Frontispiece: Looking SW from near Arroyo de Salas, April 1983.

SEDIMENTOLOGY AND TECTONICS OF THE WESTERN CAMEROS BASIN PROVINCE OF BURGOS NORTHERN SPAIN.

D. Phil. Thesis

Department of Earth Sciences

University of Oxford.

Trinity 1986.

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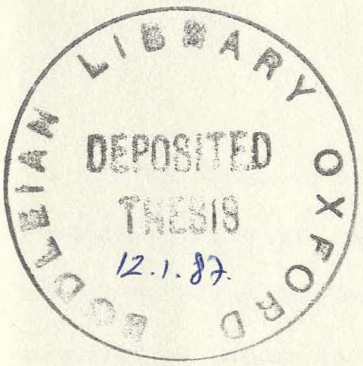

NIGEL HOWARD PLATT

Worcester College

Oxford.

To

Jo and Alan

".... I leave the problem of distinguishing between lacustrine carbonates and calcrete deposits with unsatisfactory resolution."

Colin Klappa (1978)

Morphology, composition and genesis of Quaternary calcrete in the Western Mediterranean: a petrographic approach.

Unpublished Ph.D thesis, University of Liverpool.

".... an intimate link between hinterland tectonics, fluvial activity and basinal [lacustrine] sedimentary facies is highly plausible."

Philip Allen & Maria Mange-Rajetzky (1981)

Sediment dispersal and palaeohydraulics of Oligocene rivers in the eastern Ebro Basin. p715.

Sedimentology, 29, 705-716.

"The older cover rocks are seen chiefly in the Celtiberic chains....These are two parallel structures that run south-east from Burgos to reach the Mediterranean at Valencia. They are mainly composed of gently folded Mesozoic rocks, but the exact parallelism of their structures with that of the Variscan basement indicates that they are no more than a slight reactivation. It is a neglected part of Mesozoic Europe, full of many geological delights that have yet to be published."

Derek V. Ager (1980)

The Geology of Europe, p281.

Sedimentology and Tectonics of the Western Cameros Basin, Province of Burgos, Northern Spain.

Nigel Howard Platt, B. A..
Worcester College, Oxford.
Department of Earth Sciences, University of Oxford.

Thesis submitted for the degree of Doctor of Philosophy,
Trinity 1986.

ABSTRACT.

The Cameros Basin forms the NW extension of the Iberian Chains and exhibits many geological similarities with SW Cantabria and SW Britain.

Jurassic facies and thickness changes reflected differential subsidence of blocks bounded by NE-SW and NW-SE faults. Superposed late Cimmerian unconformities are attr ibuted to successive phases of extension. Footwall uplift resulted in locally severe truncation.

The Late Jurassic - Early Cretaceous continental sediments mark a period of rapid sedimentation. The "Purbeck" sequence (Tierra de Lara Gp) comprises two formations. The Senora de Brezales Fm, consisting of red-bed sandstones and conglomerates, was deposited in small-scale alluvial fans and sandflats in a semi-arid climate. The overlying Rupelo Fm comprises fresh-water marginal and open lacustrine carbonates. Open lacustrine facies of the Rupelo Fm exhibit a fauna / flora of charophytes, ostracods, gastropods and bivalves. Marginal lacustrine facies display pedogenic features including mottling, desiccation brecciation, microkarst solution cavities, rhizoliths and laminar crusts. Rare cherts are evaