. This is a transgressive fluvial/estuarine/inner shelf sequence overlying a basal ravinement. The sandy fluvial channels at the base of the Las Penosas Fm. flowed seaward (NNE). A coarsening- and thickening-upwards sequence of estuarine tidal flats, tidal creeks, crevasse splays, estuarine, distributary and sand channels, and peritidal/lagoonal black, carbonaceous muds are capped by estuarine mouth bars. The inner shelf deposits consist of sandy tidal bars and sandwaves with interbar muds. A barrier model for this Formation is rejected as no major washover sands, beach, flood-tidal deltas exist. The transgressive nature of the sequence eliminates a tidally-dominated delta model. Such a transgressive estuarine sequence has been preserved due to rapid subsidence and major siliciclastic input during, primarily eustatic, transgression. The Las Penosas Fm. generally thickens to the west across NW Santander due to sub-basin, distension-related tilting in an opposite sense to that seen in the Wealden/Aptian. Halokinesis of the Keuper evaporites continued to operate on a smaller (5Km) scale with thickened rim synclines and thinned sequences over the diapiric palaeohighs.

The Middle/Upper Albian Barcenaciones Fm. consists of high-energy calcarenitic sandwaves at its base followed by a siliciclastic-starved, storm/wave-dominated, middle shelf, marly limestone/marl alternation (the Miradorio Mb.). Skeletal packstone tempestites were followed by wave/storm reworked carbonates deposited on a homoclinal ramp, which dominates the upper part of this formation. The Albian of NW Santander differs markedly from the rest of the Basco-Cantabrian basin in that it preserves a thin, carbonate-rich sequence. To the east, thick (1-4.5Km), tidally-dominated deltas occur around Valmaseda which pass downdip into

the coeval "Black Flysch" submarine fans/slope deposits. NW Santander is seen as a moderately stable push-up block, whereas the Valmaseda area is a major pull-apart basin linked to the sinistral, distensional tectonics that affected this margin during the rotational drifting in the Bay of Biscay.

The Cenomanian commences with the major estuarine/inner shelf, tide- and storm-dominated siliciclastic Bielba Fm. (Chapter 8). Towards the hinterland areas, to the S/SW, this formation has thick estuarine channels at the base that tend to infill diapiric palaeolows developed in the rim synclines. Further offshore a thin regressive sequence on top of the Barcenaciones Fm. is capped by condensed carbonates and oyster reefs. As the Cenomanian transgression continued, the inner shelf deposition was dominated by tidal sandbars and sandwaves with intervening bioturbated, interbar sandy mudstones. These sandbars and sandwaves are dominantly oriented offshore (NNW/NE) following the ebb tidal current, though shore-parallel sand bars running WNW/ESE (parallel to the Hercynian basement lineaments of the Basco-Cantabrian basin) do occur further up section/further off palaeoshore. Clear tidal features have been found in both the channels and the inner shelf sand waves, and stormwave HCS has been recorded from the lower parts of coarsening- and thickening-upward tidal sand bars. The Bielba Fm. thins into active diapiric palaeohighs which tend to be poor in estuarine/coastal sands but rich in shallow-marine sandbars and sandwaves.

The clastic inner shelf deposits interdigitate with an offshore marine calcarenite belt (the Altamira Fm.) that transgressed over the Bielba Fm. in the Middle/Upper Cenomanian. The Altamira Fm. has a diverse marine fauna and is dominated by moderate- to high-energy calcarenites that were reworked by tides and storms and prograded as spillover lobes during storms. NW Santander intergrates with the basinwide Cenomanian facies belt distribution that follows a general WNW/ESE trend with deeper water deposits existing to the N/NE. This lateral continuity of facies in the basin is linked to moderate, more uniform, thermal subsidence of the passive northern Iberian margin during a time of active drifting in the Bay of Biscay.

The Turonian to Middle Campanian is dominated by uniform alternations of marly limestones and marls on a dm-scale (Chapter 9). They were deposited in an outer shelfal area. These bio/biopel/pelmicrites contain benthonic and planktonic foraminifera, echinoderms, pteriomorph bivalves, quartz silt and glauconite. The basal Turonian is a hiatus with glauconitic hardgrounds and *Thalassinoides* burrows. It correlates with seismo-stratigraphic Formation 2/3 boundary offshore. There was a slight shoaling in the Coniacian and a deepening into the Santonian. Marly limestones of the outer shelf characterise the Lower/Middle Campanian.

The Upper Campanian to Middle Eocene (Chapter 10) is dominated by sandy, benthic foraminiferal, middle shelf limestones that were deposited in a shallow marine environment that lay between the Bilbao flysch to the N/NE and the more continental "Garumnian" clastics to the S/SW.

After the deposition of inner shelf limestones in the uppermost Cretaceous, there was extensive mixed-water dolomitization of the Maastrichtian/Dano-Montian sequence, especially towards the hinterland (S/SW). The Lower/Middle Eocene sequence is dominated by "nummulitic" calcarenite banks of the middle shelf. A major clastic invasion occurred in the Lower Cuisian. The uppermost Cretaceous to Middle Eocene was a relatively quiet peroid tectonically in NW Santander, with only mild diapiric rim syncline movement and basinward downwarping. In the Pyrenees and Bilbao trough to the east, compressional Pyrenean deformation had already started by this time.

The Upper Eocene/Oligocene San Vicente Group is a reef-carbonate slope-apron deposit that was deposited coevally with Pyrenean thrusting and Keuper piercement. The youngest pre-compressional formation was the La Acebosa Fm. which consists of thin-bedded turbidites in a background of outer shelf marl. The flyschoid Oyambre (Upper Eocene) and Meron (Upper Oligocene) Fms. are mainly composed of hemipelagic, periplatform ooze, neritic-derived, calcarenitic turbidites and thicker rudaceous mass flows (disorganized or graded resedimented conglomerates, debris flows etc.). These were deposited downslope of the coeval,

shallow-water, reefal Los Via Fm. which contains corallinaceae, bryozoa, "nummulitic" foraminifera and scleractinia. The deepwater Oyambre and Meron Fms. were deposited in a fault-bounded, reef-carbonate slope-apron bypass margin. Turbiditic flow was mainly axial to the structural grain of the basin. Deepwater sedimentation was coeval with active thrusting of E/W-trending thrusts, which were themselves reactivations of the Hercynian thrusts. There was also extensive Keuper piercement through to the seafloor in locations largely conditioned by major transfer structures. This active tectonism unroofed the entire Phanerozoic column, and after subaerial/littoral reworking, polymict clasts of it were resedimented in rudaceous mass flows into the Oyambre and Meron Fms. A major debris flow 90m thick exists at the Oyambre/Meron boundary (the Rupuente megabed) and contains megaclasts of brecciated, rockfall/bedded Cenomanian Altamira Fm. up to 40m wide. A Lower Oligocene hiatus can be correlated with offshore sequences/seismic stratigraphy. The initial thrusting/piercement occurred in the Upper Eocene to lower Upper Oligocene, though further, more major compressional thrusting and uplift occurred in post-Oligocene times to expose the strata as they occur today.

In conclusion, the myriad of carbonate and clastic sequences from deep-marine, shallow-marine, coastal and non-marine environments seen in the Cretaceous/Tertiary of NW Santander can be directly related to the rifting, drifting and partial closure of the Bay of Biscay.

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CHAPTER 1 - INTRODUCTION.

1.1 AIMS AND METHODOLOGY OF THE THESIS.

The objective of this project was to examine the sedimentology, and the relationship between tectonics and sedimentology, in the onland portion of the Spanish continental margin in the Basco-Cantabrian basin during the Cretaceous and Tertiary periods. In particular, the thesis considers the distinct sub-basin of NW Santander in the Basco-Cantabrian basin (Fig. 1-1).

Investigation was principally done by vertical sequence logging of the good cliff exposures in the area. Facies analysis (including petrography and geochemistry on the carbonates) led to environmental reconstructions and palaeogeographies through space and time. They were then intergrated with the active, synsedimentary tectonics of the area.

Existing work on NW Santander (see Chapter 2) has been mainly biostratigraphic or has only considered depositional environments on a very generalised, basin-wide scale. It is the aim of this research to fill this major gap by detailed vertical sequence facies analysis which is considered in Chapters 4 to 11. Tectonic control on sedimentation is considered in chapter 12.

The first volume of this thesis contains the preface and text. The figures are given in volume 2 and early part of volume 3. The latter part of volume 3 contains sedimentary logs of the sections and the IGME (= Spanish Geological Survey) maps of the area. A key to the lithological and faunal symbols used throughout the thesis are given at the start of the graphic logs section.

1.2 Introduction to the Basco-Cantabrian basin.

The Basco-Cantabrian basin forms part of the southern margin of the Bay of Biscay and the northern limit of the Iberian plate (Figs. 1-1 & 1-3). To the east of the Basco-Cantabrian basin is the Pyrenean suture (Figs. 1-1, 1-2 & 1-3). To the west of the Pyrenees, the northern Iberian plate margin passes along the

Basque Arc (Figs. 1-2 & 1-4; Amiot et al, 1983) in the Basco-Cantabrian basin. Further west still the plate margin passes into the offshore, northern continental margin of Spain (Fig 1-3).

The Basco-Cantabrian basin itself is bounded to the south by the Tertiary molasse basins of the Ebro and Duero, to the west by the Palaeozoic Asturian massif and to the east by the main Pyrenean chain (Fig. 1-2).

The tectonic history of the Basco-Cantabrian basin is closely linked to the relative movements between the Iberian and European plates and the evolution of the Bay of Biscay. The Bay of Biscay is underlain by magnetically-striped oceanic crust whose wider, western end is terminated by the north-south Atlantic Anomaly 31-32 (Soler et al., 1981; Fig. 1-1). To the NE of the oceanic crust in the Bay of Biscay is the passive, Atlantic-type, Armorican margin with Upper Jurassic/Lower Cretaceous extensional basins (Fig. 1-1; Montadert et al., 1974, 1979).

The southern, Spanish margin to the Bay of Biscay had similar Mesozoic extensional basins that were later extensively deformed and overthrusted by the compressive, Tertiary Pyrenean orogeny. The Basco-cantabrian basin is part of the southern margin of the Bay of Biscay and has similar Mesozoic extensional basins that were inverted in the Tertiary.

This thesis is primarily concerned with the sedimentary infill of this basin. Chapter 2 is provides a summary of the sedimentary infill of the whole Basco-Cantabrian basin. Chapters 4 to 11 provide detailed sedimentology of the Cretaceous and Tertiary infill of the northwestern part of the Basco-Cantabrian basin (the NW Santander area). The tectonic evolution of the basin is considered in Chapter 12.

1. 3 The structure of the Basco-Cantabrian basin.

The Basco-Cantabrian basin is infilled by a over 15km of dominantly Cretaceous-Lower Tertiary strata. At the centre of the basin is the NW-SE trending Basque Arc (Fig. 1-2). To the SW of it lies the more stable hinterlands

of the Peri-Asturian and Navarro-Cantabrian domains (Fig. 1-2). The southeastern border of the Basco-Cantabrian basin is traditionally taken at the Cinco-Villas massif/Pamplona diapiric lineament (Figs. 1-2 & 1-4). The most northeasterly part of the basin, offshore from San Sebastian, consists of a series of major thrusts with northward/northeastward transport that are a westerly continuation of the North Pyrenean Frontal Thrusts (Figs. 1-2 & 1-4).

The Basque Arc trends on about 110° and is composed of several elements. Working from the NE to the SW, it consists of the North Biscay (= Vizcaya) Anticline (NBA), the Biscay syncline (= Vizcaya syncline, =BS) and the Biscay (= Bilbao) anticlinorium (BA; Fig. 1-4). The Biscay syncline trends 110° with Cretaceous to Eocene flysch in its core. The southern flank of it is faulted (Fig. 1-4) and probably corresponds to a late Hercynian lineament reactivated as a thrust fault at depth - the Ochandiano (= Calamo, = Durango) fault of Soler et al., 1982 (Figs. 1-4 & 1-5). This fault is considered to be the suture between the Iberian (to the SW) and the Eurasian plates (to the NE) by Floquet & Mathey, 1984 and Amiot et al.,1983 amongst others. This fault (and other Basco-Cantabrian structures) continues offshore towards the NE of the Le Danois bank (Fig. 1-4) on the same orientation (Deregnaucourt, 1981). The NE flank of the Biscay syncline is more gentle than the SW one and the struct e can be traced southeastward into the Nappe des Marbres (NM). Here Albian se ments have been affected by intense, thermal metamorphism (Albarede & Michard-Vitrac, 1977) and schistocity. Further SW, the Biscay anticlinorium is a complex, arcuate, faulted anticlinorium linked to thrusting at depth (Fig. 1-4).

Along the southern margin of the Basco-Cantabrian basin, to the SW of the Basque Arc, are the frontal thrusts of the Sierra de Cantabria (SdCFT, Fig. 1-4) which thrust to the south over the Tertiary Ebro basin. The southwestern margin to the basin is the Cantabrian (= Aviles, =Ubierna, =Cardano) fault (Fig. 1-4), a complex thrust structure with a significant Tertiary, sinistral strike-slip component. The major decollement for thrusting in the southern frontal thrusts is the Upper Triassic Keuper evaporitic melange. This bipolarity of thrust directions

gives the whole basin a mega "pop-up" structure, linked to extensive basin inversion in the mid-Tertiary, Pyrenean compression.

The western Peri-Asturian part of the Basco-Cantabrian basin (Figs. 1-2 & 1-4) generally lacks the major, Tertiary thrusting of the Basque Arc and is dominated by reactivations of late Hercynian lineaments (Fig. 1-5). On land, the Basque Arc dies out just west of Bilbao, around Castro-Urdiales/Laredo (Fig. 1-4). Between there and Santander there are N/S trending structures of the Rio Miera flexure (Fig. 1-4).

The NW Santander subbasin is dominated by the E/W trending Escudo de Cabuerniga fault, N/S trending structures such as the Rio Miera flexure and WSW/ENE trending folds (Figs. 1-2, 1-4, 1-5, 1-6 & 1-7). It has been pierced by numerous Keuper diapirs (Figs. 1-6 & 1-7) and is considered in more detail in Chapter 2.

The main structural deformation of the whole basin was in the Eocene-Miocene which inverted many of the extensional structures inherited from the Cretaceous. The fill of the basin is considered below and is summarised in the "Long Abstract" and in Figs. 1-9 & 1-10.

1.4 Sedimentary infill of the NW Santander basin.

This section will provide a summary of the Phanerozoic stratigraphy of the NW Santander subbasin.

1. 4. 1 : The Palaeozoic basement of the NW Santander area.

The Palaeozoic sequence present in NW Santander is dominated by the Lower Ordovician Barrios Fm. and the Namurian (Upper Carboniferous) Caliza de Montaña limestone that forms the NE edge of the Cantabrian zone of the Asturian massif.

The Barrios quartzite is at least 550m thick (Pechon section at 53579755, base unseen) and mainly consists of a well-sorted, medium-grained, metamorphosed, micaceous quartzarenite. It was mainly deposited in a braided alluvial environment

that interdigitated with coastal and shelfal sands to the west (Aramburu & Garcia-Ramos, 1984).

The Carboniferous sequence in the Asturian massif to the west of NW Santander (Fig. 1-2) is dominated by the Namurian Caliza de Montaña (= H^{Ba-Bb} limestone. It is a generally massive, shelfal biomicrite that is about 300m thick on the Pechon section (53489764). No Upper Carboniferous sediments exist in NW Santander (Heward & Reading, 1980).

1. 4. 2 : The Permo (?) - Triassic Buntersandstone.

The Triassic of NW Santander is divided on Germanic lines into a lower Buntersandstone and an upper Keuper, with the Muschelkalk missing. The Buntersandstone (Permian (?) - mid Triassic (?)) directly overlies the Carboniferous limestones and is best exposed on the N-dipping flank of the Escudo de Cabuerniga (Figs. 1-6 & 1-7). It consists of continental, arenaceous red-bed siliciclastics that infill local, fault-controlled, extensional basins and topographic palaeovalleys left after the Hercynian orogeny (Garcia Mondejar, Pujalte & Robles, 1986). The section through the Escudo at Bielba (53709672 to 53819659) is 250m thick, but the Buntersandstone thickens sharply to the south of the Escudo to over 900m (IGMEg). From the base upwards, the Buntersandstone of the Escudo consists of three megacycles, each of which broadly fines upwards (Saiz de Omenaca Gonzalez, 1977). These are :-

1) A coarse-grained breccia/conglomeratic sandstone (0-250m thick) that grade from a basal fossil scree slope to alluvial fan/ braid stream deposits above.

2) A pebbly sandstone (5-80m thick) that was deposited in a braided stream environment capped by finer-grained meandering river deposits with 2-10m thick channels.

3) A dominantly muddy sandstone/mudstone sequence deposited in a meandering river environment. Individual channels have basal microconglomeratic sandstones that are overlain by cross-bedded sandstones which in turn pass up into extensive

overbank mudstones. Ophitic basalts and volcaniclastic sediments occur in this megacycle at the eastern end of the Escudo at Villafufre (58349627).

The Escudo Buntersandstone has yielded a typical Iberian Triassic palaeomagnetic pole (Schott, pers. comm., Univ. of Strasbourg), implying that Santander has undergone a post-Triassic anticlockwise rotation of about 35° relative to stable Europe (Van der Voo, 1969; Vandenberg, 1980). Hence NW Santander lies SW of any plate suture between Iberia and Europe.

1. 4. 3 : The Upper Triassic Keuper.

The Keuper consists of homogeneous, slightly dolomitic, red mudstones with evaporites. At the surface the only evaporites exposed are cm-thick sheets of satin-spar gypsum, through extensive sodium chloride has been proved below the dissolution zone (which is about 400m thick from drilling results quoted in Vadala, 1981). This salt has been commercially extracted by extensive solution mining of the Polanco diapir at a rate of 1.3 million tonnes/year (Vadala, 1981). Saccharoidal, laminated and fibrous gypsum have all been found along with bipyramidal quartz, rhombohedral dolomite, aragonite and pyrite cubes. The clays are about two-thirds illite and one-third chlorite (Rohou, 1984), typical for European Germanic Keuper (Jeans, 1978). An emphemeral, evaporitic playa lake which was periodically desiccated is suggested as a depositional environment.

The Keuper has been extensively remobilised into diapirs (Figs. 1-6 & 1-7) that were active during the Cretaceous-Tertiary (Fig. 1-11). Drilling through the Keuper has yielded thicknesses of 1,300m at Polanco diapir, over 132m in Orena 1 borehole (at 56349778; Rios, 1967) and 317m in Ancillo 1 at Ramales, ESE of Santander (Querol, 1973). Although halokinesis has strongly distorted the original thicknesses, a depositional thickness of the order of 200m is suggested.

1. 4. 4 : The Jurassic carbonates.

The Jurassic is composed of a rather uniform sequence of carbonates that extend laterally over much of the Basco-Cantabrian basin (Fig. 1-12; Dahm, 1966;

Perconig, 1968; Ramirez del Pozo, 1971a; Sbeta, 1976). At the base is the "Carniolas" (Rheatian (?) to Hettangian, = J_{11-12}^{O-1} of IGMEa), which represents a transgressive sequence from a basal carbonate sabkha to inner shelf limestones. The base of the sequence commonly consists of an intraformational breccia and carbonates with microdolomites, dolomicrosparites with algal laminates and gypsum pseudomorphs, pelmicrites and massive micrites. They were deposited in an evaporitic peritidal sabkha associated with very early dolomitization and recrystallisation. The upper part of the Carniolas (Lower to Middle Sinemurian) consists of medium to thick-bedded, grey (oo)biosparites and (pel)biomicrites that have a marine fauna including bivalves, brachiopods, bryozoa, algae, foraminifera, ostracods and crinoids. It was deposited in a moderately agitated to semi-restricted shelf with local calcarenitic bars.

Transgression continued into the mid Lias. In the Upper Lias (Upper Sinemurian to Toarcian (= J_{12-14}^{3-O} of IGME maps) monotonous, decimetre-bedded, marly limestone/marl alternations rich in a stenohaline fauna including ammonites, brachiopods, belemnites, ichnofauna, ostracods and foraminifera were deposited on a low-energy, mid- to outer, epicontinental shelf. In NW Santander only the basal part of the Upper Lias is present and then only poorly-exposed (eg. at Obregon diapir, Figs. 1-6 & 1-7, 58489711) and is about 200m thick.

The Middle Jurassic is only present south of the Escudo de Cabuerniga (Fig. 1-12; IGMEg). It is rather similar to the Upper Lias lithologically and consists of alternations of shelfal limestone and marl. In the *Davoei* - *Margaritatus* zones of the Pliensbachian south of the Escudo, thick (130m at Reinosa), black, bituminous shales exist (source rock), deposited in an outer shelf to basinal anoxic environment. Throughout the Basco-Cantabrian basin a prominent unconformity exits at the top of the marine Jurassic with progressively more Jurassic sequence preserved towards the NE (Fig. 1-13). South of the Escudo the unconformity is underlain by marine mid-Callovian, but in NW Santander almost all of the Middle Jurassic, most of the Upper Lias, and even most of the Lower Lias on some sections have been erosively removed (eg. Section 15 at Cuchia). This is the

pre-rift unconformity overlain by the fluvio-lacustrine Purbecko-Wealden (Figs. 1-6 & 1-9) and is discussed further in Chapter 12. The description of the Cretaceous - Tertiary sequence in NW Santander area given below is only a brief summary and is considered in more detail in Chapters 4 to 11.

1. 4. 5 : The fluvio-lacustrine Kimmeridgian-Barremian Purbecko-Wealden (Chapter 4).

The Kimmeridgian-Barremian sequence in NW Santander is a dominantly continental, fluvial siliciclastic sequence that infills extensional basins associated with the active rifting stage of the opening of the Bay of Biscay (Fig. 1-9). The older Purbeckian part of the syn-rift sedimentation is only present south of the Escudo de Cabuerniga and mainly consists of alluvial to marginal (?) marine siliciclastics with a basal lacustrine limestone.

The younger syn-rift sequence is the Wealden Pas Group (Fig. 1-9) that is subdivided into an older Barcena Major Fm. deposited in a braided-stream environment and a younger Vega de Pas Fm. deposited in a meandering river environment.

1. 4. 6 : The Aptian Urgonian carbonates of the Torrelavega Group (Fig. 1-9; Chapters 5 & 6)

The Aptian in NW Santander is characterised by extensive shallow marine, rudistid, platform carbonates with minor interdigitations of clastics in the Lower Aptian (Fig. 1-9). During the Bedoulian (= Lower Aptian) there was a progressive overstep onto the Hercynian basement. The Upper Aptian was a period of rapid sedimentation and the initiation of major diapiric activity. The Upper Aptian limestones were extensively dolomitized by meteroric waters in end Aptian times.

1. 4. 7 : The Albian/Cenomanian Supra-Urgonian clastics and carbonates of the Santander Group (Chapters 7 & 8).

The Albian was characterised by a transgressive cycle that passed upwards from a lower tidally-dominated, estuarine/inner shelf siliciclastics into an upper middle/outer shelf carbonates (Fig. 1-9). A similar transgressive cycle was repeated in the Cenomanian.

1. 4. 8 : The Turonian-Campanian Sardinero Group and the Maastrichtian-Middle Eocene Nansa Group (Chapters 9 & 10).

The Turonian-Campanian Sardinero Group is a sequence of outer shelf, hemipelagic carbonates, mainly composed of limestone-marl alternations. The Maastrichtian-Middle Eocene Nansa Group is a sequence of predominantly inner/middle shelf benthic foraminiferal limestones (Figs. 1-9 & 1-10).

1. 4. 9 : The Upper Eocene/Oligocene San Vicente Group (Chapter 11; Fig. 1-10).

The Upper Eocene/Oligocene San Vicente Group is a carbonate reef-slope-apron deposit that formed coevally with the Pyrenean thrusting and Keuper piercement. The San Vicente Group is the youngest pre-drift strata in the NW Santander area.

CHAPTER 2 - STRUCTURE OF THE NW SANTANDER SUBBASIN.

2. 1 Introduction to the NW Santander subbasin.

The NW Santander subbasin lies to the north of the Escudo de Cabuerniga, to the east of the Asturian massif, south of the coast, and to the west of the city of Santander (Figs. 1-1, 1-2 & 1-3). The N/S meridian through Santander city (the Rio Miera flexure) and the E/W Escudo de Cabuerniga fault zone were important palaeogeographic lineaments in the Mesozoic/Tertiary that set NW Santander apart, as a separate sub-basin. Although rocks from Ordovician to Upper Oligocene age are exposed in this area, the outcrop is dominated by Cretaceous and Tertiary clastics and carbonates. Continuous, relatively gentle subsidence prevailed in this sub-basin throughout most of the Mesozoic and Tertiary preserving the most complete stratigraphic sequence in the entire Basco-Cantabrian basin.

2. 2 Structural geology of NW Santander.

To the west of the city of Santander, the structure is dominated by subparallel N-S and E-W late Hercynian lineaments (Figs, 1-6, 1-7 & 1-8). The most important E-W lineament is the Escudo de Cabuerniga, a tongue of Hercynian basement/Triassic Buntersandstone that was transported southward by basement-involved thrusting in the Tertiary (Figs. 1-3 & 1-6). Within the study area, the lithologies can be subdivided into a Triassic Buntersandstone/Palaeozoic basement, a Keuper evaporitic mudstone decollement involved in halokinesis and thrusting, and the Jurassic to Tertiary cover sequence.

The geological (Fig. 1-6 and the IGME Comillas and Torrelavega sheets) and structural (Fig. 1-7) maps of NW Santander show that the area can be subdivided into three main structural zones :-

Zone 1) The zone to the west of San Vicente de la Barquera (54239758 - for location see Fig. 1-8) with an E-W lineation inherited from the underlying Hercynian structure.

Zone 2) A complex zone of thrusting and faulting to the S/SE of San Vicente de la Barquera.

Zone 3) Lying to the east of Zone 2, the majority of NW Santander has ENE-WSW trending folds which extends offshore onto the Spanish continental shelf (Fig. 1-7). This ENE-WSW trend is known as the "Santander trend" by Feuillee & Rat, 1971. This fold trend is disrupted by piercement structures that fault and fold, and locally overturn, the strata. These always have Keuper evaporitic mudstones in their cores.

2.3 Zone 1 : West of San Vicente de la Barquera.

In the Zone to the west of San Vicente de la Barquera, the strong E-W lineation of the folds (Fig. 1-7 and IGME Comillas map for cross-sections) is directly inherited from reactivations of Hercynian thrusts (dating from the Carboniferous Asturian phase) that characterise the Hercynian of the coastal ranges of the Cantabrian Mountains (IGMEk; Heward & Reading, 1980). These basement-involved, E-W oriented thrusts have a southward tectonic transport direction towards the core of the Cantabrian Mountain belt and are associated with southward vergent anticlines. The thrusts were associated with major syn-tectonic basin generation in the Upper Carboniferous. They were reactivated during the Upper Eocene to Miocene (?) compressional period of the Pyrenean orogeny, both onshore and offshore (Fig. 1-7).

The most important example of an E-W oriented thrust structure is the Escudo de Cabuerniga Thrust system which lies to the south of NW Santander (Figs. 1-6 & 1-7). This fault system was active as a horst structure during the Mesozoic with syn-sedimentary downthrow both to the south and the north (eg. during the Wealden basin-rifting phase of Chapter 4). However, during the Tertiary it was the northerly-dipping fault plane that was reactivated as a thrust.

2.4 Zone 2 : San Vicente de la Barquera thrust zone.

The complex zone S/SE of San Vicente de la Barquera (Figs. 1-6 & 1-7) is dominated by reactivation of a N/S oriented Hercynian lineament at depth which separates the stable zone 1 to the west from zone 3 to the east. It is affected by Keuper evaporitic piercement at depth and is characterised by low-angle, thin-skinned thrusting with a westerly transport direction on the western thrusts (such as the Las Arenas Thrust, Fig. 1-7). The thrusting was, at least in part, contemporaneous with the Upper Eocene-Oligocene, deep-water sedimentation of the San Vicente Group (Chapter 11).

2.5 Zone 3 : NW Santander to the east of the San Vicente thrust zone.

The region to the east of the San Vicente thrust zone is dominated by folds trending on approximately N 060° (Figs. 1-6 & 1-7). These folds are generally symmetrical, open and normal with only a slight plunge (generally less than 5°). All the anticlines have cores that have been pierced by Keuper evaporitic mudstones that have been instrumental in developing extensive, high-angle, reverse, hinge-faults around the cores of the anticlinoria. This N 060° trend - the "Santander trend" of Feuillee & Rat (1971) - is unusual in the Basco-Cantabrian basin. It probably reflects a local Hercynian lineation in the basement as seen in the inlier of Carboniferous limestone to the south of Torrelavega (Fig. 1-7).

Halokinesis has had a profound effect on the Mid Cretaceous to Tertiary geology of the area. The initial stages of Keuper diapirism (the "pillow stage" of Trusheim (1960), Seni & Jackson (1983) and Fig. 1-11) started virtually synchronously in the Upper Aptian, although slight indications of halokinetic movement can be seen even earlier. To judge from the alignment of most diapirs with regional trends, the relatively thin overburden during the pillow stage of halokinesis (as little as 500m) and the near synchronous start of halokinesis of diapirs of different sizes, a tectonic trigger (rather than a pure gravity one) to their initiation seems most probable. Indeed, throughout the western Basco-

Cantabrian basin a prominent phase of diapirism started at this time (Garcia Mondejar et al., 1982).

Once triggered into motion, the evaporitic mudstone responded to an imbalance in the overburden and/or slight basement fault reactivation and continued to flow into the diapiric anticlinoria from the Upper Aptian until the Middle Eocene. The most obvious effect of this movement is a pronounced thinning of the strata over the diapiric highs. Moderate facies changes occurred over these palaeohigh diapirs with higher-energy, arenite-rich facies locally forming over the palaeohighs and thickened, argillite-rich facies in the associated diapiric, "rim syncline" palaeolows (see Chapters 6 to 9).

During the Tertiary, Pyrenean compressive movements selected these pre-existing structures and numerous high-angle, reverse, hinge faults pierced the cores of the anticlinoria. The strata near the cores of the diapirs are nearly always oversteepened, and are locally overturned (eg. at Punta Calderon in the Caberredondo anticline at 56379789).

Complicating this picture are a number of N/S (N 000° to N 020°) lineaments that are picked out by lines of tectonic Keuper piercements and which continue offshore (eg. the Rio Miera flexure, Fig. 1-7). These acted as transfer structures between the major E/W lineaments and were active during Mesozoic/Tertiary sedimentation (Chapter 12). Onshore (Fig. 1-7) they correspond to the Bustriguado-San Vicente lineament, the Torrelavega-Miengo (57509804) lineament and the Rio Miera flexure just east of Santander.

CHAPTER 3 : PREVIOUS RESEARCH ON NW SANTANDER.

3. 1 : INTRODUCTION.

This section will provide the general background of previous work on the area and concludes with a outline of the project in relation to earlier work. This chapter will summarise the existing research framework within which this thesis was done. For those unfamiliar with general northern Iberian geology (the overall context of the area), summaries can be found in Querol (1969); Mangin and Rat (1962); Mattauer and Henry (1974); Wilson (1975); Rios (1978); Alvaredo et al. (1980) and Anonymous (1971, 1982).

3. 2 : A summary of previous research in NW Santander.

The earliest geological research on the area was by Guillermo Schulz (1837) who noted the Cretaceous and the "calcaires nummulitiques" (Tertiary) of the area. Much of the nineteenth century work on the area is summarized by Mengaud (1920) and Ciry (1940). The more important nineteenth century works are by Verneuil (1849) who described the Tertiary around San Vicente de la Barquera, Verneuil et al., (1860) who described the Cretaceous and Gascue (1875) on the Triassic around Cabezon de la Sal. Gonzalez Linares (1876, 1878, 1881) recognised *Unio* and *Paludines* as typical "Wealden" fauna and first applied this name to the non-marine, siliciclastic Neocomian.

Early structural work suggested that the area was a thrust terrain and already linked it to similar structures in the western Pyrenees (Termier (1905); Mengaud (1912); Bertrand & Mengaud (1912)).

During the early years of this century Louis Mengaud produced truly outstanding research on the area, synthesised in the key publication, Mengaud (1920). His work was primarily palaeontological/biostratigraphic and his macrofaunal descriptions of many sections remain unsurpassed to this day. His work covered the Wealden (1913)a, the Urgonian (1911, 1913b, 1913c & 1914), the

Supra-Urgonian (1920), the Upper Cretaceous (1910)a and the Tertiary (1908, 1910b, 1911, 1920).

Karrenburg (1934) covered many of the same sections as Mengaud in the area between Bilbao, Burgos and Oviedo. The very comprehensive work of Ciry (1940) describes the Mesozoic of the Basco-Cantabrian basin and illustrates the increase in marine influence on sedimentation in the Lower Cretaceous towards the NE. His work is summarized in Ciry et al., (1967) which gives a general field guide to the basin as well as a description of the Rio Saja section at Monte Barbecha between Caser (56039703) and Barcenaciones (56269712), west of Torrelavega.

Rios, Almeda & Garrido (1945) describe the Cretaceous between Bilbao and Miranda [de Ebro] (Fig. 1-4) with a brief description of the NW Santander area. Rios (1956) summarizes both the Basco-Cantabrian basin and the Pyrenees s.s..

A very comprehensive work on the Cretaceous sediments of the Basco Cantabrian basin, comparable to that of Ciry (1940), was published by Rat (1959)a. His main study was on the Urgonian as well as some work on the "sus-Aptien" or Supra-Urgonian clastics "complex". Rat (1959)a considered the geology of NW Santander only briefly and concentrated on the sections at Cuchia, Suances and Rio Saja. Further work by him has addressed the Urgonian (1959b,c; 1960 and with Magniez (1972)), Purbecko-Wealden geology (1961, 1962, 1963a, 1969) and, more recently, general problems of Basco-Cantabrian basin evolution (Mangin & Rat, 1962; Feuillee & Rat, 1971; Rat et al., 1982 and in Amiot et al., 1983).

The Pb/Zn sulphide deposits in the Urgonian carbonates around Reocin have been described by Monseur (1959, 1961, 1966, 1967) and by Monseur and Pel in 1972 & 1973. As well as giving detailed mining reports, these works propose a synsedimentary origin for the ore body. Earlier references to the mineralization are given in Mengaud (1920) and IGMEd. Adelhardt (1968) gives descriptions of the ore body and stable sulphur isotope data on it.

Feuillee (1967) discusses the transgressive Cenomanian of the whole basin. His work in NW Santander concentrates on clastic petrography and the

macrofauna of the Santander Group (Fig. 1-9) from the Comillas, Suances and Sardinero beach (98555924) sections. He also considers the wider problems of carbonate sedimentation in the Albo-Cenomanian in the basin (Feuillee, 1971).

A major work on the microfauna, microfacies and stratigraphic correlation in the Basco-Cantabrian basin was done by Ramirez del Pozo (1971a). However the study excluded any work on NW Santander. He published summarises in 1969a,b and with Aguilar Tomas (1968, 1972). He also studied the microfauna of the Aptian-Albian sections at Bielba (53779682), La Vega (54629688) and Santiago (59189762) in 1972. The Bielba section's Upper Cretaceous was studied by Carreras Suarez and Ramirez del Pozo (1971) and the Tertiary there by Ramirez del Pozo (1971b).

Aguilar Tomas (1970) subdivided the Albian of the whole Basco-Cantabrian basin into four groups. His work on the clastics was primarily petrographic/ geochemical . His work is summarized in Aguilar Tomas 1971a,b & 1975.

One of the most important works on the area was the systematic mapping of the entire area at 1:50,000 by the IGME (a,c,e,g,i,k,m). Brief memoirs are provided to each sheet that give simple lithological descriptions of the units and extensive lists of the microfauna. Each sheet is provided with supplementary stratigraphic sections available on "open-file" from IGME, Madrid (IGME b,d,f,j,l). Copies of the key IGME Comillas and Torrelavega sheets are given at the end of volume 3. This mapping is of good quality and its attendant stratigraphy and mapping were used as the basis for the stratigraphy and mapping in this work (Figs. 1-9 & 1-10). A correlation chart between the IGME mapping units and the lithostratigraphic names used in the thesis is shown in Fig. 2-1. The eight figure numbers (eg. 54369752) used in this thesis refer to the IGME reference grid as used on the maps at the end of the thesis.

Pujalte (1974, 1976, 1977, 1981 & 1985) did modern sedimentological work on the Wealden of the Basco-Cantabrian basin, including a brief study of NW Santander. Along with Garcia Mondejar (1975, 1976 & 1981) and Garcia de Cortazar (1982), he emphasised the active role of tectonism on sedimentation.

Garcia Mondejar (1979a) worked on the Urgonian to the SW of NE Santander in the southern Ason-Soba area of Santander province. The Urgonian carbonates here have extensive interdigitations of siliciclastics and have been correlated (Garcia Mondejar & Pujalte, 1981) with the Rio Saja section in NW Santander which was mentioned above. He subdivides the Urgonian (Aptian-Albian) into 4 units (U_1 to U_4) and considers their sedimentology, palaeogeographies and stratigraphy in Garcia Mondejar (1975, 1979a,b,c, 1982 & 1985) and with Pascal (1978). Garcia Mondejar et al., 1982 produced a lithostratigraphic review of the entire Cretaceous of the Basco-Cantabrian basin. As regards NW Santander, their main work was to apply formal lithostratigraphic formation names to the mapping units used by IGMEa,c etc.. These were later adapted and used by the present author (Figs. 1-9 & 1-10).

Andre Pascal has worked extensively on the Urgonian carbonates of the whole basin and has proposed a fourfold "Biosedimentary System" subdivision of the Apto-Albian (1974, 1976, 1979, 1982a,b,c, 1983 & 1985). In Pascal (1985), he gives detailed biostratigraphic descriptions of the Apto-Albian sections north of San Vicente de la Barquera, on the Barcenaciones road section, at Suances and at Cuchia. More general problems of the Apto-Albian are considered by Rat and Pascal (1979, 1982). In 1983 Amiot et al. published an extensive work on the main conclusions of the "Dijon school" including the work of Rat, Salomon, etc.. This work was summarised in Amiot et al., 1982. A summary of the tectonic evolution of the Basco-Cantabrian basin from the "Tubingen school" is given in Wiedmann et al., 1983.

Further work was done on the Pb/Zn sulphide mineralization by Sanchez (1971, 1976). Extensive work on the geochemistry and genesis of the Pb/Zn mineralization and associated dolomitization was produced by Vadala (1981). His results are summarised in Vadala et al., 1981 and further work on the dolomites associated with the mineralization is given by Barbanson et al., 1983. The biostratigraphy of the Lower Aptian was improved with the use of ammonites by Collignon et al., 1979.

The present author (Hines, 1985a) produced a systematic facies summary of the Cretaceous/Tertiary sequence of NW Santander. This work emphasises the tectonic control on sedimentation. It also provides a field guide to outcrops of the Urgonian, Supra-Urgonian and Tertiary San Vicente Group flysch (Fig. 1-10). More detailed work on the Urgonian dolomitization (1985c) and the flysch sedimentation (1985b) have also been published by the author.

3. 3 : Comments on the existing research in NW Santander.

The above review illustrates that the vast majority of earlier geological work on NW Santander has been mainly palaeontological/ biostratigraphic and generally published in local journals. It also shows that the existing work on NW Santander has only considered depositional environments on a very generalised, basin-wide scale. The following chapters (4 to 11) will consider the facies and depositional environments in more detail, primarily by detailed vertical sequence facies analysis from the good coastal exposures.

CHAPTER 4 - THE NEOCOMIAN/BARREMIAN "WEALDEN" PAS GROUP - FLUVIAL INFILL OF SYN-RIFT, EXTENSIONAL BASINS.

4.1 INTRODUCTION.

The upper part of the Jurassic and the lowermost Cretaceous of northern Spain mainly consists of dominantly non-marine clastic sediments (Fig. 1-6). Sedimentation during this period is closely connected with the main rifting stage of the Bay of Biscay which gave rise to a number of major marginal basins. On the northern margin of the Bay of Biscay, rotational half-grabens with active syn-rift infill have been well preserved and are known from seismic/drilling surveys (Montadert et al., 1979). The axial part of the Bay remained under the sea throughout the Upper Jurassic - Lower Cretaceous (Boillot et al., 1979a). In northern Spain the Asturian basin was closest to the sea (Fig 1-1) and preserved a continental to restricted marine sequence (Garcia-Ramos & Valenzuela, 1979) whilst offshore deltaic to open marine conditions persisted (Boillot et al., 1979). In the Cameros basin (Fig. 1-1), to the south of the Basco-Cantabrian basin, continental fluvio-lacustrine conditions persisted throughout the whole period (Salomon, 1982; Platt, 1986). The Santander basin, which was developed in the Peri-Asturian domain of the western Basco-Cantabrian basin (Fig. 1-2), was intermediate between the Asturian and Cameros basins.

Gonzalez Linares (1876) was the first to compare with, and name the rocks after, the Purbeck and Wealden sequences of southern England. A relatively close similarity does exist between the British and Spanish sequences with a fluvial/restricted marine Purbeck from the Berriasian to Lower Valanginian and a fluvial Wealden of Upper Valanginian to Barremian age. In essence, the Purbecko-Wealden accumulated in an elongated, E-W fault-bounded basin (Figs. 4-1 & 4-2) with clastic input mainly from the south and west.

4. 2 Major subdivisions of the Santander Basin.

The facies belts of the Purbecko-Wealden of the Basco-Cantabrian basin shows a NW-SE trend (Fig. 4-2). Three major regions can be distinguished :-

1) A western region, the Santander basin, in the Peri-Asturian domain (Fig. 1-2) which has an alluvial infill. This region includes the area of NW Santander.

2) A central region, around Bilbao, which has dark mudstones with a brackish fauna. This, and the western region mentioned above, constitute the major Purbecko-Wealden as seen on the isopach maps (Fig. 4-3A) with 2000-3000m of infill.

3) An easterly region in the Province of Guipuzcoa (around Tolosa) in which sandy, marine limestones were deposited (Rat, 1959).

The Santander basin has been subdivided by Pujalte (1977, 1981, 1985, Pujalte & Garcia Mondejar, 1981 and in Garcia Mondejar et al., 1982) into four tectono-stratigraphic sectors, labelled I to IV in Fig. 4-1. The boundary between sectors I and II is the Ubierna fault. The boundary between II and III is the Ebro reservoir/Reinosa diapiric complex (Fig. 4-1), and that between III and IV is the Escudo de Cabuerniga fault. Sequential cross-sections across the Santander basin (Fig. 4-4) show that sectors II and III experienced thicker and more continuous sedimentation in the Purbecko-Wealden. Within the Peri-Asturian domain (Fig. 1-2), the Santander basin (Fig. 4-1) has a generally trapezoidal shape. The northern boundary was at first south of the Escudo de Cabuerniga (Fig. 4-4) but, during the deposition of the Wealden Pas Group (Fig. 4-4), the boundary moved north to approximately the line of the present-day coast. The northern boundary was the up-tilted edge of a half-graben fault block (Chapter 12) whereas the southern boundary of the basin moved between the Ubierna fault (Fig. 4-1) and the present Mesozoic-Tertiary contact. To the east the basin passes gradually into the Bilbao mudstones but the western edge has been eroded due to Tertiary uplift of the Asturian massif (Fig. 4-1).

4. 3 Stratigraphy of the Purbecko-Wealden sequence.

The early stratigraphies of the Purbecko-Wealden are summarised in Fig. 4-5. Pujalte (1974, 1976, 1977, 1981 & 1985) generated a formational lithostratigraphy for the Purbecko-Wealden and applied the term "Cabuerniga Group" to the Purbeck and "Pas Group" to the Wealden (Fig. 4-6). Pujalte (1977, 1981) mapped the entire Santander basin on this stratigraphy as did IGME (a,c,e,g,i,m,o,p). This thesis essentially uses Pujalte's stratigraphy (Fig. 4-6) with the addition of the term La Revilla Fm. (Fig. 1-9) for fluvio-estuarine (?) siliciclastic Aptian sediments that occur in the western part of NW Santander.

The Purbeckian Cabuerniga Group has an overall fining-upward trend and is not present in NW Santander (except for a small outcrop around Socobio, 57939686). Its base rests on eroded Jurassic carbonate (Chapter 3.5) and its oldest formation is the Aguilar Fm. (Figs. 4-4 & 4-6; 300-1000m; Lower Kimmeridgian - Pujalte, 1985b) of micritic lacustrine limestones overlaying red mudstones. The Aguilar Fm. only occurs in sector I and the southern part of sector II. The Saja Fm. (0-100m) only appears in sector III, (Figs. 4-4 & 4-6) and is a high-energy, Donjek-type, braided river deposit (Pujalte, 1979) with palaeoflow towards the east, the area of marine sedimentation (Rios, 1965; Fig. 4-7). In sectors II and III the Aroco Fm. (Figs. 4-4 & 4-6; up to 350m thick) consists of calcareous mudstones, impure limestones and fine sandstones deposited in a fresh to brackish lagoonal environment (Pujalte, 1979). The Loma Somera Fm. (Figs. 4-4, 4-6 & 4-8) is a semi-marine equivalent of the Aroco Fm. with delta plain and tidal flat clastics, subtidal oolitic limestones and calcareous sandstones. Both the Aroco and Loma Somera Fms. are developed in sectors II and III and show increased marine influence towards the NE (Fig. 4-8).

The Wealden Pas Group (Figs. 4-4 & 4-6) is subdivided into a lower Barcena Major Fm. and an upper Vega de Pas Fm. by Pujalte (1985; Fig. 4-6). The Barcena Major Fm. (100-600m, Late Valanginian - Early Hauterivian) can be recognised in sectors II, III and the southern part of IV (Fig. 4-9). It contains stacked fluvial channel sequences consisting of vertical sequences with basal microconglomerates, (pebbly) sandstones and (cross-bedded) quartz sandstones.

Channellised sequences are commonly 2-10m thick with an average of 3.5m (Pujalte, 1985). Rare overbank fines also occur. A model of sandy braided river (South Saskatchewan model of Miall, 1977) that alternated with a sandy meandering river has been proposed by Pujalte (1981).

The Vega de Pas Fm. (100-1200m; Fig. 4-6) overlies the Barcena Major Fm. in sectors II, III and the southern part of IV but in NW Santander it oversteps the Barcena Major Fm. to rest directly on karstified Jurassic limestone as at Cuchia (Section 15 and Fig. 4-4). It has been subdivided by Pujalte (1974, 1977) into three members :

1) The "Red bed member" of infrequent fluvial channel sandstones in red mudstones.
2) The "*Viviparus* Bed member" of black, lacustrine mudstones.
3) An "intermediate" member which shares characteristics of both the above members.

The end of "Wealden" sedimentation was not synchronous with the Barremian/Aptian boundary everywhere, and "Wealden" facies siliciclastics continued to be deposited as the La Revilla Fm. during the Lower Aptian in the SW part of NW Santander (Fig. 1-9). These are best exposed on the N634 road section around the village of La Revilla (54449750). They are lithologically similar to the "Red bed member" mentioned above and are dominantly fluvio-estuarine channellised sediments in a massive red mudstone. Lacustrine limestones with *Unio* and *Viviparus* also exist. Dating on palaeomorphs (B. van Horne, pers. comm.) has yielded a Bedoulian (Lower Aptian) age from this section. The La Revilla Fm. occurs in the western stable block, west of San Vicente de la Barquera (structural Zone 1 of Chapter 1.4.1) and diachronously onlaps onto the Hercynian basement (Fig. 1-9). The age of the marine carbonates overlying the La Revilla Fm. becomes progressively younger towards the west.

4.4 Facies and environments in the Wealden Vega de Pas Fm. in NW Santander.

The Wealden in NW Santander is exceptionally poorly exposed. There are a number of small,isolated Wealden outcrops but the only continuous, unfaulted section of the Vega de Pas Fm. (Pas Group) in NW Santander is at Cuchia (Fig. 4-1; 57319829; Section 15, 33-93m). These facies associations can be distinguished :-

4. 4. 1 Fluvial channel and red mudstone facies associations.

The fluvial facies association is best seen at Cuchia (Section 15, 34.8m to 41.8m - shown in Fig. 4-10). Five facies can be distinguished :-

1) Conglomeratic sandstone facies.

Description : The conglomeratic sandstones are 10-100cm thick and consist of medium to coarse-grained sandstone with 1-50mm diameter, subrounded, subspherical, smoothed granules, generally of quartz. The pebbles tend to form isolated stringers 1-5m long. The sandstones are otherwise rather massive with local trough cross-bedding. These sandstone virtually always occur at the base channel lithosomes.

Interpretation : This facies is interpreted as coarse lag to channels. Such lags could only move during peak floods.

2) Cross-bedded sandstone facies.

Description : Cross-bedded sandstones generally occur in cosets 20-50cm thick and consist of high-angle, planar or trough cross-bedded sets of medium to fine-sandstone. This facies usually occurs above the conglomeratic sandstones within the channel lithosomes.

Interpretation : This facies is interpreted as the product of small sand waves (dunes) that developed on the base and lower part of the channel floor.

3) Parallel bedded sandstone facies.

Description : This is usually developed in medium to fine sandstone (finer sand than in the cross-bedded sandstone facies) with cosets a few decimetres thick. Mud and silt can locally cap this facies.

Interpretation : This is interpreted as a sandstone deposited from a current with flow velocity in the lower part of the upper flow regime.

4) Cross-laminated sandstone facies.

Description : This generally occurs towards the top of channel lithosomes and consists of fine to very fine-grained sandstones exhibiting both non-climbing and climbing ripple cross-lamination. The sets are trough or planar in nature and average 2-10cm in thickness.

Interpretation : This facies is interpreted as a sandstone that formed higher in the fluvial channel where ripples are a common bedform.

5) Interbedded sandstone and mudstone facies.

Description : This facies forms in the uppermost parts of channel lithosomes. It comprises interbedded fine-grained, ripple- and flat-bedded sandstones, sandy micaceous siltstones and red mudstones. Individual beds are up to 30cm thick but this facies is highly variable in thickness (20-200cm or more). The top is usually gradational into the overlying red mudstones and the facies is generally finer and thinner-bedded upwards. The siltstones have local parallel lamination and small-scale cross-lamination, but are mostly totally bioturbated with silty sandstone infilled root burrows.

Interpretation : This facies is thought to represent levee deposits accumulated adjacent to the river.

6) Overbank red mudstones facies.

Description : The mottled, red-brown mudstones are generally remarkably massive above and below sandstone lithosomes. Dull grey, green and purple are the main mottling colours. Diffuse, calcareous sandy nodules a few centimetres wide are quite common at certain horizons. Very rare quartz gravel has been found at a few localities.

The red mudstones of the Cuchia section have yielded non-marine palynomorphs (Vadala, 1981) including gymnosperm pollen of conifers (*Pinuspollenites minimus* - pine trees; *Araucariacites*-monkey puzzle tree and other pollen) and angiosperm spores. These mudstones have also yielded an ostracod fauna including *Darwinula leguminella, Neocythereidea bononiensis* and *Paracyprideis* (IGMEa). The first and latter ostracods are undoubted freshwater forms and have

direct links to Recent non-marine Ostracods (Ramirez del Pozo, 1971a; Brasier, 1980). The Wealden section at Villanuera de la Pena on the N634 road section (55939685) has yielded *Unio* bivalves, *Glauconia*, *Paludina* and *Corbula* (Mengaud, 1920). These are non-marine to slightly brackish (oligohaline) forms. *Cassiope* (an oligohaline gastropod) has been described by Pujalte (1985) from this same section. The author has found numerous Unioid bivalves on the La Revilla section (54489746) along with Viviparus gastropods. Similar fauna has been found at San Vicente de la Barquera (Mengaud, 1920).

Interpretation : These mudstones are interpreted as a suspension deposit, laid down in extensive non-marine overbank wetlands when periods of flooding in the channels allowed turbid, muddy water to invade the interchannel area. The general lack of well-preserved root structures could be linked to intense pedogenesis in semi-permanent lakes in the overbank areas. However, at the very top of the Vega de Pas Fm. on the Cuchia section (Section 15, 89-93 m) spectacular rootlet beds occur. They form subvertical, red (Fe) stained tubicals that increase in density upwards and have a prominent subhorizontal cut-off horizon at their tops. Clear bifurcation downwards and local infilling with carbonaceous woody material is found, confirming their interpretation as rootlets.

These mudstones make up most (about 75%) of the Vega de Pas Fm. in the NW Santander area. However, on some sections (e.g. south of Comillas, 55239755) the sequence is over 95% mudstone. On the measured Cuchia section, 65% is red mudstone.

4. 4. 2 Fluvial channel and red mudstones facies associations- interpretation.

The general vertical arrangement of a fining and thinning-upwards, the predominance of fines in the formation and the non-marine fauna of the overbank deposits are strongly suggestive of deposition from a meandering river sequence (J.R.L.Allen, 1964, 1965, 1970). However, the very poor Wealden exposure prevents a very detailed discussion of the fluvial sedimentology. Jackson (1976) has proposed a lithological classification of meandering rivers, and those of the Vega de Pas

Fm. would fall into his Type 1. On Schumm's (1972) classification of alluvial channels, the Vega de Pas Fm. would fall into the suspended load channel type as sands (taken somewhat arbitrarily as the bed-load component) form only 40% or less of this Wealden sequence.

The banks appear to have been muddy and therefore quite stable. Such stable banks tend to promote channel meandering (Allen, 1964). For the major incised channel at the top of the Cuchia section (Sect. 15, 80-86m with a thickness of 6m and a width of 50m) and assuming member thickness = bankfull depth and measured width = 2/3 bankfull width (Allen, 1965), this channel has a width/depth ratio of 12, which fits with the expected ratio for suspended load channels of about 10 (Schumm, 1968). The study of this single channel may be unrepresentative in that it is an isolated example and could be taken on the bend or straight reach of the channel.

The presence of both pebbles and the medium to fine sands that makes up most of the channel fill points to a bimodality in the grain size spectrum for the channels. This is probably linked to major differences at various times in the effective bed load stress and so suggests periods of severe flooding with very strong currents and longer, tranquil periods when only the small sand waves moved. Again, this suggests rivers with a strongly variable discharge that was probably climatically induced, i.e. the rivers were seasonal. Two of the main factors that induced these rivers to meander were the limited bed load and the cohesive nature of the river banks.

Thus a seasonal climate that was warm temperate to subtropical can be suggested. A similar conclusion has been reached for the Wealden of other western European areas (e.g. Allen, 19775; Allen et al., 1973; Meyer, 1976; Pujalte, 1981; Stewart, 1981; Hallam, 1985).

4. 4. 3 Lacustrine limestone facies association.

Lacustrine limestones occur sporadically within the Wealden sequence (eg. the La Reveille road section (54489746) and the Cuchia section (57319828; Section

15). On the La Revilla section the calcareous, slightly sandy mudstones grade into the red mudstones. A charophyte fauna has been found in them. They probably represent ephemeral lakes with moderate to high clastic input that subsequently suffered extensive pedogenesis and root mottling.

At Cuchia,the lacustrine limestones s.s. are white weathering, approximately metre thick and laterally extensive (over 150m) limestones (Section 15 at 42, 46 and 49m; Fig. 4-11). The lower parts of these limestones mainly consist of micritic wackestones (Fig. 4-12) with scattered reworked ostracod debris, charophyte oogonia, charophyte stems, gastropod shells *(Viviparus)* and oncoliths together with boundaries of cyanohytitic micro-stromatolites and oncoliths (Fig. 4-13). The purer (pel)micrites often have a clotted texture linked to faecal agglomeration and bioturbation.

The ostracod/charophyta (genus *Clevata*, Vadala, 1981) wackestones are generally relatively uniform. Overlying them are oncoidal boundstones with charophyte stems and freshwater cyanophyta that have precipitated calcite around, or within, their filaments, restricting subsequent infilling of voids to internal sediments. The dominant morphological types of cyanophyta are oncoliths (subspherical, locally encrusting nucleuses of broken stromatolite, shells or leaves - Fig. 4-13). Subvertical tubes 1-2cm wide and 5-20cm long are encrusted by cyanophyta and represent encrustations of charophyta stems or other plants. These encrustations are of the micro LLH-C type (Logan et al., 1964 classification).

Rare macro LLH-C lobate domes measuring 15-25cm across have been found in one block of this limestone. Towards the top of the lacustrine limestone beds there are subvertical cm-wide cavities infilled by calcareous internal sediment or sparry, non-ferroan calcite with rare geopetal structures. These are microkarsts and their method of formation has been described by Freytet & Plaziat, 1982. Associated with the microkarsts at the top of these limestones are micritic black pebbles a few centimetres wide, which are also associated with subaerial exposure (Strasser, 1984).

Interpretation : The three lacustrine limestones commonly show a general evolution from basal ostracod/charophytic micrites to oncoidal boundstones and charophyte stems and then a capping with oncoids, black pebbles and pseudo-microkarst features. Such a sequence reflects the progressive drying of the lake (Fig. 4-14). Initially, the shallow, lake was filled with an water-mass in which non-laminated micritic wackestones were deposited. As drying proceeded, the more marginal parts of the lake with mobile oncoids and charophytes encroached towards the centre of the lake. Finally, the complete dessication of the lake produced subaerial exposure and karstification of the lacustrine limestone. The black pebbles are interpreted as karstic pebbles that incorporated finely disseminated carbon probably from decayed cyanophyta (Strasser, 1984). A block diagram for this facies association is shown in Fig. 4-14.

4.4.4 Crevasse splay facies

Description : Within the overbank mudstones there are a number of thin (generally less than 1m), non-channellised fine sandstone/silt bodies. Generally, these are isolated sandstone 5-50cm thick with local parallel lamination and ripple cross-lamination (e.g. Sect. 15, 47.8m). Other sequences 50-200cm thick consist of coarsening- and thickening-upwards mudstones to fine sandstone (well shown in Sect. 15, 71-73m). These pass up from mottled red mudstones into parallel laminated muddy, black sandstones rich in mm- to cm-sized, reworked carbonaceous detritus (e.g. Sect 15, 58.5m). Above are local, small-scale ripple cross-laminations and fine, mm-scale interbeds of mudstone and sandstone that cap these sequences (e.g. Sect.15, 58.6m). The tops of these coarsening-upward cycles can show extensive root mottling (e.g. Sect.15, 63m). One coarsening-upwards crevasse splay sequence is overlain by a crevasse channel (Sect.15, 57-61m).

Interpretation : These coarsening-upward sequences are interpreted as crevasse splays (Fig. 4-17) of bedload material from the main meander channel that could be reworked by later pedogenesis.

4.5 Sandstone petrography and provenance of the Wealden clastics.

The Wealden sandstones in NW Santander are dominantly mature subarkoses (Folk, 1968) with subordinate quartzarenites and rarer feldspathic litharenites. The sand grains are commonly subangular to subrounded with a moderately developed preferred grain orientation. Monocrystalline quartz is the main detrital component and polycrystalline quartz makes up a lower percentage. Together they account for about 70-80% of the sandstone. Feldspars are the next most abundant mineral with a 2-10% content in the rock. Plagioclases dominate over potassium feldspars. Rock fragments are only a minor constituent (about 5%), mainly of schists and rarer angular volcanics. The micas are about 2% of the rock and are dominated by muscovite with rarer biotite and chlorite.

The sandstones show comparatively little diagenetic cementation. The initial, early diagenesis was a fine film of red iron oxide, probably linked to oxidised groundwaters flowing down from the pedogenic surface. There is a rare and local chalcedony cementation. A clay matrix is rare. The heavy mineral assemblage is a mature one dominated by tourmaline and zircon. Rutile, epidote and staurolite make up minor constituents. Such an assemblage suggests prolonged transport and/or polycyclic reworking of the sandstones. This supports an origin for the sandstones from far to the SW of the Basco-Cantabrian basin and/or reworking from the Buntersandstone or Ordovician Barrios Quartzite. The occurrence of staurolite is typical for heavy mineral assemblages derived (not necessarily directly) from the Asturian massif or rocks that were ultimately derived from there (Feuillee, 1967).

The relative maturity of these sandstones again points to prolonged transport and/or polycyclic reworking. The general lack of appreciable chemical solution, suturing of quartz grain contacts or fracturing of the quartz is due to the shallow burial of these sandstones (maximum burial at about 1900m in Upper Eocene times). Only the muscovite grains show appreciable breakage. A more detailed petrography of the Wealden sandstones is given by Pujalte (1977).

4. 6 Summary of the Wealden facies in NW Santander.

The Wealden Pas Group is present in NW Santander as the lower Barcena Major Fm. (in braided stream facies) and the upper Vega de Pas Fm. (in meandering river facies). The Vega de Pas Fm. is dominated by overbank, pedogenic mudstones with channel sandstones. These channel sequences have a general fining- and thinning-upwards trend. Local lacustrine limestones and crevasse splay sands also occur. The sandstones are moderately mature texturally. The tectonic control on Wealden sedimentation will be discussed in Chapter 12.

CHAPTER 5 - THE LOWER APTIAN UMBRERA, PATROCINIO, SAN ESTEBAN AND RODEZAS FORMATIONS: HIGH-ENERGY CARBONATES, CLASTIC DELTAS, PLATFORM LIMESTONES AND MARLS

Relative sea-level rise during the Aptian brought the marine carbonates of the Torrelavega Group over the Wealden peneplain (Fig. 1-9). Tectonically, the Aptian sees the changeover from syn-rift to initiation of syn-drift sedimentation, linked to the onset of sea-floor spreading during the intra-Aptian in the Bay of Biscay (Montadert et al., 1979). The most characteristic deposits of this group are the low-energy, Tethyan, rudistid platform carbonates which are similar to other Lower Cretaceous platform carbonates of southern Europe (Masse & Philip, 1981). These carbonates are generally termed "Urgonian". In this chapter the lower part of the Urgonian Torrelavega Group will be discussed.

5.1 Stratigraphy of the Torrelavega Group.

The general stratigraphy of the Torrelavega Group is shown in Fig. 1-9 Compared to other areas of the Apto-Albian of the Basco-Cantabrian basin, NW Santander has a relatively thin Urgonian sequence (mean thickness of the Torrelavega Group is only 560m) developed in laterally uniform horizons. This allows the relatively simple, layer-cake stratigraphy of Fig. 1-9 to be set up. A correlation of the stratigraphies used by various authors is given in Fig. 5-1. All authors have used a lithostratigraphy and most of the older authors a non-systematic, descriptive nomenclature (e.g. "Formation terrigène supérieure"). The IGME mapping follows a letter based lithostratigraphy (e.g. C_{16}^{23}, cf. Fig. 2-1), though their dating of the sequence has been revised by more recent authors. Garcia Mondejar et al., 1982 was the first author to propose a formalised lithostratigraphy (sensu Hedberg, 1976), and simply applied formation names to the units used by IGME. An adapted form of this scheme was used by the present author but with subdivisions of these formations into members. Two important differences exist between Garcia Mondejar's stratigraphy (Fig. 5-1) and mine (Fig. 1-9). Firstly, in the marly sequence underlying the San Esteban Fm. and overlying the Wealden, there is a high-energy, calcarenite at the base, termed the Umbrera Formation by the author (= O_1 of Mengand, 1920).

This is overlain by the marls and sandstones named by the author as the Patrocinio Formation. Secondly the sequence between the San Esteban Fm. (= Barre Urgonienne inférieur) and the Reocin Fm. (= Barre Urgonienne supérieure) was termed the Cuchia Fm. by Garcia Mondejar (Fig. 5-1). The term "Cuchia" is inappropriate for a number of reasons. In his description of this formation Garcia Mondejar links this formation with "margas con Parahoplites de Cuchia/Marnes de la Playa de Patrocinio" of Mengaund (1920) or the "calizas margosas con Douvilleiceras de Puente Arce, Reocin y Udiás" of Rat (1959). These marls/sandstones (of the Patrocinio Fm. sensu Hines, 1985a and Fig. 1-9, **underlie** (and not overlie) the San Esteban Fm. as the correlation Garcia Mondejar would suggest. His further correlations of the Rodezas Fm. (sensu Hines, 1985a) with the "Level 3-Serie détritique intermédiaire" of Rat (1959) and Ciry et al., (1967) as well as C^{1}_{153} of IGME or A_2 of Garcia Mondejar & Pujalte (1981) are quite correct. Secondly, the well exposed sandstones/marls just NW of the village of Cuchia are of the Patrocinio Fm. (sensu Hines, 1985a) and not of the Rodezas Fm. (sensu Hines, 1985a). Hence, naming a formation after a village which has a good exposure of another formation is not a good technique, especially when the marls at Cuchia (= Patrocinio Fm.) are well known in the literature (e.g. Collignon et al., 1979). The author would like to remove the term "Cuchia Fm." because of its incorrect definition and misleading connotations and use the new term: Rodezas Formation (Hines, 1985a). The author is here proposing a systematic lithostratigraphy for the area (Fig 1-9). The details of the type sections, facies, environments of deposition and tectono-eustatic controls on sedimentation will now be considered:-

5.2 The Umbrera Formation

Definition: The well-bedded, high-energy, calcarenite and calcîrudite pack/grainstones with a diverse marine fauna that occur between the Wealden Vega de Pas Fm. sandstone/red mudstones and the sandstones/marls of the overlying Patrocinio Formation.

Type section: Cuchia section, south of Punta del Cuervo at 57319828 on Section 15, 93.3m to 114.1m.

Lateral variation: This formation is relatively constant throughout the NW Santander area, though it does pinch out against the Hercynian basement to the SW. The initial, relative sea-level rise that created this formation extended only as far west as 600m west of Bustriguado (Fig. 5-7; 54559656). Typical sections through the formation are shown in Fig. 5-2 (for their location see Fig. 5-3). Based on these seven measured sections, the formation has a mean thickness of 20m ± 5m.

Age: All research workers have placed the age around the Barremian/Lower Aptian (= Bedoulian) boundary and the most recent biostratigraphic dating is uppermost Barremian/basal Bedoulian (Collignon et al., 1979), based on orbitolinid benthic foraminifera and regional correlation.

5.2.1 Lithology of the Umbrera Fm. on the Cuchia section

The type section at Cuchia (Section 15, summarised in Fig. 5-10) will first be described and then general comparisons made between this section and others within the area. The base of the Umbrera Fm. at Cuchia (Section 15, 93.3m) is a sharp, planar contact that does not appear to be erosive on an outcrop (50m) scale. The basal two metres consists of 50-cm bedded, parallel-laminated biospararenite pack/grainstones. The dominant allochems are broken, moderately sorted, marine-derived fauna, intraclasts and minor (about 10%) quartz sand. The dominant skeleta are broken rudists and other bivalves (often with extensive encrustation by bryozoa/Bacinella), echinoderms and bryozoa along with minor miliolids, biserial foraminifera and corals. The calcitic matrix is a late-diagenetic (post-compactional), ferroan, equant spar that has suffered extensive, late, ferroan, hypidiotopic/idiotopic, zoned dolomitization (Fig. 5-4).

The next 5 m (Section 15, 95-100 m) of the formation consists of high-energy, low-angle, planar, cross-bedded, orbitolinid grainstones that show a reasonably consistent palaeocurrent flow to the east (mean = 094° ± 30°, n = 8). These biospararenites are similar to those described above, except for the greater abundance of orbitolinids and intraclasts. Rare corals, rhodophyta, dasycladaceae, gastropods, ooids, serpulids, arenaceous foraminifera and biserial foraminifera also occur. The intraclasts are mainly quartz sand-rich micrites or miliolid wacke/packstones. Figure 5-6 illustrates a typical example.

From 100 to 101m on Section 15 there is the first nodular, orbitolinid wackestone with broken mollusca, brachiopods, and echinoderms as well as orbitolinids. The next 9 m (Section 15, 101-110 m) dominantly consists of well to moderately sorted, clean, orbitolinid pack/grainstones with a similar fauna and intraclasts to those described above. Rare quartz gravel with a diameter up to 4mm also occurs. The beds are dominantly 50cm-bedded with marl partings and generally are either parallel-laminated or low-angle, planar cross-bedded. The lower part of Fig. 5-6 shows this lithology. Capping this high-energy calcarenite is a more sandy, marly, bioturbated biointrasparenite/rudite with orbitolinids, bryozoa, bivalves, echinoderms, serpulids, corals and glauconite grains. This is seen as an incipient hardground.

The glauconitic, silty, marl above has a distinctly lenticular geometry, thinning out from over 2m at sea level to nothing only 100m further to the north. The Umbrera Fm. is capped by a highly bioturbated, quartz fine sand/silt-rich, marly, nodular wacke/packstone with an incipient, iron mineral-coated hardground.

5.2.2 Environmental interpretation of the Umbrera Fm. from the Cuchia section

The Umbrera Fm. clearly represents a transgressive, medium- to high-energy carbonate as it overlies the meandering river/upper estuarine(?) clastics of the Wealden Vega de Pas Fm. and is overlain by the deep-water, ammonitic marls of the Patrocinio Fm. (Canales Mb.). The formation also has a general tendency to become finer-grained, lower-energy (wacke/packstone), less well-sorted and more highly bioturbated towards the top of the formation. This is thought to reflect deposition in progressively deeper marine water. The marls at the top of the formation that interdigitate with the Patrocinio Fm. have yielded an *in situ* open-marine fauna which makes the top of the formation marine. The calcarenites below have also yielded a diverse, marine-derived fauna, including stenohaline orbitolinids, echinoderms and dasycladaceae. However, this fauna could have been derived from a marine area and then been resedimented. The high-energy nature of the physical structures and the marine nature of the fauna suggest a very shallow marine (5-10m?), subtidal, offshore calcarenitic shoal. Here winnowed calcarenites were deposited in normal marine, well-oxygenated water on a shifting substrate that was generally inhospitable to sesile and non-adapted

vagile, benthonic organisms. The terrigenous quartz sand is probably derived from reworked Wealden sands immediately below. Such high-energy shoals are widely known from the shallow, inner shelf and are similar to bioclastic submarine calcarenite sheets (or shoals)/oolitic shoals of the Bahamas (Ball, 1967). Hine, (1977) has shown that the dominant internal structures of such shoals are tidal megaripples oriented with tidal flow perpendicular to the long axis of the shoal that are capped by wave-rippled surfaces. Major reworking during storms redeposits the calcarenite/oolite as asymmetric megaripples/planar beds in spillover lobes into the back-shoal lagoon (mostly) and offshore areas (rarer events).

Such sedimentary structures clearly dominate the lower part of this formation, though a direct differentiation of these cross-beds deposited as storm spillovers as against those from tidal currents is difficult. The dominant eastward palaeocurrent direction is interpreted as an ebb flow tidal direction. The eastward flow direction is an ebb direction because the fluvial Wealden clastics were still being deposited west of Bustriguado at this time (Fig. 5-7).

South of the Escudo de Cabuerniga the facies belts have an approximate NNE/SSW orientation (Fig. 5-8) with greater marine influence and palaeocurrents to the east. A diagram showing the location of this (and other) palaeogeographies south of the Escudo de Cabuerniga is shown in Fig. 5-9.

What other environments could this formation represent? At the margin of the sea, beaches should be developed. Single beach sequences generally average 6-10m thick, with the calcarenites 2-5m thick (Shelton, 1973). This formation is much thicker than that and there is no obvious change/break after the basal 2-5m. Gentle, large-scale (metre), seaward-dipping, planar accretion sets of the foreshore and landward-dipping crossets have not been found near the base of this formation, nor has any beach cross-stratification. No indication of subaerial exposure at the top of a possible beach sequence a few metres above the base of the section has been found (in the form of rootlets, evaporites, soils, keystone or bird's-eye vugs etc.). Taking these criteria together, it appears that the basal metres are not beach sediments.

Progradational, regressive, beach sequences commonly coarsen and show less bioturbation upwards (Reinson, 1984). However, during transgression and

shoreface retreat (as here), beach sediments have low preservation potential and are usually eroded and the foreshore/upper shoreface deposits are eroded by wave action, leaving a sharp, locally scoured, surface that can be capped by a lag of coarse beachrock, intraclasts/calcarenitic skeleta (Inden & Moore, 1983; Fig. 5-20). The standard coarsening-upward offshore/shoreface/foreshore sequence is not present in this formation and neither are there any sub- and intertidal back-barrier facies underlying the high-energy calcarenites (characteristic of transgressive barrier sequences- Reinson, 1984) on the Cuchia section.

The sharp contact between the Vega de Pas Fm. and Umbrera Fm. on the Cuchia section is worthy of comment. How did the fully marine Umbrera Fm. directly overlie the non-marine/slight brackish(?) Vega de Pas Fm. with no intervening barrier, stand or beach deposit? As Sanders and Kumar, (1975) have pointed out, during relatively slow and continuous relative sea-level rise and landward migration of the barrier, shoreface retreat can lead to near or complete destruction of the back-barrier sediments by subaerial exposure and wave reworking. This would preserve a basal lag (orbitolinid/intraclast grainstone of this formation?) that was buried by the very shallow though subtidal, nearshore marine calcarenites (Fig. 5-20). The lack of extensive deltas probably links to the low quartz sand input from the mud-dominated meandering rivers combined with the relatively flat, extensive peneplain topography in top Wealden times. There would hence be no "deep" to build a delta into and the thin barrier sands could be reworked by waves. The low quartz sand input also decreases the preservation potential of the barrier system.

In conclusion, the sharp contact between the Wealden and Umbrera Formation on the Cuchia section represents the contact between coastal plain broadly fluvial sediments and shallow marine calcarenites bars. Shoreface retreat has destroyed any barrier sands/lagoonal muds that ever existed between them (Fig. 5-20).

5.2.3 Lateral variation in the Umbrera Formation.

In general, in this formation the facies are laterally relatively uniform in NW Santander, though there is a tendency for it to become more clastic-rich (quartz sand and mud) towards the W/SW (palaeo-hinterland) and thicken to the east (Fig. 5-2). Biospararenite pack/grainstones, similar to

those seen at Cuchia, are developed in the eastern part of NW Santander at Reocin (Fig. 5-2, section 9), Puente Arce (Fig. 5-2, section 12) and Campoo de la Ermita (Fig. 5-2, section 13). The Cuchia and Puente Arce sections both show wedges/beds of the Umbrera Fm. lithology interdigitating with the basal Patrocinio Fm. marls. These are interpreted as storm spillover lobes at the seaward edge of the Umbrera Fm. calcarenitic shoal building out (under storm action) in tongues into the deeper water marls (Fig. 5-20).

The Canales section (Fig. 5-2, Section 6) was deposited further towards the palaeohinterland. There the lowermost 8m of the Umbrera section consists of massive to parallel-laminated, poorly-sorted, calcareous sandstones along with sandy packstones that contain orbitolinids (common), brachiopods, miliolids, dasycladacean algae, ooliths, gastropods, bivalves, muscovite and carbonaceous fragments. These were probably deposited in a back-barrier/lagoonal environment. The upper 7m of this section is a high-energy, ebb-tidal channel with a sharp base and an infill of dominantly trough cross-bedded, clean, pack/grainstone calcarenites with a typical allochem assemblage. The near unidirectional palaeocurrent direction is to the ESE, an ebb direction. The Canales section hence represents a sequence where shoreface erosion has not totally cut though all the back-barrier deposits (as it had at Cuchia). Here, the Wealden grades up into the lagoonal sandy mudstones/mudstones and this is then cut by an erosive ebb-tidal calcarenite channel. The lagoonal deposits do not contain sharp-based, clean, storm washover sands derived from either the beach sands or the offshore calcarenite. It is dominated by crevasse splay (?) muddy, quartz sands derived from the fluvial Wealden system (Fig. 5-20).

The Umbrera Fm. wedges out against the underlying Wealden clastics 600m west of Bustriguado (54559656). The formation at Bustriguado (Fig. 5-2, section 4) interdigitates with the underlying Vega de Pas Fm. at the base. This could represent washover lobes into coastal plain deposits of the upper Wealden. They are overlain by a tidal calcarenitic channel with a typical suite of allochems.

5.2.4 Conclusions on the Umbrera Formation

The Umbrera Fm. is a very shallow marginal marine, subtidal (5-10m water depth?) shoal/bar with dominantly shallow-marine skeleta, especially orbitolinids. Tides and storms reworked the calcarenite and developed ebb-tidal channels, spillover lobes and tongues that interdigitated into deep marine water. The relative sea-level rise of the basal Bedoulian caused the bar to migrate landward (W/SW) with progressive shoreface destruction of any beach/barrier system dividing it from the meandering rivers of the Wealden facies (Fig. 5-20). The resulting high-energy, calcarenitic sheet inundated the low topography of the Wealden plain and produced a regionally correlative, cross-bedded calcarenite. It generally fines upwards and finally interdigitates, and is overlain by, low-energy, open-marine marls of middle/outer shelf (Patrocinio Fm.). The formation thins and finally pinches out in the landward zone (W/SW) of maximum updip transgression. A general palaeogeography for this formation is shown in Fig. 5-7.

5.3 The Patrocinio Formation

Definition: This is a sequence of sandstones and marls with the sandstones in mm to a few dm-scale beds that lie between the calcarenites of the Umbrera Fm. and the 'Urgonian' wacke/packstones of the San Esteban Fm.. The marls have yielded a diverse, open-marine fauna. This formation has been subdivided (this work) into an argillaceous-rich, lower member (Canales Member) and an arenaceous-rich, upper member (Lanchas Member). A generalised log of this formation is shown in Fig. 5-10.

Type section: The type section is from the Cuchia section at 57319828 and is shown graphically on Sections 15 and 15A. The section is faulted towards the top of the formation (at 57299824; Torrelavega IGME map) and tectonically repeated south of this fault. The type section starts north of the fault (Section 15, 114.1m to 197.4m) and the top of this part of the section can be correlated with 15.4m on Section 15A/4 of the strata south of the fault. The uppermost part of the type section is from 15.4m to 19.1m on Section 15A.

Lateral variation: This formation is relatively constant in facies, though the poor inland outcrop prevents detailed analysis Based on the measured sections of Fig. 5-2, the formation has a mean thickness of 95m,

thickening to the east and possibly to the south (though poor outcrop prevents more detailed description).

Age: The Cuchia section has yielded a diverse ammonite fauna (Mengaud, 1913, 1920; Rat, 1959; Collignon et al., 1979). The last author has found condensed, dwarfed, pyritised ammonites (Prodeshayesites, Deshayesites) 10m above the base of section of the Fissicostatus to Forbesi zones (Lower Bedoulian) of Casey, (1981). Calcareous/ferruginous ammonites (Chelonicerus, Vectisites) of the Deshayesi to Bowerbanki zones (Upper Bedoulian) occur in beds just 10 m above. However, according to Van Hinte, 1978 the Bowerbanki zone ought to be considered as the basal zone of the Gargasian. This dating is in remarkable agreement with Mengaud's original dating and is at odds with that of IGME (Fig. 5-1). The formation hence spans the Lower and Upper Bedoulian and could be as young as basal Gargasian depending on the zonal definition of the Bedoulian.

5.3.1 Lithology and facies of the Patrocinio Formation.

As the Cuchia section (Section 15) is the only well exposed section through this formation in NW Santander, I will describe it in some detail before briefly comparing it to other sections. Overlying the Umbrera Fm. are 44m (Sect. 15, 114m to 158m) of the Canales Mb., Patrocinio Fm. The basal parts of it are dominated by massive, dark blue/grey, carbonaceous, silty marls. They contain rare carbonaceous fragments, broken ammonites, radiolaria, mm-sized glauconite grains and late-diagenetic gypsum crystals. The Canales Mb. shows a 1 to 5 m-scale dark blue/dark grey-black banding. These marls are interpreted to be semi-anoxic, pro-coastal, hemipelagic suspension fall-out argillites, deposited on a soft-bottomed sea bed, that have been totally bioturbated to produce their present massive texture. The marls are interbedded with grey, marly wackestones (with rarer packstones) that contain cm-sized iron-coated nodules, ammonite fragments, glauconite grains, echinoderm, brachiopod, serpulid and gastropod fragments, carbonaceous detritus, exogyras, and orbitolinids. These limestones represent condensed horizons in which in-situ reworked shell beds have been reworked reflecting very slow/halted sedimentation and reworking, most probably by storms, of a suspension-feeding fauna (brachiopods, bivalves etc.) not seen in the background marly sedimentation (cf. Fursich, 1982). The high degree of

bioturbation, presence of autochthonous glauconite and biostratigraphic evidence all point to a low rate of deposition.

The upper Lanchas member (Sect. 15, 158 to 197.4m and then Sect. 15A, 15.4 m to 19.1m) is dominated by an alternation of thin-bedded sandstone "events" interbedded with black, carbonaceous, massive marl. There is a general trend towards thicker-bedded sandstones higher in the section and the mean grain size (metre by metre) also generally increases upwards. However, the average grain size of the well-sorted sand is relatively constant at medium to fine (sensu Folk, 1968). A typical microfacies of this sandstone is shown in Fig. 5-12.

The sandstones are generally thin-bedded and have a mean thickness of 13 cm ± 4 cm, (n = 68) of the cm-thick beds, though the presence of many mm-thick beds would considerably reduce this mean. Virtually without exception they are sharp based and, in the vast majority of cases, have sharp, non-erosive tops. Typical examples of these sandstones are seen in the upper part of Fig. 5-12. The bases are occasionally undulating and have local tool marks a few cm long. The thinner sandstones generally develop parallel lamination on a mm scale that can pass up into ripple bedding and local ripple cross-lamination. The ripples are dominantly asymmetric in section (current/wave-current action), though local symmetrical ripples (wave action) are also present. Most ripple cross-laminations are form-concordant. Most of the sandstones are not graded, though local normal grading does occur. It is most probable that the lack of grading relates to the relative textural maturity of these sandstones before they were deposited. Palaeocurrents are very difficult to measure but the few measured give a general eastward (offshore?) flow direction.

These thin-bedded sands could be, and some most probably are, thin-bedded turbidites developing Bouma T_{de} , T_{cde} and T_{bcde} sequences. However, there is the strong possibility that they are tempestites because:-

1) The presence of wave-generated structures is commonly associated with tempestites (Seilacher, 1982) rather than turbidites.

2) Sharp bases are given by both tempestites and turbidites, but sharp tops are more common in tempestites than in Bouma turbidites (Seilacher, 1982).

On the tempestite model, these sandstones would be seen as products of decelerating currents generated from a storm surge. The general paucity of wave structures and the relative thinness of the sands suggest deposition beyond fair-weather wave base in the outer part of that area affected by storm waves (20-100m ? water depth). Similar thin-bedded sands have been found in comparable depths in the Gulf of Mexico (sandstones 2-3cm thick that are graded with a sharp base; Hayes, 1967) and in the North Sea (normally graded, fine, parallel-laminated sandstones that thin and become less frequent further offshore; Reineck et al., 1967).

A number of beds have developed soft-sediment deformation with load casts (e.g. Sect. 15, 184.8m), ball-and-pillow structures (e.g. Sect. 15A, 5.3m in Fig. 5-11), flame structures, soft-sediment slides and slumps. Their formation is linked to the rapid (10-100 secs., Allen, 1984) input of sand on to lighter, water-charged muds. Gravitational instability induced loads and ball-and-pillow structures whereas gravitationally-induced downslope shear generated the slump and slide sheets The prominent sand ball horizon at Section 15A, 4-6m (15m below the top of the formation), is seen at this horizon on both the Canales (Fig. 5-2, Section 6) and Puente Arce (Fig. 5-2, Section 12) sections. These are the only three sections well enough exposed to examine sedimentary structures in the area. Such a correlation points to a major, synchronous event most probably induced by major storm action, a seismic trigger, or major slope failure up-dip.

Along with these sandstones and the carbonaceous marl, the other dominant facies is grey, marly wackestone with a benthic fauna described above. The uppermost 10m of the formation consists of a relatively massive, marly, orbitolinid sandstone with rare bivalves, calcareous sponges, and corals. Pervasive bioturbation makes interpretation difficult, but these sandstones probably represent sandy shoals at the margins of the internal platform (= middle shelf).

5.3.2 Environmental interpretation of and conclusions on the Patrocinio Formation.

The open-marine fauna and general coarsening- and thickening-upwards character of this formation would imply that it is a "pro-delta" s.l.

sequence or an advancing shore-line sequence.

The coarsening- and thickening-upwards trend is seen both in turbiditic pro-delta sequences (Elliot, 1978) and in tempestite sequences (Hobday & Reading, 1972; Goldring and Bridges, 1973). In both cases it is thought to reflect general shoaling. It is probable that the quartz sandstone was derived from small deltas(?)/barriers in the hinterland to the W/SW where fluvial sedimentation (and associated strand-line deposits?) persisted at this time. This formation can also be correlated with extensive areas of clastic sedimentation in the sandy alluvial fans of the coeval Rio Yera Fm. (Fig. 5-8) to the south of the Escudo de Cabuerniga.

The lower Canales Mb. (Fig. 5-1) is a condensed middle/outer shelf, semi-anoxic, marly sequence deposited below storm-wave base, at least in its lower parts. This probably represents deposition in the 50-100m range. The upper Lanchas Mb. is a coarsening-upward, thickening-upward, shoaling sequence of thin-bedded sandstones and marly limestones in marl. It too is open-marine and has tempestite and turbiditic thin-bedded sandstones. It was deposited in a shallower water (30-50m ?) pro-delta/strand-line area.

5.4 The San Esteban Formation.

Definition: A sequence of white-weathering, karstified, metre- to massive bedded, low-energy, orbitolinid/miliolid/peloidal/intraclastal/requienid rudist "Urgonian" limestones (Fig. 5-15 A&B). It is generally a very pure (98-99% $CaCO_3$) biomicritic lime mudstone/wackestone and is sharply overlain by the muddy sands of the Rodezas Fm. A generalised log of this formation is shown in Fig. 5-10. It bears a strong lithological resemblance to the limestones of the Reocin Fm. (Fig. 2-1).

Type Section: The type section is from the Cuchia section at 57299822 to 57219818 on the Punta de la Barra headland. It runs from Section 15A/4, 19.1m, to Section 15A/14, 62.8m. The nomenclature of this formation is illustrated in Fig. 5-1.

Age: The most recent dating on this formation is uppermost Bedoulian/basal Gargasian (Collignon et al., 1979). If, however, the Bowerbanki zone is considered as basal Gargasian, then this formation ought to be seen as Lower

Gargasian. Pascal's revised interpretation of the dating in the Collignon paper (pers. comm. 1985) shows that the formation is more synchronous than originally thought (Fig. 5-13). A revised stratigraphy is shown in Fig. 5-13. Earlier estimates of its age can be seen in Fig. 5-1, all of which are based on benthonic foraminifera, including orbitolinids.

5.4.1 Lithology and facies of the San Esteban Formation

The basal few metres of the Cuchia section (Section 15A, 18-20m) are orbitolinid packstones rich in quartz sand that are "transitional" between the San Esteban and Patrocinio Fms. The next few metres are dominated by rudistid, orbitolinid, coral packstone with rare sheet-like corals up to 1 m thick. These thin corals would not form an effective "barrier" to protect the slightly restricted, internal platform environment of the San Esteban Fm.

The majority of the rest of the section is dominated by a low-energy, poorly-sorted, orbitolinid/miliolid/peloidal/intraclastal/requienid rudist wacke/packstone that lacks terrigeneous clay or quartz. The rudistid fauna is a relatively high-abundance, low-diversity assemblage of thin-shelled Requiena and small-shelled Toucasia (generally 2-8cm long). Most of the shells are poorly-sorted, broken, angular, not in life position, have no preferred current orientation, and occur in a muddy substrate. Locally the shells form small banks (sensu Kauffman & Sohl, 1974) of broken to whole requienids largely reworked out of growth position, most probably by rare storm action (Tucker, 1985). Judging from the poorly-sorted nature of the rudists, they have probably been reworked only a few metres or so from their growth sites. In general, this facies is massive to metre-bedded (Fig. 5-15) reflecting the pervasive bioturbation occurring in an area of low depositional energy. Thin sections (Fig. 5-16) of this facies show the angular, poorly sorted rudistid, oyster and rarer coral fragments in a clotted, non-ferroan, totally bioturbated matrix with benthic foraminifera, peloids, intraclasts and sponge spicules. A number of allochems show well developed micritic envelopes linked to microborings by bacteria/cyanophyceae/fungi/eukaryotic green algae in shallow, warm-water, photic zones of slow carbonate sedimentation rates (Bathurst, 1971; Friendman et al., 1971; Flugel, 1982). Some of the requienids are bored by sponges. A number of

the larger allochems are encrusted by Bacinella irregularis algae(?), though the majority of grains show no degree of binding (either by coral, rudist or algal growth), suggesting that the allochems did not form wave-resistant frameworks (i.e. true reefs). Both the allochems and micritic matrix are of non-ferroan calcite. Miliolids and peloids are relatively abundant, as are high-conic orbitolinids (Simplorbitolina, Mesorbitolina). Rarer skeletal grains include intraclasts, lituolids, gastropods, dasycladaceae, echinoids, bi- and uniserial foraminifera and ostracods. The infill of the sheltered areas within larger allochems are usually finer-grained than the inter-allochem matrix.

This rudistid/benthonic foraminiferal biomicritic facies totally dominates the San Esteban Fm. in general and on the Cuchia section runs from Section 15A/5, 24m, to 15A/11, 47m. These biomicrites are interpreted as internal platform, homoclinal ramp limestones (Fig. 5-7) that were deposited in a low-energy, clear-water (little terrigenous mud) environment, below fair-weather wave base, on a muddy substrate. Similar conclusions have been reached for other rudistid platform limestones (e.g. Masse & Philip, 1981 - Type 2; Pascal, 1985; J. L. Wilson, 1975). The latter author has suggested depths of 10-15m for this facies.

Locally within this facies, there are horizons (e.g. Section 15A/9, 39m) rich in Bacinella oncoliths, black pebbles, stromatolitic debris and bird's-eye structures associated with early diagenetic, isopachous cements (meteoric phreatic?), vadose silts (Dunham, 1969), or rare meniscus cements (vadose cementation). These horizons are commonly associated with decimetre-thick, horizontal stromatolites and were most probably deposited in the intertidal to lower supratidal zones.

The next 3m above the micritic facies on Section 15A (47-50m) represents a rare terrigenous influx into the internal platform with the development of silty or nodular orbitolinid wackestones. The overlying 3m (15A/12, 50-53.5m) consists of moderate-to-high-energy, parallel-laminated and cross-bedded, calcarenitic pack/ grainstones and a small, subtidal shoal associated with local, in situ, biostromal sheet corals and caprotinids. Sheet-like sponges cap this biostrome. The caprotinid/monopleurid biohermal mounds above (Sect. 15A, 53.5-57.5m) contain in situ sponge colonies, sheet corals, caprotinids and monopleurid bouquets. Such assemblages are well known from the Albian Glen Rose,

Southern Texas (Perkins, 1974) where they occur in facies belts in the deeper, outer part of the middle shelf, relative to the requienid/oyster biostromes of the inner part of the middle shelf/inner shelf. Although the caprinid/monopleurid mounds could represent slightly deeper water than the requienid banks, other ecological changes such as better bottom-water circulation around patch reef biostromes or changes in water chemistry could equally well be responsible for the change. The top of the San Esteban Fm. on the Cuchia section consists of intensely stylotitized wackestone, cross-bedded grainstones with allochems of echinoderms, rudistids, bryozoa and intraclasts together with sandy, orbitolinid wackestones at the very top (Sect. 15A, 57.5 - 63m).

5.4.2 Lateral variation in the San Esteban Formation.

This formation outcrops as a distinctive, white-weathering, karstified limestone and can be traced with remarkable consistency across NW Santander and over much of the western Basco-Cantabrian basin (Fig. 5-18; Garcia Mondejar, 1979a, b, c, and Pascal 1982a, 1985). It is virtually always a well-bedded, lime mud/ wackestone (with rarer packstone banks and only local grainstone bars especially at its top). It also has a remarkably constant thickness, relative to the Vega de Pas or Reocin Fms., of 50m ± 10m (n = 12) based on the measured sections of Fig. 5-2. There is, however, the usual general trend of slight thickening towards the east (basinward). Only in the most westerly outcrops (Section 3, Fig. 5-2 and elsewhere) does the facies change to a high-energy, well-sorted grainstone with a diverse marine fauna (Fig. 5-19) and a well-developed, isopachous, early diagenetic, fringe cement (active meteoric phreatic cementation?). Such grainstones were probably deposited in a high-energy beach/subtidal shoal. The formation has only been extensively dolomitized in the very westernmost areas around Bielba (Figs. 5-2 & 5-14). The only other areas of dolomitization are local and fault related. The extensive western dolomitization could well be of the mixed-water type with meteoric invasion at the base of the Rodezas Fm. times. The fault related dolomitization will be considered in Chapter 6.9.

5.4.3 Environmental interpretation of the San Esteban Formation.

The moderate to low-energy wacke/packstone texture and the rudistid/benthic foraminiferal fauna point to deposition on a Tethyan internal platform. The interpretation of this formation as a platform, (rather than a shelf), is based on the major lateral extent of this facies over much of the western Basco-Cantabrian basin (Figs. 5-8 and 5-18; Pascal, 1982, Garcia Mondejar, 1979b), the lateral uniformity of this facies across the area, and the fact that neither at the base nor at the top of this formation on this or other sections in NW Santander is there any extensive development of coralgal reefs or fringing oolitic/calcarenitic shoals that are characteristic of rimmed shelves .

In general, this platform was relatively sheet-like and has only a thin 'pioneer' community of orbitolinid/corals/ quartz sandstones at its base. At the base, the key change from the Patrocinio Fm. to the San Esteban Fm. is a cut-off in the clastic input into the NW Santander basin rather than a reduction in water depth per se. A simple ramp model (sensu Ahr, 1973) has been proposed for this transition by Pascal (1982a, b, c, 1983, 1985) but the problem with this model is that the deeper Patrocinio Fm. contains abundant siliciclastic sands (Lanchas Mb.) and the potentially shallower (more landward on the Pascal model) San Esteban Fm. is a nearly pure (98-99%) carbonate with no such quartz sand. Hence quartz sand could not be supplied from the San Esteban Fm. to a contemporaneous deeper-water Patrocinio Fm. It is hence unlikely that the lithologies of these formations were ever contemporaneous in NW Santander and that the key reason for the changeover was a cut-off in siliciclastic input at the end of Patrocinio Fm. times rather than relative sea-level fall per se. The lower part of the San Esteban Fm. could well have been deposited in comparable depths of water (below fair-weather but above storm-wave base) to the top of the Patrocinio Fm.. Undoubtedly, however, a shoaling-upward sequence is seen within the Patrocinio Fm. and this continues within the San Esteban Fm.. External mechanisms such as an uplift of the Escudo de Cabuerniga ought to be invoked for the changeover to carbonate sedimentation.

Overall the San Esteban Fm. is a regressive internal platform biomicrite with a tendency towards local emergence towards the top of the formation. The high abundance/low diversity of the Toucasia wackestone, combined with the rarity

of stenohaline corals, echinoderms and abundance of miliolids points to a slight degree of restriction within this platform carbonate, simply due to its great lateral extent. It is also probable that in such shallow depths salinities/temperature could vary (high periods of run-off creating temporarily reduced salinity etc.) and that fully open marine conditions were not permanent. The low-energy, biomicritic nature of this platform is not consistent with a model of lateral progradation of calcarenitic bars and suggests that the formation primarily grew by in situ vertical accretion ("upright" model of Playford, 1980; "stationary" model of James & Mountjoy, 1983). The final destruction of this platform was by massive, allocyclic, siliciclastic input at the base of the overlying Rodezas Fm. which destroyed the ecological niche for the Tethyan fauna. This is one of the few convincing mechanisms for destroying carbonate platforms (Schlager, 1981), and avoids the paradox that Tethyan platform's carbonate vertical accumulation rate could easily outpace any relative sea-level rise.

In conclusion, the San Esteban Fm. is seen as a low-energy, rudistid/benthic foraminiferal platform limestone that spread over NW Santander (and beyond) after the clastic input of the Patrocinio Fm. had ceased. It was itself destroyed by further clastic input in the Rodezas Fm. Its lateral uniformity in facies and thickness is linked to the stability of the initial, syn-drift sedimentation in NW Santander. The Aptian tectonics are discussed in more detail in Chapter 12.

5.5 The Rodezas Formation.

Definition: The sandstones, marly limestones and marls that lie between the karstified limestones of the San Esteban Fm. and the base of the Reocin Fm.

Type Section: This is the poorest exposed formation in the entire Mesozoic/Tertiary sequence of NW Santander and there is no really suitable type section exposed within the area. A provisional type section - the Cuchia section, south of the Punta de la Barra headland on the eastern bank of the Rio Saja (57229817 to 57229815) is proposed. This section is shown on Section 15A/14, 62.8m to 15A/20, 93.2m. It is a relatively poorly exposed section, and in the long term a section ought to be defined from cores/Reocin mine through this formation. The Rodezas Fm. has been subdivided by the author into a quartz sandstone-rich lower

member (Arroyo Mb.) and a marly upper member (Calderon Mb.). The formation has a mean thickness of 66m ± 20m based on the sections of Fig. 5-2.

Age: Dating, using ammonite fauna and orbitolinid benthic foraminifera, has been done by Mengaud (1920), Rat (1959), Collignon et al., (1979), and Pascal (1985). The ammonite fauna of the Reocin section (Sect. 9, Fig. 5-2) and Cuchia section (Sect. 11, Fig. 5-2) are of Lower Gargasian age (Fig. 5-14). Such a conclusion is supported by the benthic foraminifera of the San Esteban section (Sect. 7, Fig. 5-2; Pascal, 1985) and the Puente Arce section (Sect. 12, Fig. 5-2; Pascal, 1985).

5.5.1 Lithology and facies of the Rodezas Formation.

In general the lack of exposure of this formation prevents extensive detailed description of it. The lower, sandy Arroyo Mb. is exposed on the Cuchia section and consists of isolated exposures of silty, marly sandstone that is locally bioturbated or parallel-laminated. Rare cross-bedded (Sect. 15A/18, 83 m) sandstones also occur. At Udias, Mengaud (1920) found Glauconia gastropods, possibly indicative of reduced salinity conditions.

In the overlying marly Calderon member, the lower part is dominated by a diffuse alternation of marly wacke/lime mudstones and marls on a 10-50cm scale. The bedding is rather poorly preserved due to intense bioturbation with Thalassinoides burrows. The marly pelmicrites/biopelmicrites have a macrofauna of ammonites, echinoderms, terebratulids, rhynchonellids, Pectins, sponge spicules, rarer coral debris, Myaceas, serpulids, bryozoa and relatively abundant exogyras (Aetostreon latissima). The microfauna include benthic foraminifera (orbitolinids, ostracods, textularids, miliolids) and planktonic foraminifera (radiolaria, Hedbergella).

Above the Arroyo Mb. sandstones, the marly limestones are rich in glauconite pellets, carbonaceous fragments, pyrite and detrital mica. Higher in the Calderon Mb. there are siliceous, nodular (2-10cm nodules), (bio)pelmicritic limestones in 50-100cm beds interbedded with marls. Moderate silicification of the matrix combined with calcite microsparitization has occurred, especially in the sponge spicule-rich pelmicrite beds.

Towards the top of the formation the sequence is dominated by marly pel/biomicrites that are transitional to the overlying Reocin Fm.

5.5.2 Environmental interpretation of the Rodezas Formation.

The lower sandstone-rich Arroyo Mb. was deposited in a shallow marine/deltaic(?) environment that flooded the San Esteban Fm. platform carbonate with siliciclastics. A relative sea-level rise is recorded by the onset of the marly Calderon Mb. (Fig. 4-32), which is thought to represent the outer part of the open marine middle shelf/outer shelf. In general, slightly deeper water conditions existed in the eastern part of the area where a more radiolaria/sponge spicule-rich and thicker sequence was deposited. This member was deposited in a low-energy environment with muddy (clay-rich) marine waters, soft bottoms and very little current activity. The more calcareous biomicrite lenses were colonized by exogyras and brachiopods. The absence of rhodophyta or chlorophyta suggests deposition below the well lit part of the photic zone.

Depths of deposition of the order of 40-80m are suggested. This member represents the deeper part of a homoclinal ramp to the shallower water Reocin Fm. limestones above (see Chapter 6).

5.6 General conclusions on the Lower Aptian strata of NW Santander.

The relative consistency in thickness and facies of the Umbrera, Patrocinio, San Esteban and Rodezas Formations (Fig. 5-2) points to a general cessation of rapid, differential tectonics and the creation of a relatively stable platform. Such a change probably links to the initiation of "syn-drift" sedimentation that gently drapes underlying faults. A similar change is seen in the initial syn-drift sediments of Formation 3 of northern Biscay (Fig. 4-29; Boillot, 1979b). The rapid transgression of the Aptian sea (Fig. 4-32) is, at least in part, driven by the rapid tectonic subsidence of the early drift/distensional phase. The initiation of the transgression as early as Upper Barremian is probably due to rapid tectonic subsidence accelerating a eustatic rise. Further details of the Aptian tectonics are given in Chapter 6.7.

CHAPTER 6 - THE UPPER APTIAN REOCIN FORMATION - PLATFORM CARBONATE DEPOSITION, DIAPIRISM, DOLOMITIZATION AND MINERALISATION

The Upper Aptian of NW Santander is dominated by a thick (mean 330m ± 60m based on the sections of Fig. 6-1) rudistid/benthic foraminiferal, micritic, Urgonian platform carbonate - the Reocin Formation (Fig. 1-9). It corresponds to the 'Barre Urgonienne supérieure' of earlier French authors (Fig. 5-1).

6.1 Definition and lithostratigraphic nomenclature of the Reocin Formation

The Reocin Fm. is defined lithologically as a sequence of white-weathering, karstified and dolomitized, metre to massive bedded, micritic, orbitolinid/peloidal/intraclastal/requienid limestones. It has a somewhat grad ational contact at its base with the marly limestones of the Rodezas Fm. (Calderon Mb) and is sharply overlain by the siliciclastics of the Supra-Urgonian Santander Group (mainly the Las Penosas Fm., but occasionally the Barcenaciones Fm. - see Fig. 5-1). Logs through the formation are shown in Fig. 6-1.

The formation has been sub-divided by the author (Hines, 1985a) into four members which have remarkable lithological consistency over NW Santander. The basal Barrio Member is defined as the micritic limestones overlying the Rodezas Fm. and underlying the central dolomite of the Reocin Fm. (Figs. 1-9 and 5-1). The central part (in general) of the Reocin Fm. is pervasively dolomitized and is here termed the Novales Mb.. The massive, clean, micritic, metre-bedded limestones overlying the Novales Mb. dolomite are termed the Sierra Member. A typical example of the Sierra Mb. lithology is shown in Fig. 6-3. Locally these are overlain by a sequence of marly wacke-packstone/marl alternations of the Comillas Mb..

The Novales Mb. dolomite is a diagenetic replacement of the Reocin Fm. limestone. The extent of dolomitization varies greatly laterally (Figs. 1-9, 5-1 and 6-1) from no dolomitization (e.g. Section 20, Fig. 6-1), through a typical situation of a dolomitized central portion of the Reocin Fm. (e.g. Section 7 or 8, Fig. 6-1), to near complete dolomitization of this formation (e.g. Section 15, Fig. 6-1). There is also interdigitation of the dolomite with the limestones above and below the Novales Mb.

6.2 Type section of the Reocin Formation

The type section for this formation is at Comillas (Section 8 on Fig. 6-1 and Section 5/4, 81.5m to Section 5/28, 493m). It runs down the eastern bank of the A. de Gandarias stream from 55219760 to SW of Miradorio at 55269764. The basal Barrio Mb. runs from Section 5/4, 81.5m to Section 5/7, 160m. The dolomite of the Novales Mb. extends from Section 5/7, 160m to Section 5/13, 319m and the Sierra Mb. from section 5/13, 319m to Section 5/19, 445m. The marly Comillas Mb. completes the Reocin Fm. from Section 5/19, 445m to Section 5/28, 493m.

6.3 The age of the Reocin Formation

This formation has always been broadly considered as Upper Aptian in age by previous research workers (Fig. 5-1 and references therein). This age was originally given by Mengaud (1920) and an age of Upper Aptian/Basal Albian was given by Rat in 1959 and in Ciry et al., (1967). Rat (1959) worked on the sections at Cuchia/Punta del Dichoso (57219811/57109814), Puente Arce, Reocin, San Vicente and San Esteban/Barcenaciones (Section 14 on Fig. 6-1). He gave an Upper Aptian/basal Albian age to the Reocin Fm. The Geological Survey (IGME a,c,e,g,i) produced a date of Gargasian/Clansayesian on the formation, but this was later revised by Collignon et al., 1979. The further use of benthic, orbitolinid foraminifera by Collignon et al., (1979) yielded an age of Upper Gargasian to basal Albian from the Cuchia section (Section 20, Fig. 6-1). Garcia Mondejar (in Garcia Mondegar & Pujalte, 1981; Garcia Mondegar et al., 1982; pers. comm., 1985) dated this formation from the San Esteban/Barcenaciones section (Section 14 on Fig. 6-1) as Upper Gargasian to basal Albian on Orbitolinas. The most recent biostratigraphic work is by Pascal (1985) who dated the formation as Upper Gargasian to basal Albian on the Puente Arce section (Section 22 on Fig. 6-1), the Cuchia section (Section 20 on Fig. 6-1) and the San Esteban section (Section 14 on Fig. 6-1).

In conclusion, this formation is considered to be Upper Gargasian to basal Albian in age. It is not strongly diachronous and any slight degree of diachronaity is beyond the present resolution on benthic foraminifera.

6.4 Lithology and facies of the limestones of the Reocin Formation.

The limestones of the Reocin Fm. show thirteen distinctive facies described in crudely stratigraphic order. The initial nine facies only occur in the Barrio or Sierra Mbs. and the latter four in the Comillas Mb. The dolomites of the Novales Mb. will be discussed in Chapter 6.9.

1) Aetostreon latissima wackestone (Fig. 6-4).

The fauna of this wackestone is dominated by orbitolinids (often aglutinating quartz silt), uni- and biserial-foraminifera, miliolids, textularids, rare sponge spicules, broken, angular bivalve fragments usually with well-developed micritic envelopes (cortoids) along with rare, broken ostracods, serpulids, echinodermata and brachiopods. The micritic matrix is clotted (intense bioturbation) and has been locally recrystallized to microspar (meteoric diagenesis related ?). The macrofauna is dominated by Aetostreon latissima and other ostreidae. This facies usually occurs towards the base of the Barrio Mb. (e.g. at Ruiloca, Section 10 on Fig. 6-1) on the lower part of the gradational, homoclinal ramp transition between the Rodezas and Reocin Formations (Fig. 6-19). It was deposited in a low-energy, slightly turbid-water, open-marine environment below fair weather wave base on a soft, muddy substrate in water depths of the order of 30-60m (cf. Rice, 1984).

2) Dirty, nodular, orbitolinid biomicrite (Fig. 6-11).

The nodular bedded Orbitolina biomicrite is a buff-weathering wackestone with centimetre sized, diagenetic, carbonate "lumps". It is massive to decimetre- bedded (cf. Section 5/19, 435m). This facies has a similar fauna to Aetostreon wackestone (Facies 1), though ostreids are somewhat rarer. In thin-section abundant stylolitization can be seen and the irregular bedding of this wackestone results from solution-compaction along numerous tiny solution seams concentrating clay layers between purer calcium carbonate lumps. This facies commonly occurs towards the base of the Reocin Formation under similar conditions to the Aeostreon wackestone (Fig. 6-4). However, it did receive a small amount (a few % ?) of terrigenous clay.

3) Benthic foraminiferal wackestones (Fig. 6-5).

This is one of the more abundant lithologies throughout the Reocin Fm. (as at Section 5/18, 410m). The dominant microfauna are orbitolinids, miliolids,

uni- and biserial foraminifera and textularids. These fauna are usually strongly micritised and the orbitolinids have agludinated quartz silt into their tests. The macrofauna is dominated by angular echinoderm debris (sometimes with syntaxial overgrowths), molluscan skeleta (mainly broken rudistid debris) along with rare, broken calcareous sponges and their spicules, corals, gastropods, ostracods, dasycladaceae and arenaceous foraminifera. The non-skeletal components include peloids (locally abundant), rare quartz silt and quartz-rich intraclasts. The latter can have local iron (oxy)hydroxide coatings possibly linking them to subaerial exposure. A few allochems have Bacinella encrustations. Many of the allochems show well-developed micrite envelopes linked to micro-borings by bacteria/cyanophyceae/fungi/eukaryotic green algae in shallow, warm-water, photic zones with slow carbonate sedimentation rates.

Texturally, these limestones are dominated by wackestones with rarer lime mudstones. The ubiquitous occurrence of miliolids may well indicate the original presence of prolific quantities of codiacian algae and/or marine grass. The lime mud was probably primarily produced by disintegration of rhodophyte epibionts on marine grass or aragonite algae or both. These wackestones have experienced extensive stylotization. The limestones have suffered locally extensive, non-ferroan sparitization of the micrite matrix. This neomorphic process is probably linked to active meteoric phraetic diagenesis. Late-stage dolomite and ferroan calcite drusy cements and veins are also present.

These benthic foraminiferal wackestones were deposited in low-energy, virtually open-marine areas below effective fair-weather wave base in the internal platform (5(?)-30m water depth).

4) Skeletal packstones (Fig. 6-6).

This facies is essentially a skeletal-rich version of Facies 3 with a similar micro- and macrofauna (e.g. at Section 15A/25, 144m). They too show locally extensive non-ferroan spartization of the micrite matrix (Fig. 6-6) along with zones of vadose silt (sensu Dunham, 1969). This again suggests extensive meteoric diagenesis. The skeletal packstones occur in a similar environment to the Facies 3 wackestones, though are more abundant around rudistid banks, coral patch reef or other biohermal structures.

5) **Requienid pel/biomicrites (Fig. 6-7).**

These are poorly-sorted or rounded orbitolinid/miliolid/peloidal/intraclastal requienid rudistid wacke/packstone that lack any significant terrigenous clay or quartz (e.g. Section 5/14, 3.27.5m). The macrofauna is dominated by requienids (mainly Pseudotoucasia santanderensis and Toucasia) which are generally not in life position, have no preferred current orientation and occur in a muddy substrate. They locally have clionid sponge borings (cf. Petta, 1977) which could be a further major source of micrite. Locally the shells form small banks (sensu Kauffman & Sohl, 1974) of whole to broken requienids reworked out of growth position by rare, storm action (Tucker, 1985). In general, this facies is massive to metre-bedded, reflecting the pervasive bioturbation occurring in an area of low depositional energy. In many other ways this facies is similar to that described in the San Esteban Fm. above (Chapter 5.4.1). It was deposited in low-energy, warm, clear-water conditions below fair-weather wave base on a muddy substrate (Fig. 6-4).

6) **Bacinella biomicrite.**

Bacinella is an important binding algae and has encrusted numerous allochems and locally formed Bacinella oncoliths that are up to a few centimetres wide. On the Punta del Dichoso section at Suances (57099823) there is a 4m thick Bacinella oncolite bed with cm-sized oncoliths and stromatolites. It has associated internal sediments, isopachous early diagenetic and vadose silts, birds eye structures, black pebbles and extensive areas of dissolution linked to meteoric water diagenesis. This facies represents the upper subtidal to lowermost supratidal deposition with periodic emergence. The high-energy nature of this oncolite bed suggests it is a beach deposit. It is clearly associated with the palaeohigh areas such as at San Esteban (Section 14, Fig. 6-1) or at Suances /Cuchia (Fig. 6-19). Figure 6-19 illustrates their location in marginal areas.

7) **Coral patch/table reefs.**

These rather massive, scleractinian coral reefs are relatively rare in the Reocin Fm.. These framestones are true ecological, wave-resistant bioherms and have occasional requienids or pectins within them in life position. They fall into two categories. There are the patch reefs that are a few metres high and a few tens of metres wide. They are often diamond shaped in cross-section reflecting the initial colonization, growth and final abandonment of the reef.

They generally have little or no internal sediment and the scleractinian framework has been virtually totally recrystallised. There are also table reefs which are again only a few metres high but extend for several hundreds (100-800m) of metres laterally.

The reefs tend to occur over palaeohighs such as the Comillas palaeohigh (55199761) or the Torrelavega palaeohigh (bioherms at Reocin at 56739708). The open framework of the latter has been cemented by coarse, sparry dolomite and Pb/ZnS mineralization (Chapter 6.9). These reefs formed wave-resistant, rigid, subtidal bioherms in the shallow, subtropical, limpid waters of the Reocin Fm. carbonate platform. The question of why scleractinian corals were not more abundant in this environment is slightly puzzling. It could be related to a slight degree of restriction (temperature fluctuation, high run-off-related salinity reduction ?) in the carbonate platform linked to its massive (100km) lateral extent (Figs. 6-20 & 6-21). Ecological competition from the requienids is another possibility.

8) **Sponge bioherms** (Fig. 6-8).

The calcareous sponges tend to form small bioherms up to a few tens of centimetres high and about a metre wide. Most calcareous sponges form isolated buns a few centimetres high (e.g. Section 5/4, 330m). The internal chambers are often filled by a micrite with local geopetal indicators. The sponges formed small frameworks around which skeletal packstones developed in the low to (Fig. 6-19) moderate energy, subtidal part of the internal platform. The clionid sponges are often associated with abundant peloids, and could well have been one of the key producers of such peloids. Similar developments have been seen in other Urgonian platforms (Petta, 1977) and in Holocene Jamaican reefs (Spurr, 1975).

9) **Skeletal grainstones** (Fig. 6-9).

The biosparite grainstones are generally either parallel laminated or cross-bedded (e.g. Section 15A/30, 214m). In general, they consist of moderately washed and sorted biosparites with rarer biopelsparites. The dominant fauna in the biosparites is marine and consists of orbitolinids, peloids, quartz sand-rich biomicritic intraclasts and bivalve (mainly rudistid) fragments. The latter have well-developed micritic envelopes reflecting microborings under low-energy conditions before reworking into high-energy grainstones. Minor skeletal

components include reworked corals, miliolids, echinoderms, dasycladaceae, calcareous sponges, gastropods, ostracods, serpulids and bryozoa. Rare oolitic coatings occur around some allochems. Many of the grainstones have not had all their original micrite removed and show umbrella or geopetal structures. Two main generations of spary calcite cement can be seen in these grainstones. The initial, pre-compaction thin fringing cement (Fig. 6-9) develops in a fibrous or dog-tooth habit with rare meniscus cements. These represent very-early diagenetic meteoric cementation in vadose and active meteoric phraetic conditions. Locally this is followed by a vadose silt infill. The final, later cementation is by coarse, generally non-ferroan, drusy calcite spar associated with syntaxial overgrowths on echinoderms. This spar probably formed under phraetic conditions and is most probably meteoric in origin. After these two cements, ferroan calcite and sparry (locally baroque) dolomite infill the final pore space and also occur in cross-cutting veins and faults (see Chapter 6-9).

These grainstones were deposited in shallow-water, high-energy shoals (Figs. 6-4 & 6-19) under upper flow regime velocities by wind/tidal(?) currents. It is possible that some shoals could have been shallow enough to form beach deposits, though direct evidence for this is missing in most of these calcarenites. However, at the top of the Reocin Fm. on the Cuchia section (Section 15A/31, 226 to 229m) there is a large-scale, 3.5m thick, cross-bedded, well-sorted grainstone with a sharp erosion disconformity surface above it (a major stratigraphic break between the Reocin and Barcenaciones Fms.). This represents a prograding foreshore/shoreface deposit.

10) <u>Bioturbated marl</u> (Fig. 6.10).

The bioturbated marl facies is totally confined to the Comillas Mb. (e.g. Section 5/24, 473.2m). This facies has a moderate terrigenous clay input in contrast to (Figs 6-14 & 6-15) the clean limestones of the Sierra and Barrio Mbs. which are generally very pure carbonates (98-99% $CaCO_3$ with detrital clay making up the residual component; Rat, 1959). In general, these are moderate to intensely bioturbated dirty lime mud/wackestone with no large skeletal fragments. They are usually homogeneous due to the intense bioturbation. <u>Thalassinoides</u> and <u>Rhizocoralium</u> of the <u>Cruziana</u> ichnofacies have been found in these shallow marine marls. They represent a final phase of the Reocin Fm. in which distal siliciclastic clay input invaded the carbonate platform.

Very rare exceptions do exist to the general rule that this facies does not occur within the Barrio or Sierrra Mbs. A horizon of glauconitic, carbonaceous bioturbated marl occurs in the Cuchia quarry section (at 5729811). It has yielded a diverse marine fauna of pectins, echinoderms, terebratulids, rhynchonellids, serpulids and gastropods. In the Cuchia quarry its stratigraphic position in the middle of the Reocin Fm. can be clearly seen.

11) Sandy, bioturbated, marly wackestone.

This facies is interbedded with the marls of Facies 10 (Figs. 6-10 & 6-11) and only occurs in the Comillas Mb. (e.g. Section 5/20, 450.2m). It is rich in well-sorted, subangular, fine quartz sand and has a diverse marine fauna. This facies was deposited in the innermost, subtidal platform where periodic (storm-related?) input of quartz sand occurred.

12) Sandy, skeletal pack/grainstone (Fig. 6-12).

This is a cross- or parallel-bedded calcarenite in the Comillas Mb. (e.g. Section 5/21, 455m). Characteristically these pack/grainstones are rich in well-sorted, subangular quartz sand. The dominant allochems are pelloids and cortoid bivalve fragements (mainly rudistid shells) that show well-developed micrite envelopes and drusy orthospar replacements of the original aragonite/high-Mg calcite. The microfacies also has echinoderms (with syntaxial overgrowths), biserial foraminifera, ooids, arenaceous foraminifera, glauconite, corals, and broken brachiopods. The rocks are generally cemented with a late-stage, ferroan, drusy orthospar and an even later-stage ferroan, sucrosic dolomite. These grainstones were deposited by upper flow regime wave/tidal (?) currents in the innermost carbonate platform that received moderate clastic input.

13) Requienid/nerineid/sponge/coral bioherms (Fig. 6-13).

These bioherms (e.g. at Section 5/22, 463m.) are 0.5 to 2m thick and up to several tens of metres wide. They consist of in situ solitary, branching and domal scleractinian corals, in situ requienids (mainly Pseudotoucasia, Horiopleura and Polyconites), trigonias together with minor terebratulids, rhynchonellids, echinoderms, orbitolinids, bryozoa and serpulids in a marly wacke/packstone matrix. These bioherms are small, constructional buildups in the innermost carbonate platform. The diverse fauna suggest relatively normal marine salinites, in an area of moderate clastic input. All of these bioherms are

overlain by marl or micrite horizons suggesting that the period of colonization was destroyed by major argillaceous input (as at Section 5/22m 463.3m) that killed off the skeleta.

6.5 Lateral variation of the Reocin Formation.

The Reocin Fm. shows pronounced lateral thickness changes (Figs. 6-1 and 6-20) which will be considered from west to east across NW Santander. To the west of Bustriguado (Section 5 on Figs. 6-1 & 6-2), the formation is thin (100-200m) and was deposited on the 'Western Stable Block' of Figs. 6-1, 6-2, 6-16, 6-17 & 6-20. Here the formation is rich in requienid biomicrite (Facies 5), nodular biomicrite (Facies 2) and benthic foraminiferal wackestones (Facies 3). There is progressive overstep of the Reocin Fm. to the west. At Bustriguado (54669665, Section 5 on Fig. 6-1), the Reocin Fm. rests on the Rodezas Fm. as it does up till 6km further west at the La Florida mine (54079663) and at Bielba (Section 4, Fig. 6-1). Further west on the Rio Deva section (Section 3, Fig. 6-1) it rests on Wealden siliciclastics whereas at Santillan (Section 2, Fig. 6-1) it rests directly on karstified Carboniferous Caliza de Montana.

In the area to the east of the meridian through San Vicente de la Barquera (structural zone 3 of Chapter 1.5.3 and Fig. 1-7), the Reocin Fm. is much thicker (generally 300-500m; mean 330m; Fig. 6-1). This increase in thickness is related to continued gentle movement on the San Vicente/Bustriguado transfer fault in the Upper Aptian.

Over most of NW Santander (east of the San Vicente/Bustriguado fault), the Reocin Fm. has a remarkably systematic lateral thickness variation. The Reocin Fm. thins over the Keuper diapirs (Fig. 1-7) which formed palaeohighs (Figs. 6-1, 6-16 & 6-17). The formation then thickens into the associated rim synclines/palaeolows (Fig. 1-11). The location of diapirs is shown on Figs. 1-7 & 6-18. Although the lithologies can be somewhat similar between the palaeohighs and lows, there are some higher energy deposits over the palaeohighs. These will now be described.

There is a slight thinning of the Reocin Fm. over the Comillas palaeohigh (at 55109750, ; Figs. 1-7 , 6-1:- Sections 6 to 9, Figs. 6-16, 6-17, 6-18). The Comillas sequence has some carbonaceous detritus especially towards its base, probably reflecting its proximity to land. Further

east the Reocin Fm. thicknes into the Novales palaeolow (Figs. 6-1:- Sections 10 to 13, 6-16, 6-17, 1-7). Here the Barrio and Sierra Mbs. are represented by low-energy Aetostreon latissima wackestones (Facies 1), nodular biomicrites (Facies 2), benthic foraminiferal wackestones (Facies 3) and requienid pel/biomicrites (Facies 5) with rarer skeletal packestones (Facies 4). The very top of the Reocin Fm. limestone is very slightly sandy and has only rare skeletal pack/grainstones (Facies 4 & 9). So, in general terms, the thicker Novales palaeolow preserves a lower-energy, more micritic sequence than does the Comillas palaeohigh.

To the SE of the Novales palaeolow, the Reocin Fm. rapidly thins towards the Caser palaeohigh (Figs. 6-1:- Sections 13 & 14, Figs. 6-16 & 6-17). This thinning is related to syndepositional, normal faulting around La Busta (56039720; Fig. 6-20).

The San Esteban section (Fig. 6-1, Section 14) is dominantly micritic (bio- or intrabio-) with local skeletal grainstone bars (Facies 9). Horizons with birds-eye structures (locally infilled by geopetal vadose silt), black pebbles (intertidal/lower supratidal related - Strasser, 1984), locally abundant Bacinella oncoids (Facies 6), rare stalactitic and meniscus cements together with carbonaceous detritus horizons have all been found here. These features indicate uppermost subtidal to lower supratidal deposition for these horizons. The presence of such peritidal features again points to the fact that the thinned palaeohighs (such as the Caser palaeohigh) were areas of shallower water. The Reocin Fm. remains at about 210m. thick further east towards Reocin (Section 15:- Fig. 6-1, Figs. 1-7, 6-16, 6-17 & 6-18).

To the NE of the Novales palaeolow, the Reocin Fm. again thins towards the Orena palaeohigh which is cored by Keuper evaporite mudstones (Figs. 2-4, 6-1:- Sections 16 & 17, Figs. 6-16, 6-17 & 6-18). The Reocin Fm. then thickens southeastwards towards the Santillana palaeolow (Santillana syncline of Figs. 2-3 & 2-4, Fig. 6-1:- Sections 18 & 19; Figs. 6-16, 6-17 & 6-18).

Further east still, the Reocin Fm. rapidly thins towards the Miengo palaeohigh (Figs. 1-6 & 1-7; Fig. 6-1:- Section 20, Figs. 6-16, 6-17 and 6-18). The Cuchia section (Section 15A, 93-229m) is strongly thinned (136m) and the formation is in distinctly higher-energy facies than anywhere else in NW

Santander. A number of skeletal/peloidal grainstones (Facies 9) occur towards the base of the Cuchia section that were deposited in high-energy, subtidal shoals (e.g. Section 15A/21, 104m). In the central part of this section are lower-energy benthic foraminiferal wackestones (Facies 3) or requienid biomicrites (Facies 5 - e.g. Section 15A/24, 136m). The uppermost 25m of the section (205-229m) is a continuous skeletal grainstone (Facies 9) capped by a beach sequence.

To the SE of the Miengo palaeohigh, the Reocin Fm. again thickens towards the Camargo palaeolow (Fig. 6-1:- Sections 21 to 15; Figs. 6-16 & 6-17). Around Puente Arce (58029783; Section 22 on Fig. 6-1) the basal skeletal grainstones (Facies 9) persist but pass out further east towards the centre of the Camargo palaeolow. The top of the Reocin Fm. develops a number of subtidal/intertidal bar/beach sequences and burrowed horizons indicating very shallow water sedimenation/subaerial exposure. Further towards the centre of the Camargo palaeolow the Reocin Fm. thickens (Sections 21 to 24 on Fig. 6-21) and becomes more micritic (Facies 2, 3 and 5).

In the very easternmost part of NW Santander, the Reocin Fm. again thins towards the Santander palaeohigh (Santander diapir of Fig. 1-7; Fig. 6-1:- Section 26; Figs 6-16 and 6-17). On the Pedreña section the sequence is strongly thinned, grainstone-rich with slight clastic input.

6.6 Environmental interpretation for the Reocin Formation.

The dominance of low- to moderate energy rudistid/benthic foraminiferal wacke/packstones points to deposition on a Tethyan, internal carbonate platform (Fig. 6-19). In many ways, it was deposited in a similar environment to the underlying San Esteban Fm. (Chapter 5.4.3). The lateral uniformity, low-energy of deposition and lack of any extensive development of coralgal/scleractinian reefs or frining oolitic/calcarenitic shoals all suggest that a rimmed shelf model is inappropriate for this formation (Fig. 6-19).

The base of Reocin Fm. shows a true ramp transition (sensu Ahr, 1973) from the deeper water ammonitic/ostreid marls of the Rodezas Fm., Calderon Mb. (Fig. 6-19). The top of the Rodezas Fm. was deposited on a very gentle slope that was not steep enough to initiate any sliding, slumping or carbonate mass flow deposits. The marly limestones of the top of the Rodezas Fm. pass laterally

and vertically to the basal low-energy biomicrites of the Reocin Fm. (Fig. 6-19). Generally there is a uniform, gradational transition from the marly limestones to the biomicrites of the Reocin Fm., reflecting a gentle, relative sea level fall on a homoclinal ramp. Locally, however, there are small pack/grainstone calcarenite shoals generally 0.5 to 5m thick (e.g. basal parts of the Puente Arce section, Section 22 on Fig. 6-1).

The central part of the Reocin Fm. is dominated by low- to moderate energy benthic foraminiferal wackestones (Facies 3), skeletal packstones (Facies 4) and requienid biomicrites (Facies 5). Towards the top of formation calcarenite pack/grainstone bars are developed locally. These are generally subtidal, but with some intertidal beach deposits (e.g. at Cuchia, Section 15A/31, 228m).

The uppermost part of Reocin Fm., the Comillas Mb. is only developed in the western part of NW Santander, west of the meridian through Santillana (Fig. 2-5, 56599763). This member is dominated by an alternation of bioturbated marls (Facies 10) and sandy, marly wackestones (Facies 11; Figs. 6-10 and 6-14). These were deposited in the innermost platform where "pro-delta s.l." siliciclastic clays (Facies 10) and fine quartz sand (Facies 11) could invade the area of carbonate sedimentation. Local requienid/coal bioherms (Facies 13) developed, as did high-energy, skeletal grainstones (Facies 2).

Within this overall vertical picture, there are strong lateral variations in the Reocin Fm. (Chapter 6.5). During the deposition of this formation numerous Keuper diapirs started to develop into their pillow stage (Chapter 1.5 , Fig. 1-7). This created diapiric palaeohighs (Fig. 6-20) with thinned sequences. Over these palaeohighs the formation was deposited in shallower water (a few metres above sea level to -20m of water ?) than in the rim synclines (-20 to -40m ? water depth). The deposition in shallower water is reflected in the development over the palaeohighs of higher-energy grainstones and shoals (e.g. Cuchia), the development of patch/table reefs (e.g. Comillas, Reocin) and more extensive development of intertidal stromatolites, Bacinella horizons or black pebbles (Fig. 6-19). These palaeohighs also show more significant meteoric diagenesis both in the limestones (vadose cements, vadose silts etc.) and in the development of an early diagenetic, mixed-water dolomite related to meteoric invasion over the palaeohighs (see Chapter 6.9). The

associated rim synclines have lower-energy and more micritic (Fig. 6-19) limestones. The distribution of these palaeohighs is shown in the palaeogeography for the formation (Fig. 6-20).

The mechanisms of the diapirism have already been discussed in Chapter 1.5 A relatively direct modern analogue exists between the modern Persian Gulf, which is an epeiric carbonate ramp (Purser, 1973), and the Reocin Fm. In the Gulf local diapiric highs have created seafloor topography and concentrated halos of high-energy, cross-bedded pack/grainstones and coral reefs over them. These calcarenites are rapidly lithified and are strongly reworked into 'horns' about the highs in the dominant wind direction (c.f. Fig. 6-19). Off the highs, in deeper water, the deposition becomes more micritic (wacke/packstones), lower-energy and poorer-sorted. Such a model is strongly paralleled by the Reocin Fm.. Urgonian diapiric palaeohighs with similar facies distributions around them have been recorded from the Subalpine region (Arnaud-Vanneau and Arnaud, 1976) and in the Paris basin (Masse and Philip, 1981).

The low-energy, biomicritic nature of this platform is not consistent with a model of major lateral progradation of calcarenitic bars over deeper-water facies. This formation (as well as the San Esteban Fm. below) primarily grew by *in situ* vertical accretion ("upright" model of Playford, 1980; "stationary" model of James & Mountjoy, 1983). There was a slight degree of salinity/temperature restriction in this platform linked to its great lateral extent (see discussion of this for the San Esteban Fm. - Chapter 5.4.3).

The top of the Reocin Fm. is marked by a very sharp, planar contact (on an outcrop scale) with the overlying siliciclastics of the Santander Group. This contact marks a minor biostratigraphic but major sedimentological hiatus with extensive subaerial exposure, meteoric water diagenesis and mixed-water dolomitization (discussed more in Chapter 6.9). The cessation of carbonate platform deposition in the Reocin Fm. is hence simply due to its subaerial exposure at end Reocin Fm. times because of relative sea-level fall and the associated influx of clastics in the Las Penosas Fm.. Similar relative sea-level falls are seen in the basal Albian in the southwestern part of the Basco-Cantabrian basin

(Garcia Mondejar, 1979a,c - the Fm. de Lunada of Garcia Mondejar et al., 1982 and Fig. 6-23B) but not in the central and eastern parts of the basin (Pascal, 1985). This could reflect the *very* rapid tectonic subsidence in the central and eastern areas producing a relative sea-level rise (Pascal, 1985) whereas the western part of the basin experienced a net sea-level fall probably related to eustatic fall (cf. Cooper's curve in Chapter 12).

6.7 Comparison of the Reocin Formation with coeval sequences in the Basco-Cantabrian basin.

The Reocin Fm. platform has some notable differences (and similarities) with other Upper Aptian platforms in the Basco-Cantabrian basin (Figs. 6-20 & 6-21). Firstly, the Aptian Torrelavega Group of NW Santander is much thinner than elsewhere in the basin, with a mean thickness of only 560m based on the measured sections of Figs. 5-2 & 6-1 (cf. thicknesses in Fig. 4-3B). The major increase in thickness occurs east of the "Rio Miera flexure", an active transfer fault (Ciry et al., 1967; Pascal, 1985; Figs. 4-2 & 6-21). This was a major palaeohinge zone and bounds NW Santander to the east. The Valnera-Ramales area to the east of the flexure preserves an Apto-Albian sequence that is not only thicker, but also has major siliciclastic deltas. The Rio Miera flexure could mark the boundary between the thin, relatively simple sequence in NW Santander deposited on normal continental crust (see Chapter 12) and a thicker sequence to the east of the flexure deposited on structurally-thinned continental crust (Pascal, 1985).

Secondly, the Reocin Fm. is remarkably free of siliciclastics relative to deposition elsewhere in the Basco-Cantabrian basin. An example of this is the lack of the clastic-rich U^{Ap}_{s2} (lower part) cycle (which is only seen south of the Escudo de Cabuerniga (Fig. 5-8) - Garcia Mondejar, 1979c) that thins to the thin Capa roja bed (Chapter 7) to the north in NW Santander. Figures 6-21 & 6-22 show major arenaceous deltas forming at this time south of the Escudo and west of the Rio Miera flexure. The lateral change of the Reocin Fm. into coeval marls (shown

on Fig. 6-19) occurs just to the east of Santander city (Fig. 6-21; Magniez & Rat, 1972) and to the NE of the present coast line.

Thirdly, the Upper Aptian of NW Santander has a relatively uniform, simple, layer-cake stratigraphy relative to other parts of the basin (cf. Garcia Mondejar, 1979a; Pascal, 1985). This is primarily due to its relative tectonic stability during the thermal subsidence phase of basin development (see Chapter 12).

Fourthly, the Reocin Fm. (Sierra and Barrio Mbs.) is a pure carbonate (98-99% $CaCO_3$ - Rat, 1959) suggesting that the coeval clastics to the S/SW (Figs. 6-21, 6-22 & 5-8) either could not bridge the platform carbonates to the south of Escudo de Cabuerniga or were ponded to the south of the Escudo which acted as a tectonic barrier/palaeohigh (Fig. 6-24). The latter explanation has been proposed by Garcia Mondejar (1979a, b, c, 1985) and Figs. 6-23 and 5-8 illustrate the ponding of clastics south of the Escudo de Cabuerniga during the Apto-Albian.

In conclusion, NW Santander can be seen to have had relatively simple extensional, diapiric tectonism during the Apto-Albian. The Reocin Fm. of NW Santander is isolated from arenaceous-rich input from the south by the Escudo de Cabuerniga and passes into deeper-water marls to the east and north of NW Santander. Figure 6-25 illustrates this relationship and contrasts it to other Urgonian depositional systems of the basin. The NW Santander area would fall into the extensive platform carbonate of Fig. 6-25A.

6.8 Limestone diagenesis of the Reocin Formation

After the complex depositional history and associated synsedimentary tectonics considered above, the Reocin Fm. suffered extensive diagenesis. The limestone diagenesis will be considered first, then the dolomitization and finally the epigenetic Pb/ZnS mineralization. The general sequence of diagenetic events is summarized in Fig. 6-28 with constraints of timing/depth of burial of these events given where possible.

6.8.1 Limestone diagenesis in the Reocin Formation

The main features of the limestone diagenesis will be considered sequentially (Fig. 6-28).

1) **Step 1**. The initial diagenesis involved micritization of the limestone allochems under active marine phreatic conditions (sensu Longman, 1980). Extensive micritic envelopes were produced on requienid and other bivalves. This micritization is linked to microborings by bacteria, fungi or eukaryotic chlorophyta in shallow, warm-water, photic zones of slow carbonate sedimentation rates (Bathurst, 1971; Friedman et al., 1971; Flugel, 1982). Further megaborings by sponges exist mainly on requienids. These micritic coatings are illustrated in Fig. 6-7.

2) **Step 2**. The biomicrites and grainstones both have rather quartz-sand-rich intraclasts from the Reocin Fm. These intraclasts do not show any soft-sediment deformation structures and must have been lithified to be reworked into grainstone shoals (Facies 9). Such syndepositional cementation must have occurred under active, marine phreatic conditions. Similar active marine phreatic acicular and micritic aragonite cements have been described by Schroeder, 1972. These intraclasts can locally be encrusted by *Bacinella* or bryozoa as well as having micritic envelopes. This directly confirms that cementation occurred under marine conditions. This stage is also accompanied by internal sediment infill of intra-skeletal porosity by wacke/packstone (cf. Kendall, 1971).

3) **Step 3**. Very early diagenetic cementation then occurred. This takes the form of intergranular, non-ferroan, thin, fringing cement which is usually about 50-200μm wide (Fig. 6-29). It is generally rather isopachous and it rarely develops any meniscus or dripstone features. It is moderately fibrous, grows out perpendicularly from allochems and is moderately serrated. It is probably a very early diagenetic, active, marine phreatic fringing cement. The original aragonite/high-Mg calcite cement has inverted to a low-Mg calcite retaining its isometric, subhedral, crystal shape and grain orientation.

4) Step 4. Some of the micritic envelopes show compactional fracturing that has occurred after the initial fringing cementation but before significant equant spar cementation. This probably occurred in the top few tens of metres of rock or less.

5) Step 5. Preferential leaching of the aragonitic and high-Mg calcite (HMC) parts of allochems produces mouldic porosity. There is possibly some neomorphism of high Mg-calcite to low Mg-calcite. This effect is most obvious in the internal parts of requienid shells which have been leached after developing micritic envelopes but before being infilled by equant, drusy spar. This would occur under meteoric phreatic conditions in which aragonite and high-Mg calcite would be unstable.

6) Step 6. Extensive precipitation of limpid, non-ferroan, generally equant spar then occurs which often shows a coarsening of crystal size towards the centres of inter- and intragranular pore spaces. The low-Fe, low-Mn concentrations of this equant spar generation were analysed by electron microprobe (Fig. 6-61A) and show that the Fe contents are generally under 500ppm. This cement is also associated with syntaxial overgrowths on echinodermata. It is an active meteoric cement, probably formed under phreatic conditions. The low ferrous iron content of meteoric cements has been linked to oxidizing vadose conditions by Oldershaw & Scoffin, 1967. However, this cement does not develop gravitational, meniscus or dog-tooth forms. This cementation drastically reduced the porosity at a very early stage in the diagenesis of the grainstones (Facies 9) of the Reocin Fm. Some microfacies show this stage to be associated with vadose silts, and for these a definite active, meteoric, vadose environment is envisaged (Fig. 6-30).

Geochemical support can be given to the concept of strong meteoric diagenesis in the Reocin Fm. limestones. Whole-rock trace element data have been run on Urgonian limestones of the Basco-Cantabrian basin by Pascal (1979, 1985; Fig. 6-31) using standard atomic absorption techniques. Relative to the mean values for the basin of Fig. 6-31, the values for the Reocin Fm. from the Punta del Dichoso section at Suances (57099823) and that at Cuchia (57229810) show some differences (Pascal, 1985; Fig. 6-32). The limestones of the Reocin and San

Esteban Fms. are relatively low in Sr (around 200ppm) compared to the rest of the basin (Fig. 6-31). Na is relatively standard for this facies in the basin as a whole (mean 150ppm), as it also is in Mn (mean 88ppm). However, these limestones are Zn rich (mean 60ppm) relative to the basin-wide average of 10ppm.

The interpretation of these results depends primarily on an understanding of the differences in trace element concentrations in and proportions of the aragonite, high- and low-Mg calcite that originally formed the limestone, as well as the salinity of diagenetic waters and the rock/water ratio (degree of openness of the system). This makes direct interpretation from bulk sample geochemistry of the limestones rather doubtful (Kinsman, 1969b). However, the relatively low Sr values of these limestones would fall into the class proposed by Kinsman (1969b) of those limestones derived from low-Sr, low-Mg calcite and/or affected by extensive, low-Sr meteoric water diagenesis. A similar interpretation can be placed on the Na results with values less than 1000ppm interpreted by Land & Hoops, 1973 as indicating some degree of meteoric diagenesis. The Reocin Fm. has moderate Mn values (Figs. 6-31, 6-32) whereas the Patrocinio Fm. and lacustrine limestones of the Wealden Vega de Pas Fm. (Cuchia section, Fig. 6-32) have very high values (>1000ppm). The Reocin Fm. limestones have exceptionally high Zn values (mean 60ppm) which is probably linked to diagenesis by metal-rich solutions associated with the Pb/ZnS mineralization.

7) Step 7. Sucrosic dolomitization was the next step in the diagenesis of the Reocin Fm. platform carbonate. This formed under mixed-water conditions (mixing of marine and meteoric phreatic environments) initially (see Chapter 6.9).

8) Step 8. A relatively late-stage, ferroan, equant or drusy spar is developed in some of the grainstones of the Reocin Fm. This is the final infill of inter-allochem void spaces and also occurs as a vein-fill cement. It is late-stage, "deep" burial cement linked to hypersaline(?), connate waters.

9) Step 9. Stylolitization occurs after the early sucrosic dolomitization and non-ferroan, equant spar generation. It can also form after Step 10. Stylolitization is linked to chemical solution during compaction. In some cases

(especially in the biomicrites of Facies 2), the stylolitization can be so intense as to give the rock a pseudo-nodular appearance.

10) Step 10. Local ferroan dolomitization then followed. This exists in two main types. Firstly, there was low-Fe, sucrosic dolomite which was followed by high-Fe hypidiotopic dolomite. Secondly, white sparry dolomite and associated baroque dolomite formed in open-space voids associated with faults, fractures, karst-related voids as well as reefal frameworks. The latter is associated with major Pb/ZnS mineralization at Reocin and La Florida (Fig. 6-16). This cementation is related to warm, reduced, hypersaline connate waters mixing with karst-related surface-derived waters.

11) Step 11. Recent surface weathering, karstification and caliche growth has affected these carbonates.

This diagenetic history is probably something of an oversimplification, though it represents the only systematic attempt to understand the diagenesis of this formation or other Urgonian carbonates in the Basco-Cantabrian basin.

In conclusion, the Reocin Fm. limestones underwent syndepositional marine phreatic cementation. During relative sea-level fall, these limestones were subjected to extensive meteoric water dissolution of unstable phases and related low-Mg calcite cementation. This phase is also linked to extensive mixing water dolomitization (Chapter 6.9). The phase of meteoric-water diagenesis is directly linked to the extensive subaerial exposure/karstification at the end Reocin Fm./pre-Santander Group time (upper (?) Lower Albian) - Fig. 6-28. During progressive, later burial (Upper Albian to Upper Eocene times), ferroan calcites and dolomites cemented the remaining porosity. This produced the tight limestones seen today.

6.9 Dolomitization of the Reocin Formation

The Reocin Fm. is extensively dolomitized. The central part, in general, of this formation has been pervasively dolomitized and is termed the Novales Mb.. It consists dominantly of ferroan and non-ferroan, sucrosic or limpid dolomites

with minor sparry dolomite. The geometry, petrography, geochemistry and interpretative models for the different generations of dolomite will now be considered:-

6.9.1 Geometry of the Novales Member Dolomite

Within the outcrop of the Reocin Fm. (shown as C^{2a}_{1s} in IGME a, c, e, g, i; Comillas and Torrelavega IGME maps), the Novales Mb. dolomite itself has been mapped (general map of it see Fig. 6-16). Small areas of this formation have been mapped in detail and the location of these areas (and sections through the formation) is shown in Fig. 6-33. Vertical log sections are shown in Figs. 6-1 and 6-2. In general, the Novales Mb. thickens towards palaeohighs whereas the Reocin Fm. itself thins. This picture is complicated by volumetrically-small, fracture-related, sparry dolomitization.

In the westernmost part of the area (Western Stable zone of Fig. 6-16, zone 1 of Chapter 1.5.1), around La Florida (54079664, Fig. 6-33), the Novales Mb. forms a central tongue that thins westwards (Sections 3 & 4 on Fig. 6-1/6-2, Fig. 6-34 located on Fig. 6-33). In the La Florida mine (op. cit.) there is major Pb/ZnS mineralization confined to the basal 40m of the Novales Mb. (Fig. 6-35). This is hosted in white, sparry dolomite and is primarily in four metre-thick horizons associated with coral patch reefs (Facies 7 of Chapter 6.4) and requienid banks (Facies 5). A section through the key mineralized horizon is shown in Fig. 6-36 and located on Fig. 6-33. Further east, towards Bustriguado (Fig. 6-33; 54659666), the Novales Mb. (and Reocin Fm.) continues to thicken to the east (Figs. 6-37 & 6-38; located on Fig. 6-33; Section 5 on Fig. 6-1). The upper and lower contacts of the Novales Mb. here are highly irregular. Isolated metre to ten-metre scale "pods" of the original Reocin Fm. limestone locally remain within the body of the Novales Mb. dolomite (Fig. 6-37). On the Bustriguado road section (Section 5 on Fig. 6-1), the lower 245m is a brown-weathering, fawn-coloured, sucrosic dolomite with vuggy and mouldic porosity. It is locally cut by white, sparry

dolomite. The upper 130m is a black-weathering, tight, dolomite with mm-sized carbonaceous fragments.

Further NE, around Caviedes (Figs. 6-33 & 6-38; Map 1), the Reocin Fm. is thin, virtually totally dolomitized and has only rare pods (on a tens of metre scale) of undolomitized Reocin Fm. limestone. Around Santillana (Fig. 6-16), the formation is only dolomitized at the base against a bounding thrust fault (top Jerra Thrust of Fig. 1-7, 53909764).

The Reocin Fm. in the central region of NW Santander, between Comillas and Torrelavega (Fig. 1-6), is shown in Figs. 6-16 and 6-39. Around Comillas, the Novales Mb. has a distinctly lenticular geometry (Fig. 6-39, Fig. 6-40 - located on Fig. 6-33) over the Comillas palaeohigh (Fig. 6-16). To the east this dolomite thins/is partly cut out by faulting (Fig. 6-40) until it is entirely replaced by the original, undolomitized Reocin Fm. limestone at Ruiloba (Section 10 on Fig. 6-1 and Fig. 6-40). The dolomites of the Comillas area are of the typical sucrosic and sparry types.

To the SE of Comillas is the Novales palaeolow (Figs. 6-16, 6-33 and 6-39). The southwestern part of the Novales palaeolow is intensely dolomitized (Fig. 6-39, 6-41 which is located on Fig. 6-33). Moving south from Ruiloba, the Novales Mb. dolomite initially interdigitates with the Reocin Fm. limestone (cf. Section 11 on Fig. 6-1; Fig. 6-41) then becomes progressively thicker to south until it nearly completely dolomitizes the Reocin Fm. (Section 12 on Fig. 6-1).

In the southern section of the Novales palaeolow, around Bustablado (55759702; Fig. 6-39, 6-42 & 6-43), the Novales Mb. thins rapidly to the east. This picture is complicated by dolomitization linked to faults through Bustablado (Fig. 6-42) where mylonitized, brecciated sucrosic dolomite occurs together with later, white sparry dolomite.

6.9.1.1 Novales Member at Reocin

To the SE of the Novales palaeolow is the San Esteban/Reocin area (Fig. 6-39, Sections 14 & 15 on Fig. 6-1). This is an area of extensive dolomitization

by ferroan dolomites and sparry ankerites with only rare, very late epigenetic non-ferroan dolomites (Vadala, 1981; Vadala et al., 1981). The dolomite forms a lenticular body that thickens from around San Esteban on the Rio Saja (56199708) eastwards towards Reocin where it is thickest and virtually completely dolomitizes most of the 220m of the Reocin Fm. (Figs. 6-16, 6-39; Section 15 in Fig. 6-1). The Novales Mb. then thins and interdigitates with limestone to the NE of Reocin. This interdigitation is shown on the geological map of the Reocin area (Fig. 6-44 and located on 6-33).

The top of the Reocin Fm. is marked by an extensive subaerial exposure surface (Fig. 6-45) which has locally developed karstic features such as rootlet beds or intense alteration of the top 20m of the Novales Mb. dolomite (Vadala, 1981). Overlying the karst surface are the siliciclastic sediments of the Lower Albian Las Penosas Fm. (Fig. 2-1) which are partly cemented by high-Fe ankerites. The overlying Middle/Upper Albian Barcenaciones Fm. (Fig. 2-1) has also been strongly dolomitized by ferroan dolomites. A correlated bore-hole cross-section at Reocin (Fig. 4-46) shows the high-Fe dolomite/ankerite development in the carbonates there.

Reocin is the site of major Pb/ZnS mineralization in the basal 40m of the Novales Mb. (Fig. 6-48; Monseur, 1967). The Reocin mine is estimated to have about 41 million tonnes of ore with a mean metal content of 7.3% Zn, 0.8% Pb in the crude ore (Vadala, 1981). It is one of the largest Pb/ZnS mines in Europe, and one of the youngest, major, carbonate-hosted Pb/ZnS bodies in the world. A log of the basal 40m (Fig. 6-47) shows that the mineralization is principally contained in three major beds a few metres thick. These are, from the base, the Capa Sur (Fig. 6-49), the Capa Norte and the Tercera Capa. The largest of these, the Capa Sur, is about 1.5km long, 500-800m wide and 2-10m thick (Vadala et al., 1981). These beds have been mapped out (Figs. 6-46, 6-48 & 6-50) in the Reocin area. Monseur's letter classification of the mineralized zone (Fig. 6-47; Monseur, 1962, 1967) has been used in Fig. 6-50. The Roman numerals used on this figure refer to the boreholes through the basal 40m of the Novales Fm. around Reocin located on Fig.

6-51. In the eastern part of the mine and the open pit, the ore fills palaeokarstic cavities (Fig. 6-46). The central part of the mine is barren (Fig. 6-50). In the western part, the mineralization contains roof collapse breccia blocks several tens of metres wide indicating subterranean precipitation of the ore.

6.9.1.2 Geometry of the Novales Member of the north of Reocin

To the N of Reocin, the dolomite gives out under the Santillana syncline (Figs. 6-16 & 6-52) and reappears around the Orena palaeohigh (Figs. 6-16 & 6-52). The Orena palaeohigh (Fig. 6-53 - located on Fig. 6-33) is an area of thinned Reocin Fm. but a higher proportion of the Novales Mb. (Sections 16, 17 & 18 of Fig. 6-1). To the SW of Novales, the Novales Mb. interdigitates with Reocin Fm. limestone (Fig. 6-43) and the Reocin Fm. thickens towards the core of the Novales palaeolow. Over the centre of the Orena palaeohigh the Reocin Fm. is dolomitized to the base of the formation, and nearly to the top of the formation (Fig. 6-53; Section 17 of Fig. 6-1). To the south and southeast of the Orena palaeohigh, the Reocin Fm. thickens but the Novales Mb. thins (Sections 18 & 19 of Fig. 6-1; Fig. 6-53).

Notably, at Cuchia, the Reocin Fm. section has no large-scale, sucrosic dolomitization (Section 20 of Fig. 6-1) but is cut by a few metre-thick, fault-related, ferroan dolomites (2% of the section at most). To the SE of the Miengo palaeohigh, the dolomite thickens (Sections 21 & 22 of Fig. 6-1) towards the Puente Arce thrust to the SE (57969794; Fig. 6-54). The Novales Mb. then rapidly thins towards the base of the Camargo palaeolow (Section 23 of Fig. 6-1). Around Camargo/Escobedo (58249776; Fig. 6-54) there is local, intense dolomitization around major faults (Section 24 on Fig. 6-1). There is also intense dolomitization around the eastern edge of Camargo palaeolow at Revilla/Herrera (58529774 to 58439808; Fig. 6-54). This dolomite continues in the upper part of the Peña Castillo section (Section 25 of Fig. 6-1) towards the Santander palaeohigh (Fig. 6-16). In the Sierra de Villaneuva, 10km south of Santander (Fig. 6-16, 59009740),

there is extensive dolomitization of the central part of the Reocin formation. This is dominantly a ferroan dolomite (Rohou, 1984). Conclusions on the distribution of the Novales Mb. in the light of genetic models will be given in Chapter 6.9.

6. 9. 2 Types and models of dolomite in the Reocin Formation.

The main types of dolomite will now be discussed in, approximately, the diagenetic order they were created. Figure 6-55 gives a schematic summary of the dolomite diagenetic sequence.

1A) Low-Fe, hypidiotopic, sucrosic dolomite - description.

This is one of the more abundant generations of dolomite and is generally light tan-weathering. It is a medium- to coarse-grained (about 500-1000μm), unimodal grain-size distributed, hypidiotopic (locally idiotopic), moderately to strongly sucrosic (cloudy crystal centred) dolomite. It has straight crystal boundaries, very slightly undulose extinction, some clearly rhomboidal crystal shapes despite it's general hypdiotopacy and forms a tight, interlocking mosaic with virtually no residual intercrystalline porosity that has destroyed all the primary fabric of the precursor limestone (Fig. 6-56).

This generation of dolomite preserves rare ghosts in dolomite of orbitolinids, echinoderms and miliolids together with moulds clearly after requienid rudists. All these fauna are typical of the marine Reocin Fm. limestone. Such a preservation of marine ghosts suggests that this dolomite must be a neomorphic replacement of a pre-existing marine limestone (Bathurst, 1971). On a microscopic scale, the dolomitization front can be seen to be replacing normal Reocin Fm. limestone. A general order of replacement is that initially voids and fractures are dolomitized, followed by the mouldic zones of original aragonite/high-Mg calcite (HMC), then the micritic matrix, followed by the micritized allochems (e.g. miliolids), then coarse-grained calcite cement and finally single-crystal calcites (e.g. echinodermata fragments). On a macroscopic scale, a simple interdigitation of the Novales Mb. dolomite with the Reocin Fm.

limestone can be seen locally (Fig. 6-58; Section 11 of Fig. 6-1). A few, isolated pods of rather micritic, but otherwise normal, Reocin Fm. limestone occur on a 0.5-20m scale within the body of the Novales Mb. dolomite (e.g. Section 6 of Fig. 6-1; Figs. 6-41 & 6-43). Taken together, these factors clearly show that the sucrosic dolomite is simply a replacement of the pre-existing Reocin Fm. limestone. It also indicates that this limestone must have been lithified pre dolomitization and that it was not a penecontemporaneous dolomite formed from unconsolidated ooze on the seafloor or in peritidal carbonate flats.

These dolomites show a relatively non-zoned, dull orange cathodoluminesence, with rather duller outer rims than cores (Fig. 6-57). Following Pierson (1981) and Fairchild (1983), this relates to a higher Mn/Fe ratio in the cores and an Fe content of under about 1 mole %. There are local thin (50 μm), very late-stage dolomites that rim local pore space that have a bright luminescence. These dolomites take no, or very little turquoise blue stain with potassium ferricyanide/Alizurin Red S solution (Dickson, 1966), indicating their low-Fe nature. There is no Pb/ZnS associated with this dolomite generation.

Stable isotopic values for the Reocin Fm. are given in Figs. 6-59 & 6-60. The original limestones of the Reocin Fm. have isotopic values of $\delta^{13}C_{PDB}$ = + 3.0 ‰ ± 0.4 and $\delta^{18}O_{PDB}$ = - 3.8 ‰ ± 0.3. Although these do fall within the standard field for ancient platform marine limestones (Hudson, 1977), they are slightly heavy in carbon (about +2‰) and light in oxygen (about -3‰) relative to the mean values for this type of limestone. The heavy carbon could be linked to "heavy" marine waters in general in the Upper Aptian. The Upper Aptian was an "oceanic anoxic event" (Schlanger & Jenkyns, 1976; Jenkyns, 1980) where the higher preservation potential of $\delta^{13}C$-light terrestrial organic matter in the black shales in the oceans and/or a change to non-calcifying planktonic foraminifera which strongly discriminate against ^{13}C, allows the ocean waters to become heavy in carbon, generally by about +2‰. Scholle & Arthur (1980) have found exactly the same positive shift in C by about 2‰ in pelagic carbonates. With sufficient ocean water mixing, this heavy-carbon seawater could invade marine platforms such as NW

Santander and produce the observed shift in $\delta^{13}C$. An explanation of the moderately positive $\delta^{13}C$ values by surface-related soil/karst interaction is unlikely as these produce lighter carbon limestones (Irwin et al., 1977; Hudson, 1977; Land, 1973). Similar $\delta^{13}C$ values are known from the Urgonian platform limestones of Texas which had a comparable diagenetic history (Prezbindowski, 1977). The corresponding $\delta^{18}O$ value for the limestone (mean $\delta^{18}O$ = - 3.8‰) are rather light, and most probably reflect the unavoidable incorporation in the samples of some (10-20% ?) light $\delta^{18}O$, neomorphic to equant, meteoric phreatic spar (Step 6 of Chapter 6.8.1 spar, Fig. 6-30). A mixture of this and normal lime mudstone could generate the $\delta^{18}O_{calcite}$ seen.

The low-Fe dolomites have mean isotopic compositions of $\delta^{13}C$ = +3.2 ‰ ± 0.1 and $\delta^{18}O$ = -6.8‰ ± 0.2 (Figs. 6-59 & 6-60). The most striking feature of the carbon value is that it is virtually unchanged relative to the pre-existing marine limestone value (statistically significant at 95% on a t test). On the other hand, the dolomites show about a -3‰ fall in $\delta^{18}O$ relative to the limestones. The C value is seen as a direct derivative of the pre-existing limestone. As Degens & Epstein (1964), Land et al., (1975) and Dunham & Olson (1980) amongst others have pointed out, the reservoir of carbon in marine sediments is much greater than in most diagenetic solutions, so on mass balance principles the diagenetically altered dolomites tend to inherit the vast majority of their carbon from the precursor limestone, with little effect from the light C soil/karst-derived systems.

Conversely, the reservoir of oxygen in diagenetic solutions is very large and so the $\delta^{18}O$ values of dolomites strongly reflect the compositions of these diagenetic solutions. Such strong depletions in ^{18}O are attributable to formation waters isotopically light in oxygen (such as meteoric ones) and/or precipitation at elevated temperatures (Land, 1980,1983). In this dolomite generation, the temperature can be independently estimated. As the dolomite is stratabound, has an eroded karst ***in dolomite***, was developed at Reocin at the top of this formation, and clasts of sucrosic dolomite have been found in the overlying Albian Las Penosas and Barcenaciones Fms. (Comillas section; Rohon, 1984), this dolomite formed at

near surface temperatures (such as below a karst) of about 25-30°C (the genetic origin of this dolomite is discussed below). As this dolomite formed at these sort of temperatures, the low $\delta^{18}O$ ***must*** be due to light oxygen formation waters. The relationship between the $\delta^{18}O$ of dolomite, it's formation water's $\delta^{18}O$ and temperature is, to a first approximation, known (Friedman & O'Neil, 1977). If a temperature of 25°C is taken for the formation temperature, an open system is assumed (a likely situation due to the mass balance of Mg & Ca to be moved during this dolomitization - see below), and if no subsequent recrystallisation has occurred to these dolomites, then the quadratic relation of Matthews & Katz, 1977 can be used to estimate the $\delta^{18}O_{PDB}$ of the formation waters. $\delta^{18}O_{PDB}$ values of -7.1 ‰ and -39‰ are produced. The latter figure is a mathematical artefact of the quadratic equation, and quite unreal relative to modern formation or meteoric waters (Arthur et al., 1983; Land, 1983). Similar values to the -7‰ value come from Land (1983) plot of this temperature/$\delta^{18}O_{dolomite}$ system. It is remarkable that similar "ball park" values of $\delta^{18}O_{PDB}$ are obtained if other reasonable temperatures are used (20-35°C) or if other reasonable $\delta^{18}O_{PDB}$/T relations are used (Northrup & Clayton and other systems of the other authors op. cit. above). The latter variation gives an uncertainty in $\delta^{18}O_{water}$ of about 4‰. Taken together, these arguments suggest that this generation of dolomite formed from waters with a $\delta^{18}O_{PDB}$ of about -7‰. Such values are typical of modern mid-latitude meteoric precipitation waters (Arthur et al., 1983), strongly suggesting that the negative $\delta^{18}O_{dolomite}$ values were generated in association with light oxygen-rich, meteoric-derived waters.

The major and minor elements of the Reocin Fm. carbonates were examined by standard electron microprobe techniques and atomic absorption trace element results are reproduced from Vadala, 1981 (Fig. 6-61). The precursor limestones (Lines 1 to 6 of Fig. 6-61) show that they are now typical low-Mg calcite (LMC) with neomorphic alteration of syndepositional high-Mg calcite miliolids and echinoderms to their present LMC (see Step 6 of Chapter 6.8.1). The low-Fe (≤2 mole % Fe) sucrosic dolomites (Lines 7 to 14 of Fig. 6-61) are Ca-rich (range 50-

56 mole %, mean 51%), Mg-poor (range 43-49.5 mole %, mean 47.5%) dolomites. They show a range of low-Fe values from 0.05 to 2 mole % (mean 0.96 mole%). In general, these dolomites show some zoning geochemically, with Fe-rich, Mg-poor rims to the crystals (Fig. 6-64). A slight tendancy towards Fe enrichment and Mg depletion occurred towards the more limpid, pore-space filling, later-stage, less-sucrosic dolomites. These are in fact gradational to the more Fe-rich dolomites described below.

Trace element analysis of the limestones from Vadala, 1981 on 7 values give low Sr values (mean 150 ppm, range 100 - 200 ppm) and low Mn values with a mean of 80ppm, range 20-100ppm. The low-Fe dolomites have very much lower Sr values (n=5, mean 23pppm, range 20-50ppm), but higher Mn ones (mean 687ppm, range 150-2000ppm).

1B) Low-Fe, hypidiotopic, sucrosic dolomite - interpretation.

This sucrosic dolomite is considered to be a mixed-water dolomite in which slightly reduced, meteoric-derived water invaded intraplatform diapiric palaeohighs immediately after the Reocin Fm. was deposited (Fig. 6-62). These waters were derived from the non-marine/peritidal siliciclastics of the lower part of the Albian Las Penosas Fm. (see Chapter 7) above the Reocin Fm. (Fig. 6-62). This model (an adapted "dorag" model) was proposed by Badiozamani (1973) to explain dolomitization of palaeohighs in Ordovician dolomites of Wisconsin and relied on the thermodynamic stability of dolomite over calcite in brackish water with Mg/Ca ratios greater than unity. Indirect support for the model came from the study of Tertiary aquifers in Florida (Hanshaw et al., 1971) who suggested that the mixing zone between meteoric and subjacent marine waters was a zone of dolomitization. Folk & Land (1973) suggested that dolomites are more likely to form under high Mg/Ca ratios and reduced salinity conditions and that mixing zones conserve the high Mg/Ca ratio (about 5) of seawater but have a reduced salinity. Motor (1982a) pointed out that increased alkalinity also helps promote dolomitization, as does a reduction in sulphate concentrations/less silica

diagenesis (Baker & Kastner, 1981). The mixed-water model introduces high P_{CO2} continental groundwater into seawater and reduces seawater sulphate and salinity concentrations with sulphate-poor meteoric waters.

The stratigraphic evidence that the sucrosic dolomite generally forms as lenses over palaeogeographic highs was discussed in Chapter 6.9.1 . The large size of the dolomite body precludes closed system solution/cannibalization models of dolomitization (Goodell & Garmer, 1969) on mass balance grounds. The precursor is now a LMC with a mean of 1.46% Mg (Fig. 6-61, line 6) whereas the dolomite has a mean of 47.46% Mg (Fig 6-61, line 14). As the dolomite constitutes about 45% of the 330m thick (mean) Reocin Fm., the potential cannibalization limestone would have to have been (0.45 × 330 × 0.47 ÷ 0.015)m = 4600m thick to be able to have its Mg removed to produce the quantity of Mg seen in the Novales Mb. dolomite. This excessive figure is geologically untenable, and indicates that the system must have been an open one to Mg. Although the end-Aptian Mg content of the limestones would have been higher than it is today (with HMC and aragonite) it is likely to have been no more than 2-4 mole % Mg, which would still make closed system sucrosic dolomitization impossible. Further, petrographic evidence suggests this dolomite formed *after* conversion of HMC to LMC (cf. dolomitization of LMC echinoderms, which themselves had already been converted from their depositional state as HMC - Line 4 of Fig 6-61).

The sabkha penecontemporaneous dolomitization models can be rapidly dismissed as there are no primary evaporitic minerals (or pseudomorphs after them) in the Aptian/Albian sequence, no evaporitic stromatolites and no fine-grained (10-20μm) dolomicrite. Further, the fauna of the Reocin Fm. are marine, rather than restricted peritidal ones (Chapter 6.4). Similar arguments dismiss the standard reflux model (sensu Adams & Rhodes, 1960) of dolomitization.

Primary dolomitization not associated with evaporite minerals has been reported from ephemeral alkaline lakes such as the Coorong (von der Borch, 1976, Muir et al., 1980). However, the Novales Mb. dolomite is clearly secondary. It is not fine-grained, it is not associated with alkaline minerals such as magnesite, is

not associated with peritidal/lacustrine environments (algal mats, tepees, collapse breccias etc.) and does not form as layered crust at the top of the Reocin Fm where such lakes could have occurred.

Dolomitization by the burial compaction of dewatering shales does not apply as the karst *in dolomite* clearly shows that this generation of dolomite is a very near surface one. A burial compaction dolomitization model does not explain why other Urgonian platforms of the basin were not dolomitized (shale waters should dolomitize them all equally) and does not explain the greater abundance of dolomite towards the palaeohinterland (S/SW) rather than offshore (N/NE) towards the dewatering shales. Further mass balance problems exist (too little Mg^{2+} from shale - about 30 times the volume of shale to dolomite is needed- Morrow, 1982b)

Stratigraphic arguments strongly support the mixed-water origin for this dolomite. The location over palaeogeographic high, the interdigitation of the lenses with the Reocin Fm. limestone, the development of a major karst (20m of geotitic orange soil at the top of the Novales Mb. at Reocin), the presence of an Albian flora at Reocin (Vadala 1981) and the presence of fluvial/upper estuarine siliciclastics (Las Penosas Fm.) above the Reocin Fm (Chapter 7) all point to meteoric invasion at end Reocin Fm times. Further evidence comes from the clear meteoric diagenesis seen in the Reocin Fm. limestones (Chapter 6.8.1, Steps 5 & 6). The presence of a karst *in dolomite* suggests that the initial meteoric lense could extend seaward of the coast (c.f. Florida aquifers of Hanshaw et al., 1971) and that as the regression at end Reocin Fm. times continued this lense would expand vertically downwards and horizontally seawards until the top of the dolomite lense was eroded by continued karstic denudation (Fig. 6-62). Hence, a karst *in dolomite*. The weathering of the karst to a thick (20m) soil is typical for Recent (sub)tropical karsts in dolomite.

Texturally, medium-grained hypidiotopic sucrosic dolomites such as these are commonly (though not exclusively) associated with mixed-water dolomitization. Petrographically, these dolomites bear a strong resemblance to the interlocking hypidiotopic, cloudy-centre, clear rim dolomites described by Sibley (1982) from

the Pliocene of the Caribbean where mixed-water dolomitization has occurred. Sibley (op. cit.) showed that they formed from LMC that had already been affected by meteoric water which had converted the HMC to LMC (as here). The initial dolomite formed under conditions of near LMC-saturation with few nucleation sites. The initial, fine (?) dolomite euhedra were scattered through the micritic LMC and tended to be cloudy and inclusion-rich due to inclusion of fine LMC crystals. As the system became progressively more undersaturated with respect to LMC, clearer rims developed finally to generate the fabric-destructive, medium- to coarse-grained mosaic with compromise crystal boundaries seen in this dolomite generation. The euhedral, micrometer-sized dolomite rhombs scattered in recent sediments forming on Andros Island (Gebelin et al., 1980) may record such initial steps to a later sucrosic texture. The inclusion-poor, low-Fe dolomites that occur locally with this dolomite generation are most probably true cements to initial porosity or dolomitized LMC cement. Similar, relatively limpid dolomites have been described by Sibley, 1982. It is not surprising that with such an early genesis, this generation of dolomite preserves some of the most rhombohedral, euhedral dolomites within the Novales Mb. Hence there is strong petrographic similarity between these dolomites and known mixed-water ones from the subrecent.

Geochemical data further support a mixed-water origin for these dolomites. The low-Fe dolomites show a consistent Ca-rich, Mg-poor nature which is linked to strong buffering by, and excess of calcite in, this initial dolomitization (Line 14 of Fig. 6-61). In general the very low Sr values (150ppm), especially relative to the higher Sr values in the limestones, are strongly consistent with a genesis associated with very low Sr meteoric waters. The difficulty with establishing a distribution coefficient (and its temperature dependence) for Sr in the dolomite/water system (Land, 1980) make direct extrapolation to the waters (or, more probably, generations of waters) difficult. A further problem is not knowing the water/rock ratio, though the mass balance calculations for Mg above point to this as a relatively open system. However, these Sr values do fall into the low-Sr dolomites of Veizer & Demovic (1974) and

assuming the low distribution coefficient of Katz & Matthews (1977), these Sr values would fall within the "dorag" range (Veizer, 1983).

These dolomites have moderate Mn values (mean 0.036 %, line 14 of Fig. 6.61), but an association with not-negligible Fe contents (mean 1 Mole %) suggests that these waters were slightly reduced (Harrie et al., 1985). A linear relation between the Mn and Fe contents has been described by Vadala (1981) for the Novales Mb. dolomite (Fig. 6-65). Moderate (0.1-2 mole %) to high Fe cements (see below) only form from moderate to highly reduced waters (Evamy, 1969) where the ferrous cation can exist. Similarly high Mn values are known from organic-rich brackish water sediments (such as the estuarine muds of the Las Penosas Fm.) - Berner, 1971. Continental waters (meteoric or river) are much richer in Fe & Mn than marine ones, and values of about 1ppm Fe have been found in the Recent (Durum & Haffty, 1963) but higher values occur in areas of lateritic weathering in humid, subtropical environments (the Santander area ?) where Fe is transported as colloids in oxidised waters. However, on percolating through even thin (cm/m) of organic-rich brackish/non-marine mudstones of the basal part of the Las Penosas Fm. siliciclastics, such waters will be rapidly reduced by bacterial respiration and organic matter oxidation. This will convert the iron to the ferrous state (Hem, 1970; Berner, 1971; Claypool & Kaplan, 1974). Iron can be released from the metastable iron-oxide coatings of detrital grains (Carroll, 1958; Curtis, 1967) via cation exchange by clays (Drever, 1971; Irwin, 1980) or from woody carbonaceous matter. It is unlikely that the smectite to illite transformation was a major source of Fe or Mg as it occurs at kilometres depth, not near surface (<200m) conditions. In marine sediments there is normally enough sulphide to convert all the Fe^{2+} to FeS (Taylor & Sibley, 1986), but in estuarine/brackish muds, with little S^{2-}, the remaining Fe^{2+} can build up. Hence these waters were primarily meteoric-derived but suffered surface alteration and slight-reduction before they became the dolomite formation fluids. Although reduced, the oxygen isotopic composition still strongly reflects the original, low-$\delta^{18}O$, meteoric values.

In summary, there is strong stratigraphic, petrographic and geochemical evidence for the proposed mixed-water dolomitization model. The only significant difference for the standard model (sensu Badiozamani, 1973) is that moderate reduction of the groundwater had occurred before dolomitization as the water passed through the basal parts of the Las Penosas Fm. muds, and accounts for the moderate Fe & Mn concentrations. Hence, the key change from the oxidised meteoric waters associated with the equant calcite precipitation is that these formation waters are more reduced and Fe-rich. In this model the seawater would be the Mg source. Following the Ghylen-Herzburg principle (Davies & DeWiest, 1966), the freshwater lense ought to extend below sea level about 40 times the height of water table (= meteoric phreatic environment), though, in practice, hydrologists find the lense extends only about 30 times due to mixing of seawater with fresh. This mixing zone of dolomite formation is often comparable in size to the freshwater base and is 1-100m thick in permeable coastal aquifers in Florida (Hanshaw et al., 1971). Due to continued recharge of the meteoric lense, a seaward flow of freshwater is generated which is counter-balanced by a landward flow of saline water. Thus a mixing zone is continuously supplied with Mg and freshwater. These mixing zones are highly responsive to the shore location and meteoric input rates and are known to move on a daily, annual, or longer term basis (Gebelein et al., 1980). Such variation would allow the repeated flushing of a crystal by many mixed water lenses through geological time, and so provides the 10^4-10^5 pore volumes of water necessary to allow dolomite generation (Dickson, 1985). Such lenses are known from Recent low-lying, coastal aquifers to form as lenses over domal highs (Gebelein et al., 1980; Hanshaw et al., 1971; Davies & DeWiest, 1966) as the Novales Mb ones did. In general, these lenses pass down through the permeable horizons, but are driven laterally due to lower impermeable horizons in areas of low-lying, low-dip strata. Such gentle dips were present in the Mid Cretaceous of Santander which was effected by the gentle diapiric and basement palaeohighs. The abundant diagenetic fabrics of the Reocin Fm. suggest it had a high permeability. Similar, modern micritic platform limestones of comparable thicknesses to the

Reocin Fm. do have comparable high permeabilities. Bahaman muds have about 55% porosity at 100m and about 50% at 200m (Terzaghi, 1940) and porosities of 40-50% at 100-250m have been reported by Scholle, 1977. However, underlying the Reocin Fm. are the marly limestone/marls of the Rodezas Fm. (Fig. 2-1). Such shales would have acted as an impermeable barrier, forcing the freshwater into ***laterally*** extensive lenses (cf. Allen Fm. of Land et al., 1975). Such an impermeable barrier protected the San Esteban Fm. below the Rodezas Fm. from these meteoric lenses. The Reocin Fm. is virtually identical lithologically to the limestones of the San Esteban Fm. below (Chapter 5.4), but the San Esteban Fm. is *not* dolomitized, except very locally against faults. This further proves that it was not only the primary lithology that controlled dolomitization (clean, rudistid limestones being a preferential target), but also the diagenetic solutions (meteoric waters) that the Reocin Fm. experienced as against those that affected the San Esteban Fm..

An obvious question is why isn't the thinned sequence at Cuchia (Section 20 of Fig. 6-1) dolomitized? The explanation is that the top of the sequence, unlike the Reocin Fm. elsewhere in NW Santander, is dominated by cemented grainstones formed over the palaeohigh (Section 15A/29, 205-229m). The non-ferroan, equant spar suggests that the initial marine cementation was followed by meteoric phreatic cementation which sealed up this limestone during, or immediately after, the deposition of the Reocin Fm.. The sharp truncation of the beach sequence at the top also supports synsedimentary cementation. Thus, the Cuchia sequence would be capped by an impermeable, already cemented and tight grainstone which would not permit the slightly-reduced-meteoric water lenses to enter and dolomitize the rock.

On a broader scale, why is so much of the Aptian platform limestone of the whole basin (Fig. 6-21) not dolomitized? The controlling factor is that during the Albian NW Santander experienced strong relative sea-level fall (and dolomitization) whereas the rest of the basin experienced strong, distensional subsidence (Chapter 7.10) and the Aptian limestones were covered by Albian deep-marine marls and continued to have only marine pore waters (e.g. Fernandez-

Mendiola & Garcia Mondejar, 1985). Hence, the mixed water model explains the facies, geometry, petrography and geochemistry of the low-Fe dolomite.

2A) High-Fe, hypidiotopic/xenotopic dolomite - description

This is the other dominant type of dolomite and, to some degree, there is a gradational change in geochemistry and texture from the hypdiotopic, sucrosic dolomites mentioned above. This class subdivides into transitional hypdiotopic/xenotopic-textured dolomite with moderate-Fe contents (2-4 mole % *in general*) and a truly xenotopic, high-Fe dolomite (4-6 mole % Fe *in general*). The first group are essentially texturally similar to the hypdiotopic dolomites described above and primarily differ in their higher Fe contents, lower $\delta^{13}C$ values, and slightly more xenotopic texture.

The xenotopic, high-Fe dolomites are tan to brown weathering, medium-to-coarse grained (about 400-1000 μm), unimodal grain-size distributed, xenotopic (to moderately hypdiotopic locally), generally limpid dolomites (Figs. 6-63A & B, 6-66, 6-67 & 6-68). These dolomites are tightly packed and anhedral with planar, curved lobate, serrated or irregular intercrystalline boundaries. Crystals have a moderately undulose extinction and rare crystal-face junctions. This xenotopic dolomite does not develop skeletal ghosts, though the more hypdiotopic transitional type of dolomite locally can. There is rapid change between the hypdiotopic to the xenotopic dolomites on even a mm to cm scale, and this dolomite has been seen replacing both hypdiotopic dolomite and calcite. Ghosts in the transitional dolomite suggest that it too is a neomorphic replacement of a precursor carbonate rather than a true cement. The xenotopic texture is thought to have developed by neomorphic recrystallization/limestone dolomitization above a 'critical ruffening temperature' of about 50°C (Gregg & Sibley, 1984).

Neither of these two dolomite types cathodoluminesce, primarily because their Fe content is too high (2-6 mole % Fe). However, this dolomite generation takes a strong turquoise blue stain with potassium ferricyanide/Alizurin Red S solution due to its high Fe content. Stable isotopes (Figs. 6-59 & 6-60) for the

hypdiotopic/xenotopic transition, ferroan dolomites are similar to those of other ferroan dolomites with a mean of $\delta^{13}C_{PDB}$ = -6.76‰. These dolomites are relatively stoichiometric in Ca (49.5-52 mole % Ca, mean 50.1%; line 20 of Fig. 6-61), but are strongly Mg-poor (42-48 mole % Mg, mean 46.6%). The Mg has been replaced by Fe (2-5 mole % Fe), which can substitute for the chemically similar Mg cation in a solid solution. As well as showing higher Fe concentrations than earlier carbonate generations (line 20 vs. lines 14 or 6 of Fig. 6-61), these dolomites show a correspondingly higher concentration of Mn, which has an approximately linear variation with Fe (Fig. 6-65, Vadala, 1981). This relation can also be seen in lines 15 to 20 of Fig. 6-61. Slight geochemical zoning occurs in some crystals (Fig. 6-64) with a tendency towards Fe-rich, Mg-poor rims. In general terms, the more xenotopic a phase, the more Fe-rich it is. This can be seen in the results of lines 15 to 20 of Fig. 6-61. No trace element data exists for this phase.

2B) High Fe, hypdiotopic/xenotopic dolomite - interpretation

The transitional hypdiotopic/xenotopic dolomite is thought to have formed by a similar mechanism to the low-Fe hypdiotopic dolomite, but the process of reduction and Fe-enrichment from the organic-rich muds of the Las Penosas Fm. had continued further and so produced a more Fe-rich dolomite. This could be simply due to the percolation of the meteoric-derived groundwaters through progressively thicker sequences of the Las Penosas Fm. as it was deposited through the Lower Albian. This idea is supported by the petrographic evidence that this dolomite is later-stage than the low-Fe one and that even individual crystals show a progressive Fe enrichment towards their rims. Such an increased exposure to the reducing Las Penosas Fm. muds would also account for the higher Fe & Mn contents of these dolomites, as well as the linear relation between these two elements because they originate from the same source. The calcium content (mean 50%, line 20 of Fig. 6-61) is similar to that of the less ferroan dolomites, though these are more Mg-poor because of the increased substitution for Mg by Fe.

These dolomites have a remarkably similar $\delta^{18}O_{PDB}$ values to the lower-Fe dolomites (-6.76‰ vs. -6.8‰). Although covariance of the isotopic composition of the formation waters and temperature can produce identical $\delta^{18}O$ values in the product carbonate (e.g. hotter, isotopically-heavier formation waters produced the same values as cooler, isotopically-lighter ones), such an exact covariance seems difficult to envisage for this dolomite generation, and dolomite generation from waters of similar oxygen-isotopic nature and temperature seems a much more likely explanation. This is in line with the proposed model, and suggests that, though these dolomites have a higher Fe-content, they need not have formed deeper or at higher temperatures.

The $\delta^{13}C_{PDB}$ of these dolomites are noticeably lighter than those of low-Fe dolomites (-0.44 vs. +3.2‰). An increase in temperature of formation alone would make the dolomites heavier, not lighter in $\delta^{13}C$. In addition there is little or no evidence for a higher temperature of formation. A more likely explanation is that light carbon is from surface-related karst sources. It then mixed with the wall rock values (of $\delta^{13}C$ = +3.2‰) in varying proportions depending on the water/rock ratio (Land, 1973; Irwin et al., 1977; Hudson, 1977).

The higher-Fe, xenotopic dolomites are seen as primarily a recrystallization texture, and its intimate association with more hypdiotopic dolomites suggests that it is primarily a replacement fabric of earlier dolomites.

3A) Baroque and white-sparry, very high-Fe dolomites and associated Pb/Zn sulphides - description.

The white-sparry or baroque dolomite is a volumetrically minor, later-stage (epigenetic) coarse spar that is associated with Pb/ZnS mineralisation. It is a white-weathering, unimodally and coarsely crystalline (200-10000 μm), limpid (non-sucrosic) dolomite with well-developed sweeping extinction in cross-polarized light. This dolomite generation can be subdivided into two types: an interlocking mosaic of coarse sparry dolomite (Fig. 6-69) and a void-infilling, baroque sparry dolomite (Figs. 6-67, 6-68, 6-70). The latter baroque dolomites are characterised

by angular, scimitar-like terminations, slightly curved crystal boundaries, strongly undulose extinction and local zoning by inclusions. No ghosts have been seen within the white sparry dolomite. This dolomite does not cathodoluminesce, primarily because of its very high Fe content (4.5 to 12.5 mole % Fe - Fig. 6-61). However, it does take a strong turquoise blue stain with potassium ferricyanide/Alizurin Red S solution, again due to its very high Fe content.

The white, sparry dolomites of Capa Sur at Reocin (H1, H11, H12 & H13 of lines 21 to 24 of Fig. 6-61) were precipitated as space-filling cement within the original coral framework. They have isotopic values of $\delta^{13}C_{PDB}$ = -0.7 to -1.1‰ (n = 4, mean of -0.7‰) and $\delta^{18}O_{PDB}$ = -6.7 to -7.6‰ (mean = -7.1‰). The white sparry/baroque dolomites are mildly Ca-rich (50.3 to 52.6 mole % Ca, lines 21 to 30 of Fig. 6-61), as many such ferroan dolomites are (Boles, 1978; Radke & Mathis, 1980). These dolomites are strongly depleted in Mg (36.1 to 44.5 mole % Mg, lines 21 to 31 of Fig. 6-61). The most characteristic feature of the geochemistry of these dolomites is their very high Fe content (4.5 to 12.5 mole %, mean = 7.6 %). A few of the most ferroan dolomites fall into the true ankeritic field (Fig. 6-61). This is balanced by a similarly high Mn content (mean 0.33 mole %, line 31 of Fig. 6-61), which is generally proportional to the Fe content (lines 21 to 31 of Fig. 6-61; Fig. 6-65). These dolomites, especially the void-filling baroque ones, show moderate to strong zoning by inclusion-rich zones, which tend to be Fe-rich. There is a strong geochemical zoning, especially in Fe and Mn with inclusion-rich, Fe-rich, Mn-rich, Mg-poor 100-1000μm bands alternating with those that are more limpid, Fe-poorer, Mn-poorer, and Mg-richer.

Vadala (1981) analysed the trace elements of the sparry, white ankeritic dolomites at Reocin by atomic absorption and found they had remarkably low concentrations of Sr (20-50ppm, mean 33ppm, n = 10), Pb (5-10ppm, mean 6ppm), Zn (10-150ppm, mean 43ppm, n = 9), Be (1-10ppm, mean 4ppm, n = 5) and Cu (1-5ppm, mean 2ppm, n = 10). High Mn values (600-2000ppm, mean 1170ppm, n = 10; equivalent to a range of 0.12 to 0.4 mole % Mn - cf. probe results) were also encountered.

In general terms, the sphalerites of the key mineralized horizons at Reocin are dull grey, submetallic weathering, mainly space-filling and have a colloform or locally botrioidal texture (Fig. 6-49). It is the dominant, early-formed metallic phase (Fig. 6-71). In thin section (Fig. 6-72), a colour banding/zoning in the colloform texture is seen. This is related to high Fe concentrations in the darker bands (Fig. 6-73). These sphalerites are high Fe ones (0.4 to 7.3 mole % Fe, mean = 2 %; Fig. 6-73) with a clear solid solution between Fe and Zn (i.e. higher Fe sphalerites are lower in Zn). The colloform sphalerite can form interbedded layers with xenotopic dolomite or white, sparry dolomite (Fig. 6-72). Cubic crystals of galena are found within the sphalerite, indicating their coprecipitation. Dispersed crystals of sphalerite can occur within the sparry dolomitic matrix (Fig. 6-74) and spreads out a few centimetres from the main colloform sphalerite bodies. Some late-stage, light yellow/white colloform sphalerite is present in joints and veins.

Along with the chemically precipitated ZnS mentioned above, there is some detrital, fine-grained internal sediment made of reworked crystals of sphalerite or dolomite (Monseur, 1967; Vadala, 1981; Figs. 6-75 & 6-76). Some of these clearly show internal varves on a centimetre scale (Fig. 6-76), indicating mechanical reworking of dolomite *in situ* in the cavern, followed by sedimentation and then recrystallization as a xenotopic dolomite. Most commonly the internal sediment postdates the initial phase of white sparry dolomite and is generally a brown/grey colour due to inclusion of muds/sulphides (Fig. 6-76). Local varves of sulphides and dolomite form on a mm to cm scale and can geopetally infill local depressions or dissolution cavities (Fig. 6-76). They can also be involved in soft-sediment deformation slumping (Fig. 6-75; Monseur, 1967). This indicates reworking of cavern infill by active currents/gravity, but not synsedimentary, Alpine-type, sea-floor bottom precipitation of the ore body as proposed by Monseur, 1967.

Larger dissolution cavities exist. These are the three key horizons-Capa Sur, Capa Norte and Tercera Capa of Figs. 6-47, 6-48 & 6-50 which are essentially stratiform. These have a generally sharp contact with the wall rock

xenotopic/hypidiotopic dolomites indicating a degree of wall-rock dissolution. These multi-metre cavities have "solution-breccia" blocks of wall-rock dolomites (Fig. 6-77), which were lithified before brecciation and are now coated in baroque sparry dolomite. At the base of such cavities are geopetal linings of metal-rich kaolinitic/illitic (Vadala, 1981) clay. This clay shows soft-sediment deformation features. There are noticeable differences along strike in the Reocin ore deposit. The western end is more strongly fractured, more PbS- and marcasite-rich, less mineralized overall, has more collapse breccias and has more abundant calcite than the eastern end of the ore body (Vadala, 1981). There are also major non-stratiform bodies at Reocin on a decametre scale (e.g. "Barrendera", Figs. 6-45, 6-48) with collapse breccias, colloform sphalerite and baroque dolomitization.

A relatively similar picture is seen at the mineralization at Novales (Fig. 6-16), though the mineralization is more Pb-rich, is more strongly fractured, associated with much baroque dolomite, and the mineralization has been remobilised into a local fracture stockwork. Similar mineralization occurs at La Florida (Figs. 6-34, 6-35 & 6-36).

3B) Baroque and white, sparry, very high-Fe dolomites and associated Pb/Zn sulphides - interpretation

The origins of the white, sparry dolomite and Pb/ZnS are so intimately related that they will be considered together, primarily considering the Reocin deposit as it is the most studied. The basic model (Fig. 6-78) is one of mixing of highly-reduced, organic-rich, meteoric-derived/brackish waters from the estuarine siliciclastics of the Lower Cenomanian Bielba Fm. with warm, hypersaline, Keuper-produced waters that were produced during the coeval Keuper diapirism. The mixing occurred in karstic caverns that were primarily stratiform at the base of the Novales Mb. where sulphate reducing bacteria produced sulphide for the Pb/ZnS from the sulphate of Keuper waters under strongly reducing conditions.

The timing of the mineralization/sparry dolomitization can be well constrained. Unlike the hypdiotopic dolomite, the sparry white dolomite at Reocin

is not only present in the mineralized karstic caverns of the Novales Mb. (Figs. 6-45 & 6-46) but is also present as a void-filling cement in the estuarine channel sands of the Lower Albian Las Penosas Fm. and has virtually totally replaced the Middle/Upper Albian Barcenaciones Fm. carbonates with high-Fe, sparry dolomite (Figs. 6-45, 6-46 & 7-1). Rare sphalerite has also been found within the Albian strata at Reocin (Vadala, 1981). The white, sparry dolomites of the Albian are texturally and geochemically identical to those found in the main mineralization horizons (Capa Sur etc.) at the base of the Aptian Novales Mb. (Fig. 6-79). However, there is no major mineralization or sparry dolomitization of the Cenomanian Bielba or Altamira Fms. (or younger strata). This strongly suggests that these minerals were introduced during the Lower Cenomanian (Bielba Fm. times) and does *not* link to a solely fault-related origin which would affect all the carbonates (such as the Middle/Upper Altamira Fm.) similarly. Further, spar/ore formation is clearly pre-Pyrenean deformation (Upper Eocene-Miocene) in age as it is folded in the Santillana syncline.

Meteoric recharge through the estuarine Las Penosas Fm. would provide reduced, Fe- and Mn-rich, meteoric-derived groundwaters into the Reocin Fm.. As these meteoric-derived groundwaters had to pass through both estuarine formations, the solutions generated would be enriched in Fe and Mn, accounting for the very high Fe and Mn of the sparry, baroque dolomites (lines 21 to 31 of Fig. 6-61). The solid solution between Fe and Mg in dolomites explains why such high Fe dolomites have low Mg values. The Ca concentrations are very similar to those of pre-existing dolomites and wall-rock corrosion (see below)/neomorphism could account for the Ca concentrations and source.

Other geochemical pointers support an origin, in part, from such reduced, meteoric-derived waters. The Sr values are exceptionally low (mean 33ppm), lower than the mixed-water hypdiotopic dolomite values and much less than even the equilibrium values of dolomites formed in seawater (about 500ppm, Veizer, 1983). This suggests precipitation from brackish/strongly-meteoric-influenced waters of such low Sr contents. Similar depletions occur with Cu (2ppm observed vs. 110-

300ppm in seawater equilibriated dolomite). Conversely massive enrichments are seen in surface-derived Mn (1170ppm observed vs. 1ppm predicted for sea-water equilibriated dolomite) and Fe (38,000ppm observed vs. 3-50ppm predicted). Such reductions in Sr and gains in Fe and Mn characterize the trace element geochemistry of meteoric diagenesis (Brand & Veizer, 1980; Bathurst, 1983).

The $\delta^{18}O_{PDB}$ are, surprisingly, only very slightly different from those of other dolomite generations (Figs. 6-59 and 6-60, -6.7‰ vs. -6.8‰). As the dominant controls on $\delta^{18}O_{dolomite}$ are the temperature of formation and the $\delta^{18}O$ of the formation waters, the most obvious explanation of the $\delta^{18}O_{dol}$ values is that they formed from similar isotopic waters to the other dolomites at similar temperatures, rather than a perfect covariance of these factors to mimic the original -7‰ value. Independent constraints can be put on the temperature of formation. Firstly, vitrinite reflectance of the organic material for the Reocin Fm. limestones, the hypdiotopic dolomites, the ZnS, the detrital dolomite in the Capa Sur ore body and the Albian Las Penosas Fm. all show very low values of 0.08 to 0.34 (n = 58), which indicate very low maximum burial temperatures of under 50°C (Vadala, 1981), and hence even lower temperatures in the Cenomanian. Further, the ore bodies/sparry dolomite were generated within just 200m of the Lower Albian land surface (Fig. 6-45) and so on any reasonable geothermal gradient, the temperature of the ore formation was only slightly elevated above the ambient (to about 30-50°C ?). Near-surface, very high-temperature pockets seem somewhat unlikely to the author.

Hence the ore/spar formed under low (30-50°C), near-surface temperatures. Such temperatures confirm that these dolomites formed from oxygen-light waters (cf. argument over xenotopic dolomites), and these are most likely to be the low $\delta^{18}O$ values of meteoric-derived waters. Petrographically, the massive white sparry dolomite appears to be a neomorphic replacement of an earlier carbonate, whereas the baroque dolomite is clearly a space-filling dolomite due to its euhedral terminations. The major, minor, trace and isotopic geochemistry of the sparry dolomites support an origin from hyposaline, meteoric-derived groundwaters and not

from strongly hypersaline waters as has been suggested for this baroque dolomite type by Radke & Mathis (1980). Similar doubts to the popular hypersaline model have been expressed by Morrow (1982a).

The details of the karst cavern internal sediment infill have been discussed above. But was there mixing in them? Such karst related ore/dolomite spar bodies could form from: 1) surface-derived solutions alone; 2) *in situ* dewatering of the carbonates; 3) a deep-basinal solution alone, or 4) a mixing of surface-derived and basinal solutions. To derive the Pb/ZnS from surface-derived solutions alone would have been a major problem as non/brackish-marine mudstones do not contain sufficient sulphate or sulphide to source the sulphur for the ore body. This source does supply the Fe and Mn to sparry dolomites and the Fe to the Fe-rich sphalerites. The close association of the ore and high Fe dolomites suggests that the surface could be the source for the Pb and Zn, though more research is needed on this. Support for the Albian estuarine muds as metal source comes from their high Zn (mean 250ppm Zn, n = 17; Vadala, 1981) and Pb (mean 85ppm, n = 17) contents. An *in situ* dewatering from the Reocin Fm. seems unlikely as this rock had already been lithified before sparry dolomite cementation. It also only contains about 10ppm Zn (Fig. 6-31), and so on mass balance calculations could not supply the quantity of Zn needed for mineralization.

A direct derivation from basinal brines is popular as an origin for Mississippi Valley Type (MVT) deposits (e.g. Anderson & Mcqueen, 1982). A metal source for the sparry dolomite/Pb-Zn sulphides from the dewatering of coeval deeper water shales of the Aptian (cf. Fig. 6-21) and their upward and basinmarginward migration seems unlikely as the distribution of the dolomite and Pb/ZnS ores increases towards the palaeohinterland (the S/SW) and not towards the offshore shales to the N/NE/E (Fig. 6-21). There are no local, Cretaceous igneous bodies in NW Santander to act as metal source.

A more obvious single "basinal" source are Keuper-derived hypersaline waters. Unlike our knowledge of many deep-seated sources, extensive solution mining (1.5 million tonnes/year at Polanco) has characterised these to be prolific,

warm (21-34°C), strongly hypersaline (0.5 to 7 times seawater salinity) waters dominated by dissolved NaCl (175,000ppm Cl^-) and moderate $CaSO_4$ contents (3,000ppm sulphate) derived from the evaporites in the Keuper (Chapter 3.4; Vadala, 1981). They also have a low Fe concentration (2ppm), and so are unlikely to be the source of Fe for the ferroan dolomites. They also have a neutral to slightly alkaline pH (7-9) and would be dedolomitizing due to Mg/Ca ratios of 0.1 to 0.4. Could these hypersaline brines have caused the mineralisation/sparry dolomitization by themselves? Firstly these waters have very low Zn (0.3 - 0.6ppm) or Pb (0.2 - 9ppm) concentrations and with the low Fe, are unlikely to bring in the necessary cations by themselves. These solutions have low S^{2-} concentrations, but local reduction of sulphate could allow these to be the sole sulphur source. A further problem is why only the Aptian strata was affected on a single-solution model. Directly above the Keuper are the synsedimentary collapse breccias of the basal Jurassic carbonates (Carniolas of Chapter 3.5) which would make excellent ground preparation for mineralization/dolomitization just above the Keuper source. These and other non-Apto/Albian carbonates are not affected though, making a precipitation of ore/dolomite from single-solution Keuper-derived waters an unlikely mechanism.

What of mixing Keuper-derived waters and surface-derived waters in the basal Novales Mb? This mechanism would be able to explain the meteoric-like geochemistry of the sparry dolomite which would precipitate during periods of relative meteoric-derived water enrichment. However, during periods of relative hypersaline, Keuper-water enrichment, the Pb/ZnS could precipitate (Anderson & Mcqueen, 1982) under strongly reduced hypersaline conditions. Such a mixing and dynamic variation in surface/Keuper water ratios is common in deep phreatic caverns and would immediately explain the interbedded/intimate intergrowth texture of sulphides and dolomite and the strong intercrystalline variation and intracrystalline zoning in Fe, Mg, Mn etc. in the dolomites and the ore. It would also explain the presence of the detrital dolomite as the boundary between dolomite precipitation and dissolution could be crossed repeatedly.

How were these solutions brought to the base of the Novales Mb? The surface-derived waters would be reduced and move down from the estuarine surface in the Lower Albian Las Penosas Fm. (Fig. 6-78), but would tend to move out laterally on encountering the impermeable marly limestones of the Rodezas Fm. The waters from the Keuper were probably produced by seismic pumping and then lateral outflow into the permeable caverns of the basal Novales Mb. The waters from the Keuper could come from compaction or (more probably) from even deeper sources (Hercynian basement?). Another possible source for the Keuper waters is from the Buntersandstone which would have acted as a palaeoaquifer with meteoric solutions derived from its exposure in the hinterland. This could scavenge Pb & Zn from feldspars in the Buntersandstone. Whatever their origin, these waters became strongly hypersaline by dissolving the Keuper evaporites and would transport sulphate in solution. The Lower Albian was the time of active, Keuper-related diapiric differential subsidence in the Santillana rim syncline (Chapter 8). To accommodate this differential subsidence, structurally necessary faults must have formed between the Keuper and the base of the Las Penosas Fm. These would provide ideal conduits for the hypersaline brines from the Keuper and this structural deformation would tend to enhance brine production. These could then move out into the caverns in the basal Novales Mb., but its low Mg/Ca ratio, severe pH (?) etc. would cause it to be corrosive and dissolve the dolomite wall-rock. This would explain the collapse breccias of wall-rock in the ore body and the general lack of sphelothem features in the ore body walls. This wall rock corrosion could also provide the Mg and Ca to the sparry dolomites.

The mixing of such hypersaline and hyposaline waters in a schizohaline environment is a natural location for dolomitization (Folk & Land, 1975). It was also enhanced by increased alkalinity and reduced sulphate concentrations linked to ore precipitation (see below), effects that are known to promote dolomitization (Baker & Kastner, 1981).

How was the sulphate reduced? The reduction of the Keuper-derived sulphate is unlikely to have occurred by inorganic equilibrium as this reaction is

very slow *indeed* at such near-surface temperatures and pressures (Berner, 1971). Also inorganic sulphide precipitation does not accord with the stable sulphur isotopes of the Reocin ore body (Adelhardt, 1968; Fig. 6-80). The $\delta^{34}S$ values for ZnS at Reocin run from -2‰ to +17‰ with a distinct increase in $\delta^{34}S$ to the west (Fig. 6-80). As Rye & Ohmoto (1974) have pointed out, it is not necessarily the absolute value of the $\delta^{34}S$ that indicates an inorganic or bacterial origin for the ore as this depends on the sulphate/sulphide ratio. However, under inorganic equilibrium, the difference between coexisting sulphides $\delta^{34}S$ values is inversely proportional to the absolute temperature. Adalhardt (1968) gives values of $\delta^{34}S_{ZnS} - \delta^{34}S_{PbS} = +4.4‰$ and +3.9‰. These correspond to temperatures of 160°C and 195°C respectively (method from Rye & Ohmoto, 1974). Such temperatures are clearly inconsistent with the known temperatures of ore body genesis (see above), and suggest inorganic equilibrium did not apply.

Under such near-surface (<85°C), clearly reducing conditions, obligate anaerobic sulphate-reducing bacteria such as *Desulforibrio* are very active and digest organic matter exothermically. Such fine-grained, humic organic matter has been seen in some dolomite generations (e.g. Fig. 6-66). It would be derived from an abundant carbon source in the estuarine black mudstones and is transported with the reduced, meteoric-derived waters. The key reaction is (Zobell, 1963):

$$RCH_2 + SO_4^{2-} \rightarrow H_2S + CO_3^{2-} + RO$$

where R is a complex alkyl radical. Such a reaction produces sulphide which will rapidly scavenge any metal ions to form metal sulphides such as ZnS or PbS. The increase in CO_3^{2-} is removed from the aqueous system by dolomitization. Both of these encourage the precipitation of dolomite. A source of carbon or sulphur from petroleum is unlikely as no dead oil or native sulphur has been found in the ore body. This would provide a local source of sulphide for ore precipitation.

The carbonate anion derived from such a bacterial reduction would be isotopically light (Hudson, 1977; Mattes and Mountjoy, 1980), but the final $\delta^{13}C$ of the sparry dolomite would be buffered by that derived by corrosion/dissolution of the heavier ($\delta^{13}C$ = + 3‰) wall-rock dolomite. The final sparry dolomite value

would depend on the rock/water ratio. As more of the bacterially-derived, very light $\delta^{13}C$ mixes with that of the wall-rock dolomites's carbon, a progressive mixing line will be set up towards lighter C values from a C value no heavier than the initial non-sparry dolomite value. This pattern is clearly seen in the isotopic values for the dolomites (Figs. 6-59 & 6-60). This will occur at relatively constant $\delta^{18}O$ values as neither the temperature or nature of the hyposaline groundwaters has altered. It is likely that the latest, pore-filling baroque dolomites are most Fe-rich and have the lowest $\delta^{13}C$ values (cf. Woronick & Land, 1985).

Hence a low-temperature (25-50°C ?) MVT deposit formed by the mixing of the hypersaline, Keuper-derived waters and reduced, hyposaline, surface-derived waters. The origin of the other mineralization zones is thought to be similar, though there has been some later remobilization of galena especially at Novales.

4) Fracture-related sparry dolomites - description

These are a latter stage of dolomitization. They are coarse, epigenetic white sparry to baroque dolomite that clearly follow fractures, veins or other tectonic features. Locally the fracture-related dolomite can have a halo (mm to about 1m thick) of xenotopic dolomite. This dolomite postulates the more massive spar mentioned above and is similar to that described by Choquette (1971) or Mattes & Mountjoy (1980). This dolomite has the most variable major and minor element concentrations of any dolomite generation (lines 32-36 of Fig. 6-61). It is generally Ca-rich due to strong buffering/corrosion of the volumetrically-large wall rock calcite and correspondingly Mg-poor. It shows a wide distribution of Fe and Mn values. It varies in whether it takes a turquoise blue stain with potassium ferricyanide/Alizurin Red S solution (Fig. 6-63B) and whether it shows a spectacular zoning under cathodoluminesence (Fig. 6-81). This dolomite generation contains the lightest $\delta^{18}O_{PDB}$ values (-7.7 ‰ to -8.1‰; Figs. 6-59 & 6-60) and typical sparry dolomite $\delta^{13}C_{PDB}$ values (+0.5 to 1.5‰).

4B) Fracture-related sparry dolomite - interpretation

This is a fracture infill dolomite that formed epigenetically. At localities such as Comillas it is connected with the Pyrenean fault/thrust structures and veined clasts of Novales Mb. dolomite have been found in the resedimented flysch of the Upper Eocene-Oligocene San Vicente Group (Chapter 11). The variability of Fe/Mn contents points to a complex, multiphase history to fracture infill, as does the cathodoluminesence zoning. It is difficult to say whether the more negative $\delta^{18}O$ values relate to a more negative meteoric input or to elevated temperature.

5) Stylolite-related dolomites

These are very locally developed and consist of hypdiotopic to euhedral crystals of dolomite that are typically 150-500μm wide. There is a haphazard neomorphic replacement of the calcitic matrix or spar by rather irregular, patchy, cloudy or clear dolomites that mimic stylolite topography and decrease in density away from the pressure-solution interface. As dolomite is more resistant to pressure solution than calcite, previously formed dolomites tend to accumulate as pressure-insoluble residue along stylolitic surfaces.

6.10 Conclusions on the Reocin Formation

The Upper Aptian/basal Albian Reocin Fm. was a low- to medium-energy, rudistid/benthic foraminiferal platform carbonate with a homoclinal ramp transition to deeper-water coeval marls. It was dominated by requienid/orbitolinid/miliolid lime mudstone/wackestones with local rudistid packstone banks, coral patch reefs and cross-bedded skeletal grainstones over palaeohighs. The Upper Aptian saw the initiation of Keuper-related diapirism with thinned Reocin Fm. over palaeohighs and thickened, micrite-rich sequences in the associated diapiric rim synclines. The key feature of the carbonate diagenesis is exposure to meteoric-derived waters from the subaerial horizon at its top. The initial cementation was synsedimentary marine phreatic micritization and isopachous, fringing, fibrous cementation. An

early generation of non-ferroan equant cement was then precipitated under active, meteoric conditions and was associated with leaching of HMC/aragonite or neomorphic transformation of HMC to LMC.

Massive invasion of slightly reduced, Fe-rich, meteoric-derived groundwater over palaeohighs generated extensive lenses of Novales Mb. sucrosic dolomite in the seawater/groundwater mixing zone. These surface-derived groundwaters became progressively more reduced and enriched in Fe and Mn and precipitated a dolomite light in $\delta^{18}O$. A more local phase of karstification brought in further estuarine-mud-derived groundwaters and precipitated sparry, baroque dolomite. In karstic caverns at the base of the Novales Mb. at Reocin these hyposaline groundwaters mixed with hypersaline, sulphate-rich, Keuper-derived waters and precipitated Pb/Zn sulphides by bacterial sulphate reduction.

CHAPTER 7 - THE ALBIAN LAS PENOSAS AND BARCENACIONES FORMATIONS: ESTUARINE/INNER SHELF TIDAL CLASTICS AND SHELFAL CARBONATES.

The Albian sequence in NW Santander is both complex and laterally variable. In general, it is represented by a lower transgressive siliciclastic formation (the Lower Albian Las Penosas Fm.; Fig. 1-9), deposited in a fluvial/estuarine/inner shelf environment, and an upper middle to outer shelf limestone (the Middle/Upper Albian Barcenaciones Fm.). The Albian strata show strong lateral thickness and facies variations (Fig. 7-1 : sections located on Fig. 7-2) which can be broadly linked with the period of distensional tectonics associated with the early phase of drifting in the Bay of Biscay. Relative to the rest of the Basco-Cantabrian basin, NW Santander preserves a very thin Albian sequence that varies from 20m to 347m, with an average of only 140m (Fig. 7-1). The NW Santander area also shows a major reversal of the Wealden/Aptian eastward tilting of this sub-basin (Figs. 5-2 & 6-1) to a westward tilt in the Albian (Fig. 7-1). This can be related to NW Santander acting as a "push-up" (sensu Mann et al., 1983) block with a thinned sequence that shows a reversed sense of tilting relative to the pre-Albian position because it is on the western, upthrown side of the synsedimentary fault - the Rio Miera Flexure (Fig. 7-65) which exists to the east of NW Santander. The other major, but more local (5kms wide rather than 50kms) tectonism affecting the NW Santander Albian is the continued growth of the Keuper diapirs that had largely started in the Upper Aptian.

7.1 Stratigraphy and dating of the Albian sequence in NW Santander

The stratigraphy of the Albian is one of the most complex in the column due to the strong lateral facies and thickness changes as well as the confusion of lithostratigraphic and biostratigraphic units. A summary of the stratigraphy used by workers is shown in Fig. 7-3. An accurate stratigraphy for the Albian was first set up by Mengand (1920), who found Albian *Knemiceras* ammonites in the lower Barcenaciones Fm. Both he and Rat (1959 and in Ciry et al., 1967) considered the Albian of NW Santander as "Supra-Urgonian". However, later authors (Garcia

Mondejar, 1979a; Pascal, 1985) have considered the Albian to be part of the Urgonian. All authors agree that the Lower Cenomanian Bielba Fm. (Fig. 7-3) is "Supra-Urgonian". Mengaud and Rat defined the term Supra-Urgonian as the major siliciclastic input that destroyed the Urgonian platform limestones. In NW Santander there is major, tidally-dominated, siliciclastic sequence in both the Lower Albian Las Penosas Fm. (Fig. 7-3) and the Lower Cenomanian Bielba Fm.. However, rudistid shelfal limestones do occur in the upper Torriente Member of the Middle/Upper Albian Barcenaciones Fm. Differences between the Barcenaciones Fm. and typical Urgonian platform limestones are the development of a marly lower Miradorio Member and extensive tidal calcarenitic bars. Hence, it differs from the type Urgonian from around Orgon in Provence (Rat & Pascal, 1979) in being younger (Albian as opposed to Aptian/Neocomian), in having major siliciclastic input and high-energy calcarenites (compared to the Reocin Fm. which is typically "Urgonian"). Although these Albian facies can be correlated across the Basco-Cantabrian basin (Garcia Mondejar, 1979a; Pascal, 1985), it does seem more logical to include the Albian in the Supra-Urgonian in NW Santander. The author (Hines, 1985a) has proposed the term "Santander Group" (Fig. 7-3) for the Supra-Urgonian in NW Santander.

Feuillee (1967, Fig. 7-3) gave zonal ages to the Upper Albian/Cenomanian sequence of NW Santander. IGME (a, c, e, g, i) mapped the Albian of NW Santander lithostratigraphically as a lower C'_{1e} (= Las Penosas Fm.) and an upper C^{2a}_{1e} (= Barcenaciones Fm.). Garcia Mondejar & Pujalte (1981) subdivided the N634 road section at Barcenaciones (Section 19 on Fig. 7-1) into the major clastic/carbonate units (e.g. A_4 of Fig. 7-3). Tentative correlation of these with equivalents south of the Escudo de Cabuerniga is given in Figs. 5-8, 7-3 & 7-56. Garcia Mondejar et al., (1982) first suggested formalised lithostratigraphic names for the units mapped by IGME.

Pascal (1985) suggested that there was a major disconformity in the Lower/Middle Albian between the top of the Aptian Reocin Fm. (= his 2eme Systeme Biosedimentaire) and the Upper Albian (= his 4eme Systeme Biosedimentaire). The

present author (1985) essentially followed the lithostratigraphic nomenclature of Garcia Mondejar et al., (1982) and erroneously introduced the term Suances Fm. that ought to be abandoned. In this work I have subdivided the Barcenaciones Fm. into three members. At its base there is a local, high-energy calcarenite that interdigitates with the Las Penosas Fm. and Miradorio Mb. Following the nomenclature of Mengaud (1920) and Pascal (1985), it is termed the Santa Catalina Member. This is overlain by the low-energy marly limestone/marl alternation of the Miradorio Mb.. This corresponds directly with the "Marnes a Myaces" of Mengaud (1920) and Pascal (1985) and C_3 of Garcia Mondejar & Pujalte (1981). It is this member that has yielded the *Knemiceras* ammonites. The upper part of the Barcenaciones Fm. is a relatively massive, medium- to high-energy limestone that developed during a slight relative sea-level fall relative to the Miradorio Mb. and is here termed the Torriente Member. It is capped by the *Caprina choffati* beds. The Torriente Mb. corresponds with the " Gres á Orbitolines et calcaires coquilliers" of Feuillee (1967) and C_4 of Garcia Mondejar & Pujalte (1981). Around the junction of the Miradorio and Torriente Mbs. there is a correlatable reddish, mottled, fine sandstone bed (1-20m), termed the "Capa roja" bed by the author. It corresponds with the "Couche rouge" of Mengaud (1920), Rat (1959), Pascal (1985) and A_4 by Garcia Mondejar & Pujalte (1981).

The exact age of the Albian sequence is somewhat controversial. All authors to date agree that the top *Caprina choffati* beds are uppermost Albian (= Vraconian, = *S. dispar* Zone; Mengaud. 1920; Feuillee, 1967; Garcia Mondejar & Pujalte, 1981; Pascal, 1985). The age of the Torriente Member is considered to be Upper Albian. Feuillee considered the top of this member to belong to the *S. dispar/M. inflatum* Zones. Pascal (1985) considered the entire Barcenaciones Fm. to fall into the Upper Albian on the presence of *Caprina choffati* (at the top), *Knemiceras uhligi* Choffat, *Neorbitolinopsis conulus, Dissocladella* & *Neoiraqia convexa*, which are all Upper/Uppermost Albian fauna. Hence, the Torriente Mb. is of this age. Along with the Knemicerids found by Mengaud (1920), I have found *Knemiceras flexiloculosum* Renz non Basse (kind identification by H.G. Owen, British

Museum) in the basal parts of the Miradorio Mb. of the Miradorio section, which is dated as early *Lyelliceras lyelli* Subzone of the early Middle Albian *Hoplites dentatus* Zone (Renz, 1982). If Mengaud's (1920) identifications are correct then the *Cnemiceras (Placenticerfas) Ebrayi* (= *Parengonoceras ebrayi* (de Lorrol)) and the *Sonneratra* cf. *cleon* of the Miradorio Member would fall into the late Lower Albian *Douvilleiceras mammillatum* Zone (the former ammonite into the *Cleoniceras floridum* Subzone of this Zone). This would make the age of the Miradorio Member (at least in its type area at Miradorio, east of Comillas) as upper(most) Lower Albian *D. mammillatum* Zone to (at least) early Upper Albian (*K. uhligi* Choffat of the *varicosum* Subzone of the *D. cristatum* Zone). The age of the Barcenaciones Fm. is hence from the uppermost Lower Albian to the uppermost Upper Albian, which is in remarkable agreement with the dating by Mengaud (1920) or IGME (a, c). The presence of *K. flexiloculosum, K. (Placenticeras) Ebrayi* and *Sonneratia* cf. *cleon* makes Pascal's (1985) claim for a hiatus in the Lower/Middle Albian seem very doubtful.

The age of the Las Penosas Fm. is difficult to define precisely as it has not yielded any biostratigraphically useful fauna. Its local interdigitation with the Barcenaciones Fm. but sharp base suggest that it is closer in age to the Barcenaciones Fm. than to the Reocin Fm. If the Reocin Fm. does extend up to the basal Albian (see Chapter 6.3), then the Las Penosas Fm. could well represent the middle part of the Lower Albian. Its base most probably represents a minor(?) hiatus and its top is no younger than *Cleoniceras floridum* Subzone of the *D. mammillatum* Zone (uppermost Lower Albian).

7.2 Definition and type section of the Las Penosas Formation

Definition: The Las Penosas Fm. is made of siliciclastic mudstones, siltstones and sandstones that lie between the top of the Reocin Fm. (Comillas or Sierra Mbs.) and the base of the Barcenaciones Fm. carbonates. It commonly contains pyrite, carbonaceous matter and has virtually no macrofauna. It has strong lateral thickness variation (Fig. 7-1) from 0 to 194m, with a mean of 70m (Fig. 7-1).

Type section: The type section is from the west side of the Miradorio headland, 1km east of Comillas (from 55269763 to 55279768). It corresponds to bed 3 ("Gres a lignites ...") of Fig. 33 of Mengaud (1920). Here the type section runs from the top of the Comillas Mb. (493m on Section 5/29) in the slightly overturned siliciclastics in the south to the central fault of this section (Fig. 6-14) at 552m on Section 5/41. The section continues north of the fault in the right-way-up strata from Section 6/1, 0m to Section 6/12, 59.8m. There is then an interdigitation of the Santa Catalina Mb. of the Barcenaciones Fm. between 59.8m and 64.8m on Section 6. This Santa Catalina Mb. is itself overlain by an intertongue of the Las Penosas Fm. (64.8m to 74.5m). It is then overlain by the Miradorio Mb.

7.3 Lithology and facies of the Las Penosas Formation

The Las Penosas Fm. grades from fluvial sandstones at the base to estuarine deposits and then inner shelf siliciclastics at the top. Its facies will be described in approximately stratigraphic order, representing progressively deeper water facies. A general facies model for the Las Penosas Fm. is shown in Fig. 7-64.

1) Facies 1: Sandy river deposits

Description :These characterise the basal parts of the Las Penosas Fm. in the Comillas area. Here the vertical sections 8, 9, 10 and 11 of Fig. 7-1 are close enough spaced to allow the lateral variation of individual facies to be seen. This shows that the sediments are made up of a 15-20m thick, coarse sandstone sheet with rare, thin (dm-m) mudstones (Figs. 7-5 & 7-6). The internal structure of the sandstone sheet is dominated by stacked, clean, trough cross-bedded sands (Figs. 7-7 & 7-8) with scouring at their bases that combine to produce multi-metre thick sand packages (Section 5/29, 493 to 508m). These packages show local, basal lags of quartz granules and non-armoured mud flakes which are overlain by trough cross-bedded sandstones in lenticular cosets. These sands do not develop significant

mud drapes. The sand is coarse grained and feldspathic which is unusual for the Las Penosas Fm..

Interpretation : The stacking of the multi-metre packages in thinning-upward sequences that show a slight degree of fining-upwards suggests that these sands represent multiple stacked fluvial channel sands, that have formed a sand sheet due to repeated lateral channel migration and/or rapid arenaceous input during sea-level low-stand. They have no extensive mud-draping, a strongly unimodal palaeocurrent direction (mean of N 022° ± 11°, n = 20, see Section 10/1, 0 to 11m), subaerial rootlets locally (Section 5/33, 511m; Fig. 7-9), a generally constant, coarse grain size in the channel fill and no heterolithic sand/mud interbedding (dm-scale). These factors suggest that these are fluvially-dominated channellised sands that were essentially non-marine.

2) Facies 2: Storm (?) sheet sands

The next 20-30m of the Miradorio section (Section 5/33, 512m to 535m) generally coarsens upwards and is mud-dominated with dm-m scale sandstone beds. The sands are generally slightly muddy, sharp-based and topped, are often ripple- or locally planar-laminated and have mud-draping of the laminations. Most beds are laterally planar on an outcrop scale (1-10m), but are demonstrably lenticular on a larger (100m) scale (Fig. 7-5). These sandstones can be subdivided into metre-scale coarsening-upwards sequences, fining-upwards sequences and sheet sandstones.

Coarsening-upward sequences - description : The coarsening-upward sequences (e.g. Section 5/33, 512-513m) are of impure sands alternating with black, carbonaceous mudstone. The lower, thinner bedded sands (cm-scale) are commonly wavy bedded. The upper, thicker-bedded sandstones are flaser, parallel and cross bedded. They are usually rich in woody carbonaceous fragments 0.2-20cm long and can be iron-stained with pyrite or have sulphurous coatings.

Coarsening-upward sequences - interpretation : These sequences are thought to represent storm-derived (?), sheet sands that flooded into a mud-dominated lower

supratidal/upper intertidal flat. The upper supratidal flat with its extensive vegetation, rootlet beds etc. does not seem to be well developed here.

Fining-upward sequences - description : They are generally sharp and erosionally-based, have small-scale cross-bedding at the base and are generally less than 1m thick (Fig. 7-10). The tops of the sands are generally rippled.

Fining-upward sequences - interpretation : They are thought to represent small tidal drainage channels on the intertidal/lower supratidal flat.

Sheet sands - description : Muddy sheet sands less than a metre thick also occur. These are extensively mud draped with flaser or wavy bedding locally (Fig. 7-11). They are generally sharp based and topped with no vertical change in mean grain size. They can be totally bioturbated as in Fig. 7-12 (located at Section 5/35, 522m).

Sheet sands - interpretation : They are again thought to be arenaceous storm input into the peritidal flat. The possibility of them being washover fans in a back-barrier environment is unlikely because of their muddy nature (lack of clean beach sand), association with carbonaceous fragments, lack of transported open-marine shells and repeated mud/flaser draping internally (indicating a lack of sudden, storm injection of sand). The typical landward-dipping planar lamination of barrier washover fans has also never been found in this formation. This makes a barrier island model for the Las Penosas Fm. generally unlikely. Rare, cleaner sands could represent washovers from local spits. These flood sands are clearly ultimately derived from the fluvial source upstream and do not come from a marine shelf, which was an area of carbonate sedimentation at the time.

3) Facies 3: Convoluted sheet-flood sands

Description : These are generally about a metre thick and are interbedded in black mudstone. They consist of muddy fine sandstone that shows extensive soft-sediment deformation in the form of slump folds, sand-ball horizons, contortion around carbonaceous fragments etc. (Fig. 7-13).

Interpretation : These features indicate rapid sedimentation into an area of unconsolidated mud. Their location directly beneath the estuarine channel on

Section 11/11, 53-55m in the estuarine sequence suggests they formed in a peritidal flat.

4) Facies 4: Peritidal mudstone

Description : These black, carbonaceous mudstones are relatively unconsolidated and dominate the lower part of the Las Penosas Fm. above the basal fluvial sheet sandstones. The mud is black and rich in carbonaceous fragments that are 0.5-20cm in length. Locally, thin, starved-ripple wavy bedded sands occur within the muds with 1-10% sand (e.g. Section 5/34, 519-521m). In sequences over 2m thick these ripples can be entirely unbioturbated, probably linked to an environment hostile to colonization such as an intermittently submerged, generally brackish water, soft-bottomed peritidal flat whose salinity varied on scales of hours to years. In other areas the peritidal muds are totally bioturbated. These muds lack any marine macrofauna (or any macrofauna) but have yielded arenaceous foraminifera (IGMEa) which occur in brackish-water (Ramirez del Pozo, 1971a).

Interpretation : These are interpreted as estuarine peritidal mudstones. This indicates a mud-rich upper estuary with a turbidity maximum in the upper estuary. This is due to the intense flocculation of muds as they first encounter saline waters. On the tidal Klang River delta (Coleman et al., 1970), the macrofauna do not colonize the tidal flats due to the very high sedimentation rate and unconsolidated nature of the surface there. This would explain the lack of bioturbation in some of the estuarine muds. The thinner mudstones (up to 2-3m) are interpreted as supratidal and upper intertidal flat deposits, whereas the very thick sequences (up to 5m) were deposited in lagoonal embayments.

5) Facies 5: Brown palaeosol

Description : These occur about 10m above the basal fluvial sandstones. They are extensively bioturbated, brown-weathering, powdery clays with a high degree of oxidation. They have developed abundant, late-diagenetic gypsum crystals (e.g. Section 5/34, 518m). Local rootlets occur within them.

Interpretation : They are interpreted as ancient upper supratidal estuarine palaeosols.

6) Facies 6: Inter/supratidal dendritic drainage channels

Description : These occur in the Comillas area (Sections 8-11, Fig. 7-1) within the interbedded sandstone/mudstone sequences of the peritidal flat [Facies 4] (Fig. 7-14, Section 5/35, 524m). They consist of soft-sediment deformed blocks of interbedded sandstone/mudstone that have slumped into small, channel-like depressions generally 20-50cm deep and 50-200cm wide.

Interpretation : Similar structures have been described from dendritic channels (= runoff channels) from the North Sea flats by Weimer et al. (1982) and in Pleistocene estuaries (Clifton, 1983) where they formed by side-wall collapse of minor channels rather than the lateral migration of deeper tidal creek channels.

7) Facies 7: Distributary estuarine channels

Description : These are major, heterolithic channels that are 2m to 11m thick, with an average of 6m. The best example is from the Punta Calderon section (Section 11/12, 56 to 66m; Fig. 7-15). They are characterized by a sharp, locally erosional base cut into underlying peritidal mudstones (Section 11/12, 56m). A basal lag usually exists of cm-sized carbonaceous fragments and armoured ironstone conglomerate mudflakes (derived from erosion of the cemented supratidal flat during lateral migration of the channel). The lower metres of the channel are dominated by 50cm-scale, trough cross-bedded, medium to fine grained, unbioturbated sandstones in lenticular cosets. These cross-beds show local mud-draping. Only very locally is herringbone cross-bedding found in the lower parts of these channels (Fig. 7-20). The upper third of the channel is an alternation on a cm to dm-scale of mud-draped/flaser bedded sandstone and black carbonaceous mudstone. The 6m sequence at Punta Calderon fines upwards. The lower part of this 6 metres (Fig. 7-16) has decimetre scale, flaser bedded sandstone interbeds with wave and current ripple tops. These are interbedded with lenticular bedded mudstone with starved sand ripples. The uppermost 3m of the channel is progressively finer-grained and increasingly bioturbated towards its top (Fig. 7-17).

The palaeocurrents of these channels show a reasonably wide dispersion with numerous flow directions running E/W (perpendicular to the offshore direction

of N/NE). This E/W flow is due to the strong meandering and looping of the estuarine channels in the supra/intertidal area (Sections 8, 9 & 10 of Fig. 7-1; cf. Fig. 7-64). A rose diagram of the palaeocurrents in these channels is shown in Fig. 7-34. It is this very meandering and directional variation of estuarine channels that allows them to migrate laterally and erode the inter-supratidal flat deposits during relative sea-level rise (Clifton, 1982). This is especially true if the overall sea-level rise is punctuated by minor stillstands. This gives the estuarine channels a high preservation potential in this part of the estuary (Fig. 7-5).

Interpretation : These fining-upward sequences are interpreted as distributary estuarine channels. The lower, arenaceous part of the channel sequences interpreted as the active channel infill. The upper mixed, arenaceous/argillaceous part is thought to represent the lower intertidal/upper subtidal flat deposits that laterally prograded over the active estuarine channel infill as the channel meandered or was abandoned.

These clastic channels bear a close resemblance to the mesotidal, lower-estuarine channels of the Rhine : the Haringvliet (Terwindt, 1971) or the Oude Mass (Van Beek and Koster, 1972). The distributary estuarine channels of NW Santander do not develop thick (several metres plus) sequences of rippled/flaser or parallel bedded sands between the small sand waves of the lower channel fill and the interbedded sandstone/mudstone. This implies that extensive, lower intertidal sand flats were not developed here, only mixed and mudflats of the intertidal. Such sand flats are considered to be one of the most diagnostic features of macrotidal as against mesotidal estuaries (Klein, 1970; Dalrymple et al., 1985) and so the Lower Albian tidal range is unlikely to be macrotidal. The presence of extensive tidal channels and tidal sand waves and lack of wave-related barriers suggests that the range was not microtidal, leaving mesotidal as the most likely tidal range in the Lower Albian.

8) Facies 8: Epsilon-cross-bedded intertidal creek channels

Description : Epsilon cross-bedded muddy sandstones occur as highly lenticular channel deposits (Section 9 of Fig. 7-1). They have a basal lag of carbonaceous fragments and pyrite nodules. This is overlain by a 1.5m to 2.2m thick channel sequence that shows channel-wide epsilon cross-bedding with couplets of moderately bioturbated muddy sandstone alternating with mudstone on a 5cm scale (Fig. 7-18). The channel is at least 30m wide. The sandstone is mud-rich, moderately bioturbated and is rich in carbonaceous fragments. The direction of epsilon cross-bedding is perpendicular to the mean offshore channel direction of N 020°.

Interpretation : These are interpreted as the products of lateral migration of the inner point bar of an estuarine tidal creek channel. The alternation of sandstone/mudstone in this channel reflects the storm-winnowed lag of ebb/flood tidal sands alternating with suspended load mudstones that have numerous thin layers of carbonaceous detritus.

Intertidal creek channels have been used to produce an estimate of the palaeotidal range (Klein, 1971). As neither the subtidal cross-bedding sediments beneath, nor the rootlet/subaerial-exposed supratidal sediments above are present, any estimate based on this example is only a minimum. Although the syndepositional basin subsidence is negligible, the post-depositional compaction of the mudstones is not. Estimating the channel to contain 35% mud, then a *minimum* estimate for the palaeotidal range, based on the observed thickness of 1.5m to 2.2m and assuming a compactional thinning of the muds by 50%, would be 2m to 3m. Although decompaction makes the exact values somewhat questionable, this channel clearly suggests a mesotidal range, as did Facies 7.

9) Facies 9: Peritidal flats

Description : The peritidal flat deposits are characterized by an alternation of flaser bedded, medium- to fine-grained sandstones and wavy bedded mudstones on a millimetre to decimetre scale (Fig. 7-22). In a number of cases there is a clear vertical sequence on a 1.5 to 2.5m scale that grades up from flaser bedded sands with abundant mud draping of ripples to wavy bedded sands and muds (Fig. 7-22). The top is generally made of lenticular bedded mudstone with starved ripple

sandstone laminae on a centimetre to millimetre scale. The top of this sequence can be slightly/moderately bioturbated.

Interpretation : This sequence is interpreted as an aggrading sand flat/mixed flat/mud flat sequence (Reineck & Singh, 1980). Much of the sand is introduced onto the flat during storm floods of the estuarine/fluvial channels and is then reworked by wave action. The peritidal flat sequence can be more complex and included small tidal gullies with heterolithic cross-bedded sands (Fig. 7-23). Abundant millimetre to centimetre-graded woody carbonaceous detritus also drapes the sands. This is derived either from fluvial input from the humid, subtropical hinterland or from erosion of peaty supratidal flats (Buller & Green, 1976). Abundant fetal pellets occur as drapes. The total lack of rootlet structures in even the uppermost intertidal flat sediments points to a largely unvegetated flat.

10) Facies 10: Ironstone conglomeratic sandstones

Description : These occur towards the top of the Las Penosas Fm. (Sections 8, 9 & 10 of Fig. 7-1). They are matrix-supported ironstone conglomerates in a muddy sand matrix and the clasts are 0.5-7cm in diameter (Fig. 7-24). The mud chips are often armoured and are probably originally derived from channel erosion of the lower supratidal flats.

Interpretation : Their occurrence towards the top of Las Penosas Fm. suggests that they could be a transgressive lag preserved by shoreface retreat as waves destroyed beaches on local barriers, mouth bars etc.. Such local transgressive lags between the estuarine sequence and the fully marine sequence above have been noted by Swift (1968), who termed them the "upper ravinement".

11) Facies 11: Estuarine mouth bars

Description : These occur above the intertidal/estuarine channel sequence of this formation but below the inner shelf tidal sand waves. The best exposed example is on the Miradorio section (Section 6/1, 0 to 10m). It is a coarsening-upward, thickening-upward sequence (Fig. 7-25) that grades up from a lower part of interbedded laminated/cross-laminated sandstone and wavy/lenticular bedded mudstones (Section 6/1, 0-3m; Fig. 7-26) to cross-bedded sandstones above. The

upper part of the sequence is dominated by parallel lamination and trough cross-bedded, metre-bedded sandstones, which have a similar palaeocurrent direction to those below. The top of the sequences locally has an abraded, reworked marine fauna of orbitolinids, echinoderms and molluscan shells in a slightly coarser, parallel and trough cross-bedded sandstone. The sequences are also associated with major, land-derived carbonaceous logs which are up to 80cm in diameter and 3m long (Fig. 7-29). The top of this bed is truncated by well developed wave ripples (Fig. 7-28). A rose diagram for the palaeocurrents from the mouth bar and tidal sand waves (Fig. 7-35) shows the predominance of the WNW-directed currents with a minor, bipolar, tidal reversal current trending ESE.

Interpretation : These are interpreted as estuarine mouth bars. From the general orientation of the basin (Fig. 7-65) and the direction of the hinterland channels running offshore towards NNW, the WNW and ESE directions are seen as shore-parallel tidal currents of the inner shelf, and as the WNW flows out of the Albian Basco-Cantabrian basin (Fig. 7-65), it is interpreted as an offshore ebb current. A remarkably similar change from shore-perpendicular ebb and flood tidal currents in the estuary to shore-parallel inner shelf tidal/wave currents has been described by Terwindt (1973) from the Recent estuarine mouths of SW Holland. He also noted that the inner estuarine sediments only go down to about 15m water depth. The cross-lamination in the lower parts of the mouth bar shows a dominantly WNW direction, reflecting the parallel to shore, stronger tidal ebb current reworking the mouth bar (Section 6/1, 0-5m). Abundant starved ripples are present in the interbedded lenticular bedded mudstones of the lower part of the mouth bar sequence (Fig. 7-27).

This sequence bears a very strong resemblance to estuarine mouth bar sequences described from Recent estuaries/tidally-dominated deltas (e.g. Colorado river delta - Meckel, 1975; Mahakam delta - Allen et al., 1979). Such mouth bars may have beach/spit sequences at their tops, but here these have been destroyed during the Albian transgression. It is probable that the wave reworked top and offshore skeleta are all that remains of the former subaerial beach systems that

capped the mouth bar at the termination of a major estuarine channel. Another possibility is that this sequence represents a barrier bar that has been truncated at the top by transgressive shoreface retreat. However, lack of a basal erosion surface (over a potential lagoon) and, more importantly, the extreme lateral discontinuity of this sequence (Sections 8, 9 & 10 of Fig. 7-1) suggest it was a lobe-like mouth bar rather than a laterally continuous beach/barrier system. A further difficulty with a barrier island model punctuated by tidal inlets (Hayes, 1980) is the lack of landward-directed (SSW) flood-tidal inlets and the presence of major fluvially-dominated river channels at the base of the formation. A system of seaward-directed tidal sandwaves (NNW) and wave-related landward migrating swash bars (SSW), characteristic of tidal deltas (Hayes, 1975), has again not been found in Santander.

12) Facies 12: Inshore tidal sand bar

Description : This facies is dominated by sheets of cross-bedded and parallel-laminated sandstones often with stringers of ironstone conglomerates or carbonaceous fragments 1-8mm long. These are interbedded with heterolithic, decimetre-scale interbedded cross-laminated/flaser bedded sandstone or wavy bedded mudstone (e.g. Section 6/8, 16-18m; Fig. 7-30). The dominant currents are to the WNW with a significant number of crossbeds in the opposite direction (ESE) (e.g. Section 6/7, 30-43m), testifying to the tidal domination of these currents. Further support for tidal domination comes from the presence of herringbone cross-lamination in the heterolithic interbedded facies (Fig. 7-31).

Interpretation : This facies is thought to be the product of shore-parallel sandstone bars and sheets that dominate the area just beyond the mouth bar of major, sand-rich estuaries. As they form slightly above the estuarine mouth bars (e.g. Section 6/7, 30-43m), in a generally transgressive sequence they would represent deeper-water bars. In this sequence clear vertical trends in grain size or bed thickness are hard to identify. A similar system with sand waves and a sand-dominated interbar area is known from the Klang River delta (Coleman et al., 1970; Coleman & Wright, 1975).

13) Facies 13: Inner shelf interbar mud facies

Description : It is dominated by black silty clay that may be completely bioturbated. It is locally lenticular bedded with mm to cm-thick, starved-ripple, fine sand laminae which only show a slight degree of bioturbation. Soft-sediment deformation occurs locally. A good example of this facies is Section 6/10, 49.5 to 52m.

Interpretation :This is a mud-dominated, low-energy facies that occurs between the tidal sand waves and bars of the inner shelf (Section 8 of Fig. 7-1). The thin bedded sand laminae are interpreted as storm wave sands and are not dissimilar from those described from the Patrocinio Fm. of Chapter 5.3. Extensive soft-sediment deformation and vertical sand balls in some of the thicker (dm) storm sands is due to very rapid deposition (e.g. Section 6/11, 53.4m). Similar mud-dominated, interbedded sandstone/mudstone sequences are well known from estuary-related, intersandbar areas (Stride, 1982; Stubblefield et al., 1984).

14) Facies 14: Inner shelf tidal sand wave

Description : A well-exposed example of this facies occurs on the Miradorio section (Section 6/15, 70-74.5m; Fig. 7-32). It is 4.5m thick and is strongly lenticular on a 500m scale in the direction parallel to tidal flow direction of WNW. It dominantly consists of trough cross-bedded sandstones with extensive mud draping in lenticular cosets 10-40cm thick. In general, the palaeocurrent direction is relatively unidirectional, being oriented towards the WNW (the parallel-to-shore dominant tidal current direction). A number of the cross-beds are capped by cross-lamination which is only rarely reversed in direction relative to the trough cross-beds. The trough cross-beds show mud-draping, sigmoidal cross-bedding with down coset decrease in angle of repose, local double drapes, rare herringbone cross-bedding, numerous reactivation surfaces and rare hanging set boundaries. At the top of the sequence is a current rippled sandstone. Both the base and top of the sequence are sharp and it is overlain by the open marine marls of the Miradorio Mb. (Barcenaciones Fm.) with *Rhizocorallium*, *Thalassinoides* and *Knemiceras* ammonites together with planktonic and benthic foraminifera. Another

sequence occurs between 59.8m and 69.8m on Section 6/12 at Miradorio (Fig. 7-33). It is a mixed quartz-sandy calcarenite with moderate bioturbation towards its top (65-68m).

Interpretation : The sequence shows a remarkably close resemblance to the internal structures described from ancient tidal sand waves (de Raaf & Boersma, 1971; Levell, 1980; Stride, 1982, amongst others). The structure are very similar to the type V sand wave of Allen (1980) which has considerable but not excessive sand input and a moderately asymmetric tidal ellipse. Here the net bed transport is too small to create metre-scale, so-called "sand wave" slopes, steep enough for flow separation (Allen, 1980). Winnowing in the interbar area creates a basal erosion surface (E_1 of Allen, 1980), accounting for the sharp base to these sands. The internal structure lacks large foresets or bottomsets and consists of gently inclined, several-decimetre-spaced, second order (E_2) erosion surfaces with internal cross-lamination and cross-bedding (dipping in master bedding direction?) These represent small sand waves (= megaripples) that migrated over the large sand wave in the direction of dominant tidal current transport (WNW; Stride, 1982).

Very similar internal structures of translational sand waves occur in the intertidal zone of the Bay of Fundy (Dalrymple, 1984). Such detailed observation is not available as yet from the subtidal inner shelf sand waves, though Houbolt (1968), shows that small-scale cross-stratification is the dominant internal structure of these waves in the North Sea as it is here. The tops of the sand waves too show local rippling, which is probably linked to the slight bevelling of the sand wave during the moribunding stage before complete burial in mud.

7.4 Lateral variation in the Las Penosas Formation

Remarkable lateral variation in both thickness and lithology characterises this formation. The most northwesterly section at Punta Ninera (Section 1 of Fig. 7-1) shows no development of the Las Penosas Fm. lithology between the Gargasian Comillas Mb. limestone/marl alternation and the calcarenitic tidal sand bars of the Santa Catalina Mb. (basal Barcenaciones Fm.). This section

has been transported southwards in the hangingwall block of the Jerra Thrusts (Figs. 2-3 & 2-4, Chapter 2.2.1, Chapter 11.3) and so the Las Penosas Fm. was absent north of the present Punta Ninera. During the Lower Albian, the present offshore area N/NE of NW Santander did not receive the major buildout of the Las Penosas Fm. siliciclastics from the south (Fig. 7-65; Boillot, 1979a).

A relatively uniform (53-100m) autochthonous development of the Las Penosas Fm. occurs on the northern flank of the Escudo de Cabuerniga west of the Bustriguado/San Vicente fault (Sections 2 to 5 of Fig. 7-1). Rare channels are seen near the base of the formation, but very poor outcrop prevents more detailed work. Around Comillas, closely spaced coastal sections allow a detailed lateral correlation to be built up. At the base of Sections 7 of Fig. 7-1 (= measured Section 10/1), 8 of Fig. 7-1 (= measured Section 5/1 & 6/1) and 10 of Fig. 7-1, there is a laterally correlatable (over 2.5km) fluvial sheet sand. In the muddy supra/intertidal estuarine sequence of Section 8 of Fig. 7-1, the sequence is dominated by storm sands of the muddy upper estuary. In this sequence, direct correlation of beds is virtually impossible on a 500m or even 100m scale (e.g. correlation of beds between Sections 9 and 10 of Fig. 7-1). Such a lack of correlation is typical of estuarine sequences with their extreme lateral heterogeneity. This is especially true of transgressive estuaries where reworking by estuarine channels etc. during the transgression disrupts lateral correlation.

Higher in Section 8, the estuarine mouth bar is laterally equivalent to the much muddier estuarine channel sequence just 500m to the east in Sections 9 & 10. This reflects either the random development of an estuarine mouth bar around Comillas (Fig. 7-36) or a sand-rich sequence developed over the diapiric Comillas palaeohigh. Similar sand-rich are developed over Texan diapiric palaeohighs (Seni and Jackson, 1983).

There is rapid thickening of the Las Penosas Fm. towards the Cobreces diapiric palaeolow rim syncline (Fig. 7-36). The Trasierra section (12 of Fig. 7-1) preserves the thickest Las Penosas Fm. in the entire area (194m). The formation then thins east towards the Orena diapiric palaeohigh (Figs. 2-4, 6-16 &

7-36; Sections 15, 16 & 17 of Fig. 7-1). Over the Orena high, Section 16 (31m) shows only the peritidal mud with storm sands overlain by a distributary estuarine channel. This reflects the low preservation potential of the estuarine mouth bar sequence over palaeohigh structures. The Las Penosas Fm. then thickens to the east into the Santillana palaeolow rim syncline (Sections 18 & 19 of Fig. 7-1). The section at Reocin (Section 20 of Fig. 7-1) has ankeritic cements to the muddy estuarine channels of the Las Penosas Fm. and thick estuarine muds. The Miengo palaeohigh shows a remarkable thinning of this formation to zero or a few metres at most (Sections 21-23 of Fig. 7-1). A thickened sequence is developed in the Camargo palaeolow (Sections 24 & 25) is followed by a zero thickness over the Santander diapiric palaeohigh (Section 26 of Fig. 7-1).

Hence, on a 5km scale, the lateral thickness of this formation is strongly controlled by the active diapirism of the early drifting/final rifting stage of the opening of the Bay of Biscay. However, on a 50km scale, the NW Santander sub-basin shows a general thinning to the east up to the Rio Miera flexure (Figs. 7-36 & 7-65; Sections 1-26 of Fig. 7-1). This is the reverse of the general Aptian thickening of the sequence eastwards towards the centre of the Basco-Cantabrian basin (Figs. 5-2 & 6-1). This reversal in the tilting of the sub-basin can be attributed to strong upthrow to the west of the Rio Miera flexure (Figs. 7-1 & 7-65) that relatively uplifted NW Santander to the west of the flexure. Major facies and thickness changes occur across this flexure and the Santiago section (27 on Fig. 7-1) shows a strongly thickened (> 347m), deeper-water limestone/marl sequence.

7.5 Environmental interpretation of the Las Penosas Formation

Clearly, the Las Penosas Fm. represents a transgressive, fluvial to marginal marine, siliciclastic sequence. Its basal contact with the shallow-marine carbonates of the Comillas or Sierra Mbs. (Reocin Fm.) is always sharp, but not demonstrably erosional. The top of the Reocin Fm. beneath the Las Penosas Fm. records the maximum regression with with subaerial-exposure causing meteoric-

related mixed-water dolomitization, meteoric phreatic cements and local karstification (Chapter 6.9). Any unconsolidated regressive clastics that were deposited between the end of marine Reocin Fm. deposition and the minimum Albian sea-level stand would be cut down through by the fluvio-estuarine channels of Facies 1, so that these channels now rest directly upon the indurated and cemented Reocin Fm. carbonates. Such sharp bases to transgressive coastal sequences are widely known (Fisher, 1961; Swift, 1968; Kraft, 1971; Hobday & Jackson, 1979) and have been termed "ravinements" by Swift (1968). The presence of the major karst at Reocin (Chapter 6.9) allows the approximate distance from the palaeoshoreline to be estimated. The Las Penosas Fm. hence formed around the coast itself, to no more than 10-20km offshore even for its most marine parts.

The basal sheet sands of the Las Penosas Fm. are interpreted as fluvial (Facies 1) and is then followed by a peritidal sequence of the upper estuary (Section 5/33, 512-552m; Section 8 on Fig. 7-1; Facies 2 to 6 and 9). This is cut by numerous estuarine channels (Facies 7) and is capped by estuarine mouth bars (Facies 11). In the inner shelf (5-20m water depth ?) there are inshore tidal sand bars (Facies 12) and slightly further offshore tidal sand waves (Facies 14) with intervening mudstones (Facies 13).

Three main types of estuaries have been reported in the literature (Hayes, 1975): microtidal estuaries that have long barriers, "mesotidal" estuaries with short barriers and major flood and ebb tidal inlets and funnel-shaped "macrotidal" estuaries. Another type of estuary are tidal embayments with minimal arenaceous input (e.g. the Wash - Evans, 1965). The Las Penosas Fm., with its major terrigenous-sand, fluvial input clearly cannot be a tidal embayment.

The presence of major estuarine channels and a mesotidal range (see Facies 8) suggest that a microtidal estuarine model is also inappropriate. The key difference between a mesotidal barrier island/inlet model and a mesotidal, funnel-shaped estuary model is the presence, or lack, of major barriers. As the shoreface, beach or aeolian sands of a barrier are normally destroyed during transgression anyway, one of the key differences in the models will be in the

lagoonal areas. This formation does not seem to show well-developed, clean washover sands or landward-dipping, laminated, washover fans. Another major point is in the lack of flood-tidal deltas (sensu Hayes, 1980) in the Las Penosas Fm. Both funnel-shaped estuaries and barrier-island systems have ebb-tidal deltas, though these are often termed (as here) estuarine mouth bars, shields or shoals. However, in barrier systems the flood-tidal delta has landward-directed palaeocurrents and has high preservation potential relative to the barrier (Hayes, 1975). Such coarsening-upward deltas with landward-directed palaeocurrents (SSW) have *not* been found in this formation, suggesting they were never present as their preservation potential is high. A distinctive feature of funnel-shaped estuaries as against barrier systems is the presence of major fluvial sands beneath the estuary in a transgressive system (Elliot, 1978). These are present in NW Santander as Facies 1. Taken together, these points suggest that this estuary was dominantly non-barred and mesotidal.

Another possible model for this formation is a tidally-dominated delta. Although this and an estuarine model share many of the same tidal facies associations (.e.g tidal channels, intertidal flats, etc.), the key feature of such deltas is that their progradation produces a regressive, coarsening- and thickening-upwards sequence on a scale comparable to the vertical thickness of the Las Penosas Fm. (Coleman & Wright, 1975). Such a sequence is not seen in this formation, and the general absence of a macro-scale fining-trend is comparable to Recent studies of estuarine/inner shelf sands in the North Sea (Terwindt, 1981). The lateral spread of these coastal, tidal clastics along over 100km of the Basco-Cantabrian basin (Fig. 7-65) suggests multiple source input rather than a single sourced delta.

Studies on Recent barrier/estuarine coasts have shown that the transgressive siliciclastic sequences are, in general, strongly thinned relative to the regressive sequences and extensively modified by the shoreface retreat destruction of barrier islands [Fisher (1961); Oomkens (1970); Meckel (1975); Weil (1977)]. It is clear that at Comillas the estuarine sequence is better preserved

than the Holocene transgressive sequences. However, a 30m thick transgressive estuarine sequence has been described by Wilkinson & Byrne (1977) who had fluvial rivers overlying an oxidised Pleistocene sequence. Above the fluvial infill were dominantly estuarine muds with rare sands. In this situation (and here), the dominant, source of the mud was the fluvial input, which also supplied carbonaceous fragments in both these estuaries. Similar Holocene estuaries have been described by Tucker (1973) and tidal facies by Allen et al. (1979). A schematic model for the estuary is shown in Fig. 7-64.

The preservation of such a thick, transgressive, sequence must have meant that the rate of relative sea-level rise just exceeded that due to vertical aggradation of the siliciclastic wedge. The primary reason for this preservation was the degree of tectonic subsidence experienced by particular areas. In the areas with higher subsidence in the west (Sections 2-20 of Fig. 7-1), a reasonably complete estuarine sequence is present. In detail, local diapirism has preserved thicker and more stratigraphically complete sequences in the rim synclines (e.g. Section 12 of Fig. 7-1). However, over diapiric palaeohighs such as Section 16 of Fig. 7-1 the entire estuarine mouth bar/inner shelf siliciclastic sequence has been destroyed by the basal marine transgression disconformity at the top of the estuarine distributary channel (Section 11/14, 66m). Such a strong control by the underlying topography on transgressive shoreline deposits has been noted in both modern (Kraft & John, 1979) and ancient sequences (Berg, 1976; Heward, 1981). It is probable that the areas of greatly thinned (few to zero metres) sediments in the east (Sections 21-23 & 26 of Fig. 7-1) had siliciclastics deposited but these were probably destroyed by shoreface erosion.

Although greater local subsidence was the primary control on the preservation of a thickened transgressive wedge, other factors were of some importance. Amongst these was the rate of sediment supply. The Basco-Cantabrian basin was experiencing strong distensional tectonics (Austrian orogeny) at this time with coeval zones of rapid uplift and of rapid basin subsidence (Chapter 12). The uplifted strata (to the S/SW of the basin) included the relatively

unconsolidated Wealden sands as well as clays from the Wealden or Jurassic marls. A derivation of the sand from polycyclic reworking is supported by the relative textural maturity of the sand (well sorted, mature heavy mineral assemblage - Feuillee, 1967). The climate was a humid, subtropical one because of the Tethyan coral, rudistid, orbitolinid subtropical marine water fauna; major Albian freshwater deltaics; abundant carbonaceous matter and logs; lack of evaporites or flack breccias; abundant meteoric water diagenesis and the active high rainfall-related pedogenesis of major karsts e.g. Reocin or bauxite development east of Tolosa. There was also an extensive fluvial system to the south of NW Santander (Fig. 5-8). Rapid weathering of these easily erodable sediments could easily supply large quantities of siliciclastic material northward to NW Santander. The major input of clastic material from the south (Fig. 7-36) was a further important factor in generating a thick but transgressive Las Penosas Fm..

Another possible intrabasinal control is the large-scale longshore transport of sand which is known to be a major factor in modern coastal sand budgets (van Andel, 1960). As the dominant offshore current was to the WNW (Fig. 7-35), and major deltas in the Albian Basco-Cantabrian basin lay to the SE of NW Santander (Figs. 7-65 & 7-66), Chapter 7.10), longshore drift ought to bring sand parallel to coast from the SW of NW Santander. However, the area in the eastern part of NW Santander (Sections 21-23 & 26 of Fig. 7-1) shows virtually no Las Penosas Fm. clastics, suggesting a lack of major longshore sand transport.

Although there was rapid aggradation of siliciclastics in the Las Penosas Fm., relative sea-level rise just outpaced it. As has been discussed above, the tectonic subsidence varied enormously across NW Santander. A more likely mechanism for the relative sea-level rise is eustatic transgression. The Lower Albian is a period of notable eustatic sea-level rise in western Europe. In southern England, a similar, broadly transgressive succession occurs in the Lower Greensand (Sandrock Fm. and Carstone Fm.) to Gault clay transition in Upper Aptian to Lower Albian times (Anderton et al., 1979). This transition is coeval with the Las Penosas Fm. to Barcenaciones Fm. evolution in NW Santander. A similar basal

Lower Albian lowstand and Lower Albian transgression is recorded by Cooper (1977), Hancock & Kauffman (1979) and Garcia Mondejar (1979a)).

7. 5. 1 Las Penosas Fm. model.

The Las Penosas Fm. was a fluvially-attached estuarine/inner shelf system (Fig. 7-36) that was transgressed in a tectonically active basin with both active diapirism and distension-related sub-basin tilting. During relative sea-level rise (primarily eustatically induced) the estuarine mouth would be choked by its own abundant, fluvially-derived sand and flocculation-related fines. Storm flooding (cf. Facies 2) and base-level rise in the lower parts of siliciclastic river systems tends to generate funnel-shaped "ria" estuaries (cf. Trinity River, Texas - Le Blanc & Hodgson, 1961) which act as excellent sediment traps. This leads to the coeval middle shelf shallow marine sequence being siliciclastic sediment starved. This would then explain the nature of the greatly condensed, sand-poor Miradorio Mb. carbonate above the Las Penosas Fm. with its abundant marine fauna, high degree of bioturbation, carbonate concretions, slow-sedimentation-rate-related glauconitic limestone/marl alternations, and concentrated, unreworked fauna. Similar processes have been noted by McCave (1969) in the Devonian Catskill Fm./shelf transition and in the transgressive hiatuses of the Cretaceous seaway of North America (Ryer, 1977).

The source of the siliciclastic quartz sand of this formation is clearly derived from the fluvial area to the S/SW (Fig. 7-36). It cannot be derived from the purely carbonate Reocin Fm. beneath it or from the Barcenaciones Fm. which was partly coeval with the Las Penosas Fm. as transgression occurred. Such a terrigenous/estuarine source of quartz sand is at odds with the offshore source seen in the modern, tidal North Sea where major sands are relict from the glacial lowstand and are being reworked from offshore into the estuaries (Swift et al., 1971). This contradiction is explained by the unusual situation in the Recent, with extensive glacial sand offshore rather than an unusual situation in the Las Penosas Fm. Rapid subsidence allowed the transgressive estuarine/inner shelf to be

preserved, especially in the western part of the area. The preservation of such sequences with tidal sand waves in rapidly subsiding basins has been noted in numerous basins by Nio (1976). The regional setting of the Las Penosas Fm. will be considered in Chapter 12.

The transgression itself has had a major effect on what is preserved in the ancient record of this clastic estuarine/inner shelf system. Unlike regressive sequences, the subaerial parts of the estuary (beaches, barriers, cherniers etc.) have a low preservation potential and are not seen here. However, subtidal channels, mouth bars and sand waves tend to be preserved. If basins have a suitable high sediment input and tectonic subsidence, then estuarine systems such as this can be preserved. Somewhat comparable sequences are known from the British Lower Greensand (Narayan, 1971; de Raaf & Boersma, 1971; Stride, 1982), though these are less complete than the Las Penosas Fm. with its sequence from karstified basement to middle/outer shelf marls.

7.6 Definition and type section of the Barcenaciones Formation

Definition: The carbonate limestones, marls and limestone/marl alternations that lie between the top of Las Penosas Fm. siliciclastics and the base of the Bielba Fm. siliciclastics. This formation shows considerable lateral thickness variation (Fig. 7-1) from 20m to 117m in NW Santander, but time equivalent strata is much thicker to the east of the Rio Miera flexure (Section 27 of Fig. 7-1; Fig. 7-63).

Type section: The type section for this formation is 100% exposed coastal section between Punta Calderon and Punta de Carrastrada, north of Torriente (56209786 to 56149784; Section 16 of Fig. 7-1). A graphic log of this section is shown in Section 11/14, 66m to 148m. This formation has been subdivided into the local, high-energy Santa Catalina Mb. at the base which is overlain by the low-energy limestone/marl alternation of the Miradorio Mb. (uppermost Lower Albian to basal Upper Albian). This is overlain by the massive, medium to high energy Torriente Mb. limestone (Chapter 7.1, Fig. 7-3).

7.7 Lithology and facies of the Barcenaciones Formation

The basal Santa Catalina Member (Fig. 7-3) can be subdivided into two facies (Facies 15 & 16) :-

15) Facies 15: High-energy, calcarenitic sandwave

Description : These are 2-15m thick and are dominated by planar and trough cross-bedded, calcarenitic grainstones that are well sorted and are in 10-50cm thick lenticular cosets (Fig. 7-37). The clean, clast-supported grainstones have a well sorted and rounded oobiosparite/biosparite marine fauna that includes broken bivalves, echinoderms, orbitolinids, serpulids, ooids, gastropods, dasycladacae, miliolids, rare corals along with nonbiogenic ooids, cortoids, peloids, intraclasts and rarer carbonaceous fragments, glauconite pellets, quartz sand and stylolites (Figs. 7-38 & 7-39). A conspicuous feature of this facies is the oolitic coating of grains. The initial cementation was by a thin (20-50 μm) isopachous, non-ferroan, fibrous fringing cement. This is a typical active marine phreatic cement in which the original aragonite/high-Mg calcite has now neomorphosed to low-Mg calcite. This is followed by a blocky, ferroan, medium-grained (50-200 μm), anhedral calcite. Although this could be a meteoric phreatic cement, the marine nature of both the formation and the earlier fringing cement suggest that this ferroan cement is also an active marine phreatic one.

The general vertical sequence consists of a sharp, slightly erosional basal contact over argillaceous facies. In rare cases a basal coarsening-upward, thickening-upward sequence is seen (Section 1 of Fig. 7-1). The dominant internal structure of the calcarenitic sand wave is stacked planar and trough cross-bedded sets in 10-50cm cosets (Figs. 7-37 & 7-40). Although the cosets are generally lenticular (Fig. 7-41), the central and upper parts of the sequence can contain relatively planar cosets with planar sets (Fig. 7-40). The set boundaries can be rippled with packstone drapes (Fig. 7-40, cf. mud drapes).

In a few cases there are multi-metre-scale, cross-bedded, usually 2-15m thick (Fig. 7-41). The multi-metre coset has bounding planar horizons at its top and base (E_1 of Allen, 1980). The major sets to this coset are 10-40cm thick and

bounded by erosion surfaces (E_2 of Allen, 1980). Internally these sets contain an even smaller scale of cross-bedding.

In general, the palaeocurrent directions of individual units are relatively unidirectional. However, when the currents from a number of units (e.g. those on Section 1 of Fig. 7-1 in Fig. 7-43 or those from the Miradorio Sections 8-11 of Fig. 7-1 in Fig. 7-44) are plotted out a bipolarity is seen, with a dominant current to the W or WNW (Figs. 7-43 & 7-44). These palaeocurrents directions show a remarkable similarity to those of the inner shelf of the Las Penosas Fm. (Fig. 7-35), and are considered to be the products of similar shore-parallel, dominantly ebb tidal currents (cf. discussion in Facies 11 of Chapter 7.3).

Interpretation : These are interpreted as shallow-marine, calcarenitic sandwaves that were probably formed by a similar process, in similar orientations and have similar internal structures to their slightly shallower-water siliciclastic equivalents in the Las Penosas Fm. (Facies 14). The decimetre-scale cross-bedded sequences are interpreted as type V sandwaves (Allen, 1980). The multi-metre scale cross-bedded calcarenites closely correspond to Class IVA sandwaves of Allen (1980) with moderate time-velocity asymmetry, flow separation over the wave and a steep, progradational sandwave slope. These waves have relatively well-developed bottom and topsets. They represent the largest, tidally-dominated sandwaves in the Santa Catalina Mb. which formed in the outer part of the inner shelf.

It could be possible that these structures are tidal sand bars or ridges. Virtually all the wave/storm-related linear sand bars/ridges from the Cretaceous Western Interior Seaway (e.g. Brenner & Davies, 1973; Walker, 1983) show well-developed coarsening- and thickening-upwards sequences with hummocky cross stratification in the interbar zones. None of these features are seen in this member. The bipolar palaeocurrents, association with the tidal estuarine/inner shelf Las Penosas Fm., the similarity of the internal structure of these sandwaves with those of Allen (1980), the correlation of palaeocurrents here with those in the top Las Penosas Fm., the numerous reactivation surfaces (Fig. 7-41) and

packstone drapes all suggest a primarily *tidal* control on sandwave development, though minor storm modifications are most probably present. Shore-parallel calcarenitic shoal belts along the western margin of the Persian Gulf are a good modern analogue (Loreau & Purser, 1973). An ebb calcarenitic channel occurs in the basal part of the Barcenaciones Fm. on Section 16 of Fig. 7-1 (Section 11/14, 66-70m).

16) Facies 16: Interbar calcarenite/marl facies

Description : This is a marl-dominated facies consisting of quartz sand rich wacke/packstones or calcareous sandstone interbedded on a cm to dm scale with a moderately bioturbated marl (Fig. 7-46). The starved ripple "sands" show both wave and current ripples.

Interpretation : This facies occurs between offshore tidal clastic and calcarenitic sandwaves, and is thought to be the product of minor storm reworking of the bar tops resedimenting sandy calcarenite into the interbar lows by storm. Similar facies are widely known in other offshore interbar areas (Walker, 1983; Stubblefield et al., 1984; Tillman & Martinsen, 1984).

The Miradorio Mb. of the Barcenaciones Fm. can be divided into the following nine facies:-

17) Facies 17: Marl

Description : This is one of the dominant lithologies of the Miradorio Member and is characterized by a grey, homogeneous marl (e.g. Section 11/15, 82m). It is intensely bioturbated with total destruction of all primary structures. Locally *Thalassinoides*, *Rhizocorallium* and *Chondrites* ichnofauna of the *Cruziana* ichnofacies (Frey & Pemberton, 1984) occur. This marl has yielded glauconite pellets, *Knemiceras* ammonites, benthic and planktonic foraminifera (including *Hedbergella sp.* and *Tritaxia*) as well as rare oysterid bivalves.

Interpretation : It represents a open-marine, hemipelagic suspension fallout deposit that has been intensely bioturbated due to its slow deposition. The lack of fine-fraction reworking suggests that it was deposited beneath fair-weather wave base, but not necessarily beneath storm wave-base. The shallow water nature of the

Barcenaciones Fm. (grainstone, ooids, etc.) and the abundance of algae in these limestones suggests that the dominant source of micrite is algal, whereas the terrigenous clay is pro-estuarine.

18) Facies 18: Marly nodular limestone and limestone/marl alternations.

Description : This is the typical lithology of the Miradorio Mb. and corresponds to the "Marnes a Myaces" of Mengaud (1920) and Pascal (1985; Fig. 7-3). The marls are highly bioturbated nodular wacke/packstones with 2-20cm nodules in a more bioturbated, marly matrix (Fig. 7-47). This nodular lithology grades into a decimetre-scale limestone/marl alternation. This lithology consists of an alternation of marls (Facies 17) and marly biomicritic/pelmicritic limestone (Fig. 7-49). There are numerous decimetre-scale nodules developed in the marl (Fig. 7-49), and these can coalesce to form nodular limestones such as in Fig. 7-50. The limestones consist of medium- to low-energy, poorly sorted bio- and pelmicrites with a skeleta of broken molluscan fragments (mainly of rudists with calcite rims and recrystallized calcite layers to their shells), uni- and biserial foraminifera, miliolids, rhodophyta, dasycladacae, echinoderm fragments, orbitolinids, sponge spicules, bryozoa, gastropods and arenaceous foraminifera. The dominant macrofauna include *Cardium, Glauconia, Trigonia*, Myaces and Naticids.

Non-skeletal elements are dominated by quartz silt and peloids with minor glauconite grains, micritic envelopes and stylolites. A typical example of this facies is shown in Fig. 7-48. This shows the extensive microsparitization of the matrix that is probably linked to meteoric diagenesis.

Interpretation : These were deposited on an open-marine, low-energy, shelf which periodically received increased fine-grained clastic input (ie. clay). The nodular nature probably links to early diagenetic cementation of the nodules in clay-poor horizons that were picked out by differential weathering and late diagenetic stylolitization. The limestone/marl alternation probably reflects primary differences in pro-estuarine terrigenous clay runoff. The clay-poor horizons were then preferentially cemented, primarily during their early diagenesis. This early

diagenetic cementation of the marly limestone horizons is indicated by the preferential development of burrows in the marls, with some burrows avoiding penetration of limestone nodules. Such a low-energy deposit was clearly deposited below effective fair weather wave base. However, the winnowing of lags in the packstones could be due to storm wave reworking of the substrate.

19) Facies 19: Skeletal (wacke)/packstone tempestites

Description : Individual wacke/packstone beds are relatively abundant in the marly Miradorio Mb. and appear to be generally 20-700cm thick, though some of the thicker ones are amalgamated. They are sharp to locally erosionally based and have a slight tendency to fine upward into wackestones. The dominant skeleta is a poorly sorted or rounded hash of molluscan fragments, ostreidae, echinoderms, gastropods, along with orbitolinids, rhodophyta, pellets, dasycladacae, (*Dissocladella* etc.), codiacian chlorophyta, bryozoa, miliolids, arenaceous foraminifera, chlorophyta, local *Bacinella* and lituolids (cf. Fig. 7-51). Glauconitic grains and infills are common as are carbonaceous fragments, intraclasts, quartz sand and Fe-coating of allochems. These are contained in a micritic matrix and the unstable carbonate phases have been replaced by a non-ferroan calcite cement. The upper parts of these beds are finer-grained and more bioturbated. Sometimes the whole packstones can be totally bioturbated. The tops often show wave/combined flow rippling, Fe-rich hardground generation (Fig. 7-52) or ostreidae colonization. These packstones (e.g. Fig. 7-52) weather out of the generally marly Miradorio Mb. (Fig. 7-54) in a packstone/wacke-mudstone alternation. An example of this facies is shown in Section 11/17, 104-110m.

Interpretation : The presence of calcrudite grade, poorly sorted skeletal hash in muddy shelf sediment is suggestive of rapid, episodic transport probably by storms and that the deposits are tempestites (Aigner, 1985). The hardgrounds/ostreidae probably represent post-storm colonization. The fact that skeleta are not in position yet are not imbricated suggests that storm action was moderate.

20) Facies 20: Isolated skeletal grainstones

Description : These are texturally and compositionally similar to the grainstones of Facies 15, but occur as isolated 15-40cm thick cosets in the Barcenaciones Fm. (e.g. Section 11/16, 88.4m). They mainly show parallel lamination or cross-bedding and can be rich in quartz sand or glauconite grains.

Interpretation : Such isolated high-energy events in marly sequences suggests rare storm action on the sea floor.

21) Facies 21: Hummocky cross-stratified calcarenites

Description : This is a rare structure and consists of beds with a sharp, slightly erosional base overlain by laminated calcarenite that forms low-angle antiformal and synformal hummocks with internal truncation surfaces. The tops of the calcarenites show combined flow/wave ripple laminations.

Interpretation : These are interpreted as storm wave calcarenites generated by a combined flow mechanism below effective fair weather wave base. An example of swaly cross-stratification is shown in Fig. 7-55.

22) Facies 22: Graded calcarenitic tempestites.

Description : These are generally composed of broken, poorly sorted or rounded skeleta forming graded beds that are 20-40cm thick. Their key characteristic is a well developed fining-upward sequence internally.

Interpretation : They may be tempestites or even turbidites related to bottom flows such as occur in the North Sea (Aigner, 1985) or the Atlantic shelf (Swift et al., 1983).

23) Facies 23: "Capa roja" fine sandstone bed

Description : This bed is useful for correlation because it is well exposed in the sections of the western part of NW Santander (Sections 8, 9, 10, 11, 16 & 19 of Fig. 7-1). Its stratigraphic position can be seen in Fig. 7-3. It is a rather massive, silty, slightly calcareous, fine quartz sandstone with abundant red, Fe-stained burrow mottling (Fig. 7-57). Vague decimetre bedding and rippling is locally present, though intense bioturbation has generally destroyed most primary structures. It is wedge shaped and thins from about 18m in its centre (Sections 9, 10, 11 & 19 of Fig. 7-1) to about 1m on the peripheral zones (Sections 8 & 16

of Fig. 7-1). The Capa roja bed can be correlated with a major siliciclastic episode south of the Escudo de Cabuerniga (Fig. 7-56; lower part of U_4 of Fig. 5-8 correlates with the Portillo de las Escalerucas Fm. (upper Middle Albian of Pujalte et al., 1982) of Figs. 6-23A, B). It represents the distal, tapering wedge of a major clastic event south of Escudo, and its thickness was strongly controlled by the subsidence pattern in NW Santander.

Interpretation : It represents a shallow marine sand that has been intensely bioturbated.

24) Facies 24: Calcareous quartz sandstones

Description : These can be subdivided into three types :- 1) Mm-scale interbeds of marl and calcareous sand with abundant carbonaceous fragments, Fe stained allochems and moderate bioturbation; 2) Flaser bedded sands with mud drapes; 3) Massive calcareous sands 20-200cm thick that can be planar bedded at base, but become ripple laminated and bioturbated towards their tops.

Interpretation : The first two types of sand, on account of their fine grain size, intimate association with fines and location in an area below fair weather wave base, are thought to be quartz sands that have been moved, in suspension, into the middle shelf by wave/storm action. The massive calcareous sands, similar to the Capa roja, are also moved by storm waves. Although some beds are clearly amalgamated and represent numerous storm events, they could also reflect either major storms (winter hurricanes?) or seismites triggered by faults.

25) Facies 25: Clean, skeletal, bedded wacke/packstone

Description : This is the dominant lithology of the Torriente Mb., the upper part of the Barcenaciones Fm,. It is generally 10-60cm bedded and has a similar skeleta to that of the wacke/packstone tempestites (Facies 19). A noticeable feature of these carbonates is the abundance of algal thalli, including codiacians, rhodophyta (Melobesia), *Bacinella* and lituolids. The typical bedding in this facies (Figs. 7-58 & 7-59) is rather irregular with a superposition and amalgamation of thin skeletal sheets and wedges. The bedding has been strongly affected by pressure-solution (Fig. 7-58) and bedding-parallel stylolites have generated mm-

thick marl-partings, the "stylo-cumulates" of Logan & Semeniuk (1976). The bedding is wedge-shaped to gently undulating (Fig. 7-59) and is cut by broad, gentle scours. Such a bedform is strongly suggestive of swaly cross-stratification (Leckie & Walker, 1982) produced by storm wave reworking of thin calcarenitic banks in the middle shelf. The internal laminae of the beds conform either to lower bounding surface, but flatten to conform to the upper surface at the top of a bed. Local, small-scale cross-lamination and bedding occur, though this is virtually always formed at less than the angle of repose. Concave-up swales can pass laterally into convex-up hummocky surfaces (Fig. 7-59).

Interpretation : This facies was deposited on a carbonate middle shelf that was relatively cut off from terrigenous clay input. The local rudaceous debris in a low-energy micrite also support a storm-dominated ramp-like middle shelf interpretation. A possible analogue is the muddy *Halimeda* banks of South Florida formed by storms and waves (Aigner, 1985).

26) Facies 26: Local coral bioherm

Description : Isolated coral bioherms up to 3m thick occur in the middle shelf calcarenites of the Barcenaciones Fm. (e.g. Section 11/19, 122 to 125m). This boundstone consists of amalgamated domal corals (5-50cm diameter) in a packstone matrix rich in a diverse marine fauna including Pectins, bryozoa, sponges, hydrozoa, chlorophyta, gastropods and rudists.

Interpretation : Such buildups are interpreted as small patch reefs that lack zonation and occur in the top few tens of metres of water of the open-marine middle shelf (James, 1983).

27) Facies 27: "Caprina choffati" beds

Description : These are present in the top 5-10m of the Barcenaciones Fm. and are seen on numerous sections (Sections 1, 5, 16, 19, 21, 22 & 23 of Fig. 7-1). In general, this packstone facies is similar to Facies 25, but has an abundance of rudists, mainly requienids, caprinids and monopleurids (Figs. 7-60 & 7-61) and some, local sucrosic dolomite. The most famous of these is *Caprina choffati* (Fig. 7-61). These beds locally have abundant stromatoporids.

Interpretation : Such an assemblage of rudists suggests middle shelf conditions, where caprinids are known to occur (Perkin, 1974).

28) Facies 28: Quartz sand interdigitations

Description : The very top 15m of the Barcenaciones Fm. contain a number of 0.5-8m thick interdigitations of quartz sand (e.g. Section 17/6, 141m), or multi-metre thick marly, glauconitic sands (Fig. 7-62). The marly sands of the Trasvia section (Fig. 7-62) show a distinct swaly stratification.

Interpretation : These are the initial interdigitations of the Bielba Fm. quartz sands and represent a very thin regressive sequence (see Chapter 8.6).

7.8 Lateral variation in the Barcenaciones Formation

The lateral variation of this formation shows that the main members and facies remain relatively constant over much of NW Santander, but vary greatly in thickness (Fig. 7-1). The most northwesterly section at Punta Ninera (Section 1 of Fig. 7-1) is a typical sequence of Santa Catalina Mb. calcarenites, overlain by limestone/marls of the Miradorio Mb. The top of the *Caprina choffati* beds are thrusted out on this section. The northern flank of the Escudo de Cabuerniga develops a relatively uniform sequence in the Albian (Sections 2 to 5 of Fig. 7-1), though the proportion of clastics to carbonates varies. This is in part due to poor outcrop preventing differentiation of the Las Penosas Fm. and the Miradorio Mb. Further east, the sections around Comillas (Sections 6 to 11 of Fig. 7-1) show a normal Barcenaciones Fm. This formation only slightly thickens into the Cobreces diapiric palaeolow rim syncline (Section 12 of Fig. 7-1), and it is dominantly the clastic Las Penosas Fm. that accommodates the thickness increase into this palaeolow. A relatively constant Barcenaciones Fm. continues towards the Orena diapiric palaeohigh (Sections 13 to 16 of Fig. 7-1). There is then a moderate thickening into the Santillana palaeolow rim syncline (Sections 17 to 19 of Fig. 7-1). The Reocin sequence (Section 20 of Fig. 7-1) is exceptional in that the carbonates of this formation have been totally dolomitized (see Chapter 6.9).

The Barcenaciones Fm. shows marked thinning and facies changes over the Miengo palaeohigh (Sections 21 to 23 of Fig. 7-1). These sections lack the Las Penosas Fm., the Santa Catalina Mb. and the Miradorio Mb., and so the contact with the Aptian Reocin Fm. must represent a hiatus. The Albian of the Punta del Dichoso section (21) is about 12m thick and characterized by high-energy, rudistid, orbitolinid, pack/grainstones with ooids, rudists (*Caprina, Toucasia, Polyconites*) glauconite and clastic interdigitations (Facies 31) (Pascal, 1985). The 20m thick section at Rio Saja/Cuchia (Section 22 on Fig. 7-1; Section 15A/21, 229m to 249m) consists of a marly, bioturbated wacke/packstone and local tidal shoals, overlain by a rudistid shelfal bank rich in requienids, orbitolinids, miliolids, monopleurids and calcareous sponges. A similar section is seen at Punta de Sumocueva (Section 23 of Fig. 7-1; Section 17/6, 140m to 160m). A slight thickening occurs into the Camargo palaeolow (Sections 24 & 25 of Fig. 7-1), though this formation rapidly thins towards the Santander palaeohigh (Section 26).

The Santiago section (Section 27 of Fig. 7-1), to the east of the Rio Miera flexure, is much thicker (>347m) and is dominated by a marly limestone/marl alternation rich in planktonic (and benthonic) foraminifera. The westward tilting of NW Santander, active Keuper diapirism and role of the Rio Miera flexure have been discussed at the end of Chapter 7.4.

7.9 Environmental interpretation of the Barcenaciones Formation

The detailed facies analysis of this formation (Chapter 7.7) points to a marine middle shelf environment that was developed on a homoclinal ramp (sensu Ahr, 1973) with no major fringing shoals or reefs to the shelf. This homoclinal ramp lacks distal deepening, rimmed shelves or steep slopes as no slope facies has been found (cf. Read, 1982). Continued relative sea-level rise until the uppermost Lower Albian diachronously moved the siliciclastic estuarine/inner shelfal Las Penosas Fm. landward to the S/SW. The tidally-dominated, shore-parallel currents inherited from the inner shelf rework the calcarenitic sands into major, shore-parallel tidal sand waves and bars of the inner part of the middle shelf (15-30m

water depth ?). The transgression continues to its Albian maximum through the Middle Albian and deposited a condensed marly limestone/marl sequence with ammonites in the outer part of the middle shelf/innermost part of the outer shelf (50-100m water depth ?). The Upper Albian, Torriente Mb. consists of storm-dominated packstones and marls that reflect a minor relative sea-level fall. The latter was deposited above storm wave base, and because of the presence of codiacian algae and swaly cross-stratification , near to fair-weather wave base. Such storm domination of open, homoclinal ramps is relatively normal (Aigner, 1985) as there was no major rim to absorb wave/storm attack to the shelf. The dominance of shore-parallel WNW/ESE currents and land to the SE, suggests that the dominant storms came in from the Bay of Biscay to the NW (Fig. 7-63). Parallels hence exist with the alongshore wave/storm approach documented by Swift et al. (1983) from the Atlantic shelf of North America. In general, such a shelf with active diapiric highs, a homoclinal ramp geometry, coralgal skeleta, shallow marine skeletal banks and deeper water tempestites/marly limestones and marls bears a remarkable resemblance to the subtidal parts of the Holocene Persian Gulf (Purser, 1973).

7.10 Summary of the Albian of NW Santander

The thin Albian sequence of NW Santander (mean = 140m) commences with the siliciclastic Lower Albian Las Penosas Fm. This consists of a transgressive fluvial/estuarine/inner shelf sequence overlying a basal ravinement. The sandy fluvial sands are followed by a coarsening- and thickening-upward sequence of estuarine tidal flats, tidal creeks, storm sands, estuarine, distributary and sand channels, peritidal/lagoonal black muds capped by estuarine mouth bars. The inner shelf deposits consist of sandy tidal sand bars and sand waves with interbar muds. The Las Penosas Fm. generally thickens to the west across NW Santander due to sub-basin, distensional-related tilting, but is also affected on a smaller (5km) scale by Keuper diapirism with thickened rim synclines and thinned sequences over diapiric palaeohighs. A barrier model for this formation is rejected as no major

washover sands, beach, flood-tidal deltas occur, and the transgressive sequence eliminates a tidally-dominated delta model. Such a transgressive estuarine sequence has been preserved due to rapid subsidence and major siliciclastic input during, primarily eustatic, marine transgression.

The Middle/Upper Albian Barcenaciones Fm. consists of high-energy calcarenitic sand waves at its base followed by siliciclastic-starved, storm-dominated middle shelf marly limestone/marl alternations. Skeletal packstone tempestites followed by wave/storm reworked, homoclinal ramp middle shelf carbonates dominate the upper part of this formation. It has somewhat similar lateral variations to the Las Penosas Fm.

CHAPTER 8 - THE CENOMANIAN BIELBA AND ALTAMIRA FORMATIONS - ESTUARINE/INNER SHELF TIDAL CLASTIC AND CARBONATE SAND BANK.

The Cenomanian of NW Santander consists of two formations - the Bielba Fm. below and the Altamira Fm. above (Fig. 1-9). The Bielba Fm. (Lower Cenomanian) is a transgressive siliciclastic estuarine/inner shelf tidal deposit with a local, thin regressive sequence at its base. The top of the Bielba Fm. interdigitates with the overlying Altamira Fm. (Middle to mid Upper Cenomanian) which is a uniform, medium- to high-energy, calcarenitic pack/grainstone deposited on a middle shelf bank. It is capped by a hardground/hiatus within the upper part (?) of the Upper Cenomanian and is overlain disconformably by the basal glauconitic marly limestones/marls of the Turonian Sardinero Group (Chapter 9). In essence, the overall Cenomanian sequence is a repetition of that seen in the Albian, though there are some notable differences. The Bielba Fm. is notably thicker, more quartz sand-rich and is dominated by marine inner shelf clastics rather than the estuarine deposits of the Las Penosas Fm. Also, the basal regression of the Bielba Fm. did not extend as far inland as that of the basal Las Penosas Fm. An important consequence of this was that the most non-marine fluvial facies of the Las Penosas Fm. (Facies 1 of Chapter 7.3) did not extend as far north (seaward) in the Bielba Fm. as it had in Las Penosas Fm. times. This explains why karstification of the underlying Barcenaciones Fm. limestone only occurs in the most inland areas (e.g. Reocin), whereas it is more extensive at the base of the Las Penosas (cf. Chapter 6.9). The Altamira Fm. carbonates are higher-energy and much more uniform than those of the Barcenaciones Fm.

8.1 Stratigraphy of the Cenomanian of NW Santander

The Cenomanian stratigraphy is shown in Fig. 7-3 which primarily relies on the work of Mengaud (1920), Feuillee (1967) and IGME (a, c, e). Most authors have broadly followed the zonal stratigraphy proposed by Feuillee (1967), though

the top of his "Calcaires roux d'Altamira" is now thought to be within (and not at the top of) the *C. naviculare* Zone. Basically, IGME termed the Bielba Fm. C^{1}_{21} and the Altamira Fm. C^{23}_{21}, though these are combined to C_{21} in the San Vicente area due to the poor exposure/thinned nature of the Bielba Fm. clastics. On the eastern Torrelavega sheet (IGME c), the unit C^{9-1}_{16-21} is used. This area has no Albian Las Penosas Fm. and a remarkably thinned Albian Barcenaciones Fm. (Fig. 7-1) which has been mismapped by its inclusion within C^{23}_{15}. The area mapped as C^{9-1}_{16-21} is, in fact, the basal Santillana Mb. of the Bielba Fm. as seen from the correlation sections (Fig. 8-1).

The Bielba Fm. has been subdivided into three members by the author (Hines, 1985a). The basal Santillana Mb. is sand-rich and represents estuarine/innermost shelf deposits. This is overlain by the more bioturbated inner shelf deposits of muds, marls, tidal bars and sand waves of the Herran Mb.. This is equivalent to the "Calcaires greseux" of the upper part of the *M. mantelli* Zone (Feuillee, 1967). The eastern part of NW Santander (Torrelavega sheet - IGME c) has a major tongue of Altamira Fm. lithology calcarenite - the Cerrazo Mb.. This occurs within the Herran Mb. This has been mapped by IGME as Cc^{1}_{21}. Because of the uniformity of the Altamira Formation it has not been subdivided into members.

8.2 Definitions and Type Sections of the Bielba and Altamira Formations

Bielba Formation Definition: This is defined as the siliciclastic mudstones, siltstones and sandstones that lie between the top of the Barcenaciones Fm. limestones and the base of the Altamira Fm. The top of the Bielba Fm. is somewhat more difficult to define as it interdigitates with the calcarenites of the Altamira Fm., but is here taken as the base of the major continuous body of calcarenitic limestone of the Altamira Fm..

Altamira Formation Definition: This is defined as the well-bedded, decimetre-bedded calcarenites whose base is taken from the point where there are no more major interdigitations of Bielba Fm. clastics. The top is sharp on all sections

and the contact with the marls/marly limestones of the Sardinero Group is usually marked by a hardground with glauconite that is a biostratigraphic hiatus.

Type Section: The type section for the Cenomanian is taken from the 100% exposed coastal section at Punta de Carrastrada/Punta Calderon, 3km due N of Novales (56149784 to 56019779; Section 6 on Fig. 8-1). A graphic log (Section 11) shows the Bielba Fm. running from Section 11/21, 148m to Section 11/47, 351m. The Altamira Fm. runs from Section 11/47, 351m to Section 11/50, 429.4m.

The Bielba Fm. is 230m thick on average, though it does show strong lateral variability and ranges from 34m to 344m thick (Fig. 8-1). The Altamira Fm. has a mean thickness of 84m with a range of 20m to 130m. In general terms, the Cenomanian strata thin to the east across NW Santander (50km scale) but show thickening into active, diapiric rim synclines on a smaller (5km) scale (Fig. 8-1 & 8-68).

8.3 Lithology and Facies of the Bielba Formation

The lithologies of the Bielba Fm. were studied from the excellent coastal exposure (e.g. Fig. 8-3) that are summarized in Fig. 8-1. Inland exposure of the formation is very poor indeed. The Cenomanian sequence (summary in Fig. 8-4) is clearly a transgressive one, except for a thin regressive sequence at the base of the thinned sections of the Miengo/Sumocueva palaeohigh (Fig. 8-68; Sections 9-13 of Fig. 8-1). The facies of the Bielba Fm. will be described in approximately stratigraphic order, working from the less marine ones in the basal parts of the westernmost sections to more open marine at the top of the easternmost ones.

1) Facies 1: Estuarine tidal channels

Description : These are major sandy channels that are well developed on the basal parts of the Trasvia (Section 4 of Fig. 8-1) and Playa de Luana sections (Section 5 of Fig. 8-1; Section 7/2, 21.5m to 7/5, 38m). Both of these sections have three stacked channels which range from 3m to 9m thick, with a mean of 6m. They are characterised by sharp, locally erosive bases that cut down into the underlying peritidal heterolithic deposits (Fig. 8-5). A basal lag (Fig. 8-6) usually exists

of cm-sized carbonaceous fragments or armoured ironstone conglomerates (themselves probably derived from channel erosion of its peritidal banks). The estuarine channels of the Playa de Luana has a basal hash of marine skeleta, brought in from offshore. The lower part of the vertical channel sequences are dominated by multidecimetre-scale, planar to trough cross-bedded sandstone in lenticular cosets (Figs. 8-7 & 8-8). These cross-beds show moderate to abundant mud and fine-carbonaceous fragment draping (Figs. 8-8 & 8-9). The cosets can be sigmoidal with a decrease in angle of repose downcurrent. Figure 8-9 clearly shows the double mud drapes present on the sets (indicating subtidal megaripples - Visser, 1980) which often thicken down coset. The thin sand between such mud drapes locally has reverse-flow cross-lamination ripples climbing up the main set. Figures 8-8 & 8-9 show the occurrence of abundant reactivation surfaces, avalanche foresets and hanging set boundaries.

Channels show a clear fining- and thinning-upwards trend and are more bioturbated towards their tops (Fig. 8-11). The base of the upper 1/3 of the channel often has a heterolithic cm to dm interbeds of clean sand and black mud (Fig. 8-12). The very top of channel sequence is muddy and moderately to intensely bioturbated (Figs. 8-13, 8-14 & 8-15). Direct evidence for reduced salinity comes from the occurrence of *Cerithium tenouklense* Coquand, a brackish water mesogastropodid. Within the general fining-upwards, thinning-upwards sequence, thicker, decimetre sandstone beds occur (Fig. 8-16). These probably represent storm sands into the subjacent peritidal flat derived from estuarine channel floods. The tops of individual channel muds/muddy sandstone is erosionally incised into by the next channel sand/conglomerate stacked above (Figs. 8-5, 8-6, 8-7, 8-8, 8-11 & 8-13).

The sands of the Bielba Fm. are, in general, mature, well sorted, orthoquartzites to subarkoses. They have a mature heavy mineral assemblage (Feuillee, 1967) dominated by tourmaline (50-85% of heavy minerals), zircon (1-25%), anatase (1-12%) and rutile (1-6%). A significant feature of the heavy mineral distribution is the presence of metamorphic minerals such as staurolite,

garnets, epidote and andalusite (Feuillee, 1967), which are commonly present in the Lower Cenomanian sands of the western Basco-Cantabrian basin (Fig. 8-10).

In general, individual channels have relatively unimodal palaeocurrent patterns, with a clear predominance of palaeoflow towards the NNW to NNE (Figs. 8-17 and 8-18). Although the majority of channels flow broadly northwards, rare palaeocurrents are diametrically opposed, towards the SSW (flood-directed-lower part of Playa de Luana - Section 5 of Fig. 8-1; Section 7/6, 40-45m). Herringbone cross-stratification is a rare structure in these channels (Fig. 8-19A).

Interpretation : These channels bear a remarkable resemblance to the mesotidal, lower-estuarine distributary channels of the Haringvliet (Terwindt, 1971) , the Oude Mass (Van Beek & Koster, 1972), or the Mahakam tidal delta (Allen et al., 1979). Here the cross-bedded sands occur as channel margin-attached bars that can form extensive sheets, much wider than the width of the original channel as the channel migrates across the peritidal flat. It is this very migration that explains the abundance of channel deposits over tidal flat ones in this (and other) ancient sequences. The bioturbated interbedded sandstone/mud facies at the top of channels (Fig. 8-12) has been described by van der Berg (1981) from the Oosterschelde Mouth as being deposited winter/spring arenaceous storm/river flood run-off interbedding with summer argillaceous low-stage bioturbation. Very similar current orientation patterns to those in Santander have been described by van der Berg (1981) from this Recent estuary in Holland.

The Hercynian grain to the Basco-Cantabrian basin runs WNW-ESE and this was the orientation of facies belts in the Lower Cenomanian (Fig. 8-76). Coeval, high-energy, braided river sands existed to the south (Fig. 12-15) and this suggests that the broadly northward flowing tidal currents in these estuarine channels are the ebb ones, transporting sand offshore.

The sandstone petrography testifies to the erosion of the crystalline Hercynian basement to the S/SW of the basin (Fig. 8-17), or from rocks derived from such erosion. The textural maturity suggests that these were polycyclic sands with a small first cycle component from the basement. The light mineral

fraction is dominated by muscovite (1-2% of the rock), which is often concentrated on bedding planes. The association with black, carbonaceous mudstone (Figs. 8-8 & 8-9) is also rather diagnostic of a tidal environment (slack water stage).

2) **Facies 2: Bioturbated Sandy Mudstone**

Description : This consists of moderately to intensely bioturbated sandy mudstone (Fig. 8-19B), that is usually so bioturbated that primary sedimentary structures have been totally destroyed (eg. Playa de Sumocueva, Section 17/20, 224m or Playa de Luana, Section 7/14, 81.2m). The bioturbation comes from a shallow-marine *Cruziana* ichnofacies of *Chondrites*, (Fig. 8-20), *Thalassinoides* and *Rhizocorallium*, amongst others. The open marine fauna that have been described from the Bielba Fm. (Mengaud, 1920; Feuillee, 1967, IGME a, c, e) has primarily come from this facies (author's descriptions). The microfauna is dominated by benthonic, marine orbitolinids (abundant *Orbitolina aperta* and *O. concava* along with biserial foraminifera (*Flabellamina goodlandensis* amongst others), textularids (*Textularia aggutinans* amongst others, lituolids (*Pseudocyclammina rugosa*) and rare planktonic foraminifera (*Tritaxia tricarnata*). Marine ostracods (*Cythereis sp.* and *Cytherella sp.* amongst others) have also been found (Feuillee, 1967). This facies has yielded numerous shallow marine ostreidae, mainly exogyras (*Exogyra columba, E. flabellata*) and oysters (*Ostrea, Arctostrea carinata*). The mudstones contain moderate to abundant mm-sized carbonaceous fragments along with echinoderm fragments, glauconite pellets and muscovite.

Interpretation : The sandy bioturbated mudstone facies is seen as a low-energy, inner shelf marine mudstone that occurred in quiet water between tidal bars or sand waves (Fig. 8-42).

3) **Facies 3: Interbedded sandstone/mudstone of interbar Facies**

Description : This consists of interbedded sandstone and mudstone with bedding on a few mm to 10cm scale (Figs. 8-21 to 8-24). Typical examples occur at Playa de Luana, Section 7/11, 66m; Playa de Valdearenas, Section 16/4, 16m or Playa de Sumocueva, Section 17/27, 258.5m. The sandstones in this facies are sharp topped

and bottomed in general (except where bioturbated). They usually have rippled tops together with somewhat rarer ripple cross-lamination. There is an abundance of wave-generated ripples with near symmetrical bedforms, locally opposed cross-lamination, swollen lens sets, local offshooting and foreset draping or irregular/uneven set and lamination boundaries (de Raaf et al., 1977). Clear unidirectional, form concordant ripples, interpreted as current ripples, have also been found. Individual sands are laterally extensive for few to or a few tens of metres and give the rock a distinctly "pinstripe" wavy to lenticular bedded appearance. Local "packages" of sand-rich wavy bedding with cm-thick sands occur and these alternate with mud-rich packages that are lenticular bedded. The degree of bioturbation varies from low to moderate (e.g. Figs. 8-21, 8-22, 8-23 & 8-24). The fauna is very similar to that described in Facies 2 of Chapter 8.

Interpretation : The size and scale of interbedding of the sedimentary structures are remarkably similar to rather distal, low-moderate energy, suspension-derived storm wave deposits described from the North Sea basin (Aigner & Reineck, 1982; Aigner, 1985) or offshore Texas (Hayes, 1967). The most likely origin of the sand is from the adjacent topographic high of tidal sandbars/sandwaves (Fig. 8-42) with storm waves catastrophically reworking the sands from the top of the bars into the interbar areas. Small-scale (20-100cm) vertical fining- or coarsening-upward sequences are seen in this interbedded facies, in a very similar way to those from the estuarine/inshore deposits of SW Holland (Terwindt, 1981).

4) <u>Facies 4: Hummocky Cross Stratified Sandstone</u> (HCS)

Description : This facies consists of hummocky sandstones (Fig. 8-25) with slightly erosional lower set boundaries that have a low-angle, undulatory form. The internal structure consists of laminated sand parallel to basal set boundary that are cut by other such packets at low angle (1-15°). The top of the HCS is locally parallel laminated or, more commonly, wave ripple laminated. These isolated HCS sands occur within the interbedded sandstone/mudstone sequence, are not amalgamated, and built clear antiformal hummocks that are commonly 5-30cm thick (at the centre) but taper out to zero over wavelengths of 2-6m (Fig. 8-26).

This HCS bears a remarkably close resemblance to HCS described by Dott & Bourgeois (1983), though the basal parallel lamination of their typical sequence is rarely present here. The HCS at other localities (e.g. Playa de Sumocueva - Section 12 of Fig. 8-1; Section 17/23, 239.5m) is somewhat thinner (a few cms) and is interbedded with black mudstone (Figs. 8-27, 8-28 & 8-30). This has a strong resemblance to the "micro hummocky cross-stratification" of Dott & Bourgeois (1983).

Interpretation : Hummocky cross-stratification is considered to be a product of waves/storms product (Dott & Bourgeois, 1983), though other authors have questioned the relative effect of each of these processes. The micro HCS is considered to form from weaker storms or by more distal wave action in areas that were relatively sand starved. The abundance of sand or frequency of storms was higher at other localities/times (e.g. Playa de Sumocueva, Section 17/20, 222.5m) and amalgamated HCS was generated (Fig. 8-29). Here successive storms would have initially scoured out any background hemipelagic fallout mudstone before depositing a thin (<10cm) HCS sandstone from the waning storm/combined flow.

HCS is generally considered to be a product of storm/wave action, usually in the inner shelf. Independent evidence here suggests that it formed in the interbar areas in the outer parts of the marine inner shelf (10-30m water depth?). The karst/subaerial exposure at Reocin during Bielba Fm. times (Chapter 6.9) allows the distance from shore to be constrained to no more than about 10-20km, though the exact distance to shore cannot be given. The sand source for the HCS is most likely to be storm wave erosion of the tops of the sand bars or sand waves (Fig. 8-42). These tidal sandbars or sandwaves have sharp, erosive tops. As they formed between storm and fairweather wave base, storm waves could periodically erode them and redeposit the sand as HCS/interbedded sandstone-mudstone in the interbar lows.

5) **Facies 5: Black Mudstone**

Description : Black mudstones characteristically have very little (0 to 5%) sand, and range from a few tens of cm to 6m thick (e.g. Playa de Sumocueva, Section

17/14, 194m to 200m). They vary from having little or no bioturbation (Fig. 8-31) to being strongly bioturbated or massive. They are similar to, but less sandy than, the bioturbated sandy mudstone of Facies 2, though they do locally have thin (mm-cm) sandstone laminae.

Interpretation : These massive mudstones are the product of continuous sedimentation and bioturbation (Potter et al., 1980) and represent a pro-estuarine mud fallout deposit. As these mudstones are interbedded with marine clastics, they cannot be muddy peritidal flat deposits. Thick (about 5m), non-bioturbated mudstone with lenticular bedded, thin (mm-cm) sands have been described from inshore mud banks of the Surinam coast by Rine & Ginsburg (1985) where their lateral migration as subtidal mud deposits makes them too unconsolidated to support a significant fauna or flora. Such a location, in the innermost shelf of a mud-rich estuarine coast with a humid, subtropical climate, is very similar to the Playa de Sumocueva section.

6) <u>Facies 6: Tidal Sand Bar Facies</u>

Description : This facies consists of sandstone units arranged in coarsening- and thickening-upwards sequences in which there is a decrease in bioturbation up the unit. From 18 such bars in the Bielba Fm. (Fig. 8-1), they range in thickness from 3 to 23m, with a mean of 10m (eg. Fig. 8-32).

These sand units generally lie upon bioturbated sandy mudstones (Facies 2) or interbedded storm wave sandstone/mudstone alternation (Facies 3). From the base there is either a gradational (usually) or sharp (rarely) transition to flaser bedded sandstone that can be interbedded with thin (1-10cm), wavy bedded mudstone horizons (Fig. 8-33). There is usually slight (Fig. 8-33) to moderate (Fig. 8-34) bioturbation in this subfacies by *Cruziana* ichnofauna.

The upper part of the sand bar usually has planar, rippled (up to 5cm high) flaser bedded sandstone that passes gradationally upwards into low- to high-angle, trough cross bedded sandstone (e.g. Playa de Valdearenas, Section 16/15, 70-75m; Fig. 8-32). Minor, planar cross bedded sandstones with tangential bottomsets can replace the trough cross bedded sandstone locally. These crossbeds show mud-

draping, rare double mud drapes, down coset decrease in angle of repose and rare reverse-flow ripple cross-lamination climbing up the main sets (Fig. 8-35). The very top of the sand bar is often made of a rather massive calcareous sandstone with orbitolinids, echinoderm and molluscan debris. It can have some calcite cement and be topped by an iron-rich hardground (Figs. 8-36 & 8-37, Section 7/12, 71m to 75m; Section 7/14, 84.6 to 87.7m; Section 16/17, 80m to 84.4m). The top is sharp and is overlain by the next marine facies, often the base of another sand bar (e.g. Section 6 of Fig. 6-1).

Interpretation : These features are strongly suggestive of, though not exclusive to, deposition from a tidal bar. The presence of such mud drapes and muddy interbeds suggests that even the tops of the bars were still below fair weather wave base, though above storm base. The tops of the bar can show undulatory erosional surfaces (Fig. 8-35) that are traceable over 30m or more. These are associated with major erosional events, possibly storms. The tops of the bars pass up into further bioturbated marine mudstones or interbedded sandstone - mudstone facies. Although this fine grained rock is lithologically similar to lagoonal/tidal flat sequences, its marine fauna disproves such an interpretation. The lack of beach swash cross-stratification or lagoonal deposits at the top of these bars prevents a barrier island/lagoonal interpretation for them.

The calcareous sandstone cap is considered to be an abandonment facies in which fauna colonize the bar and the currents switch elsewhere (cf. abandonment phase of estuarine/shelf sands of Nio, 1976). However, the concentration of a poorly sorted or washed hash of skeleta not in life position suggests storm reworking. As these skeleta are not present elsewhere in the Bielba Fm., it is most likely that they are *in situ* accumulation of fauna on the bar top that have been storm reworked after bar abandonment, not dissimilar to the swell or storm lags described by Brenner & Davies (1973) from Oxfordian sand bars of Wyoming.

These sand bars clearly represent prograding, shoaling-upward sequences in which the higher energy cross bedded facies progressively prograde over the finer grained muds. The generally coarser, thicker-bedded and less bioturbated

nature of the upper parts of the bars indicates that they were formed from higher velocity/more persistent marine currents. The tops of the bars show no indication of subaerial exposure though, and the marine fauna in the mudstones supports an inner marine shelf location for the bars.

The type of marine currents that deposit marine sand bars, especially the upper sandy cross-bedded parts of them, is still highly controversial, with both tide and storm action suggested from the Recent and ancient analogues (Walker, 1984). A serious difficulty with interpreting the ones in Santander is the lack of along strike exposure inland, hence making it impossible to prove elongation relative to the palaeoshoreline. Most of the small-scale sedimentary structures are, to some degree, equivocal. The mud drapes of the cross sets could be products of fair weather fallout alternating with storm set progradation as much as they could be tidal ebb/flood drapes. Angle-of-repose cross beds up to 50cm in thickness or so have been recorded by Swift et al. (1973) from the Recent Atlantic shelf as being present in storm-generated sand ridges as well as being well known from tidal systems (Allen, 1980). Reactivation surfaces can be similarly explained in fair- to storm-weather changes and/or minor changes in the larger bed form migration as opposed to tidal reversals. However, the lack of bioturbation in these cross bedded sands suggests multiple (semi-permanent?) reworking, and unless storms are very frequent (days to months?), then tidal currents seem more obvious as a mechanism than storm ones.

The palaeocurrent pattern of these sand bars is most instructive (arrows against lithologies in Fig. 8-1). Individual bars show considerable spread in their current patterns, though are broadly unidirectional. These unidirectional patterns differ from one bar to another and dominantly run N/NE (Sections 4A, 6, 9B of Fig. 8-1) but with a significant number running nearly diametrically opposite (to the S/SSE - Sections 6, 9 & 11 of Fig. 8-1). Another grouping run approximately perpendicular to this axis, towards the WNW (Sections 9B, 11 & 12 of Fig. 8-1) with a few opposite to this on E/ENE (Sections 9B, 11 & 12 of Fig. 8-1). A further point is that these directions (especially the NNW one) are subparallel

to those derived from the demonstrably tidal sand wave facies (Facies 7) or estuarine channels (Facies 1), confirming the activity of tides in these directions. These directions are seen as the dominant directions in the rose diagrams of the Bielba Fm. (Fig. 8-17: Trasvia section (Section 4 of Fig. 8-1); Fig. 8-18: Playa de Luana (Section 5); Fig. 8-38: Punta de Carrastrada (Section 6); Fig. 8-39: Playa de Los Locos (Section 9); Fig. 8-40: Playa de Valdearenas (Section 11); Fig. 8-41: Playa de Sumocueva (Section 12)). The dominant palaeocurrent direction, to the NNW/NNE is perpendicular to the palaeoshoreline (cf. Figs. 8-68 & 8-76), and not parallel to it as many storm-generated sand bars are in the Recent (Swift et al., 1973) or ancient (Berg, 1975; La Fon, 1980; Boyles & Scott, 1982; Hobson et al., 1982). The bars with orientations on the WNW/ESE axis are noticeably stratigraphically higher in the Bielba Fm. and/or occur on the more seaward sections to the NE (Fig. 8-1). This could well reflect a progressive change from flow perpendicular to the coast in the estuary / innermost part of the shelf to more shore parallel further offshore. Such a trend is seen on most modern day tide- (Stride, 1982) or storm-dominated coasts (Swift, 1973).

Hence, these sand bars are seen as a storm - tide interaction in which thin storm sands are deposited in interbar lows whilst dominantly tidal currents are able to (semi-continuously) rework the upper parts of the bars to produce high-energy physical structures. The very tops of the bars seem to have been reworked by storms to generate the sands for the interbedded sandstone - mudstone facies, or HCS, or broad undulatory surfaces. A rather condensed lag of sand with coarse, broken skeletal hash can be left by storm winnowing. A generalised model for their generation is shown in Fig. 8-42. The vertical sequence through the sand bar is often more complex than the continuous coarsening-upward sequence suggested above. This is due to storm spillover lobes or surge channels within the bar sequence (Fig. 8-42).

A number of further points need to be answered about these sand bars. Firstly, what was their sand source? In tidal sand ridges/banks of the North Sea (Belderson & Stride, 1966) or storm ridges formed by erosional shoreface retreat

in the Middle Atlantic Bight (Swift et al., 1972), the sands are primarily autochthonous reworkings of the Pleistocene lowstand outwash sands. Such an autochthonous source is unlikely (or only minor) in Santander as 1) the underlying Barcenaciones Fm. is a clean carbonate shelfal limestone, devoid of quartz sand for the tide/storm currents of the Bielba Fm. to rework and 2) the reworked/palimpsest sands of the Holocene transgression are generally less than 20m thick (Stride, 1982) and a possible regressive sand sheet formed similarly in base Bielba Fm. times simply could not supply sufficient sand to the whole of the Bielba Fm., which is over an order of magnitude thicker. Hence the fluvio-estuarine channel system to the S/SW (Fig. 8-68) was the dominant ultimate source of the sands in the Bielba Fm. How was the sand transported onto the shelf and why are no direct mouth bars seen in the sequence? These two questions are interrelated. Sand would be transported towards terminal mouth bars of the estuaries that prograded over the marine mud/interbar facies of the innermost shelf. These would become abandoned due to sediment building up to fair weather wave base, feeder channel switching etc. During transgression, these would become isolated offshore and reworked by storm/tidal processes to produce the bars. Other mechanisms for offshore sand transport such as rip currents are unlikely as no rip channels exist, and only estuarine channels have been seen. Underflow currents also transport sand offshore (cf. Hurricane Carla - Hayes, 1967) but thin underflow "event" sands are unlikely to carry out the sand required for this formation.

As to a modern analogue to these deposits, the key problem is the lack of lateral control in this ancient sequence and the (near total) lack of multi-metre scale vertical sequences from Recent shelves. The most obvious analogues are the tidal sand ridges/banks of the North Sea (e.g. the Norfolk Banks - Johnson, 1979; Stride, 1982) or the linear sand ridges of the Middle Atlantic Bight that become detached during transgression (as here). These have comparable vertical dimensions and facies locations to those in Santander. The author's preference is for an analogue with the more tidally influenced ridges of the North Sea because of the strong tidal influence with the upper parts of the bars themselves and within

the formation as a whole. The tops of the banks show medium-scale, cross bedded sandstones with thin mud drapes/laminae (Houbolt, 1968), consistent with the tops of these bars. Whether these Recent banks have metre- to decametre-scale coarsening-upward sequences is still unknown. The long axes of these banks subtend a small oblique angle relative to the peak tidal flow direction and have "cats-back" megaripples on their sides that migrate along the axis of the bank and towards its top (Caston, 1972). Their overall geometry is usually asymmetrical and their progradation would produce the predominant directions seen in internal cross-bedding. The ebb and flood usually take opposite courses around the bank, and depending on which is more dominant, this direction will be preserved in the megaripples as the bank migrates. This would explain the N and S dominant directions of the different bars in Santander, and suggest they formed from N/S oriented sand banks, isolated just offshore of the estuaries by the marine transgression (Fig. 8-68).

7) <u>Facies 7: Tidal Sand Waves</u>

Description : These are dominated by clean, fine- to medium-grained, quartz sandstone in trough cross-bedded sets and lenticular cosets (Figs. 8-43 & 8-44). The only mud that occurs is as thin (mm-cm) mud drapes to set or coset boundaries (Figs. 8-43 & 8-44). These are high-energy sandstones with no, or virtually no, bioturbation and a very uniform grain size. The key distinguishing feature of these sand waves are that they have : 1) a sharp, though not demonstrably erosional base; 2) internally they have a very monotonous sequence of medium-scale (5-60cm), cross-bedded sandstones; 3) they neither thicken/thin nor fine/coarsen upwards and 4) they have sharp tops (Fig. 8-45). They range in thickness from 2m to 18m, with a mean of about 10m. The total lack of inland outcrop prevents their lateral distribution being accurately determined, but the lack of continuity of individual sand waves between closely spaced (100-1000m plus) sections (e.g. Sections 4A/4B, 5/6, 9A/9B and 11/12 of Fig. 8-1) suggests that these sand waves have a highly lenticular, "wave-like" plan geometry on a scale of few hundreds to thousands of metres. Locally, as at Punta de Carrastrada (Section 6 of Fig. 8-1),

a larger-scale dune structure to these sand waves can be seen (Fig. 8-60). These have a 2-5m height and 50-200m wavelength and have very low (a few degrees or so) slope angle (unlike the angle of repose megaripples).

Interpretation : These are interpreted as tidal sand waves and are very similar to the type V sand waves of Allen (1980) described in Facies 14 of Chapter 7.3. The trough cross-beds are mud-draped, often with clear double mud drapes, linked to the diurnal tidal variation of a thick sand bundle from 1) the stronger tide →2) mud drapes →3) thin sand bundle from the subordinate tide (with local reverse flow cross lamination) →4) to another mud drape. These are clearly seen in Figs. 8-44 & 8-47. Sigmoidal cosets (Fig. 8-46) with oblique concave surfaces which converge down coset tangentially into thin horizontal bottomsets (Mutti et al., 1985) indicate several successive episodes of sand deposition alternating with slack water mud deposition. These could well record lunar, or higher order (i.e. equinox) tidal cycles.

Reactivation surfaces are abundant in some cross-beds, probably because of the dominance of either the ebb or flood in any, relatively unidirectional, sand wave and because of their moderate, decimetre size (Mowbray & Visser, 1984). These decimetre sized cross-beds are more easily dominated by a particular tidal current than larger sand dunes due to their smaller size. The preserved subordinate flow structures (backflow ripples within the double mud drape) are rather rare. Locally ripples are smaller and are located on the upper part of the sandwave. These generate topset hanging reactivations. Mild reverse flow with downset increase in preserved mud draping is also seen (Fig. 8-48) and there is reactivation in the topsets. Local strong reactivation with complete destruction of the mud drapes and strong angular discordances between the reactivation surface and the set boundaries are seen in Fig. 8-44.

Horizontally, along an individual cross-bed, a "sigmoidal" geometry can be seen on a 5-30m scale (Fig. 8-48) from a basal set mud drape, → to thickening sets and higher angle of repose sets (acceleration stage), → to maximum set thickness (sand wave full vortex stage) → to thinning and flattening sets

(deceleration stage). This is typical of tidal influence of sand wave migration in moderately rapid (>60cm/sec) tidal sand wave currents where one tidal direction is dominant (Boersma & Terwindt, 1981).

The neap/spring lunar month tidal cycles were investigated in a manner similar to those from subtidal shoals of the outer parts of the Recent Oosterschelde estuary (Visser, 1980) where bundle sequences have been recognised. The author measured the thickness of *all* sandstone sets in a number of 15-40cm thick cosets from the tidal sand waves at the Playa de Los Locos (Figs. 8-49 & 8-50) and Trasvia (Fig. 8-51). The measuring of *all* sandstone sets allows both the dominant and subordinate currents to be seen, and the thick/thin pairs seen in Figs. 8-49 to 8-51, and from observation (Figs. 8-44, 8-46, 8-47 & 8-52) confirms the existence of double mud drapes, a unique indicator of the subtidal (as against intertidal) environment (Visser, 1980). Some of the bundle sequences (such as Fig. 8-50B or 8-51B) show an approximate periodicity of about 28-30, which could be directly linked to the neap/spring cycle. The other bundle sequences show good suggestions of cyclicity, but there is rather strong "noise" to the cycles. Considering all the other tidal indicators in this facies, this deviation from a simple 28 cyclicity is explained as variants on the standard tidal cycle. Firstly, for one of the tidal currents to be strongly dominant requires a high degree of asymmetry to tidal ellipse. In type V sand waves (Allen, 1980) there is only moderate asymmetry to the ellipse, so the 28 day cycle of the dominant tidal current is confused by the smaller, though important, subordinate one. Secondly, it is clear that this inner shelf was affected by storm action which could produce HCS and rare thick sets of Figs 8-49 to 8-51. This storm action would introduce a random element to the cycles. Further complications such as the prevention of mud deposition by waves or changes in sand wave orientation also exist.

Although dominated by medium-scale, trough cross-bedded sandstones, some vertical changes within the sand wave can be seen. The base is universally sharp, indicating basal (storm tide?) erosion of the interbar area before deposition of the sand wave. Such basal erosion surfaces to sand waves have been

termed E_1 by Allen (1980). There is no basal lag, possibly because of the very good sorting of the primary sand. The lower part of the sand wave tends to be slightly higher-energy with well developed megaripples in lenticular cosets (Figs. 8-43, 8-44, 8-46, 8-47, 8-48, 8-53 to 8-55). The upper part have more tabular cosets (lower energy? - Figs. 8-56 & 8-57) and can have smaller megaripples (Fig. 8-58) or are more mud rich (Fig. 8-59). Similar muddy megaripples have been described from the fallout stage of the neap tide of intertidal sand waves by Boersma & Terwindt (1981), though the general vertical sequence and description above prevents these from being intertidal. The increase in mud could well be linked to the sand wave becoming moribund and preserved into the ancient record. The very top of the sand wave is again (knife) sharp, probably indicating storm wave(?)/tide peneplanation of the sand waves, possibly at the fair weather base where waves prevented any further upward growth.

The almost total lack of bioturbation is due to the exceptionally rapid tidal currents and high rates of deposition. However, a rare exception to this is a systematic, leaf-like, burrow structure (Fig. 8-64), which formed in the high-energy surface of the sand wave.

The large sand dune structures described above (Fig. 8-60) are similar to the sand waves of the North Sea (Stride, 1982). It is interesting to note that the internal structure preserved from such large dunes is *not* multi-metre-scale cross bedding, but decimetre-scale, trough cross bedding. What little work has been done on Recent sand waves (e.g. Houbolt, 1968), confirms this picture from the ancient (as do theoretical studies - Allen, 1980b). The megaripples of the sand waves probably represent "cats-back" megaripple forms on the backs of larger dunes of the whole sand wave. Such a hierarchy of bed forms on sand waves has been repeatedly observed (Stride, 1982). In general, the internal structure of these sand waves is very similar to type V sand waves of Allen (1980) - cf. Facies 14 of Chapter 7.3 or the translational sand waves described by Dalrymple (1984) from the Bay of Fundy.

The palaeocurrent pattern of an individual sand wave is relatively unidirectional. *Very rare* examples of herringbone cross bedding do occur (Fig. 8-65). The mean palaeocurrents for individual sand waves (Fig. 8-1) show a general dominance of NW to NE directions. However, oppositely-directed flows to the SSE occur at the base of the Punta de Carrastrada (Section 6 of Fig. 8-1), on the Playa de Los Locos (Section 9B of Fig. 8-1) and in the upper part of the Playa de Valdearenas sections (Section 11 of Fig. 8-1). These palaeocurrents are shown in the overall ones for the Bielba Fm. (Figs. 8-17, 8-18 & 8-38 to 8-41).

On general basinal and palaeogeographic considerations and the orientation of the estuarine channels discussed above, the northerly direction is considered to be the ebb one of the inner(most), subtidal shelf and the southerly the flood. These sand waves are therefore a swarm of ebb-dominated, innermost shelf sand bodies, not that dissimilar to those in the present Bristol Channel (Stride, 1982). Similar inshore/estuarine mouth parabolic sand shoals (= waves) have been described by Ludwick (1974) where the flood and ebb follow mutually evasive channels that build out into mutually evasive sand shoals. Some of the sand waves higher up section/further to the NE were deposited in deeper water and have a tendency to rotate to a more coast-parallel direction on WNW-ESE (Sections 11 & 12 of Fig. 8-1). Similar changes from rather parabolic bedforms to more coast parallel ones further offshore are seen in the present-day North Sea (Ludwick, 1974; Stride, 1982). It may be possible that the tidal sand bars are the distal fringes of the sand waves and that these two facies are part of the same bedform. The lack of subaerial or peritidal features at the tops of the sand waves suggests they formed around fair weather wave base. The bases of the sand waves probably formed in 5-40m of water, and the presence of 18m thick sand waves constrains the depth to be at least that in the upper part of the Playa de Los Locos section.

7A) <u>**Facies 7A: Sand wave associated subtidal channels.**</u>

Description : These are multi-metre thick (mean = 4m) sandstone units with an erosional base with lag of bored, carbonaceous logs up to 40cm long. It fines- and

thins-upwards with decimetre-sized megaripples towards its base passing up into interbedded flaser bedding/megaripples and finally flaser-wavy-bedded sandstone/mudstone.

Interpretation : These are interpreted to be ebb-oriented channel and formed as a feeder channels to an ebb-oriented sand wave/bar as suggested in Fig. 8-68. Similar subtidal channels have been described by Ludwick (1974). This facies occurs within the tidal bar/tidal sand wave deposits sequence (e.g. at Playa de Los Locos, Section 9B of Fig. 8-1).

8) <u>Facies 8: Oyster Patch Reef</u>

Description : An oyster patch reef of whole, locally *in situ* specimens occurs on the Playa de Sumocueva section (Fig. 8-66; Section 12 of Fig. 8-1). The oysters are in a matrix of calcareous sandstone. The reef is strongly lenticular and thins from 3.5m thick to zero over 20m.

Interpretation : This is interpreted as an oyster patch reef. Oysters indicate brackish marine conditions of the outer estuary, and represent the most regressive, least marine lithology of the Sumocueva section.

9) <u>Facies 9: Tidal Calcarenitic Bars (of the lower part of the Bielba Fm).</u>

Description : Centimetre- to metre-thick, quartz sandy limestone or calcareous, skeletal sandstones occur towards the base of the Bielba Fm. on the condensed sequences over palaeogeographic highs such as the Playa de Los Locos, Valdearenas or Sumocueva sections. They are not interdigitations of the Altamira Fm. which will be considered below. They are of two types.

The majority are rather massive quartz sandy wacke/packstones or skeletal sandstones with a diverse marine, rather poorly sorted or rounded skeleta of orbitolinids, molluscs, echinoderms, rudists (requienids, caprinids), rare corals, brachiopods and gastropods (e.g. Section 17/10, 172 to 174.5m). Some peloids and glauconite also occur. There is a moderate degree of calcite cementation to the skeletal quartz sandstones and an anomalous 33cm-diameter granite boulder on the Playa de Los Locos section. The other type are parallel to cross-bedded, quartz-sandy calcarenites that have a similar poorly sorted, marine skeleta.

Interpretation : These are tidal calcarenitic bars. Their marine origin is confirmed by *Thalassinoides* burrows. The rather poorly sorted or rounded nature of the skeleta suggests storm reworking, possibly from further offshore. On the Sumocueva section (12 of Fig. 6-1) this facies occurs both above and below the oyster patch reef (Section 17/10, 172 to 194m). A regressive→transgressive model for the calcarenite→oyster patch→calcarenite sequence is envisaged.

10) Facies 10: Thick Sand Ball Horizons

Description : A pair of metre-thick major sand ball horizons occur on the Playa de Sumocueva section (top of Fig. 8-26; Section 12 of Fig. 8-1; Section 17/20, 223-224m and Section 17/22, 233-234m). They are cm-laminated and form balls a metre thick and 1-4m wide. They are associated with minor soft sediment deformation structures such as sand flames or microfolding. The tops have locally abundant fluid escape structures.

Interpretation : Such soft sediment-deformed sands point to very rapid introduction of a single, thick sand body that underwent sand balling due to its greater density relative to the marine, black mudstones. Although these could be tsunami deposits, the lack of lateral continuity to the Playa de Valdearenas section just 2km away suggests a more local source, most probably seismically induced failure of a sand body in this tectonically active area (active Keuper diapirism).

11) Facies 11: Massive, Orbitolinid Sandy Mudstone

Description : This is the most abundant lithology in the upper part of the Bielba Fm. where it starts to interdigitate with the Altamira Fm. (e.g. Section 17/32, 282 to 285m). It is dull grey, rather massive mudstone formed by intense bioturbation of thin sand ripples/sand patches etc. in mudstone (Fig. 8-6). It is characterised by an orbitolinid fauna. These muds are dirty, rather massive and have mm- to cm-sized carbonaceous fragments, pyrite nodules, glauconite and rare echinoderm debris (Fig. 8-67).

Interpretation : These are interpreted as inner shelf, low-energy mudstones which were totally bioturbated by organisms that could tolerate the soft, clay-rich bottoms of the outer part of the inner shelf.

12) Facies 12: Thin, Muddy Sandstones

Description : These occur interbedded with Facies 11 and generally consist of cm-thick fine sandstones with slight to intense bioturbation. The tops are both current and wave rippled, as are the bases. Local internal cross-lamination occurs within individual sandstones. Typical examples occur on Section 17/29, 265 to 268m, and are shown in Fig. 8-67.

Interpretation : These sands probably represent the distal ends of sand sheets or sand patches (cf. North Sea - Stride, 1982).

The Altamira Fm. will be considered in Chapter 8.6.

8.4 Lateral Variation in the Bielba Formation Facies

The Bielba Fm. shows strong lateral variation in facies. The most obvious variation is the considerable lateral changes in thickness (Fig. 8-1). The Bielba Fm. thins progressively to the west in the western stable block, on the northern flank of the Escudo de Cabuerniga (Figs. 2-3 & 2-4; Sections 1 to 3 of Fig. 8-1; Fig. 8-68). This sequence has NNE-flowing estuarine channels at its base. To the north, in the Comillas/San Vicente area (Figs. 2-5 & 8-68), the Bielba Fm. thins, reflecting its southerly derivation. The Trasvia section (Section 4 of Fig. 8-1) shows a series of stacked estuarine channels (Facies 1) at its base which are overlain by offshore tidal sandbars, sandwaves, interbar mudstones and finally interbedded sandy marls and marly wacke/packstones of the transition to the Altamira Fm. A somewhat similar sequence is seen at the Playa de Luana (Section 5 of Fig. 8-1). Above basal estuarine channels, this has well-developed calcarenitic/skeletal sandstone tops to the tidal sand bars of the lower part of the Bielba Fm. (Section 7/5, 38m to 87m). Above the tidal sand wave of the upper, eastern part of this section is the usual interdigitation with a rather dirty Altamira Fm. lithology.

The section at Punta de Carrastrada (Section 6 of Fig. 8-1) is thinned relative to the sections east or west of it as it formed over the Orena diapiric palaeohigh. The most noticeable features are absences of basal estuarine channels

and the abundance of high-energy, subtidal sand waves/stacked tidal sand bars. One explanation for the lack of estuarine channels is that it was too far off palaeo-shore (ie. too NE) for the estuarine channels to have prograded out this far. An alternative explanation is that the estuarine channels have a natural tendency to avoid palaeohighs and be concentrated in topographic lows, such as rim synclines. Hence, stacked estuarine channels occur in the Playa de Luana section of the Novales/Cobreces palaeolow, infilling palaeo valleys. Offshore marine currents repeatedly rework detritus over palaeohighs and produce sand-rich sequences (cf. Seni and Jackson, 1983 - sand rich over diapiric highs; Tillman & Martinsen, 1984 - stacked sand bars over palaeohighs of the Cretaceous Western Interior basin). Such a concentration of subtidal sand bars and waves is clearly seen on the Punta de Carrastrada section.

Further SE, the Bielba Fm. thickens rapidly (Sections 7 & 8 of Fig. 8-1) into the Santillana rim syncline. These are the most spectacular thickness increases into the Santillana syncline (cf. Fig. 7-1), indicating that it was highly active in the Lower Cenomanian. The formation then thins again towards the Miengo/Sumocueva palaeohigh (Sections 9 to 12) of Fig. 8-1). These palaeohigh sections are rather sand-rich and do not develop estuarine channels at their bases as 1) they were palaeohighs and 2) they were further offshore, i.e. further NE than the Playa de Luana/Trasvia/Escudo de Cabuerniga area. These sequences also preserve a thin regressive sequence at their bases between the top of the Barcenaciones Fm. carbonate shelf and the least marine strata (i.e. the oyster patch reef - Facies 8). The standard transgressive sequence then follows above. The more offshore (further NE) sequences are probably able to preserve something of their regressive sequence as they do not have the estuarine channel complex of the more hinterland sections. These estuarine channels would have cut down through any regressive clastics at the maximum sea level low-stand, and so no regressive sequence is preserved there. The base of this inland erosion would have been the resistant Barcenaciones Fm. limestone, which would prevent any further downcutting. The lower topographic gradient and lack of channels further offshore

would allow a thin regressive sequence to be preserved. It is hence probable that this regressive sequence (with its tidal calcarenitic bars/oyster reefs) is, in fact, time correlative to the much thicker sequence of stacked estuarine channels to the SW.

The Playa de Los Locos sequence (Section 9 of Fig. 8-1) has a number of well-developed tidal sand wave sequences as they formed over the Miengo palaeohigh (e.g. Figs. 8-44, 8-47, 8-54, 8-55 & 8-56). The Playa de Valdearenas and Sumocueva sections (Sections 11 & 12 of Fig. 8-1; Section 16/1 & 17/1) consist of stacked sand bars and waves, but the interbar mudstones are thicker, probably because these are sequences formed further offshore (to the NW) and, hence, more depleted in estuarine-derived sandstone.

8.5 Environmental Interpretation of the Bielba Formation

The Bielba Fm. represents an estuarine/inner shelf environment. The base of the formation in the more southwesterly sections was deposited more towards the palaeohinterland on an erosional ravinement (Swift, 1968; cf. base of the Las Penosas Fm. - Chapter 7.5). The preservation of a thin regressive sequence further offshore has been discussed above. The fact that *thick* transgressive sequences are generated yet thin or non-existent regressive ones exist in the Santander Group (Fig. 2-1) is unusual and relates to the competent nature of the platform carbonates that underlie both the Las Penosas and Bielba Fms. and to the rapid rate of clastic input, rather than instantaneous eustatic regressions and slow transgressions *per se*. Clastics deposited on rather flat, laterally extensive carbonate platforms during the regressive phase will tend to be reworked at the minimum sea level stage, especially by the estuarine channels cutting down into them. This differs from a sloping shelf/delta situation where clastics can prograde during regression into deeper water and generate thick regressive cycles (Ryer, 1977). After the basal erosion at minimum sea level stand, the *very* rapid influx of clastics (cf. Valmaseda Fm. of Chapters 7.10 & 8.7) would rapidly build up the thick pile of clastics in the Bielba Fm., especially if relative sea-level

rise just outpaced vertical aggradation. This sea-level rise is seen across the entire Basco-Cantabrian basin through the Cenomanian (Feuillee, 1967; Figs. 8-76 & 8-77) and is probably truly eustatic in origin, as it marks the initiation of the great Upper Cretaceous transgressive cycle (Fig. 4-33; Cooper, 1977; Hancock & Kauffman, 1979; Vail et al., 1977).

The presence of major, sand-rich estuarine channels - Facies 1 - points to a meso- or macrotidal coast, though the lack of a definable intertidal zone makes differentiating between these two possibilities difficult. This was not, at least primarily, a barrier island coast for many of the same reasons as were discussed in Chapter 7.5. The lack of inter/supratidal lithologies (beaches, washovers) and faunas (brackish fauna/rootlets) at the tops of the coarsening-upward tidal sand bars suggests that they are offshore bars rather than coastal barriers. Their lateral discontinuity does not support a laterally continuous barrier island model for them, and the lack of landward-flowing flood-tidal deltas supports the same conclusions. The lack of a regressive, coarsening- and thickening-upwards cycle on the thickness scale of the entire Bielba Fm. dismisses the tidal-dominated delta model here. The dominance of NE/N/NW-directed palaeocurrents in the sand waves and bars, just offshore from the estuary mouths, suggests that the dominant sand source was from these estuaries rather than longshore drift. However, the upper parts of the more seaward (NE) sections (Sections 11 & 12 of Fig. 8-1) with their more shore-parallel palaeocurrents, could be receiving significant longshore input.

The transgression modified what was preserved of the estuarine sequence with the subaerial parts (beaches, barriers, cheniers etc.) being reworked by waves during shoreface retreat, but the high preservation potential of the channels allows them to be retained into the Bielba Fm. The transgression preserved offshore tidal sand waves and bars (cf. Nio, 1976) as they became moribund under mud blankets. A general summary of the environment is shown in the palaeogeography of the Bielba Fm. (Fig. 8-68).

8.6 Facies and Environment of the Altamira Formation

The definition and type section of this formation were considered in Chapter 8.2. This formation shows a rather subdued version of the thickness variation pattern (Fig. 8-1) seen in the Bielba Fm. and varies from 20m to 130m (mean = 80m, n=13 of Fig. 8-1). In general, this formation is remarkably laterally continuous both vertically and laterally across NW Santander. It is universally dominated by well bedded, tan-weathering, decimetre (20-80cm) bedded, well sorted and rounded calcarenitic pack/grainstones (Figs. 8-67, 8-69 & 8-70). The dominant skeleta are marine cortoids/aggregate lumps and orbitolinid benthic foraminifera (Fig. 8-71) along with codiacian chlorophyta, bryozoa, mollusca, echinoderms, and dasycladacae. Further marine macrofauna present include ostridae fragments, broken, moderately sorted and rounded rudistid fragments (including *Ichtyosarcolites* and radiolitids), gastropods, brachiopods, rare corals and broken *Inoceramus* shells (Fig. 8-72). An abundant microfauna includes orbitolinids, miliolids, textularids, arenaceous foraminifera, lituolids, prealveolinids, ostracods, trocholinids, cuneolinids and *rare* planktonic foraminifera (stomiospheres, Globigerines, *Hedbergella*, Rotaliporids). Details of the species identification are given in Feuillee (1967) & IGME (a, c, e). The non-skeletal component (under a few % of the rock) is dominated by quartz silt along with rare glauconite pellets and rare oolitic coatings to rounded, reworked skeleta. The areas of micritic matrix (packstones) or sparry matrix (grainstone) form relatively distinct layers or zones on a mm to 1cm scale, reflecting periodically moderate or high energy currents respectively. The earliest diagenesis to these limestones is rather intense micritization by bacteria, cyanophyceae, fungi or eukaryotic chlorophyta in a shallow, warm-water, limpid, photic zone (Bathurst, 1971; Friedman et al., 1971; Flugel, 1982). This occurred under active marine phreatic conditions. The earliest sparry cement is rather irregular to locally isopachous, rather fibrous, non-ferroan calcite cement (Fig. 8-73). This intergranular cementation most probably formed under active marine phreatic conditions too and the initial aragonite/high-Mg calcite of this cement has later recrystallized to low-Mg calcite while

preserving the crystal form. Further porosity destructive cementation is by a coarse equant to drusy, mildly to strongly ferroan, orthospar calcite (Fig. 8-73) which formed under phreatic conditions, possibly under marine phreatic conditions because of the higher Fe content of this latter cement (Oldershaw & Scoffin, 1967). A latter phase cementation is destructive, very coarse (0.5-2mm crystals), blocky calcite cement that locally contains floating spar-supported remnants of skeleta (Fig. 8-74). Such a fabric-destructive spar with dog-tooth corners is typical of active meteoric diagenesis (Flugel, 1982). Late-stage calcite veins cross-cut this limestone and local dolomitization has occurred on bedding planes, stylolites and burrows.

Although these pack/grainstones overwhelmingly dominate the Altamira Fm., other lithologies do occur. The first of these are rare, domal, coral reefs that vary from under a metre to over 10m high. One of the best example occurs on the eastern cliff of the Playa de Luana (Section5 of Fig. 8-1). This bioherm grew at the same rate as the skeletal sands aggraded as there is no draping of the sands over it. Similar, isolated patch reefs that have been inundated by carbonate sands have been described by Hine & Newmann (1977) from the Recent Little Bahama Bank, the Bahamas. Another rare lithology are decimetre-thick microbreccia horizons of mm/cm-sized, clast-supported autochthonous fragments of lithified pack/grainstone in a coarse, calcite spar. Rare fauna, such as orbitolinids, are present within the breccia but are well rounded. The microbreccias are probably *in situ* breakups of lithified hardgrounds. Under warm water, early, marine phreatic cementation of high permeability grainstones, such grainstones can be rapidly lithified to form local hardgrounds, which could be broken up to form these microbreccias. Towards the centre of the formation certain horizons have moderate to abundant oolitic coatings to the skeleta, and rare oobiosparite also occur.

The bedding, type of skeleta and texture of the pack/grainstones all suggest deposition in a subtidal, carbonate skeletal sand bank or belt. Its offshore marine nature is further confirmed by thin (cm-dm) interbeds of marl/marly limestone that have an *in situ* ammonite fauna. The occurrence of the

inner shelf tidal siliciclastics of the Bielba Fm. beneath the Altamira Fm. and the outer shelf marly limestones of the Sardinero Group above the Altamira Fm. constrain this formation to deposition in an open marine, middle shelf environment. Following the classification of Ball (1967), this calcarenite would be a marine sand belt with Recent analogues from the Lily Bank, Little Bahama Bank (Hine, 1977) or the Cat Cay ooid shoal (Ball, 1967). The most characteristic features of these marine sand belts is their elongation parallel to shore and their strong tidal/storm processes. The Santander Altamira Fm. extends laterally in remarkably similar facies right across NW Santander for over 50km parallel to palaeoshore and for over 15km perpendicular to palaeoshore. The most dominant depositional unit within these skeletal banks are extensive spillover lobes that are fed by tidal channels. These spillovers are laterally extensive for 100's-1000's of metres and are a few decimetres to metres thick. Such large, sheet-like sand beds (on an outcrop scale) account for the very well-bedded nature of the Altamira Fm. (Fig. 8-69). These spillover lobes commonly prograde landward (Hine, 1977), and would account for the interdigitation of the Altamira with the Bielba Fm. as major spillover lobes into the clay-rich inner shelf (Fig. 8-67). The most major, polysequential, spillover lobe was the Cerrazo Mb. (Fig. 7-3), which is of typical Altamira Fm. pack/grainstone lithology. It represents a series of stacked spillovers during a short period of reduced clastic input/relative sea-level rise. These spillovers are dominantly generated during hurricane storms after prolonged reworking of the calcarenite by tidal action (Hine, 1977; Aigner, 1985). Internal sedimentary structures *within* individual decimetre beds are rarely preserved here (diagenetically destroyed?), though rare parallel lamination and cross bedding have been found. Individual beds are either finer or coarser upwards. The skeleta layers are moderately sharp-based with occasional scours (storm erosion). The fining upward sequences are more bioturbate upwards, suggesting post-storm recolonization of the graded unit (cf. Safety Valve bank's similar sequences - Aigner, 1985). The tops of beds are usually capped by a thin (mm-cm) marly limestone, possibly representing post-storm fallout. These marly limestones

thicken towards the top of formation, reflecting greater depositional depth. The coarsening-upward beds probably represent *in situ* winnowing of the calcarenite with the coarsest skeleta left as a lag at the top of the bed. These factors, together with the lateral continuity of beds, suggest storm spillover lobe deposition.

Although the very high-energy banks are dominated by very clean ooids in large sand waves, the rather more moderate- to high-energy banks are dominated by peloids, cortoids, aggregate lumps, benthonic foraminifera and calcareous algae in the Recent (Halley et al., 1983) and here in the Altamira Fm. These more moderate-energy marine sand belts are characterized by persistent tidal reworking with current velocities of the order of 30-50cm/sec and usually have their surfaces stabilized by sea grasses and chlorophyta (Halley et al., 1983). These moderate velocities account for the lack of extensive ooid or channel development and the presence of surface binding flora is confirmed by the abundant codiaceae and dasycladacae described above. Such flora suggest shallow, moderate-energy, agitated, limpid marine waters over the banks.

The moderate energy structures, moderate degree of micrite still present in the matrix, encrusted aggregate lumps and lack of channels or metre-scale cross-bedding suggest deposition below fair weather but above storm wave base. Especially on the windward margins of marine skeletal banks such as Little Bahama Bank, marine sand belts can form in rather deep waters of about 30-50m (Halley et al., 1983). This would be a reasonable estimate for the Altamira Fm.. The depth would have been slightly (5-10m?) less over the active diapiric palaeohighs and would have generally increased through the Cenomanian. A general summary of the palaeogeography is shown in Fig. 8-75.

8.7 Conclusions on the Bielba and Altamira Formations

The Cenomanian of NW Santander commences with the major estuarine/inner shelf, tide- and storm-dominated siliciclastic Bielba Fm. In the more hinterland areas (to the S/SW) this has thick estuarine channels at the base that tend to

infill diapiric palaeolow rim synclines. Further offshore a thin regressive sequence on top of the Albian Barcenaciones Fm. is capped by condensed carbonates and oyster reefs. As the Cenomanian transgression continued, the inner shelf deposition above was dominated by tidal sand bars and waves with bioturbated, interbar sandy mudstones between them. These are dominantly oriented due offshore (NW/NE) following the ebb tidal current, though shore- parallel sand bars running WNW/ESE (parallel to the Hercynian basement lineaments of the Basco-Cantabrian basin) do occur further up section/further off palaeoshore. Clear tidal features have been found in both the estuarine channels and inner shelf sand waves, and storm wave HCS has been recorded from the lower parts of coarsening- and thickening-upward tidal sand bars. The Bielba Fm. thins into active diapiric palaeohighs which tend to be poor in estuarine/coastal sands but rich in shallow-marine sand bars and waves.

The clastic inner shelf interdigitates with an offshore marine calcarenite belt of the Altamira Fm. that transgressed over the Bielba Fm. in the Middle/Upper Cenomanian. This has a diverse marine fauna and is dominated by moderate- to high-energy, calcarenites that were reworked by tides and prograded out as spillover lobes during storms. This facies distribution fits with a general WNW/ESE facies belt trend for the basin in the Cenomanian, with deeper water to the N/NE. This lateral continuity of facies in the basin is linked to moderate, more uniform thermal subsidence of a passive margin of the actively drifting Bay of Biscay in the Cenomanian.

CHAPTER 9: THE UPPER CRETACEOUS SARDINERO GROUP; MIDDLE TO OUTER SHELF CARBONATES OF THE TURONIAN TO MIDDLE CAMPANIAN.

The Upper Cretaceous (post-Cenomanian and pre-Upper Campanian) of NW Santander (Figs. 9-1 & 1-6) is dominated by outer shelf, low-energy marls/marly limestones of the Sardinero Group (Fig. 1-9). The term "Formación del Sardinero" was first proposed by Garcia Mondejar et al., 1982 to cover this stratigraphic interval. Summary sections giving the stratigraphy, lithology and thicknesses of the formations are shown in Fig. 9-2.

A) THE TURONIAN - CONIACIAN PEDRUQUIOS FORMATION.

This is defined as the uppermost Cenomanian/Turonian to Coniacian marly limestone/marl alternation of the lower part of the Sardinero Group that is bounded top and bottom by prominent disconformities/hardgrounds. In general, this is a monotonous sequence of marly limestone/marl alternations on a decimetre-scale (usually 5-100cm) that forms highly laterally continuous beds on the scale of even the largest of outcrops (50-300m; shown in Fig. 9-3). There are also rather nodular zones (such as on Fig. 9-4) and omission surfaces which are burrowed by *Thalassinoides, Phanolites and Chondrites.* These omission surfaces can develop into rather nodular decimetre horizons (Fig. 9-5) with burrows clearly avoiding the early-formed nodules and locally generating pseudobreccias as allochthonous material is brought in round the nodules by the burrowing organisms (arthropods etc.). Both incipient and true hardgrounds (Fig. 9-6) are developed and the latter have been locally encrusted by *Spondylus* etc. or have local boring fauna such as those by calcareous sponges, serpulid worms or pteriomorph bivalves. They are usually well burrowed (commonly by *Thalassinoides*) and have been strongly mineralized by glauconite usually within the tests of skeleta, on burrow walls or rather irregularly into the body of the micrite. These hardgrounds are a direct result of the rapid transgression in uppermost Cenomanian/basal Turonian times

when the area was sediment starved and local upwelling brought in the cations for glauconite generation.

The marly limestones are always micritic and are dominantly biomicrites, biopelmicrites or pelmicrites. The stenohaline fauna is dominated by planktonic foraminifera (*Hedbergella, Pithonella, Globotruncana, Lenticulina* etc.), benthonic foraminifera, ostracods, echinodermal debris, pteriomorph and other bivalves (chiefly *Inoceramus*), rare ammonites and belemnites. There is a rare to moderate quartz silt to fine sand fraction, glauconite pellets/fragments and rare phosphates. The fauna and lithology of the Pedruquios Fm. suggest an outer shelf environment of deposition.

B) THE SANTONIAN - MIDDLE CAMPANIAN CABO MENOR FORMATION.

Lithologically, the Cabo Menor Fm. is similar to the Pedruquios Fm.. The base of the Cabo Menor Fm. is usually a biostratigraphic hiatus (Carreras & Ramirez, 1971) with significant zones within the Coniacian missing. The basal disconformity is usually marked by a hardground horizon with extensive glauconitization and an abundance of fauna including *Micraster, Spondylus,* ammonites, *Inoceramus, Holaster, Pinna, Terebratula* and calcareous/silicious sponges. The fauna is unfragmented indicating a low-energy environment

The petrography, lithology and environment of deposition are all similar to the Pedruquios Fm.. However, the Cabo Menor Fm. does contain some thin (decimetre), poorly cemented, lithic to quartz arenites. Their rare occurrence in such an outer shelf environment suggests that they are rather distal tempestites. The lateral distribution of the Cabo Menor Fm. is somewhat similar to that of the Pedruquios Fm. (Fig. 9-2) and shows a similar strong thinning from the southern to the northern limb of the San Román syncline. A somewhat similar palaeogeography for the basin as a whole is envisaged for the Santonian as for the Turonian (Fig. 8-78B) with a massive carbonate ramp that deepened towards the palaeo-Biscay to the NNE.

CHAPTER 10: THE UPPER CAMPANIAN TO MIDDLE EOCENE NANSA GROUP - SHALLOW MARINE "NUMMULITIC" CARBONATES

Upper Campanian to Middle Eocene deposition was dominated by shallow marine, benthic foraminiferal limestones of the Nansa Group (Figs. 1-9 & 1-10). This Group is confined to the core of San Román syncline and to an E/W strip south of San Vicente de la Barquera (Fig. 1-6). A summary of the lithostratigraphic sequences and correlations is shown in Fig. 10-1.

10.1 Nansa Group Formations and Environments

The first six formations described below come from the San Vicente area and the latter three from the San Román syncline (IGME Comillas and Torrelavega maps; Fig. 9-1):-

1) The Upper Campanian/Maastrichtian/Palaeocene Muñorrodero Formation

This occurs in the San Vicente area and is defined as the conglomerates, sandy dolomites and miliolid limestones above the Campanian marls/marly limestones (the Cabo Menor Fm.) and below the alveolinid packstones of the Ilerdian Estrada Fm. The term was first proposed by Garcia Mondejar et al., (1982). The base of the formation consists of a local, thin (2-5m) breccia of subjacent clasts and sandy limestones. Above this are rather sandy, marly limestones with ostreids and bivalves. These pass up into metre-bedded, pale yellow-weathering, sandy, benthic foraminiferal/echinodermal/bryozoan pack/grainstones (e.g. Fig. 10-2). The microfauna of these packstones has yielded an Upper Campanian age (*Orbitoides tissoti* Zone, Carreras & Ramirez, 1971). Other fauna include bivalve cortoids, rotalids, arenaceous foraminifera, gastropods, rhodophyta together with glauconite grains. Local graded beds occur within this calcarenite (probably storm deposits) and the lithology, fauna and stratigraphic position suggest it was deposited in a middle shelf calcarenitic belt into which considerable siliciclastic sand was coming from the hinterland to the S/SW.

Higher in the Muñorrodero Fm., the carbonate becomes progressively more dolomitized until it is entirely replaced by a quartz-sandy, dolo(micro)sparite (Fig. 10-3). Rare ghosts of rotalid forams or echinoderms exist within this dolomite but no evaporites (or their moulds) have been found. Laterally, the formation is progressively more dolomitized towards the west (palaeohinterland), but to the east, in the coeval Cabo de Lata Fm. of the San Román syncline (Fig. 2-1), there is virtually no dolomitization (see below). The upper parts of the dolomite have moulds after cerithid gastropods (brackish water fauna). Taken together, all these observations point to a mixed-water dolomitization model. Fresh/brackish waters would be introduced in the marine limestone (with echinoderms) from the end-Cretaceous/basal Palaeocene regression (accounting for the brackish fauna at the *top* of the dolomite). Increased dolomitization towards the palaeohinterland and up section to the exposure horizon are characteristics of this model, as is the sparry texture (Fig. 10-3) and the replacement of a *marine* limestone. This model differs markedly from the sabkha model proposed by Garcia Mondejar et al. (1982).

The top of the Muñorrodero Fm. (Palaeocene) is a sandy, miliolid/cortoid pack/grainstone (Fig. 10-4) that marked the return of shallow marine, shelfal carbonate deposition to NW Santander. The fauna include, along with the miliolids and cortoids, echinoderms, bryozoa, alveolinids, rhodophyta, ooids, bivalves and biserial foraminifera. Mature quartz sand makes up about 10% of the rock. The diagenetic history appears to be similar to other limestones in this formation (described above).

2) <u>The Ilerdian Estrada Formation</u>

Sandy limestones rich in alveolinids and nummulites dominate the basal part (Ilerdian) of the Eocene (*Aveolina ellipsoidalis* to *A. trempina* Zones - Hottinger, 1960) in both the San Vicente and San Román areas. The formation generally thickens to the east from about 40m west of the Rio Deva section (Fig. 10-1) where it is fauna-poor and sandy, to about 120m along strike due south of San Vicente. It is dominated by quartz-sandy, benthic foraminiferal pack/grainstones that is white-weathering and generally 50-100cm bedded. Calcareous quartz sands

exist at the bottom of the Estrada Fm. and the lower parts of it are quartz sand-rich in general. Slightly higher in the formation it is a sandy micrite with about 20% fine, angular sand. Rare multilocular and rotalid foraminifera exist within it. The upper part of the formation is a sandy calcarenite with a diverse assemblage of alveolinids, nummulites, assilinids, operculinids, orbitolitids and other benthic foraminifera. The uppermost 20m of the formation contains numerous, broken corallinaceae fragments (*Lithothamnium, Lithophyllum*).

The poor preservation of sedimentary structures in the foraminiferal banks makes precise interpretation difficult, though the winnowing of fines and lack of tidal features does suggest storm wave reworking of mobile, middle shelf banks.

3) <u>The Lower Cuisian Hortigal Formation</u>

This is a spectacular sequence of coarse, rather massive, poorly cemented quartzose sands with microconglomeratic lenses, calcareous quartz sands and rare sandy limestones. It is equivalent to the "Banc à dragées" of Mengaud (1920). The unit is remarkably massive above a sharp basal contact but grades at the top into the overlying Peña Saria Fm. It varies from being about 5m thick in the west to over 40m in the east. The dominant lithology is of well-sorted, moderately rounded, coarse, quartz sand which is remarkably constant across the area. This sand occurs in trough cross-bedded cosets or is parallel-bedded with conglomeratic lags to coset bases. No obvious vertical changes occur through the sequence, though it is interbedded with calcareous sandstones in the east that contain alveolinids, orbitolitids and nummulites (IGMEa). The occurrence of flattened *N. planatus* (gamont and schizont forms) is typical of these inner shelf marine sands (cf. the Cuise sands of the Paris basin - Pomerol, 1982). This suggests that at least the eastern end of the formation is a high-energy, inner shelf, clastic sand bar. The western end of the formation possibly represents a sandy braided stream deposit.

4) <u>The Cuisian Peña Saria Formation</u>

This is dominated by sandy, benthic foraminiferal wacke/pack/grainstones that are fawn weathering and 30-100cm bedded. The lower parts are rather marly at the base but then pass up into benthic foraminiferal wackestones. Higher in the

formation, it is dominated by packstones that are less quartz sand/silt rich than the rest of the formation below. It has been dated as *Alveolina oblonga* (non-base ?) to *A. daniellii* Zone by Hottinger (1960) and IGME(a). A different biostratigraphic age has been proposed by Clavell & Reguant (1977) who consider the Peña Saria Fm. as Lower Cuisian, the Serdio Fm. as Middle Cuisian and the Colombres Fm. as Upper Cuisian from *Assilina* population sizes (probably incorrect though). The fauna in the calcarenites (Fig. 10-5) is dominated by nummulites, alveolinids, assilinids, discocyclinas and echinoderms. Minor glauconite grains, broken ooids, bivalve cortoids, asterodisci, rare rhodophyta and bryozoa occur in an angular-quartz-silt rich matrix that has been partially neomorphosed to microspar. This formation thickens to the east, and is about 90m on the El Barcenal section (Fig. 10-1). It too was deposited in a middle shelf foraminiferal bank, periodically reworked by storm waves.

5) <u>The Lower/Middle Lutetian Serdio Formation</u>

This formation is rather similar lithologically to the underlying Peña Saria Fm., though is characterised by its large (3cm plus) assilinids (*A. exponens, A. mamillata, A. spira* etc.). Its base is rather nodular with 30-60cm bedding, but this passes up into foraminiferal packstones which are very similar to those of Fig. 10-5. This formation varies from above 20m to 60m across the area, and is thicker to the east. It was deposited in a middle shelf foraminiferal bank.

6) <u>The Upper Lutetian Colombres Formation</u>

Most of this formation is lithologically similar to the underlying Serdio Fm. and is mainly made of slightly sandy, benthic foraminiferal packstones that are decimetre to metre bedded. These packstones (Fig. 10-6) are dominated by nummulites, alveolinids, asterodisci, discocyclinas, operculinas and assilinids. There are also bivalve cortoids, glauconite pellets, rare bryozoa, pectins, miliolids, echinoderms, ostreas, terebratulids, *Schizaster*. These calcarenites were deposited in middle shelf foraminiferal banks, locally reworked by storm waves. There is a general alternation on a metre-scale between pack/grainstones with a microspar matrix and micritic wacke/packstones, reflecting high- and moderate-energy

conditions respectively. Towards the top of the formation it becomes progressively richer in flatter discocyclinas, micrite, rhodophyta, glauconite pellets, quartz silt and has a generally more fragmented, broken fauna. These are all indicative of deeper water deposition, formed towards the outer part of the middle shelf. The formation is about 90m thick on the El Barcenal section (Fig. 10-1), though it does thicken to the east. To the west it is generally slightly more dolomitized.

The three formations within the San Román syncline, NW of Santander city, will now be considered:-

7) <u>The Upper Campanian/Maastrichtian Cabo de Lata Formation</u>

The sandy calcarenites and rather local calcareous quartz sands of the Upper Campanian/Maastrichtian were first termed the Cabo de Lata Fm. by Garcia Mondejar et al., (1982). On the type section (157m thick; southern limb of the San Román syncline) the lower 37m are Upper Campanian and consist of a fawn-weathering, sandy, pack/grainstones rich in benthic foraminifera (orbitoides etc.), bryozoa, echinoderms and cortoids together with rarer ostreids, pectins, corals, glauconite pellets, *Lithothamnium* and silica nodules.

The Maastrichtian sequence (120m on the northern flank of the San Román syncline) is lithologically similar to the underlying Upper Campanian sequence though is generally more dolomitized, especially on the northern limb section. This formation was deposited in middle shelf calcarenitic banks that received considerable terrigenous sand input. Through the Maastrichtian a general regression occurred ending in inner shelf/coastal deposits. The formation thins considerably from 157m on the southern limb to only 52m on the northern limb of the San Román syncline.

8) <u>The Dano/Montian San Juan Formation</u>

This is a strongly dolomitized formation and can be subdivided into a lower quartz-sandy, rather massive Danian dolomite and an upper, slightly dolomitized, Montian, wacke/packstone calcarenite. The lower, Danian, dolomite is lithologically similar to that described from the Muñorrodero Fm. (Fig. 10-3). At its very top there is a *Microcodium* horizon, probably formed by calcification of

mycorrhizae, symbiotic associations between soil fungi and cortical cells of higher plant roots (Klappa, 1978). Such clear indications of subaerial exposure, combined with all the evidence for this end-Cretaceous/basal Palaeogene dolomitization presented under the Muñorrodero Fm. (which apply to the Danian of the San Juan Fm.) suggest that these are mixed-water dolomites with meteoric invasion in end Danian times into shelfal limestones.

The Montian calcarenites are slightly dolomitized, sandy and 20-100cm bedded. They contain a diverse marine assemblage of rotalids, rhodophyta, operculids, bryozoa, echinoderms, nummulites, alveolinids, *Quinqueloculina* and miliolids. It was deposited in middle shelf, marine environment.

9) <u>The Thanetian Sancibrian Formation</u>

This formation is only about 7m thick on either limb of the San Román syncline, and is lithologically and environmentally similar to the underlying San Juan Fm. It is, however, noticeably richer in both glauconite and rhodophyta. Above it in the San Román sections are the Estrada and lower part of the Peña Saria Fms., which are described above. They noticeably thicken towards the southern limb (Fig. 10-1), indicating that this rim syncline (and others ?) were still active during Nansa Group times.

CHAPTER 11 - THE UPPER EOCENE/OLIGOCENE SAN VICENTE GROUP - A CARBONATE SLOPE-APRON-REEF DEPOSIT COEVAL WITH THRUSTING AND KEUPER PIERCEMENT

The San Vicente Group (Fig. 1-10) only occurs on the Cap Oyambre headland and to the SW of San Vicente de la Barquera (Fig. 1-6; see IGME Comillas map). Continuous exposure is confined to three coastal sections on the Cap Oyambre syncline (Fig. 11-1 and located on Fig. 11-2). The Group is over 630m thick (top unseen). Similar deepwater Upper Eocene/Oligocene reef/carbonate slope/apron deposits do not occur elsewhere in northern Spain, and have never been studied from a sedimentological/tectonic approach except by the author (Hines, 1984, 1985a, b). All earlier work is biostratigraphic and includes studies of the macrofauna by Mengaud (1920), *Nummulites* by Karrenburg (1934), rhodophyta by Miranda (1935), *Nummulites* by Maldonado et al. (1970) and Mary (1979), planktonic foraminifera by IGMEa, b and bryozoa by Buge & Reguant (1974). They are coeval with Keuper piercement, thick- and thin-skinned thrusting and hence give zonal dating to the initial, compressional tectonics in NW Santander. The main compressional deformation occurred after the San Vicente Group was deposited (i.e. post Oligocene), and was the time of the main Pyrenean deformation in the area (and rest of the Basco-Cantabrian basin). A general summary section of these flyschoid deposits from the Playa de Meron section is shown in Fig. 11-3.

11.1 Stratigraphy of the San Vicente Group (Fig. 2-2)

The type section for this Group is the Playa de la Jerra (Fig. 11-1) which has the formations indicated on it. The oldest formation of the deepwater San Vicente Group is the La Acebosa Fm. which is an off-white, silty marl (Fig. 11-4) with rare thin-bedded turbidites on the Cap Oyambre sections. The Playa de la Jerra section has yielded a diverse, planktonic foraminiferal assemblage (IGMEa, b) which clearly gives a Biarritzian age (= uppermost Middle Eocene) of the *Truncorotaloides rohri* BOLLI Zone (= P14 zone of the standard DSDP stratigraphy of

Blow, 1969). The presence of *Nummulites.* aff. *biedai* SCHAUB (Reguant & Truyols, 1968) in a disorganized resedimented conglomerate near the top of the La Acebosa Fm. (Fig. 11-1) confirms a Biarritzian age (cf. Mary, 1979).

Flysch sedimentation of the red marls/carbonate mass-flow deposits of the Cap Oyambre headland starts with the Upper Eocene Oyambre Fm. (Fig. 1-10). The section on the Playa de la Jerra between the start of flyschoid red marl deposition (Fig. 11-1, Playa de la Jerra - 160m) and the base of major resedimented conglomerates/calcarenites (255m) has yielded Upper Eocene (= Priabonian *s.l.*) planktonic foraminifera of the *Globigerapsis seminvoluta* KEIJZER (=P 15), *Globorotalia cerroazulensis* COLE (=P 16) and *Cr. centralis* CUSH. & BERM. (=P 17) Zones (IGMEa, b). An Upper Eocene macroforaminiferal assemblage has been described from here by Mary (1979).

A major sedimentological and biostratigraphic hiatus then occurs on all the sections (Fig. 11-1) - at 245m on the Playa de la Jerra, at 262m on the Playa de Meron and at 148m on the (syn-sedimentary) faulted base of the Playa de San Vicente section. This corresponds with the Rupuente megabed on the Meron section - a major, 90m thick debris flow (Facies 12 of Chapter 11.2) and major mass flows on the other sections. This is the base of Oligocene Meron Fm., which is lithologically similar to the Oyambre Fm. However, biostratigraphically, it has been given an Oligocene age by Mengaud (1920), Maldonado et al. (1970) and Mary (1979) on macroforaminifera and an Upper Oligocene age by IGMEa,b, Buge & Reguant (1974) and Reguant pers. comm. 1986. The Upper Oligocene age is based on planktonic foraminifera and IGME(a, b) have dated the Playa de San Vicente section (Fig. 11-1) as *G. opima opima* BOLLI and *G. ciperonsis ciperonsis* BOLLI Zones at the base (147m to 360m; = P 21a & b) and the upper part as uppermost Oligocene (*Catapsidrax dissimilis* CUS. & BERNG., = *G. angulisuturalis* Zone, = P 22) with the incoming of the first *Globigerinoides* (*obliqua* BOLLI). Hence, there is a biostratigraphic hiatus in the Lower Oligocene with the *G. tapuriensis* (= P18) to *G. ampliapertura* (= P20) Zones missing.

Just 3km south there is an autochthonous outcrop of a pink-weathering, massive, reefal boundstone with shallow marine rhodophyta, bryozoa, larger benthic foraminifera and scleractinia (= "caliza rosada con Políperos" of Mengaud, 1920) that I have termed the Los Via Fm. (Hines, 1985a). Mengaud (1920) and IGME(a) considered this rock to be Upper Eocene in age, but undoubted Oligocene foraminifera (Morley Davies, 1971) of *N. intermedius* ARCH., *N. boucheri* HARPE, *N. vascus* JOLY-LEYM. have been found by Karrenburg (1934). The base of the Los Via Fm. is certainly Upper Eocene in age as it conformably overlies the Biarritzian La Acebosa Fm. SW of Lamadrid (54449723) and clasts of it have been found in the basal rudaceous mass flow deposits of the Upper Eocene Oyambre Fm. of the Playa de la Jerra (Fig. 11-1, 188m). The top of the formation is certainly Oligocene as Oligocene *Nummulites* (op. cit.) have been found in the reefal strata around Los Via and in clasts in the Meron Fm. It is possible that no Upper Eocene Los Via Fm. is preserved in the southern, autochthonous area due to erosion or tectonic removal (though more research is needed on this). In conclusion, after the deposition of the Biarritzian La Acebosa Fm., the area differentiated into a deepwater, flyschoid Oyambre Fm. and a *coeval* shallow water, Los Via Fm. reef. A major hiatus exists in the Lower Oligocene of the flyschoid sequence but Upper Oligocene flysch in the Meron Fm. coexisted with the youngest part of the Los Via Fm. (Fig. 11-43).

11.2 Facies and Facies Interpretation of the San Vicente Group

The facies of the San Vicente Group can now be considered in roughly stratigraphic order. Only the initial three facies are found in the La Acebosa Fm.

Facies 1) Massive, off-white marl (Fig. 11-4)

The massive, off-white, quartz silty marl of Fig. 11-4 is the dominant lithology of La Acebosa Fm., accounting for over 95% of the section on the Playa de la Jerra (Fig. 11-1, 0m to 160m). Towards the palaeohinterland (S/SW at La Acebosa - 54189742) the La Acebosa Fm. becomes more clastic rich and slightly dolomitized, especially at the base. Rare carbonaceous fragments, *Plicatula*, gastropods and pteriomorph bivalves also occur here. The marl contains 40-70%

$CaCO_3$ (IGMEb) with the rest of it dominated by either quartz silt or terrigenous clay. It has an abundant, diverse, planktonic foraminiferal assemblage including *Globigerina*, *Globigerapsis* and *Truncorotaloides* (IGMEa). Mixed with them are benthic foraminifera, with *Cibicidoides* which is known to be an outer shelf/slope Barbadian foram. (as opposed to a shallower Rupelian foram. - Miller, 1982) and exists in stenohaline depths of about 250m-2000m. The marl is remarkably massive due to intense bioturbation. In the slightly less bioturbated parts of the basal Playa de la Jerra section, *Zoophycos* has been found.

The marly nature of the formation, foraminifera and ichnofauna suggest outer shelf deposition in a basin relatively starved from terrigenous or resedimented detrital input that extended in rather uniform manner across NW Santander. It was deposited by hemipelagic fallout of both biogenic (e.g. planktonic foraminifera) and terrigenous (e.g. clay) materials by vertical settling with a moderate contribution from land run-off.

Facies 2) Thin-bedded, fine sand/silt turbidites

Rare, thin-bedded turbidites exist within the La Acebosa Fm. (e.g. Fig. 11-4). They are characterised by sharp bases that have numerous flute structures on them, along with rare prode marks or feeding trails. They show only slight to no grading internally. Most are 0.5-10cm thick and are dominated by parallel lamination on a mm- to cm-scale internally, though rare current ripple cross-lamination has also been found. These beds are fine-grained, thin-bedded turbidites of the Bouma T_{bcde}, T_{cde} and T_{de} type, and are not muddy enough to be considered under either Stow & Shanmugam's (1980) or Piper's (1978) muddy turbidite subdivisions. They are dominantly composed of subangular, quartzose fine sand/silt (Fig. 11-5) and are hence allochthonous. The flute structures show a unimodal flow direction towards the ESE (Fig. 11-6), away from the palaeohinterland and axial along the structural grain of the basin (see below).

Facies 3) Disorganized resedimented nummulitic conglomerate/breccia

Towards the top of the La Acebosa Fm. on the Playa de Meron (Fig. 11-1: 103m) and the Playa de la Jerra (122m) sections there is a 2-3m thick

disorganized, resedimented "nummulitic" conglomerate (nomenclature after Walker, 1975). It is a clast-supported, non-graded or stratified bed with subangular to subrounded, poorly-sorted, 1-20cm long (mean about 5cm), rather monogenetic clasts of nummulitic pack/grainstone in a quartz sandy, nummulitic pack/grainstone matrix. It has a sharp top and bottom and has clasts of rhodoliths, echinoderms and cortoid bivalves. The foraminifera are a *coeval* shallow-water assemblage including ***Nummulites*** aff. *biedai* SCHAUB, *N. striatus* BRUG, *A. fragilis* HOTT, *N. aturicus* LEYM and *N. crusafonti* REG.-CLAV. (Reguant & Truyols, 1968). These are a *coeval* Biarritzian neritic fauna (Mary, 1979) and indicate long-distance transport from neritic zones (to the S/SW/W?) by high-density turbidity currents/debris flows with final rapid resedimentation by frictional freezing of the basal traction carpet (Walker, 1975; Nardin et al., 1979; Hiscott & James, 1985). This indicates an initial "instability" in the source area, though only coeval, neritic, lithified carbonate and unconsolidated, shoal-derived foraminifera were available to be resedimented in the Biarritzian. This indicates that the main, compressional unroofing of the underlying strata had not started by the end of the Biarritzian, but *does* start in the initial, Upper Eocene flyschoid sediments of the Oyambre Fm., described below. Facies 4 to 15 belong exclusively to the Oyambre and Meron Fms..

Facies 4) Massive, red/brown, hemipelagic marl (Fig. 11-7)

This is the dominant lithology of the Oyambre and Meron Fms. and accounts for between 30% (minimum) and over 98% (maximum) of the flyschoid sediments, though the proportion does vary considerably up, and between, sections (Fig. 11-1). It is remarkably uniform in composition, and is always a massive, dull red/brown-weathering marl that is millimetres to several metres thick between mass flow deposits. In general, it is a silty clay grade marl with only a few % biogenic sand-grade detritus and a $CaCO_3$ content of between 10 and 90%, though generally between 20% and 60% (IGMEb). Intense bioturbation has completely homogenized the marl, though rare *Zoophycos* has been found in the marl-rich uppermost Oligocene sequences of the Playa de la Jerra (544m) and the Playa de San Vicente (531m). It has yielded a diverse, abundant planktonic foraminiferal

assemblage (IGMEa,b) including *Globigerina, Globinella, Globorotalia, Catapsidrax* and towards the top of the Playa de la Jerra section, *Globigerinoides*. Such diverse and abundant assemblages are typical in waters deeper than the shelf/slope break (at around 200m depth) and the occurrence of *Globigerinoides obliqua* BOLLI indicates a low-latitude, tropical belt of deposition (Olson, 1982). The benthic foraminifera include numerous *Uvigerina, Cassidulina, Gyroidinoides, Amphistegina,* miliolids and textularids of a deepwater Barbados fauna (as opposed to a shallower Rupelian fauna) which occur in an outer shelf/slope/bythal depths of 250m-2000m (Miller, 1982) by comparison with Recent faunas from the Red Sea and Caribbean (Sen Gupta, 1982; Miller, 1982).

This marl is a hemipelagic, "background" sediment which formed by slow accumulation by vertical settling, with substantial terrigenous clay, carbonate mud and foraminiferal input. Such bioturbated marls demonstrably close to coeval neritic deposits have been termed periplatform ooze (Schlager & James, 1978; McIlreath & James, 1984). On the slopes of the Recent Bahama Bank, the overproduction of the neritic shelf is transported (semi)permanently onto the deeper water slope/apron (Boardman & Neumann, 1984) and sequences with highly comparable percentages (about 60%) of hemipelagic ooze to those here have been recorded from there (Mullins, 1978). The red/brown colour of hemipelagic marl deposits has been noted from Recent (and ancient) sediments deposited below the CCD (Stow & Piper, 1984). However, the high $CaCO_3$ % (>>10%) and the preservation of calcareous foraminifera, lack of *Nuttallides* foraminifera and the presence of uncorroded skeleta in the mass-flow deposits indicates that these beds were not deposited below the CCD. The red colour is probably due to coeval erosion of incompetent red, Keuper evaporitic mudstone which are known from the San Vicente piercement (54309753) and elsewhere. This idea is supported by the occurrence of Keuper-derived, euhedral, inclusion-rich quartz crystals in the rudaceous mass flow deposits (see below).

<u>Facies 5) Sandy, calcarenitic, pack/grainstone turbidites</u>

After the hemipelagic marl, these are the most abundant deposits of the Oyambre and Meron Fms., constituting about 30% of the sequence. In general, they form as rather monotonous interbeds with the red/brown hemipelagic marls (Figs. 11-7, 11-8 & 11-9). Their mean thickness is 6cm on the Playa de Meron section (Oyambre Fm.) and they have been informally subdivided into thin-bedded calcarenitic turbidites less than 10cm thick and thick-bedded ones over 10cm thick. These sandy calcarenites always have sharp, virtually planar bases that have abundant ichnofauna (positive moulds), rare flute, prode or groove marks, very rare parting lineations and exceptionally rare soft-sediment deformation load-structures (Fig. 11-10). The Oyambre Fm. flute marks show a uniform flow direction to the ESE (Fig. 11-6).

The ichnofauna on the base of the calcarenites contains straight or branched *Granularia sp.* (very common - Fig. 11-10), *Paleodictyon sp.* (Fig. 11-11), *Palaeodictyon regulare* Sacco, *Palaeodictyon minimum* Sacco, *Helminthoida crassa* Schafhautl (Fig. 11-12), *Scolicia sp.*, *Scolicia prisca* De Quatrefages, *Sabularia simplex* Ksiazkiewicz, *Megagrapton irregulare* Ksiazkiewicz and *Fustiglyphus* Vyalov amongst others (identifications kindly made by Peter Crimes, Liverpool Univ.). Such an assemblage clearly falls into the *Nereites* ichnofacies of Seilacher (1967), his deepest ichnofacies. There is a total lack of shallow-water *Skolithos/Cruziana* etc. ichnofauna and a remarkably consistent *Nereites* assemblage throughout the Oyambre and Meron Fms., indicating relatively constant ecological/depth conditions throughout flysch sedimentation (cf. Zumaya flysch of Crimes, 1977). This flyschoid setting is typical for the *Nereites* ichnofacies (Frey & Pemberton, 1984) where quiet, stenohaline, oxygenated waters support a complex grazing and patterned feeding-dwelling structures which reflect highly organized, efficient behaviour in a low-stress environment. Spreiten structures are typically nearly planar and mainly post-depositional (Fig. 11-11 & 11-12). Numerous crawling-grazing, intrastratal traces and sinuous faecal castings (e.g. *Helminthoida*) exist with the fauna as deposit feeders or scavengers, although many formed microbe cultures with virtually permanent, open graphoglyptid domiciles (e.g. *Paleodictyon*). The relatively high

diversity and high abundance of this *Nereites* association suggests rather proximal, bathyal, flysch deposits, rather than sedimentation in more distal regions in truly abyssal depths. Although the *Nereites* association is clearly a deepwater one, Wetzel (1984) has pointed out from studies of the Recent and ancient that a short recurrence time between turbiditic emplacement and a significant non-biogenic content/high sedimentation rate in the associated marls are more important direct controls on *Nereites* development than depth *per se*. This idea is directly supported in this sequence as the turbidite-rich strata that make up most of the Oyambre and Meron Fms. have a uniform *Nereites* ichnofauna whereas only in the turbidite-poor marls at the top of the sequence is the "deep" burrower *Zoophycos* found. Wetzel (1984) also points out that at depositional rates less than about 400m/Ma there is complete bioturbation of deep-sea marls/muds. The mean, compacted depositional rate here is about 640m/16Ma = 40m/Ma, and so accounts for the intense bioturbation of the marls. Such slow depositional rates are highly comparable to compacted modern or other ancient (hemi)pelagic deposital rates (Scholle et al., 1983) and also indicate that the mass flow deposits have not been thick/frequent enough to generate the exceptional rates seen in some other flysch sequences.

Internally these calcarenites are relatively massive, with only local parallel lamination on a mm- to cm-scale in some beds. Clear fining upwards trends are generally rare, though are present in some of the thicker calcarenites. The tops are generally sharp and slightly to intensely bioturbated.

Petrographically, these are moderately sorted, quartz sandy calcarenites (Fig. 11-13). The dominant allochems are subangular to subrounded quartz grains, broken corallinaceae, bryozoa, echinoderms and larger benthic foraminifera (*Nummulites, Discocyclina, Asterodiscus, Actinocyclina, Fabiana, Campmanina* and in the Meron Fm. these and *Lepidocyclina*). Minor quantities of broken bivalves (with rare micritic envelopes), echinoderms, broken gastropods, glauconite pellets, quartz sandy micrite extraclasts and rotalids also exist. The proportion of quartz sand varies considerably between beds and is locally the dominant allochem, making the

rock a skeletal sandstone. These skeleta are clearly derived from an allocthonous, shallow-water (<50m? depth), warm-water, coralgal setting. These calcarenites contain hardly *any* autocthonous, slope-derived clasts except for *very rare*, granule-grade red marl rip-up clasts. The skeleta are *not* lithified lithoclasts and have age ranges coeval with the ages derived from the planktonic foraminifera of the interbedded marls (Mengaud, 1920; Mary, 1979 and IGMEa,b). Such shallow water, coeval skeleta can only be derived from neritic shoals of uncemented calcarenite (Fig. 11-43). These calcarenites received considerable terrigenous, quartz sand input and clasts of shoal-derived calcarenite have been found within the Los Via Fm. and in derived clasts of calcarenite coated in Los Via Fm. rhodophyta in the Meron Fm. (Fig. 11-15). The shoals are repeatedly reworked by storms and tides in bank margins (Hine, 1977; Hine & Neumann, 1977) and could easily be reworked off the edge of the narrow shelf (Fig. 11-43) and generate these "event" deposits (see below).

The quartz grains of these beds can be subdivided into three genetic groups. The dominant quartz grains are single crystals with slightly undulose extinction. Polycrystalline quartz (Fig. 11-20) has rather sutured grain contacts and is clearly ultimately derived from a metamorphic source, though was involved in some neritic shoal reworking in this case (see below). A third, rather minor group are euhedral to subhedral quartz grains rich in inclusions. This is authigenic quartz derived from the first-cycle erosion of Keuper clays, which does contain such quartz crystals (Cater, 1984).

These sandy, biomicritic pack/grainstones generally have a sparry matrix with equant microspar and local drusy spar infilling larger voids (Fig. 11-14). This spar varies from non-ferroan to ferroan. The latter is similar to that found in tectonic, calcite veins cross-cutting the strata, and is clearly a late diagenetic cement (Fig. 11-14).

The neritic calcarenite composition of these beds and their sedimentary structures (*op.cit.*) clearly indicate that these are not *in situ* carbonates but "event deposits" - carbonate contourites, tempestites or turbidites. These beds are

unlikely to be contourites as they 1) are ubiquitously sourced from the shallow-marine area *up* (not parallel) to slope and have virtually no slope-derived material; 2) are laminated and, on rare occasions, normally graded; 3) are hardly ever bioturbated internally (no syn-sedimentary bioturbation) and show little degree of internal reworking and 4) the vertical sequence is *not* highly irregular. These criteria (and others) were used by Stow (1979) and Stow & Piper (1984) to positively identify turbidites as against contourites. Both types of event deposit can flow parallel to slope, as the flyschoid ones here did.

Carbonate tempestites have a number of distinctive differences from their turbiditic counterparts (Aigner, 1982, 1985; Seilacher, 1982; Walker, 1985). The key small-scale sedimentary structure that is distinctive of storm-wave-derived tempestites are wave oscillation ripples which have *never* been found in this flysch. Further criteria that positively identify turbidites as against tempestites include: 1) no spill-over ripples have been found; 2) numerous washed-out burrows of *Granularia*, *Paleodictyon* etc. are present though; 3) a remarkably consistent unidirectional tool and flute cast direction; 4) the presence of flute but no pot and gutter casts; 5) the presence of climbing ripple cross-lamination and asymmetric current ripples; 6) complete lack of hummocky or swaly cross-stratification; 7) a lack of rip channels; 8) the presence of *deepwater* (slope/apron - see below) ichnofauna, benthic foraminifera, planktonic foraminifera but a lack of indigenous, shelfal skeleta/deposit-feeding ichnofauna suggest deposition beyond the shelf slope break/storm wave base where event flows, even if initiated by storms, turn into turbiditic flows (Seilacher, 1982; Walker, 1985). The weight of evidence does not seem to support a tempestite origin for these beds but suggests a turbiditic origin for them.

The key problems with the turbiditic model are the relative lack of normal grading in the calcarenites and lack of well-developed Bouma sequences (characteristic of low-density turbidity currents). These features have been noted from other ancient calcarenite event deposits (e.g. Cook & Mullins, 1983). The lack of normal grading could well be due to a rather uniform grain size distribution in

the original sandy calcarenite of the neritic shoals and the fact that these are probably high-density, unconfined turbidity currents (sensu Lowe, 1983), which usually lack well defined Bouma-type vertical sequences.

Facies 6) Siliciclastic turbidites (Fig. 11-16)

These turbidites are confined to the Meron Fm., especially the sequence just above the basal megabeds of the formation (Playa de la Jerra - 270m, Meron - 371m, San Vicente - 391m). There is a genuine gradation between sandy calcarenitic turbidites to calcareous sandstone turbidites. The sandstone turbidites are dominated (>80%) by medium to fine quartz sand and have rare carbonaceous fragments in them (land source close by - i.e. a narrow shelf). Although they are basically similar to their more carbonate-rich counterparts, they do generally have better developed sedimentary structures than them. They range from rather massive T_a, T_{ab} turbidites (Fig. 11-16) to those that show perfect T_{abe} Bouma sequences (Fig. 11-17). A few examples of climbing ripple cross-lamination have been found in the sandstone turbidites of the Playa de San Vicente (386m), indicating high suspension load deposition of Type 2 ripples (McKee, 1965). Interestingly, the cross-lamination is flowing northward, indicating a direct source from the shallow shelf to the south (Fig. 11-43).

Facies 7) Cross-bedded and parallel-bedded calcarenites (Fig. 11-18)

These are unique to the lower part of the Oyambre Fm. on the Playa de la Jerra section (199m). They consist of well-bedded, parallel-laminated/bedded and cross-bedded calcarenites that are well-sorted calcarenitic grainstones with a similar fauna and texture to the calcarenitic turbidites described above (Facies 5). The cross-bedding flow direction is consistently towards the ESE, parallel to that from the flute structures of the calcarenitic turbidites. Such cross-bedding is only formed in the upper part of the lower flow regime, under moderate current velocities. It is now known (Andrews et al., 1970) that deepwater carbonate slopes can experience semipermanent or sporadic high velocity currents on the seafloor that are capable of moving calcarenite dunes, especially in confined, narrow basins (as here). Similar deepwater currents have been recorded from the Monterey

Canyon, off California (Shepard et al., 1979). These currents would have been able to erode the turbidite/marl alternation, winnow out the fines and redeposit the calcarenite as cross-beds. This evidence, and the virtual total lack of intraformational, tabular turbiditic calcarenite rip-up clasts suggest that the calcarenites were not rapidly cemented on the sea floor (too deep - see below).

Facies 8) Granule calcarenite/rudite turbidites (Fig. 11-19)

These are several decimetres thick, clast-supported, normally graded (rarer disorganised examples do occur) resedimented granule calcarenite/calcrudites. In general, they show sharp bases, followed by a relatively granule-rich, grain-supported base that grades up into a calcarenite with floating granule grade clasts (Fig. 11-19). Rare examples of coarse-tail, inverse grading at the very base of the flow occur locally. The calcarenitic part of the flow can be rather massive or locally laminated with only poorly to non-existent grading within the calcarenite grade material (cf. calcarenitic turbidites). The granules are commonly of quartz/quartzite which have been very well rounded and smoothed (Fig. 11-20) before incorporation into the flow. The calcarenite is of similar composition to that in the calcarenitic turbidites. A number of the other granule turbidites contain other rounded and smoothed extraformational clasts, mainly of Los Via Fm., Aptian Novales Mb. dolomite, Aptian Reocin Fm. limestone or Upper Cretaceous Sardinero Gp. sandy marls (see below). These granule flows are rather sheet-like at outcrop (10m) scale (Fig. 11-16), though some examples of small channelised gullies of granule calcarenite (Figs. 11-19 & 11-10) have been found on the Playa de Meron section (Fig. 11-1; 160m-190m). These gullies are only 20-60cm deep and 2-5m wide, and are more small gullies than major channels. Such flows are thought to be the products of high-density turbidity flows, with minor grain-to-grain or matrix support in the basal, granule-rich part of the flow (Davies & Walker, 1974; Walker, 1975; Pickering et al., 1986).

Facies 9) Normally graded conglomeratic calcarenite (Fig. 11-21)

These are fundamentally similar to the granule calcarenites (Facies 8), though are coarser grained with clasts up to cobble grade (25cm) generally, though

rare outsize boulders up to about 4m diameter do occur (Fig. 11-21). In general, these flows show basal coarse-tail inverse grading with a sheared, traction carpet locally (Fig. 11-22). The coarse clasts are concentrated in the lower part of the bed which rapidly gives way upwards into fine pebbles and calcarenite. The upper calcarenite is either massive or parallel-laminated/bedded (Fig. 11-21). Such beds have been termed "normally-graded resedimented conglomerates" by Walker (1975) or "stratified, normally-graded breccias" by Krause & Oldershaw (1979). They too are considered to be primarily products of high-density turbidity currents (Lowe, 1982; Pickering et al., 1986), though the presence of such large outsize clasts suggests some degree of matrix support, at least in the lower part of the bed (poorly developed traction carpet ?). These outsize clasts do not develop any accumulations of coarser than average clasts to their sides (upstream or downstream of them), suggesting that they finally accumulated by vertical sinking as the flow "froze". A particularly common lithology of the outsize, metre-sided boulders are Los Via Fm. limestones, which have been previously interpreted as *in situ* bioherms (Maldonado et al., 1970), but are clearly resedimented clasts within conglomeratic mass flow deposits.

The rudaceous clasts of these flows are *exceptionally polymict* with clasts of the *entire Phanerozoic* column seen locally in NW Santander present. The only exception to this are lithologies (e.g. the Keuper) that are too incompetent to form lithified clasts, but these are most probably present as disaggregated sand, mud etc. The following lithologies have been found as rudaceous clasts: the Ordovician Barrios Quartzite, the Carboniferous Caliza de Montaña limestone, the Triassic Buntersandstone, the Jurassic Carniolas dolomite, Liassic limestone, the Wealden sandstone, the Aptian Umbrera Fm., the San Esteban/Reocin Fm. rudistid limestone, the Las Penosas/Bielba Fms. tidal sandstone, the Barcenaciones Fm. limestone, the Cenomanian Altamira Fm., the Sardinero Gp. marls, the Palaeocene Muñorrodero Fm. dolomite, the Pena Saria/Serdio/Colombres Fm. nummulitic limestone and the Los Via Fm. reefal limestone. Another, somewhat minor, clast type are angular, red marl rip-up clasts that usually weather out to give the rock a

vesicular appearance (Fig. 11-22). The origin of this polymict assemblage is from local unroofing of the Phanerozoic sequence by coeval Keuper piercement and thick- and thin-skinned thrusting. The thrusting must include reactivations of the Hercynian thrusts - the Jerra Thrusts (Figs. 2-3 & 2-4) and the Escudo de Cabuerniga Thrust - as clasts of basement from beneath the Keuper (hence not brought up by Keuper piercement) have been found in the flows. This further demonstrates that the main deep-seated thrusting was thick-skinned, basement-involved reactivations of Hercynian thrusts in NW Santander, rather than the normal thin-skinned thrusting on the Keuper décollement that is normal in the Southern Pyrenean belt (Williams, 1985 amongst others).

A remarkable feature of these clasts is that they are overwhelmingly well rounded (/subrounded) and very well smoothed (Figs. 11-21, 11-23 & 11-26) despite their indurated nature. Such a degree of rounding and smoothing is highly unlikely to be a product of the mass flow process *per se* (with a considerable degree of matrix support). The attrition most probably occurred in alluvial/littoral areas to the S & W of Cap Oyambre (Fig. 11-43) where the underlying formations were tectonically exposed, eroded, repeatedly reworked/rounded and finally resedimented into the deepwater flysch trough rudaceous flows.

<u>Facies 10) Disorganized clast-supported conglomerates</u> (Fig. 11-23)

These are granule to boulder grade, poorly-sorted, clast-supported conglomerates with a sandy calcarenitic matrix. The clast lithologies show a similar diversity to those of the graded conglomeratic calcarenites (Facies 9), though do include a number of subangular clasts. Most clasts, however, do show a considerable degree of reworking. This is a thick-bedded facies ranging from under 1m thick to over 9m. They have sharp, flat bases that often show well-developed coarse-tail inverse grading (Figs. 11-24 & 11-25) and sheared traction carpets (Fig. 11-22). Internally the flow is disorganized, non-stratified, non-imbricated and non-graded (Fig. 11-26). The upper surfaces are also relatively planar with occasional clasts projecting above the bed. They are considered to be

the product of high-density debris flows/turbidity currents in which there was considerable matrix support within a plug flowing over the sheared basal traction carpet. Such features are commonly observed in debris flows (Middleton & Hampton, 1976), though a calcarenitic matrix is less common than a mud one. It is most probable that during transport the matrix was water-rich and able to support the large rudaceous clasts in these flows, but that the very dewatering caused the flow to freeze rapidly (beyond the hydraulic jump?) and deposit these disorganized beds.

Facies 11) Disorganized rudaceous marlstone debris flows

These are 40cm-400cm thick, marl-supported, rather massive, very poorly sorted, disorganized flows with random distributions of generally subrounded clasts showing the usual polymict lithologies described above. Very few marl rip-up clasts or tabular clasts of lithified turbidites have ever been found in this facies. Large cobbles and boulders are dispersed through the matrix, but rarely do project through the top of the unit, indicating matrix strength to the flow. They are classic debris flows in which the matrix strength of the cohesive flow could transport large rudaceous clasts, but deposition occurred due to rapid dewatering and "freezing" of the flow (Middleton & Hampton, 1976; Nardin et al., 1979).

Facies 12) The "Rupuente megabed" (Fig. 11-28)

The Rupuente megabed (*new name*) is the most spectacular horizon in the San Vicente Group and is some 90m thick on the Playa de Meron section (Fig. 11-1: 267m-357m). Here, the Rupuente megabed is essentially a very large debris flow with matrix-supported, multi-decametre sized megaclasts (Fig. 11-28) in a contorted red marl/thin-bedded calcarenitic turbidite matrix (Fig. 11-29). The largest megaclast in the Rupuente bed at Meron (Fig. 11-28) is over 40m long by 15m thick by 20m wide. It is made of Cenomanian Altamira Fm. lithology. The very core of this (and other) megaclasts is relatively well-bedded, "normal" Altamira Fm. orbitolinid calcarenite (Fig. 11-30). The bedding orientation varies between individual Altamira Fm. megaclasts and is strongly discordant with respect to that within the Cap Oyambre syncline. Towards the edge of these megaclasts the bedding is progressively destroyed and the rock is a clast-supported, highly angular,

virtually monomict breccia of decimetre-sized, Altamira Fm. clasts (Figs. 11-28, 11-30, 11-31 & 11-32). Such a texture is typical of rockfall deposits (Hsu, 1974; Pickering et al., 1986), which probably formed as a tectonically-derived talus/scree fringing relatively steep submarine cliffs or slopes. The matrix to these monomict breccias is an angular sand of Altamira Fm. limestone. An unroofing of the Altamira Fm. by thrusting or Keuper piercement would produce such a tectonic breakup of this limestone and talus production. This talus slope breccia was then itself broken up into megaclasts by faults that produced slickensides on the sides of the megaclasts (Fig. 11-33). They were finally resedimented in the Rupuente megabed debris flow to their present position. A four-metre diametre clast of the Los Via Fm. has also been found on the Meron section (Fig. 11-28).

The matrix is a highly contorted thin-bedded turbidite/marl (Fig. 11-29) that tends to wrap around the megaclasts. Especially towards the edge of the flow there has been intense folding of the marl (Fig. 11-34), boudinage of the calcarenite beds and calcite spar cementation within the megaclasts. The beds just below (<10m) the bed itself have been slightly folded, probably due to rapid overpressuring as such a thick bed was virtually instantaneously introduced. To the sides of the lower part (30m) of the Rupuente megabed, one passes along strike into "normal", undeformed calcarenitic turbidites/marls. Between the Rupuente megabed and the undeformed turbidites there is a zone (about 10m wide) with intense folding of the turbidites (Fig. 11-35), with limb orientations that generally indicated lateral pressure away from the core of the megabed. A close analogue are the mass flow deposits from the Bilbao trough near Zumaya (Hanish & Pflug, 1974) where the Keuper is also thought to have pierced through to the seafloor during Upper Cretaceous times.

Similar Altamira Fm. megaclasts (Fig. 11-32) have been found in a major debris flow on the Playa de San Vicente (Fig. 11-1, 274m-295m), about 3km along strike (measured along the bedding and allowing for folding). This part of the same flow is thinner (21m), and also has more extensive disorganized conglomerates just below it. The correlation of the Rupuente megabed on the Jerra section (Fig.

11-36) is at 255m-275m (Fig. 11-1). This consists of a set of amalgamated clast-supported, graded conglomeratic calcarenites (Facies 9; Fig. 11-36), disorganized resedimented conglomerates and massive calcarenite beds (2m thick) with clasts of the Altamira Fm. (Fig. 11-36). Rare fluid-escape structures exist within it. It is capped by a fining-upward sequence of thick-bedded, and locally cross-bedded, sandy calcarenites (Fig. 11-1; 270m-285m). Basal striations on these flows and the cross bedding in this flow indicate flow to the ENE/ESE, though its character is somewhat different to the flows on Meron and San Vicente playas. The only directional indicators on these latter two are: 1) a very large fold limb on the San Vicente section (15m thick and 60m long) indicating flow to the north and 2) flute marks on the base of the disorganized conglomerate at the base of the bed which also indicates flow to the north. Hence, the Playa de San Vicente flow does seem to be a *lateral* continuation along strike of the Playa de Meron Rupuente megabed, but that on the Playa de la Jerra could well be the product of a different flow, though closely related in time due to its coarse nature, presence of Altamira Fm. clasts and occurrence at the base of the Meron Fm.

Somewhat similar carbonate megabeds have been recorded from the syn-thrusting South Pyrenean basin during the Eocene (Rupke, 1976; Johns et al., 1981; Seguret et al., 1984). These Pyrenean flows differ in that the clasts are primarily derived from *coeval* carbonate platforms, and have a general fining-upward, "megaturbidite" trend. Here most of the megaclasts were derived from much older, lithified formations. Indeed, there is a remarkable similarity between the Altamira Fm. megaclasts in the flyschoid sequence and the brecciated Altamira Fm. *in situ* near to the thrust-faults just south of Cap Oyambre headland on the Rioturbio (54839758). The Rupuente megabed is somewhat smaller than the Roncal Unit of Johns et al. (1981), though still has a huge volume exceeding 200 million m^3. Such a size of flow in such a small basin (Fig. 11-43) would still mean a catastrophic failure of much of the slope. It can be termed a "seismite" because of 1) its enormous size, 2) the brecciated and slickensided nature of the

megaclasts and 3) the derivation of most of the megaclasts from deeply buried, ancient lithologies (Altamira Fm. over 1km down at end Middle Eocene times).

Facies 13) Angular cobble/boulder breccia (Fig. 11-37)

These are very rare compared to rudaceous conglomeratic flows and consist of angular cobble and boulder clasts that can be as polymict as their conglomeratic counterparts, but are usually more monomict (Los Via Fm. clasts are most common). They are clast-supported, with either an angular rock flour or (rarely) marl matrix. Individual clasts do have slickensides on them, as well as locally intense calcite veining and fracturing. These clearly indicate a tectonic origin for these breccias, either from the coeval Los Via Fm. or from lithologies exposed by active thrusting/Keuper piercement.

Facies 14) Folded slump beds (Fig. 11-38)

Folded slump beds are very rare in this flyschoid sequence and are confined to a complex example from the Playa de la Jerra (Fig. 11-38, Fig. 11-1 at 420m) and to a very large, northward-vergent slump nose associated with the Rupuente megabed on the Playa de San Vicente (Fig. 11-1: 280m). The Jerra slump (Fig. 11-38) is 5m thick and consists of semi-coherent folded and contorted strata of medium-bedded calcarenitic turbidites that have locally broken up into isolated arcs of folds. The base is planar (glide plane?) but the top is more irregular. The San Vicente slump nose has already been described above.

Facies 15) Intraformational debris flows

A single, thick (over 47m), debris flow occurs at the top of the Playa de la Jerra section (Fig. 11-1: 559m-596m). It consists of a contorted red marl with rare (<<1%), isolated skeletal fragments (rhodophyta, foraminifera, coral etc.), rounded quartzite pebbles/granules and rare turbiditic calcarenite tabular lithoclasts (up to 4m diameter). Multimetre-scale slump noses and folds exist along with intraformational red-marl clasts (up to 18m in diameter) and glide scars. The general lithology is similar to that in Fig. 11-34. Unlike the rudaceous debris flow of Facies 11, the clasts within this debris flow are virtually exclusively intraformational and are mainly red marl ones. They show a

considerable degree of internal contortion/folding concordant to the external morphology of the clast indicating that they were only semi-lithified during flow.

These are the main facies of the flyschoid La Acebosa, Oyambre and Meron Fms. The coeval, shallow-water Los Via Fm. will now be considered:

Facies 16) Los Via Fm. - neritic, rhodophytic reef (Fig. 11-39)

This outcrops autochthonously around the village of Los Via (54489733) but allochthonously NE of La Revilla (54569751). It is a pink/red weathering, rather massive, reefal boundstone mainly composed of neritic, encrusting corallinaceae, bryozoa, larger benthic foraminifera and scleractinia (Fig. 11-40). The most dominant flora are corallinaceae, mainly *Archaeolithothamnium, Lithothamnium, Lithophyllum* and *Melobesia* (Miranda, 1935). Separate thalli do exist, though most of the corallinaceae are encrusting/fenstrate and were the dominant reef binders and builders. The bryozoa are rather ramose and locally take on a binding form. They are dominated by vinculariiformes (60-90% of the population - Buge & Reguant, 1974) with minor cellariiformes (5-30%) and subordinate (<10%) cideoniformes, reteporiformes or lichenoporiformes. The abundance of cyclostomes concords with an Oligocene age for the formation. The dominance of vinculariiformes is indicative of fully open marine conditions with an "oceanic" influence and not of epicontinental inland seas. This is clearly consistent with the picture of the Los Via Fm. as a fringing reef to the shelf open to an oceanward slope (Fig. 11-43). The scleractinia are dominated by numerous astreid and meandrinid hermatypic corals. Specimens from the astrocoeniina and faviina suborders are abundant (*Stylina, Turbinolina, Flabellum, Cerototrochus, Cyathoseris, Heliastraea* and *Stylocoenia* - Mengaud, 1920).

The reef core facies of this formation is characterised by domal, clastic-poor, massive and encrusting, hermatypic, astreid and meandrinid scleractinia, encrusting corallinaceae, ramose bryozoa and an abundance of encrusting benthic foraminifera (e.g. *Fabiania*), along with larger benthics such as *Lepidocyclina, Nummulites, Operculinoides* and rare *Discocylinas* (e.g. Figs. 11-39 & 11-41). Such communities and growth forms are typical of high-energy wave attack

reef core environments in tropical, Eocene-Oligocene reefs (Frost, 1977; James, 1983). Although no *in situ* outcrops of a back-reef environment have been found, resedimented clasts in the Meron Fm. could well represent this environment. They are composed of low-energy, gastropod-rich biomicrites with *Cerithium*, *Turritella*, *Solarium*, *Scalaria* along with ostreids (*Pycnodonta*, *Chlamys*), pectins and echinoderms (*Psammechinus*). Seaward and landward of the reef core were extensive shoals (and beaches?) of skeletal, calcarenitic, pack/grainstones derived primarily from the reef complex itself, but with a moderate, terrigenous quartz sand input. Corallinaceae-coated clasts of this calcarenite have been found in the Oyambre and Meron Fms. (Fig. 11-44), though the calcarenite was mainly resedimented as unconsolidated detritus into the calcarenitic turbidites (Facies 5) of the flyschoid formations.

Differentiating the reef core environment from the fore-reef (area between the reef and top-of-the-slope) is difficult, though the fore-reef does seem to be more *Discocyclina*-rich and coral-poor than the core facies. The neritic area produced mm to cm sized rhodoliths (Fig. 11-42) which were mainly bound by *Lithothamnium*, *Lithophyllum* and *Archaeolithothamnium* (Fig. 11-40). These can be bound together into pebble/cobble-sized aggregates (Fig. 11-42) and have been resedimented as clasts in the Meron Fm.

11.3 Environmental Interpretation of the San Vicente Group

Sedimentation of the San Vicente Group was initiated by the relative sea level rise between the Upper Lutetian Colombres Fm. and Biarritzian La Acebosa Fm. This rapid transgression at the end of Middle Eocene times is known throughout most of western Europe (Peyreblanque limestone of Biarritz; the upper Blaye limestone of Gironde; the Upper Bracklesham beds of the Isle of Wight; the Falun de Foulangues of the Paris Basin and in the Transdanubian Mountains of Hungary) and globally (top of TE2.1 of the Vail et al., (1977) sea-level curve). This suggests that the rise was primarily eustatic in origin. The possibility of it being foreland basin thrust loading with coeval relative sea-level rise seems unlikely as

the sequence is marl-dominated, there is *only one* conglomeratic bed of coeval skeleta (ie. little coeval tectonism) and this sea-level rise can be correlated over much of western Europe.

The lithology, benthic foraminifera and rare sandy turbidites all support an outer shelf model for the La Acebosa Fm. which was relatively starved of detrital input. There is a general increase in its clastic content towards the S/SW/W (palaeohinterland), but an overall similarity in facies between the outcrops around La Acebosa (54309733) and those of Cap Oyambre (54729779). This indicates that major differentiation of the shallow- and deepwater areas had not started by the end Biarritzian.

A general model for the Oyambre and Meron flyschoid Fms. is shown in Fig. 11-43, which shows the differentiation into the shallow-water Los Via Fm. and the deepwater, flyschoid Oyambre and Meron Fms.

The general sedimentary facies of the Los Via Fm. have been discussed above. Such neritic reefal bodies in which corallinaceae are the dominant reef builders and binders are well-known from modern, tropical Brazil (Jindrich, 1983), Bermuda and other Caribbean islands (Mesolella et al., 1970; Ginsburg & Schroeder, 1973; Adey, 1976), Pacific atolls (e.g. Emergy et al., 1954) and elsewhere (Racz, 1979; Wray, 1979). The fringing reef cores are dominated by corallinaceae and hermatypic corals. They generally exist in the high-energy zones of wave attack in water depths less than 20/30m (Macintyre, 1972 and references therein; Adey, 1976; Wray, 1979) and range right up through to the intertidal zone. The reefal zones are often associated with high-energy shoals of reef-derived skeleta (corallinaceae, coral, benthic foraminifera - Mesolella et al., 1970; Adey, 1976; Macintyre, 1972). These are a direct analogue to those in Santander. The only difference is in the type of skeleta present - Santander is especially rich in corallinaceae but poor in *Halimeda* (and other codiaceae) as these algae were not dominant until the Mio-Pliocene. The modern algal reefs commonly grow rapidly, produce much internal sediment, cement rapidly (mainly with HMC micrite), are dominated by corallinaceae with minor corals, bryozoa, serpulids, hydrozoa and

gastropods (Ginsburg & Schroeder, 1973; Jindrich, 1983). This bears a remarkably close resemblance to the Los Via Fm. Modern corallinaceae reefs are usually associated with rhodolith production (Mesolella et al., 1970; Newell & Bloom, 1970; Bosellini & Ginsburg, 1971; Adey & MacIntyre, 1973; Bosence, 1983). Rhodoliths generally form in between 5m and 90m water depth (Bosence, 1983), though relic or resedimented ones go deeper. They have a lamellar and/or globular growth form (cf. Fig. 11-42) and the species of corallinaceae in the Los Via Fm. indicate tropical reef sedimentation. They are usually reworked by storms and can generate beaches littered in pebble to boulder sized clasts (Newell & Bloom, 1970). These clasts have also been resedimented downslope in NW Santander as the reef was a shelf-edge fringing one and storm action could move them over the shelf-slope break. The corallinaceae also encrust any other pebble/cobble clasts in the littoral zone, and clasts of such coated grains have been found in the Meron Fm. (Fig. 11-44). Rounded, occasionally corallinaceae-coated granules/cobbles of extraformational clasts mixed with reef-derived calcarenite are known from the Recent Fernando de Noronha, Brazil (Jindrich, 1983), and these have a close analogue to the rudaceous clasts, calcarenites and conglomeratic calcarenites (Facies 9) of these flyschoid formations.

Between the reef core and the shelf slope break is the fore-reef which seems to have had similar skeleta to the reef core, but probably had numerous reef-derived skeletal banks (cf. Belize fore-reef - James & Ginsburg, 1979; Enos & Moore, 1983) and rudaceous clasts (Mutter et al., 1977). Fringing reefs tend to have narrow fore-reef zones and pass seaward via steep slopes into deeper water. Abundant examples of these slopes exist in the Recent (Enos & Moore, 1983 and references therein), though such steep to vertical slopes have only rarely been found in the ancient (e.g. Cathedral Escarpment - McIlreath, 1977). Unfortunately, any such escarpment has been thrust over in Santander, and it is only reasonable conjecture to say that such an escarpment existed (see below for fault nature of escarpment though).

The deepwater, flyschoid Oyambre and Meron Fms. can now be considered. In general, the flow of thin-bedded turbidites was to the ESE (most of the data from the Oyambre Fm. though), parallel to the axis of the basin (Fig. 11-43). There are a few indications of the more rudaceous/siliciclastic-rich flows coming from the shallow-water area to the south and west, though no indications of flow from the north or east exist. The dominant easterly-directed flow is clearly (sub)parallel to the dominant Hercynian trend for the basin, and so the lineation direction of the Hercynian thrusts was of primary importance to basin geometry.

Within this tectonic context, what was the depositional environment of these resedimented flows? The two main environments for resedimented carbonates are slope-aprons or submarine fans (Stow, 1986). The distinguishing features between these can now be considered (McIlreath, 1977; Nardin et al., 1979; Cook & Mullins, 1983). Submarine fans normally have a single, canyon-fed source as against a line source in slope systems. The lateral variability of flows and variation within an individual flow (e.,g. the Rupuente bed) suggests that different types of sediment were being supplied from different parts of the slope. This would also explain the very poor lateral correlation between beds seen only a few kilometres apart. There is no direct evidence for extensive channelling anywhere within the Oyambre or Meron Fms., except for a few, rare, very small (<60cm deep) gullies (Facies 8). Slope sequences tend to be rich in slide/mass-flow deposits, as the slopes tend to be steeper than on fans and there is no systematic, downslope pathways. In sequences that preserve the upper part of the slope with moderate cementation-lithification, numerous slump and slides exist (McIlreath, 1977; Nardin et al., 1979). However, further down the slope in areas with much less cementation, the flows rapidly pass their hydraulic jumps and break up into debris flows/resedimented conglomerates/high-density turbidity currents. This latter situation of a poorly-cemented lower slope flows is clearly what applies in the Santander case. Massive sands, often found in upper fan facies (Walker, 1978) are absent in this Santander slope system.

One of the most distinctive differences between the fans and slopes is the presence, or absence, of systematic middle fan organisation. Middle fans generate coarsening- and thickening-upward sequences from lobe progradation (Walker, 1978; Ricci-Lucchi, 1975) and fining- and thinning-upward channel sequences on a 10m scale. Such organized sequences have *not* been seen in this flysch sequence (Fig. 11-1). The only exception to this are local, poorly-developed, packets of thicker-bedded calcarenites. Occasional packets (0.5-3m thick) are to be expected from even a random sequence (Hiscott, 1981) and could also be the product of progressive footwall failure of a slump scarp that develops until the source slope becomes stable again and thicker-bedded turbidite deposition stops. This principle can be applied on a larger scale (50-150m) to the entire Oyambre and Meron Fms. (Fig. 11-1; e.g. Playa de la Jerra, 187m-254m) which has numerous such, 100m-scale, fining- and thinning-upward sequences (Fig. 11-1). These probably relate to an initial, tectonic disturbance of a stable slope, followed by rudaceous mass-flow deposits which thin/fine-upwards as the slope restabilises. There is no evidence for these being *channellised*, mid fan units. In conclusion, a slope-apron model would seem more appropriate than a submarine fan one.

Within the slope-apron model, a number of important subdivisions can be made. The clear differentiation between the shallow-water Los Via Fm. and the flyschoid Oyambre and Meron Fms. indicates that a "depositional carbonate slope margin" (sensu McIlreath & James, 1984) is inappropriate. The presence of the reefal Los Via Fm., peri-platform talus blocks of Los Via Fm. and skeletal calcarenites indicate that a "reefal by-pass" margin is a much better analogue. Mullins & Neumann (1979) have further subdivided by-pass, slope-apron margins. They point out that those margins facing the open ocean, windward side tend to be more marl rich relative to those on the leeward side as storm waves tend to rework the skeleta landward, so less is taken downslope on the windward side. They also point out that around the Bahamas the dominant bottom currents are parallel to the slope (as here) and not down it and that these can rework the bottom sediments

(cf. cross-bedded calcarenites - Facies 7). Hemipelagic marls are only preserved when not winnowed out by bottom currents, overdiluted by mass flow deposits or deposited below the CCD (none of which happened here). In the "open ocean windward (Type B)" model of Mullins & Neumann (1979) there is a steep slope to 150-200m, followed by a muddy upper slope with hemipelagics, intraformational debris flows and slumps. On the lower slope (600m-1000m depth) there were numerous biolithoclastic turbidites (up to 2m thick), much of whose detritus is made of margin by-pass, shallow-water allochems (cf. Santander). Hence, the flyschoid San Vicente Group is a windward-facing, fault-controlled, reef-carbonate slope-apron, by-pass margin.

A further set of rather direct analogies with modern carbonate slope systems can be made. In modern carbonate slopes the shallow, neritic area is the dominant source of the hemipelagic marl, especially in windward-facing margins (Boardman & Neumann, 1984 - cf. Santander). Carbonate turbidites with neritic skeleta are well known from Caribbean slopes (Davies, 1968; Bornhold & Pilkey, 1971; Bennetts & Pilkey, 1976). In the Columbus Basin, Bahamas (Bornhold & Pilkey, 1971), they are roughly tongue-shaped and extend for up to 25km from the shelf break (cf. lack of correlation in Santander). Distal turbidites are finer-grained, thinner, have less neritic skeleta, a greater HMC/LMC ratio, and less of the lower parts of the Bouma turbidite subdivisions than more proximal turbidites. In general, they carry little of the fines into the deep basin (cf. most fines in Santander are hemipelagic fallout and *not* fine tails of turbidites).

Numerous descriptions of more rudaceous, modern, mass-flow deposits exist, mainly from the Caribbean (Jamaica - Moore et al., 1976; Land & Moore, 1977; Schlager & Chermak, 1979; Belize - James & Ginsburg, 1979; Bahamas - Crevello & Schlager, 1980; Mullins et al., 1984). Beyond the base of the shelf/slope break cliff, at the top of the upper slope, there is a zone of coarse debris and neritic-derived, calcarenite in Discovery Bay, Bahamas (Moore et al., 1976) which extends down to about 200m depth. The proximal fore-reef slope off Belize (James & Ginsburg, 1979) is similarly coarser than the deeper, distal part and consists of

coalescing fans (sic) with limestone blocks up to 10m wide with coarse calcarenite. Similar large coralgal blocks (locally termed "haystacks") have been found on the upper slopes off North Jamaica (Land & Moore, 1977). However, on many other slopes, the upper part of the slope is essentially non-depositional and has abundant slump scars in hemipelagic mud. Examples include the gullied upper slope of the Tongue of the Ocean (Schlager & Chermak, 1979) in 300m-500m depth, or the canyons of the upper slope (200m-900m depth) of Little Bahama Bank (Mullins et al., 1984) with a hummocky topography and downslope slumps. In the cases of a non-depositional upper slope, the lower slope/apron receives abundant resedimented marl, by-pass turbidites and rudaceous flows (Crevello & Schlager, 1980; Mullins et al., 1984). This grades off seaward into a finer-grained apron with progressively thinner-bedded turbidites, fewer rudaceous mass flows and increased bioturbation of the marl. Differentiating between sequences with rudaceous-rich upper slopes or lower slopes is difficult, as the mechanisms that produce either of them in the Recent are still poorly understood. However, the general lack of slump scars in the argillaceous-rich lower part of the Oyambre Fm. (e.g. Playa de Meron on Fig. 11-1, 110m-150m) or the upper part of the Meron Fm. (e.g. Playa de Meron on Fig. 11-1, 490m-630m), the presence of thick marl sequences themselves, lack of cemented hardgrounds, and the ubiquitously disaggregated nature of the flows suggests that the deposition of the rudaceous portion of the flyschoid sequences are parts of lower (and *not* upper) slope and that the marl-rich parts of the Oyambre/Meron Fms. are the apron seaward of the lower slope. The author described these thin-bedded turbidites/marl-rich sequences as "base of slope" in Hines (1985a). A reasonably good analogue is from the lower slope of Little Bahama Bank (Fig. 11-45; Mullins et al., 1984) which has a "hummocky", depositional lower slope rich in rudaceous flows. The key problem with this analogue is the comparative lack of tectonic activity in the Recent Bahamas relative to that in Santander with its extraformational clasts. The Bahaman hummocks of the lower slope are mainly chaotic debris flows or coarse-grained carbonate turbidites (cf. Santander) with little cementation to generate hardgrounds. At the top of the lower slope (about

1100m depth), the coarsest flows occur with about 60% gravity flows and 40% periplatform ooze. This would have rather direct analogue with the coarser parts of the flyschoid sequence on the Playa de la Jerra (Fig. 11-1: 186m-208m; 255m-277m; 360m-372m), Playa de Meron (Fig. 11-1: 262m-420m) and Playa de San Vicente (Fig. 11-1: 138m-145m; 258m-323m). The rudaceous debris flows in the Bahamas are, at least, 2-5m thick (Crevello & Schlager, 1980; Mullins et al., 1984). They are a few to few 10's km long, distinctly tongue shaped in plan and wedge-shaped in section perpendicular to flow. Direct analogue to this exists in the Rupuente megabed which is thickest in its central section (Playa de Meron - Fig. 11-1) but thins to its lateral margins as modern debris flows do. Indeed there is usually a more rapid variation in thickness perpendicular to flow direction on a debris flow than parallel to it (Crevello & Schlager, 1980). On the modern Bahaman slopes the transition from the lower slope to the basinal apron is usually taken at the abundant rudaceous/debris flow to abundant turbidite/marl transition, as it is here in Santander. Hence all the sequences not described in the coarser lower slope are seen as basinal apron deposits.

The depth of deposition of this lower slope/apron sequence can be approximately estimated. The abundant planktonic foraminiferal assemblage of the Oyambre and Meron Fms. (IGMEa, Chapter 11.2) are typical of waters in the outer shelf/beyond the shelf slope break (around 200m depth) (Olson, 1982). The Barbadian benthic foraminiferal assemblage only exist in a similar outer shelf/slope/bathyal depths of 250m-2000m (Miller, 1982; Sen Gupta, 1982). The *Nereites* ichnofauna are characteristic of bathyal depths (200m-1000m plus). The flyschoid sequences lack wave- and storm-generated sedimentary structures such as HCS which only occur above storm wave base, which is around 100m (Walker, 1984). The resedimented flows have *not* been encrusted by corallinaceae algae. The tops of such flows would form an ideal, rocky substrate for colonization, but have not been so, because of the great depth of water. Virtually no living corallinaceae are known below about 150m depth (Adey & MacIntyre, 1973; about 100m - Bosence, 1983) and this is a *minimum* estimate for depth. The clear evidence for (lower)

slope/apron deposition can be used to estimate depth, though the local tectonics might make such estimates somewhat doubtful. Such slope sequences are generally in over 100m of water, but range down to over 1000m (Enos & Moore, 1983; Mullins et al., 1984). In conclusion, the flyschoid sequences are clearly deepwater and were deposited in depths exceeding 100m-200m. The maximum depth is harder to estimate, but clearly the deposits are not abyssal and were deposited above the CCD. Most modern analogues exist in under 1000m depth. A depth range of 200-500m seems most reasonable, though there is some leeway in this figure.

11.4 Tectono/Eustatic Controls of Oyambre/Meron Formation Sedimentation

Allochthonous controls were paramount in the development of the San Vicente Group flyschoid sequence. The key external control was tectonic. The basin was bounded to the north by emergent Jerra Thrusts, reactivations of Hercynian thrusts of the Coastal Ranges of Asturias (Figs. 11-46, & 1-6). Palinspastic section balancing on these thrusts from commercial seismic data indicates that they have a net southward displacement of about 5km in the Tertiary (J. Hossack, pers. comm. 1986). A similar sense of transport occurred on the Escudo de Cabuerniga Thrust (Fig. 11-46). These have their sole thrust décollements deeper than the Ordovician Barrios quartzite, and so the Escudo de Cabuerniga Thrust would have been thick-skinned and able to uplift the entire Phanerozoic sequence south of the Los Via Fm. fringing reef (Fig. 11-46), exposing a sequence not that dissimilar to that seen today south of the Los Via Fm. outcrop (Fig. 1-6 and the IGME Comillas map). This would explain the highly polymict nature of the source of the rudaceous mass flows of the Oyambre and Meron Fms. These thrusts were also the primary control on basin geometry, and created an E/W elongated trough.

This picture of a funnel-shaped, E/W oriented basin is complicated by the presence of the San Vicente-La Madrid-Bustriguado complex tectonic zone (Zone 2 of Chapter 1.5). Basically, this is a transfer structure that was created during the rifting, Wealden stage of the opening of the Bay of Biscay. It was reactivated

as a Keuper piercement during the Pyrenean compression. To the west of the central complex zone of the La Madrid area (IGME Comillas map), there was a sequence of westward-transporting imbricate thrusts (Fig. 11-49) which are summarised in Figs. 11-47 & 11-48. These are a series of thrusts cut by E/W-oriented faults (Fig. 11-47). A typical example of this thrusting is shown in Fig. 11-41, which shows Cenomanian Altamira Fm. being thrusted over Los Via Fm. on a décollement of Sardinero Group marls. A summary tectonic history of thrust emplacement is shown in Fig. 11-48.

But was this thrusting/Keuper piercement coeval with flysch sedimentation? Critical evidence comes from the base of the frontal, Las Arenas Thrust (Fig. 11-47). This has angular, broken breccia clasts of the hangingwall thrust lithology (Cenomanian Altamira Fm.) in a matrix of pink, Los Via Fm. micrite (Fig. 11-50). The micrite is so penetrative into voids, major- and hair-line cracks in the Altamira Fm. breccia that the base of the thrust *must* have been emplaced while the Los Via Fm. micrite was still unconsolidated. Normal breccia clasts of the Los Via Fm. do not exist, and the clasts of the Altamira Fm. can sometimes be fitted back together again (Fig. 11-50). Zones of brecciated Altamira Fm. also exist elsewhere. Rhodophytic micrites are known to lithify *very* rapidly on the sea floor, usually within the top metre of sediment (Ginsburg & Schroeder, 1973) or a few 1000 years or less (Jindrich, 1983) due to rapid cementation by HMC micrite/algal boundstone action. For this micrite to be unconsolidated, it implies that the Las Arenas Thrust *must* have cut through right to the seafloor to incorporate unconsolidated Los Via Fm. micrite. This would have exposed the hangingwall sequence, transported and uplifted by the Las Arenas Thrust, to subaerial erosion. Reworked clasts of the hangingwall lithologies would be resedimented into the rudaceous mass flows of the Oyambre and Meron Fms. This intruded breccia rock is only found within the lower 10m or so of thrust planes, and the breccia clasts are always those of the lithology being thrust over the Los Via Fm. The most important conclusion from this is that it dates the time of

thrust emplacement to that of the Los Via Fm. (Upper Eocene/Oligocene), directly coeval with the polymict rudaceous flows in the flyschoid Oyambre and Meron Fms.

The San Vicente ria (54309752) is underlain by a Keuper piercement, related to its location on a major transfer structure. The small diapir indicated on the IGMEa map on the Playa de Meron (54599775) is *not* a structural diapir, but is the Rupuente megabed. The fact that the bedding on either side of it is parallel and simply part of the Cap Oyambre syncline and the fact that the Rupuente megabed can be traced onto other sections (e.g. Playa de San Vicente, Fig. 11-1: 275-295m) suggests that it is simply a large flow and not a true piercement structure. The intense distortion of the bedding within the flow probably accounts for its misinterpretation as a diapir by IGMEa.

The San Vicente piercement (and others) were active during Oyambre and Meron Fm. sedimentation because first-cycle, euhedral quartz crystals from it have been found in the resedimented flysch flows (Chapter 11.2). It is most likely that these pierced through onto the seafloor and the local piercement could well be the origin for the rockfall deposits of the Cenomanian Altamira Fm. megaclasts in the Rupuente megabed. Remarkably similar evaporitic piercements have been seen in the modern Hellenic Trench (Le Pichon et al., 1982) where vertical fault scarps with slickensides, slope breccias, rockfall deposits and tectono-sedimentary melangés have been found (cf. megaclasts of the Rupuente megabed).

The vertical evolution of the Oyambre and Meron Fms. can now be considered in terms of its tectono-eustatic controls. The change from the La Acebosa Fm. to the Oyambre Fm. is marked by the change in colour from off-white to dull red in the hemipelagic marl and an increased frequency of calcarenitic turbidites. The colour change is due to erosion of red-coloured Keuper inland. The increased frequency of, and abundance of neritic skeleta in, the calcarenitic turbidites indicates the establishment of well-developed, rather narrow shallow-water zone to the south and west. Similar high rates of turbidite generation (about 3-10 per metre) have been recorded by Crevello & Schlager (1980) for turbidites from the Recent Bahama Bank into the Exuma Sound. However, on the

Playa de la Jerra section (Fig. 11-1: 187m-230m), there is a 40m-scale fining- and thinning-upward sequence that commences with coarse, rudaceous, disorganised resedimented conglomerates with polymict clast assemblages. This indicates that thrusting/piercement was active even by the start of the Oyambre Fm. (Upper Eocene) and that it formed *very local* (<2km wide) rudaceous flows. The presence of Los Via Fm. clasts indicates that shallow water fringing reefs had also established themselves by the Upper Eocene. The overall fining- and thinning-upwards sequence shows no indication of channeling and is likely to be the response to initial tectonic activity with thick, rudaceous flows, followed by progressively lower-"energy" flows as the slope reestablishes itself. This would generate fining-upward sequences on this scale. The presence of rounded extraclasts suggests that a simple slope progradation model due to relative sea-level variation is inappropriate. Such sea-level changes would also be expected to produce more than very local (<2km) lithology changes. These factors suggest that relative sea-level change is *not* the dominant control here.

The base of the Upper Oligocene Rupuente megabed (Fig. 11-1) is marked by the incoming of a major debris flow (Rupuente megabed) and the start of thick (300m plus) fining- and thinning-upward sequences (Fig. 11-1: Playa de la Jerra, 255m-278m; Playa de Meron, 263m-425m; Playa de San Vicente, 269m-323m). Such fining-upward sequences have a similar origin to the one described above. The Rupuente megabed (Facies 12) is clearly a seismite linked to active Keuper piercement/thrusting. The incoming of the Upper Oligocene Meron Fm. above the Upper Eocene Oyambre Fm. (see Chapter 11.1) marks a major biostratigraphic hiatus (which is *not* a fault contact). It seems difficult to imagine that in such a tectonically-active area a deepwater basin would not receive any sediments in the Lower Oligocene. The hiatus is hence a question of preservation, not deposition. One possibility is that the major slope failure of the Rupuente megabed moved the entire Lower Oligocene sequence basinward in a major mass flow and that nothing of it is preserved here. A thin slump scar is all that would mark the hiatus. Another mechanism of hiatus generation is by bottom current winnowing of the

Lower Oligocene. The Lower Oligocene is a widely recognised hiatus in deepwater sequences in the Bay of Biscay (Montadert et al., 1979; Laughton et al., 1972) and has been used as a marker between seismic stratigraphic Formations 1 and 2 across the Bay of Biscay (Chapter 12). The lower part of Formation 1 (= 1b of Chapter 12) consists of turbidites, and carbonate-poor hemipelagic marls that have been reworked by bottom currents (cf. San Vicente Gp.). It is clear that this offshore stratigraphy has a direct, onland analogue in the San Vicente Group. This hiatus affected much of the North Atlantic (Roberts & Montadert, 1979) and has been termed the "great sculpture event" on the western Atlantic margin (Tucholke & Mountain, 1978). It also acted globally (Davies et al., 1974; van Andel et al., 1975; Moore & Heath, 1977). It is considered to be a product of scouring by deep ocean currents which were initiated by a global circulation involving a circum-Antarctic current and major cooling (recorded in $\delta^{18}O$ values) due to the start of Antarctic glaciation (Shackleton & Kennett, 1975; Kennett, 1977). The opening between Antarctica and Australia and between the Norwegian and Greenland seas at this time resulted in increased latitudinal temperature gradients and the initial generation of cold-region-derived bottom waters in the North Atlantic (Roberts & Montadert, 1979). These scouring bottom waters could well have removed the Lower Oligocene sequence in Santander. Unlike the Oligocene of the rest of northern Spain, which is syn/post-thrusting continental molasse (e.g. Puigdefabregas & van Vlient, 1978 or Allen et al., 1983), the direct analogues to this flysch sequence lie offshore, in deepwater sequences. The DSDP Sites 399-402 have mainly yielded orange/brown chalks from the Oligocene (Roberts & Montadert, 1979). Samples of extraformational conglomerates of shallow-marine limestones/dolomites of Jurassic to Lower Cretaceous age in a marly, chalky matrix of Upper Eocene/Lower Oligocene age have been dredged from the Armorican margin (Auffret et al., 1979), along with other polymict conglomerates of metamorphic and igneous Palaeozoic basement. Similar mass flow transport mechanism moved these polymict conglomerates into orange/brown marls just as they affected the Oyambre and Meron Fms. These flows

have also been related to the effects of the Pyrenean orogeny reactivating older Hercynian(?) and rifted margin lineaments (Roberts & Montadert, 1979).

The origin of the Lower Oligocene hiatus links to the lack of preservation of Lower Oligocene strata as 1) down slope removal by mass flow occurred at the start of the Upper Oligocene when the Rupuente megabed formed (itself associated with the same catastrophic slope failure) and 2) probable syn-depositional erosion of the Lower Oligocene strata by deepwater, scouring, bottom currents that were widespread in the Bay of Biscay (and beyond) at this time.

The Meron Fm. is, on a 100's m scale, clearly a fining- and thinning-upward sequence with fewer turbidites or rudaceous mass flow deposits up section (Fig. 11-1). Could such a system be the product of eustatic changes? The Oligocene was a period of relative sea-level fall throughout northern Spain (Pomerol, 1982), and a major eustatic fall occurred at the start of the Upper Oligocene (T_b/T_c fall - Vail et al., 1977). The sea then gently rose during the rest of the Oligocene. During sea-level lows, there is a low production of carbonate turbidites and a low sedimentation rate (Lynts et al., 1973; Schlager & Chermack, 1979; Hine et al., 1981; Boardman & Neumann, 1984; Droxler & Schlager, 1985) despite the (erroneous) work of Shanmugam & Moiola (1984). This is the opposite of siliciclastic fan systems as the neritic, carbonate skeleta are generated *in situ* on the shelf whereas quartz sand (etc.) has to be moved across the shelf from coastal areas. A major fall at the base of the Upper Oligocene (= base of the Meron Fm.) followed by a gentle rise *ought* to produce a paucity of carbonate turbidites at the base of the Meron Fm. and an increasing number of turbidites up formation on a purely eustatic/carbonate-turbidite model. This is *clearly* not the case (Fig. 11-1), suggesting that local tectonics outweighed the effects of sea-level here. A further problem with the eustatic model is its inability to generate deep-seated, ancient clasts (e.g. of Carboniferous basement) for incorporation into the rudaceous flows.

The primary control on the Meron (and Oyambre) Fm. was local tectonism. The generation of the Rupuente megabed was seismically triggered due to Keuper

piercement/active thrusting. The fining- and thinning-upward sequence above it (Fig. 11-1) is primarily a response to decreased seismic activity and gradual slope reestablishment. Hence the change from the rudaceous lower slope deposits of the lower part of the Meron Fm. to the marl-rich apron sequence of the upper part of the Meron Fm. is *not* a relative sea-level rise-related overstep but records a period of higher tectonic activity at the base of Upper Oligocene that wanned through to the end of the Upper Oligocene. The decreased frequency of turbidite/rudaceous flow generation in the upper part of the Meron Fm. (Fig. 11-1) was probably due to thrust-related subaerial uplift of the hinterland during the lower Meron Fm. times. In the latest Oligocene there was a temporary cessation in tectonic activity, which explains the lack of rudaceous mass flow deposits/siliciclastic turbidites and an abundance of argillaceous deposition.

11.5 Post San Vicente Group Tectonism and the Thrusting of the Basco-Cantabrian Basin

The main structural deformation of NW Santander occurred after the sedimentation of the Upper Oligocene Meron Fm. This involved folding of the Meron Fm. in the Cap Oyambre syncline, further thrusting, and piercement of the Keuper diapirs (see Chapter 1.5). It was also marked by a general uplift of the area to expose the sequence much as we see it today. Pleistocene/Holocene drift above is mainly confined to river terrace gravels, coastal deposits and altered carbonate debris (IGMBa, c).

11.6 Conclusions on the San Vicente Group Flyschoid Carbonates

The youngest pre-orogenic formation in NW Santander is the La Acebosa Fm. which consists of thin-bedded turbidites in an off-white, outer shelf marl. The flyschoid Oyambre (Upper Eocene) and Meron (Upper Oligocene) Fms. are dominated by hemipelagic, periplatform ooze, neritic-derived, calcarenitic turbidites and thicker rudaceous mass flows (disorganized or graded resedimented conglomerates, debris flow etc.). These were deposited downslope of the coeval,

shallow-water, reefal Los Via Fm. which contains corallinaceae, bryozoa, "nummulitic" foraminifera and scleractinia. The deepwater Oyambre and Meron Fms. were deposited in a fault-bounded reef-carbonate slope-apron bypass margin. Turbiditic flow was mainly axial to the structural orientation of the basin. Deepwater sedimentation was coeval with active thrusting of E/W-oriented thrusts, themselves reactivations of Hercynian thrusts. There was also extensive Keuper piercement through to the seafloor in locations largely controlled by major transfer structures. This active tectonism unroofed the entire Phanerozoic column, and after alluvial/littoral reworking, polymict clasts of it were resedimented in rudaceous mass flows into the Meron and Oyambre Fms. A major debris flow some 90m thick exists at the Oyambre/Meron Fm. junction (the Rupuente megabed) and contains megaclasts of brecciated, rockfall/bedded Cenomanian Altamira Fm. up to 40m wide. A Lower Oligocene hiatus can be correlated with offshore sequences/seismic stratigraphy. The initial thrusting in the area can hence be dated as Upper Eocene to lower Upper Oligocene, though further, more major compressional thrusting and uplift occurred in post-Oligocene times to expose the sequence as it is seen today.

CHAPTER 12 : TECTONO-SEDIMENTARY EVOLUTION OF NW SANTANDER.

12. 1 : Tectonic evolution of the northern Iberian plate.

The Hercynian orogeny introduced the basic structural elements that controlled the later Mesozoic/Tertiary tectonic evolution of the NW Santander area. This was a period of dextral, strike-slip motion (Arthaud & Matte, 1977; Heward & Reading, 1980) and formed the a number of E/W structures including the Escudo de Cabuerniga (Fig. 1-6). The Hercynian structures were reactivated during the Permo-Triassic as extensional structures with the creation of grabens and half-grabens which were filled with alluvial clastics (Garcia Mondejar et al., 1986; Saiz de Omenaca Gonzalez, 1977; Chapter 3). Similar Triassic-Lower Liassic basins were generated on the Armorican margin and in the Aquitaine basin (BRGM et al., 1974; Curnelle, 1983; Curnelle et al., 1982; Brunet, 1984) and in offshore Galicia (Mauffret & Montadert, 1987). The Middle Lias to end Middle Jurassic was a period of relative tectonic calm with the deposition of shelfal carbonates in the Basco-Cantabrian basin (Chapter 3).

The main tectonic periods will be considered in more detail below. Each section will cover tectonic models for the movements of the Iberian plate, the degree to which the sedimentary infill of the basin conforms to such models and lastly extrabasinal controls such as eustacy or climate.

12. 2 : A review of the Cretaceous/Tertiary tectonic evolution of the northern Iberia and the Basco-Cantabrian basin.

The magnetic anomaly fit allows a relatively accurate reconstruction of the Iberian plate motion during the Cretaceous/Tertiary. The initial rifting in the Bay of Biscay started in the Late Jurassic (Kimmeridgian) on the northern margin of the Bay of Biscay (Montadert et al., 1979). The onset of seafloor spreading in the Bay of Biscay is dated as Late Aptian-Early Albian (119 Ma) in the western part of the Bay of Biscay (Montadert et al., 1979, Olivert et al., 1984).

Spreading continued in the Bay of Biscay until between anomaly A_{34} (Early Campanian, 80-84Ma) and anomaly A_{33} times (Early Maastrichtian, 73-74Ma). Spreading most likely ceased in the Middle Campanian (77-78Ma), if a constant spreading rate is assumed (Engeser & Schwentke, 1986). The spreading of the Bay of Biscay is generally considered to have been accompanied by extensive sinistral transtensional tectonism in the Basco-Cantabrian/Pyrenean belt (Engeser et al., 1984).

The partial closure of the Bay of Biscay commenced towards the end of the Cretaceous but was most active during the Palaeocene-Miocene. The total compression across the Pyrenean belt has been estimated by a number of authors to be in the range of 50-100Km (eg. Seguret & Daignieres, 1986; Williams, 1985). Within the Pyrenean belt this compression is accommodated by extensive thin-skin thrusting in the forelands to the north and south of the Pyrenean chain (Williams, 1985; Fisher, 1984) and basement-involved thrusting within the axial parts of the Pyrenean chain (Parish, 1984). To the west of the Pyrenean chain the compression is accommodated by limited oceanic subduction of the Bay of Biscay towards the south beneath the northern Spanish continental margin (Boillot et al, 1979; Malod et al., 1979).

12. 3 The Purbecko-Wealden (Kimmeridgian - Barremian) extensional phase.

This was a period of alluvial infill of extensional half-grabens linked to the rifting stage of the opening of the Bay of Biscay (Fig. 4-4).

12. 3. 1 : Extensional tectonics of the Purbecko-Wealden period.

On the comparatively undisturbed northern continental margin of the Bay of Biscay opposite the Basco-Cantabrian basin, seismic profiles and DSDP drilling have shown that this has a sequence of tilted and rotated extensional fault blocks formed during the Kimmeridgian-Barremian (Figs. 12-1 & 12-2). These fault blocks are bounded by listric faults which delineate half-grabens created during rifting phase and are subsequently covered by post-rift sediments (Montadert et al., 1974, 1979; Fig. 12-1 & 12-2). This area uses a standard seismic stratigraphy for Biscay

(Fig .12-3) that has been applied on the northern margin of the Bay of Biscay (Montadert et al., 1974; Deregnaucourt & Boillot, 1982a), around Galicia Bank (offshore NW Iberia - Montadert et al., 1974; Boillot et al., 1979b), in the Western Approaches Trough/Goban Spur area (Roberts et al., 1981) and in the seamount boreholes J-118 & J-119 of the Spanish marginal trough (Soler et al., 1981). The offshore seismic formations (1 to 4 - Fig. 12-3) show an excellent correlation with the "Group" stratigraphy of the onland NW Santander area. The pre-rift Jurassic limestone are partly bevelled off over the tops of rotated fault blocks (Figs. 12-1 & 12-2). The syn-rift sediments (Formation 4 of Fig. 12-3) onlap towards the crests of half-graben fault blocks, indicating syn-rift sedimentation. Similar rift features are seen in the Armorican continental margin (Roberts et al., 1981), Celtic Sea, Bristol Channel and Western Approaches basins (Allen, 1975; Hallam & Sellwood, 1976; Ziegler, 1982). The ODP Leg 103 drilling in offshore Galicia (Fig. 12-3; Mauffret & Montadert, 1987) has also found rotated half-grabens in seismic formation 4 that are linked with active rifting between Iberia and Canada. The major change to post-rift sedimentation occurs in the intra-Aptian (Montadert et al., 1979a) in the Bay of Biscay.

The fault geometry of seismic formation 4 (Figs. 12-1 & 12-2) strongly suggest a rift basin (Balley, 1982) formed by roughly N/S extension linked to the extension that opened the Bay of Biscay.

12. 3. 2 : Basin geometry of the Purbecko-Wealden infill of the Santander basin.

This section will examine the Purbecko-Wealden stratigraphic infill of the Santander basin that formed coevally with the offshore extensional basins described in Chapter 12.3.1.

In the Santander basin, E/W structures were reactivated as down-to-the-north half-grabens (Fig. 12-4, with location map in Fig. 12-5). Within the NW Santander area a N/S cross-section shows a distinctly wedge-shaped , half-graben-like infill (Fig. 12-6, located on Fig. 12-7). The Wealden thickens to the south from only 60m on the Cuchia section (Sect. 15) to about 565m just north of the

Escudo de Cabuerniga. This would suggest that the northern margin of the Escudo de Cabuerniga ridge was a down-to-the-north extensional fault during Purbecko-Wealden times. The geometry of the NW Santander infill is very similar to the pattern seen on other N/S sections from the Santander basin (Fig. 12-4).

An E/W cross-section across NW Santander is shown in Fig. 12-8. In the area to the west of San Vicente de la Barquera (Zone 1 of Chapter 2) the Wealden onlaps onto the Hercynian basement to the west (Sections 5, 6, 7 & 8 of Fig. 12-8). The N/S Bustriguado fault was active during sedimentation with a down throw to the east (Figs. 12-8 & 12-9). A west-tilted fault block existed between the Bustriguado and Rio Saja faults (Figs. 12-8 & 12-9). To the east of the Rio Saja fault another west-tilted fault block existed (Sections 11, 12 and *Ancillo 1* of Fig. 12-8). The Bustriguado and Rio Saja faults are considered to be synsedimentary transfer faults because of their high angle (NNE/SSW) orientation relative to the E/W-oriented Escudo de Cabuerniga half-graben fault (Fig. 12-9). Similar NNE/SSW structures have been recognised further east in the onland Basco-Cantabrian and Pyrenean basins (Amiot et al., 1983).

12. 3. 3 : Stratigraphy of the Purbecko-Wealden extensional basin infill.

Within the Santander basin (Figs. 4-4, 4-6 & 12-5) there was a progressive fill of extensional half-grabens through the Kimmeridgian to Barremian period (Fig. 12-4). At the base of the sequence there is a prominent unconformity representing the pre-rift to syn-rift transition (Figs. 1-9 & 12-4). There is greater erosion of the Jurassic in the areas where Purbecko-Wealden infill is thinner (Fig. 12-4). The Kimmeridgian Aguilar Fm. (Fig. 4-6) lacustrine limestones (Figs. 4-4 & 12-4) mark the initial fill of the extensional basin. The alluvial infill of the basin flowed east, an axial flow direction (Figs. 4-7, 4-8 & 4-9). During the Portlandian to Mid Valanginian the alluvial Arcera Fm. onlaps towards the NE and is wedge-shaped (Fig. 12-4). This is itself overlapped to the north by the semi-marine Aroco and Loma Somera Fms (Figs. 4-4 & 12-4).

The Wealden cycle extends from the mid-Valanginian to the Barremian (Fig. 4-7). The Barcena Major Fm. has a wedge shaped geometry in the respective half-grabens of the Santander basin (Fig. 12-4). In NW Santander, it wedges out against the Jurassic basement (Fig. 12-6). The overlying Vega de Pas Fm. is a mud-dominated, meandering river sequence that is the most laterally extensive and youngest syn-rift deposit in the Santander basin (Fig. 12-4).

12.3.3 : The rift basin model for the Santander basin Purbecko-Wealden.

This section will consider how well the rift basin model fits the geology of the Purbecko-Wealden. Before the extensional rifting, the Santander basin was undergoing shallow marine carbonate sedimentation in the Middle Jurassic. The initiation of extension in the Kimmeridgian caused extensive uplift and erosion of the mid/late Jurassic sequence. This uplift and erosion generated the extensive Callovian to Kimmeridgian unconformity (Fig. 1-9), which is seen as a pre-rift unconformity. Such uplift is characteristic of areas where the uplift effect of lithospheric thinning outways the subsidence of crustal thinning (Jarvis, 1984), suggesting that the crust was thinned at this time in northern Spain. Noticeably thinned crust has been noted on the seismic surveys over the passive margins in offshore Galicia (Mauffret & Montadert, 1987) and the Armorican margin (Montadert et al., 1979).

The Purbecko-Wealden infill was (near) exclusively continental, unlike the marine Jurassic carbonates beneath the Purbecko-Wealden or the overlying Aptian Urgonian carbonates. The primary cause of the continental nature of the Purbecko-Wealden was the regional elevation due to lithospheric thinning at depth. The classic half-graben rift geometry has been illustrated in Chapter 12.3.2 and in Figs. 12-4, 12-6 & 12-8 which shows that most of the extensional faults were down-faulting towards the rifting Bay of Biscay to the north. These figures also show that there is a general relationship between the syn-rift infill and the pre-rift strata. In general, the younger and thinner the syn-rift Purbecko-Wealden sequence,

the older are the pre-rift Jurassic/pre-Jurassic sequence. This is a typical rift basin feature (Balley, 1982).

A further typical feature of rift basins are transfer faults that compartmentalise the rift half-grabens (Balley, 1982). Figure 12-8 shows the direct tectonic evidence for such transfer faults. In NW Santander, these transfer faults run approximately NNE-SSW (Fig. 12-9), and appear to be connected with transfer structures further offshore (eg. the Santander Spur is an offshore continuation of the Rio Miera flexure). The transfer faults in the Santander basin are directly analogues to those that compartmentalise the Armorican margin (Montadert et al., 1979), the Goban Spur (Masson & Roberts, 1981) or offshore Galicia (Mauffret & Montadert, 1987).

In NW Santander, the Wealden Barcena Major Fm. is an arenaceous, braided-stream deposit (Chapter 4). Such high-energy, high-sand-content sedimentation points to major sand input and steep topographic gradients in the alluvial system. This could be produced by very active extensional faulting with hangingwall uplift to generate areas of rapid sand erosion and steeper gradients due to faulting in the footwall areas.

The overlying Vega de Pas Fm. is a muddy meandering river deposit (Chapter 4). The change to finer-grained sedimentation probably relates to a fall-off in the rate of fault movement with lower topographic gradients and sand input.

The end of rifting occurs in NW Santander with the start of the Aptian Torrelavega Group (Fig. 1-9) and more rapid subsidence in the Upper Aptian. This timing is remarkably close to the time when rifting ceased offshore (the intra-Aptian according to Montadert et al. ,1979).

12. 3. 4 : Other tectonic models for the Purbecko-Wealden and eustatic/climatic models in the Purbecko-Wealden.

The Purbecko-Wealden Santander basin does not fit at all well into a strike-slip pull-apart basin model. It lacks demonstrable Lower Cretaceous strike-slip fault margins, any mis-match over source area for the basin, end echelon folds,

zones of contemporaneous, fault-induced push-ups and pull-apart basins (sensu Mann et al., 1983) or the generally coarse-grained localised infill normally associated with pull-apart basins.

Figure 12-10 shows eustatic curves for the Cretaceous/Tertiary from Vail et al., (1977), Cooper (1977) and Hancock & Kauffman (1979). In NW Santander there is a change from marine sedimentation in the Callovian to continental sedimentation in the Purbeckian. This does not correspond with any eustatic drop at this time. The Late Jurassic is a period of relatively high eustatic sea-levels (Vail et al., 1977; Haq et al., 1987) that does not directly equate with the generally non/marginal-marine nature of the Purbeckian. The slight unconformity between the Purbeckian and the Wealden in the mid-Valanginian may correspond with a eustatic fall at this time (Haq et al., 1987). The non-marine nature of the Wealden does not correspond with the moderately high sea-level stand in the Late Valanginian-Barremian. In conclusion, eustacy has had little direct effect on the rifting phase as tectonic activity was the primary control on basin evolution.

Climatic variation within Upper Jurassic/Lower Cretaceous has traditionally considered that there was a relatively arid Purbeck and a more humid Wealden (Rat, 1959, Hallam, 1985). However, such a variation could not explain the thickness variations in the Purbecko-Wealden or the cycles *within* the Pas or Cabuerniga Groups. It may well explain the presence of evaporites in the Purbeckian but not in the Wealden (Chapter 4). The palaeocurrent patterns (Figs. 4-7, 4-8, 4-9 & 12-4), heavy mineral analysis, conglomerate composition, facies belt distribution and sandstone petrography all support the idea of a long-lived, relatively stable source area to the W/SW of the basin, so changes in provenance for the basin can be discounted as a major controlling mechanism.

In conclusion, eustacy, climate and changes in provenance are seen as second-order variables that slightly modified the primary tectonic control on the evolution of the Purbecko-Wealden rifting phase.

12. 4 : Drifting stage (Upper Aptian - Middle Campanian) in NW Santander.

The start of the Upper Aptian saw a major change in tectonic style of sedimentation in NW Santander/Basco-Cantabrian basin. This continued until the initial pulses of the Pyrenean orogeny in latest Cretaceous times (Middle Campanian ?).

12. 4. 1 : Drifting stage tectonics of the Basco-Cantabrian basin.

After the rifting phase the Bay of Biscay opened due to seafloor spreading. This lasted from approximately the mid-Aptian until about the Middle Campanian (Montadert et al., 1979; Engeser & Schwentke, 1986). The drifting stage also involved an anticlockwise rotation of Iberia by about 30°-35° (Van der Voo, 1969; Vandenberg, 1980; Schott & Peres, 1987). Because of this rotational component in the opening of the Bay of Biscay, the drifting stage has a strike-slip component to it and gave rise to a period of generally transtensional tectonics in the Basco-Cantabrian basin.

12. 4. 2 : Stratigraphy and basin tectonics of the drifting stage in NW Santander.

The detailed stratigraphy of the drifting stage has been considered in Chapters 5 to 9 and Fig. 1-9.

12. 4. 3. : Aptian stratigraphy and basin tectonics.

The Aptian was a period of comparatively rapid, thermal subsidence (560m in 6 Ma = 93m/Ma in NW Santander in the Upper Aptian) linked to the initiation of the drifting stage in the Bay of Biscay. This is the greatest rate of subsidence seen in the Cretaceous/Tertiary in NW Santander. This rapid subsidence is also reflected in the incoming of transgressive marine carbonates (the Urgonian carbonates of Chapters 5 & 6). During the Aptian there was a onlap onto the basement by progressively younger Aptian carbonates (Chapter 5), producing a "stags-horn" geometry (Sections 1 to 4 of Fig. 5-1; compare Fig. 5-7 vs. Fig. 5-14; Bailey, 1982). Rapid subsidence, deeper-water sedimentation and a stages-horn onlap

pattern are all typical characteristics of the initiation of thermal basin subsidence in the early phases of drifting basins (Balley, 1982).

The palaeogeography at the end of the Aptian (Figs. 5-8 & 6-21) was a mosaic of extensive carbonate platforms separated by faults, homoclinal ramps and drop-off margins from areas of coeval deep-marine marls. The existence of such extensive platform carbonates can also be linked to the fundamental tectonic control of thermal subsidence. The creation of a continental margin thermal downwarp would create an extensive area of subsidence and so reduce the areas with pronounced topographic relief, leading to reduced terrigenous clastic input in the Aptian. The more regional subsidence produced by thermal subsidence would moved the clastic shoreline towards the palaeohinterland to the SW (compare Figs. 4-2 & 6-21), so cutting off clastic input to NW Santander and allowing platform carbonates to form. Rudistid platform carbonates can only form in areas of very low clastic input - Masse & Philip, 1981.

A further feature of tropical (Tethyan) platform carbonates is their ability for rapid vertical aggradation (up to 1Km/Ma - Schlager, 1981). Such rapid aggradation allowed the platform carbonates to proliferate during a period of very rapid basin subsidence. On the standard seismic stratigraphy of Biscay's marginal basins (Fig. 4-29), the Urgonian carbonates of NW Santander are a shallow-water, shelfal equivalent of the black shales of Formation 3 (Fig. 12-3).

An additional important feature was Aptian diapirism that is discussed in detail in Chapters 5 & 6. This is very typical of the drifting stages of continental margin evolution (Woodbury et al., 1980; Chapter 6.5).

By comparison with the rest of the Basco-Cantabrian basin during the Albo-Aptian, NW Santander was a relatively stable tectonic zone. The areas around Bilbao, the western part of the Ebro sector (Sector III in Fig. 6-24) and the Valnera area (Sector II in Fig. 6-24) experienced very strong, differential subsidence (thicknesses on Figs. 4-3B and 6-23 - located on Fig. 6-24). The basin had a pronounced NW-SE facies trend and in the eastern Basque arc (Fig. 1-3) the Bilbao synclinorium was created (Fig. 1-3). A smaller, similar oriented trough

occurred around Vitoria (Fig. 4-3B). SW-NE trends for the basin occur in the Ebro trough (Sector III in Fig. 6-24), near Bilbao, on the Navarro Diapiric lineament (Fig. 1-3) and along the Silio-Soba-Valmaseda valley. Finally, the E-W Hercynian trend of the Escudo de Cabuerniga controlled the northern margin of the "Southern Santander" trough (Sectors II and III of Fig. 6-24). The Albo-Aptian is over 4,000m thick in the Soba graben (Sector II, Fig. 6-24) and the Bilbao (Figs. 4-3B and 6-21) area. During deposition the depocentres migrated north (Fig. 6-25B). To the east of the Bilbao meridian, platform carbonates and isolated offshore banks occur (Fig. 6-21). The latter locally had tectonically-active carbonate slope margins (Garcia Mondejar, 1985). In the western Basco-Cantabrian basin, extensive diapirism started in Albo-Aptian times, mainly in the Upper Aptian as in NW Santander (Garcia Mondejar, 1979; Reitner, 1982; Antiguedad et al., 1983; Pascal, 1985; Hines, 1985a).

On a larger scale, the location of the main Urgonian buildups and related synsedimentary faults in the Basco-Cantabrian basin and their offshore equivalents (Deregnaucourt & Boillot, 1982) show a strong parallelism with those on the "European" plate margin (Fig. 6-26). It is probable that the Southern Santander trough was connected via the Bilbao trough to the offshore Santander canyon (Fig. 6-26). The major lineaments of the basin run NW-SE (inherited from the Wealden rifting stage - Chapter 4) and SW-NE.

In conclusion, during the Aptian NW Santander had relatively simple extensional thermal subsidence, modified by diapiric tectonics. However, elsewhere in the basin, major tectonic lineaments were strongly reactivated and thick, laterally complex facies were deposited (Fig. 6-22). These can be linked with similar trends on the "European plate" margin with complex strike-slip (?) motions mainly centred on the Bilbao and Soba troughs (Fig. 6-27; Pascal, 1985).

12. 4. 4 : Eustacy and climate during the Aptian in NW Santander.

Worldwide eustatic curves show a very poor correlation with the inferred depth of water during deposition of the Aptian in NW Santander (Fig. 12-10). Haq et al., 1987 show a relative eustatic fall from the Barremian into the Aptian. This

is the exact opposite of what is observed in NW Santander where the Aptian marine Umbrera Fm. overlies the non-marine Barremian Wealden Vega de Pas Fm. This is interpreted to be the result of greater tectonic subsidence outweighing any eustatic effect.

The climate in the mid Cretaceous in NW Santander appears to have been a warm, humid, sub-tropical one that was relatively constant throughout the mid Cretaceous.

12. 4. 5 : Albian basin tectonics.

The Albian tectonics of NW Santander are dominated on a small scale (5 Km) by the continued, active pillow-stage diapirism of the early drifting stage of the opening of the Bay of Biscay (Fig. 7-1; Chapter 7). However, on a 50Km scale, the NW Santander sub-basin thins to the east up to the Rio Miera Flexure (Figs. 7-36 & 12-12; Sections 1-26 of Fig. 7-1). This is the reverse of the general Aptian thickening of the sequence eastwards towards the centre of the Basco-Cantabrian basin (Figs. 5-2 & 6-1). This reversal in the tilting of the sub-basin can be related to the strong downthrow to the west of the Rio Miera Flexure at the eastern end of NW Santander (Figs. 7-1 & 12-11) that relatively uplifted NW Santander to the west of the fault. To the west of the flexure a much thicker, deeper-marine Albian sequence exists (Santiago section > 347m, Section 27 of Fig. 7-1). This suggests that the Rio Miera Flexure continued to act as a transfer structure during the Albian. Such a complete inversion of the basin tectonics in the Albian suggests active transtensional tectonism.

The Albian was the period of maximum distensional tectonics throughout the basin (Figs. 12-11 & 12-12) with large rhomboidal(?) pull-apart basins. Areas of very rapid subsidence existed such as the Valmaseda delta which had over 4.5km of Albian subsidence and decompacted depositional rates of over 1km/Ma. In contrast there were also push-up blocks such as NW Santander with sequences as thin as 20m in the Albian or the uplifted Palaeozoic strata on the Biscay High (Fig. 12-12). Submarine volcanism started along the key NW-SE Bilbao-Durango or the

Urquiola-Ochandian faults. Thermal metamorphism and schistosity developed in the Nappe des Marbres (Fig. 1-3) and fault-related slope breccias were generated around the Cinco Villas massif (Fig. 1-3; Feuillee, 1971). Such synchronous development of both rapid subsidence and rapid uplift in a continental marginal basin during a period of rotational opening in Biscay (Deregnaucourt & Boillot, 1982a) strongly points to sinistral distension as the dominant kinematic control on Albian tectonics. The Albian would hence correspond to the early phase of distensional drifting in the Bay of Biscay. Relatively similar distensional tectonics and push-up/pull-apart sedimentation has been described from the Albian Black flysch of the Western Pyrenees (Choukroune & Mattauer, 1978; Peyberres & Souquet, 1984; Souquet et al., 1985).

12. 4. 6 : Albian stratigraphy.

During the Albian NW Santander showed some similarities but major differences in facies and tectonics compared with the Basco-Cantabrian basin as a whole. In general terms, NW Santander preserves a thin sequence that is only 140m thick on average (Chapter 7) compared to up to 4,500m in the Valmaseda area (Figs. 4-3C, 12-11 & 12-12). The NW Santander Albian is dominated by shallow marine (middle shelf) carbonates of the Barcenaciones Fm., whereas in the rest of the basin "Urgonian" carbonates were progressively inundated by clastics (Figs. 12-11 & 12-12). NW Santander did not develop any algal mud-mounds or carbonate slopes/drop-off margins, as other parts of the basin did (Pascal, 1985) but had a relatively shallow-water sequence with abundant karstification, meteoric water diagenesis, codiacian algae and oolitic/skeletal shoals.

In most of the basin, the Albian starts with a major relative sea-level rise, whereas the exact opposite occurs in NW Santander with meteoric diagenesis at the top of the Reocin Fm. and the incoming of the estuarine Las Penosas Fm. (Chapters 6 & 7). NW Santander acted as a palaeohigh sub-basin (Fig. 12-11), bounded to the east by the Rio Miera flexure, which was a major, active lineament throughout the Albian (Chapter 7.4 & 7.8). The relative sea-level fall that created

the Las Penosas Fm. is probably linked to a eustatic fall in the mid Lower Albian (Haq et al., 1987). Within the Basco-Cantabrian basin the major carbonate build-ups and the overall southwestward passage to more continental conditions persisted through from the Upper Aptian (cf. Figs. 6-21 & 12-11 and Fig. 6-22). One of the most significant changes in these platforms was the major development of algal carbonate mud-mound platform margins, on the edge of rapid drop-offs into the coeval deeper-water (semi-) anoxic *Tritaxia* marls (Garcia Mondejar, 1985; Pascal, 1985). The encrusting algae (stromatolites, *Bacinella* etc.) of the mud-mounds generated lenticular, 100m plus thick mounds over tectonic lineaments (Fig. 6-23A) and these mounds were locally exposed to high-energy reworking and meteoric diagenesis. Transitional between the mounds and shales were slope facies with slumps and megabreccias (e.g. Garcia Mondejar, 1985; Pascal, 1985). Offshore, isolated, Bahaman-type, carbonate banks (e.g. Aralar, Aitzorri) occurred on halokinetic highs. These were often surrounded by tectonically-active carbonate slope deposits (Reitner, 1982; Garcia Mondejar, 1985).

The Upper Albian (Figs. 6-22 & 12-12) was a time of major siliciclastic input into the Valmaseda area (Fig. 12-12) where a sequence 1000-4500m thick was deposited over more than 5,000km^2 in a tidally-dominated delta (Pascal, 1985; Pujalte & Monge, 1985). The delta was fed by "Utrillas" fluvial sands from the S/SW (200-300m thick) and passed NE into the coeval deep marine "Black flysch" carbonaceous mudstones, turbidites and mass flow deposits of the Durango and Deva Fms. (600-1500m). The "Black flysch" was deposited in a rapidly-subsiding, tectonically-active submarine fan and slope system (Feuillee, 1967; Crimes, 1976; Garcia Mondejar et al., 1985). Further north, the offshore Biscay High was composed of uplifted Palaeozoic basement/Urgonian carbonates (Fig. 12-12; Voort, 1964). It discharged fan/slope detritus south (e.g. at Arminza, Ea, Ondarroa - Garcia Mondejar, 1985). This palaeoflow turned WNW/NW in the axis of the Bilbao flysch trough (Fig. 12-12).

Rudistid atolls continued to persist over diapiric palaeohighs (Fig. 12-12) and are associated with growth-fault slope breccias at Villasana de Mena and

Murguia. The only other major development of carbonates in the Upper Albian was in NW Santander (Torriente Mb.), which was a marginal area protected from siliciclastic input primarily by the Escudo de Cabuerniga fault and the Rio Miera flexure. The general eustatic rise throughout the Albian probably accounts for the incoming of the marine Barcenaciones Fm. carbonates over the marginal-marine Las Penosas Fm.

In general, the Albian was a period of very active transtensional tectonism in the Basco-Cantabrian basin linked to the rotational opening of the Bay of Biscay. Distinct, coeval pull-apart basins and push-up areas were formed that had radically different thicknesses, stratigraphy and sedimentology in their Albian sequences. Within the NW Santander area, there was relatively little subsidence and so a broadly eustatic control on facies development is seen.

12. 4. 7 : Cenomanian basin tectonics and stratigraphy.

In general, the Cenomanian was a time of more uniform facies belts than those of the Upper Albian, probably linked to the mature drifting stage in the Bay of Biscay (Deregnaucourt & Boillot, 1982b) and more uniform, moderate thermal subsidence in its marginal basins. In NW Santander, there was continued pillow-stage diapirism (Chapter 8.4) with arenite-poor, thickened rim syncline sequences continuing into the Cenomanian.

During the Lower Cenomanian the facies pattern (Fig. 12-13) inherited from the Upper Albian continued across most of the Basco-Cantabrian basin (Figs. 12-11 & 12-12). In the E and NE of the basin the Upper Cretaceous transgression began to replace the coarse terrigenous clastics of the "Black Flysch" with marls and marly limestones. However typical "Black Flysch" submarine fan/tectonically-active slope deposits continued to be formed around the Cinco Villas massif (Fig. 12-13; Feuillee, 1967; Amiot et al., 1983), N of Bilbao (Garcia Mondejar et al., 1985) and to the east in the Western Pyrenees (Souquet et al., 1985).

The Valmaseda delta (Fig. 12-13) continued during the Lower Cenomanian, but progressive transgression towards the S/SW reduced the amount of arenaceous supply to it. The Valmaseda Fm. contains some sandy, orbitolinid limestones towards

the top of the Lower Cenomanian. This facies belt of coastal and shallow marine clastics continues WNW into the NW Santander area (Fig. 12-13) where the tide and storm-dominated estuarine/inner shelfal Bielba Fm. was deposited (Chapter 8). The ENE/WNW direction of both the basinwide facies belts and the underlying Hercynian lineaments within the basin is clearly reflected in the dominant directions of the estuarine channels (Facies 1 of Chapter 8.3) and the offshore sand waves and bars (Facies 6 & 7 of Chapter 8.3). Hence, the arrival of major mid-Cretaceous clastics was diachronous, and although it was well established around Valmaseda by the Upper Albian, it was not until the Lower Cenomanian that these clastics invaded NW Santander. This was because NW Santander was protected from clastic input from the south by the Escudo de Cabuerniga fault and to the east by the Rio Miera Flexure fault in the Upper Albian and by the intense subsidence in the Valmaseda delta area (Fig. 12-12) diverting clastic influx towards it. To the SW of the shallow marine Valmaseda delta belt there is an interdigitation with coastal/delta plain environments that gives way further SW to thin "Utrillas" braided river coarse sandstones (Figs. 12-11 & 12-13).

During the Middle and Upper Cenomanian marine onlap continued towards the S/SW across the basin with much wider development of carbonates (Fig. 12-14). This is probably the result of the Cenomanian eustatic transgression (Fig. 12-10; Vail et al., 1977; Haq et al., 1987) and continued thermal subsidence. Clastic deposition was limited to the south of the Basco-Cantabrian basin in the area SW of Burgos (Fig. 12-14). Black Flysch clastics/schists were replaced by "planktonic calcschists" of Feuillee (1967), the "Flysch calcaires" of Mathey in Amiot (1983). They are about 1km thick and consist of a rather monotonous alternation of calcarenites and marls on a cm-dm scale. The calcarenites were deposited as thin-bedded turbidites (Mathey in Amiot, 1983) that flowed mainly NW/W/SW along the axis of the Bilbao trough (Fig. 12-14). Slightly to the south, north of the Sierra de Aralar, thick resedimented carbonate conglomerates and thick-bedded turbidites/debris flows reflect active tectonism (Feuillee, 1967). This was the period of volcanism in the Biscay trough. By the Upper Cenomanian high-energy

carbonate breccias, conglomerates and turbidites are found in the "Flysch de Mixe" around the Cinco Villas massif (Fig. 12-14). Thick sandstone beds are found in the calcareous flysch around Saint-Jean-de-Luz (Feuillee, 1967), together with microconglomerates of basement/Cretaceous strata derived primarily from the N (Fig. 12-14).

Further SW, in a vast facies belt running WNW/ESE through Vitoria (Fig. 12-14), a monotonous, rather nodular alternation of marly limestone and marls was deposited that constitutes the outer shelf "Flysch a boules" - Feuillee (1967). To the S/SW of this were extensive shallow-marine platform limestones with slightly sandy, glauconitic, muddy limestones. In NW Santander, the moderate- to high-energy Altamira Fm. was deposited as an offshore, subtidal calcarenitic bank strongly affected by both storms and tides (Chapter 8.6; Fig. 12-14). To the south of the Ebro an extensive area of internal platform marly limestones, sandstones and mudstones with oysters existed.

In conclusion, the Cenomanian was a period of rather more moderate subsidence in the Basco-Cantabrian basin with the development of relatively wide facies belts across the region. This facies belts moved progressively towards the SW/S through the Cenomanian in response to thermal subsidence and eustatic transgression.

12. 4. 8 : Tectonics and sedimentation during the Turonian to Mid Campanian.

The Upper Cenomanian/Turonian contact is a major hiatus across the entire Basco-Cantabrian basin (Feuillee, 1967; Ramirez, 1971a) and can be linked with the Iberian-margin wide contact between seismo-stratigraphic units 2 and 3 (Fig. 12-3). The Middle/Upper Cenomanian unconformity is widely seen offshore - in offshore Galicia (Group Galice, 1979; Mauffret & Montadert, 1987), on the Armorican margin (Montadert et al., 1979) and on the northern Spanish margin (Deregnaucourt & Boillot, 1982). This seismic contact corresponds with the Santander Group/Sardinero Group contact in NW Santander (Fig. 1-9).

In general terms the Turonian to Campanian sequence of the Basco-Cantabrian basin followed the same WNW/ESE facies belt pattern inherited from the Middle/Upper Cenomanian (Fig. 12-14; Amiot et al., 1983). This period is dominated by a vast carbonate ramp that deepened to the NNE. The marly Pedruquios Fm. (Chapter 9) shows a general thickening to the east across NW Santander (Fig. 9-2), which is probably explained by continued gentle (thermal ?) downwarping towards the core of the Basco-Cantabrian basin.

12. 5. 1 : Tectonics and sedimentation during the Upper Campanian to Middle Eocene.

In broad terms, the Upper Campanian to Middle Eocene saw the initiation of compressional tectonics across much of the Pyrenean/Basco-Cantabrian chain (Puigdefabregas & Souquet, 1986) but this compression did not greatly effect the NW Santander area. In general, the Upper Campanian to Middle Eocene Nansa Group in NW Santander (Fig. 1-10) is dominated by sandy, benthic foraminiferal inner shelf limestones that fit in as a shallow marine facies belt between the Bilbao flysch basin to the N/NE and the continental facies to the south (the "Garumnian" facies - Fig. 12-15).

The Campanian/Maastrichtian was a period of regression relative to the Upper Santonian. This regression was probably the combined effect eustatic fall (Haq et al., 1987) and the initiation of uplift associated with compressional tectonics. Rejuvenation of the Asturian and Ebro massifs produced increasing terrigenous influxes into the western part of the Basco-Cantabrian basin during the Upper Campanian to Maastrichtian. In the more south/southeastern parts of the basin relatively clastic-free carbonates persisted through the Senonian. This pattern is clearly reflected in NW Santander in the change from the outer shelf Cabo Menor Fm. of the Sardinero Group to the Upper Campanian parts of either the Muñorrodero or Cabo de Lata Fms. (Fig. 1-9).

During the Upper Campanian to mid Maastrichtian terrigenous sandy flysch returned to the Bilbao trough (Mathey in Amiot et al., 1983). It was up to 1700m thick at Derva/Zumaya. It was dominated by clastic turbidites in submarine fans

whose source lay mainly to the N/NE of the Bilbao Trough but which then turned and flowed W/WNW along the axis of the trough. Local resedimented conglomerates occur, especially in the Upper Campanian, which contain clasts of biomicrites and even Palaeozoic basement derived from the NE of the Bilbao trough. The Late Santonian-Maastrichtian saw the start of compressive transpression in the Southern Pyrenean basins (Puigdefabregas & Souquet, 1986).

The mid Maastrichtian/basal Palaeocene is marked by a marly limestone sequence in the Bilbao trough (the "calcaires rosés du Danien") of alternating marly limestones and marls on a decimetre scale, which contain the Cretaceous/Tertiary boundary. During the Upper Maastrichtian clastic turbidites ceased in the Bilbao trough (Mathey in Amiot et al., 1986).

By the Upper Palaeocene, turbiditic terrigenous sands reappear in the eastern part of the Bilbao trough (Fuenterrabia area - Kruit et al., 1975; van Vliet, 1978). The incoming of active submarine fan sedimentation has been linked to coeval compressional tectonics in the western Pyrenees (Puigdefabregas & Souquet, 1986). Alveolinid limestones dominate the marginal areas to the N and S of the Bilbao trough (similar to Fig. 12-15). NW Santander is the northwesterly continuation of this southern belt of foraminiferal limestones and develops the Montian/Thanetian Muñorrodero, San Juan and Sancibrian Fms. (Fig. 1-10).

The Ilerdian (Fig. 1-10) was a period of major transgression of benthic foraminiferal limestones over the continental areas and was remarkably constant in facies, timing and thickness (100-600m) over much of the central part of the Basco-Cantabrian basin (Mangin & Rat, 1962) and further east in the southern Pyrenean basins (Pomerol, 1982; Puigdefabregas & Souquet, 1986). This is directly represented in NW Santander by the transgression of the shallow marine Ilerdian Estrada Fm. over the Palaeocene marginal marine carbonates (Fig. 1-10). It was the period of extensive, sandy submarine fan sedimentation (> 1km thick) in the Bilbao trough, including the clastic submarine fan sequences around Zumaya and Derva (Crimes, 1973, 1976; Kruit et al., 1975; van Vliet, 1978). These are the youngest sediments preserved in the Bilbao syncline.

The Lutetian was a period of major inversion in the Bilbao trough and similar inversion was occurring throughout the Southern Pyrenean basins by the Lutetian (Puigdefabregas & Souquet, 1986). Here the axial zone was uplifted and southward moving thrusts were progressively thrusted over the Ebro basin. Marly flysch basins were generated along with coeval sandy/conglomeratic molasse. However, in NW Santander benthic foraminiferal limestones persisted through the Lutetian (the Serdio and Colombres Fms. - Fig. 1-10). This suggests that the initiation of major thrusting in the Pyrenean belt was diachronous and progressively later to the west.

12. 5. 2 : Compressional tectonics and sedimentation in the San Vicente Gp. flysch.

The San Vicente Group flysch (Chapter 11) was strongly controlled by active thrusting (such as the southward-moving Escudo de Cabuerniga Thrust) and Keuper piercement (see discussion in Chapter 11.4). As the only syn-thrusting sediments in the Basco-Cantabrian basin are either Oligocene continental molasse on the Sierra de Cantabria (Fig. 1-4) or the San Vicente Group flysch, these must be used to date the regional compression. The enhanced biostratigraphic resolution in the flysch sequence date two periods of compression - one in the Upper Eocene to basal Upper Oligocene and another in the post-Oligocene (Miocene?). The San Vicente Gp. flysch responded to a southward-transporting set of Hercynian-reactivated thrusts. This transport direction differs from the N/NNE direction of the thrusts that existed on the northern side of the Iberian/European plate suture, such as those in the Basque Arc (Chapter 1.3; Figs. 1-2, 1-3 & 1-4). However, the Escudo de Cabuerniga and Jerra Thrusts both strike E/W and transport to the south as do those of the Sierra de Cantabria (Fig. 1-4 & 1-7) and its eastward continuation into the Southern Pyrenean basins (Williams, 1985 amongst others). Such a southerly transport is typical of those thrusts on the Iberian plate in the Basco-Cantabrian basin and confirms the sedimentological and palaeomagnetic interpretation of NW Santander as part of the Iberian plate. The westerly transport direction on the Las

Arenas Thrust is unusual, and is linked the structure of the Bustablado transfer structure (Fig. 1-7).

Further offshore a pattern of N/S-oriented transfer faults exists with comparable Upper Eocene - Miocene thickness and structural trends to those seen in the San Vicente Group flysch (Deregnaucourt & Boillot, 1982b). Here there is a major seismic contact between the top of seismic stratigraphic unit 2 and the base of 1 in the end Middle Eocene (Fig. 12-3). On the Spanish continental margin seismic stratigraphic unit 2 has been extensively deformed by the Pyrenean orogeny (Deregnaucourt & Boillot, 1982b; Boillot, Dupeuble & Malod, 1979; Groupe Galice, 1979). However, in NW Santander the major compressional period of the Pyrenean orogeny postdates (not predates) the end Middle Eocene. This suggests that the Pyrenean orogeny was progressively later towards both the south and the west across the Pyrenean/ Basco-Cantabrian chain. This would explain why the youngest syn-thrust flysch sequence occurs in NW Santander. Final compressional uplift and erosion of the NW Santander area occurred in the Miocene- ? Pliocene and is overlain by Pleistocene-Holocene drift deposits in NW Santander (IGMEa, c, e).

In conclusion, the myriad of carbonate and clastic sequences from deep-marine, shallow-marine, coastal and non-marine environments seen in the Cretaceous/Tertiary of NW Santander can be directly linked to the rifting, drifting and partial closure of the Bay of Biscay.

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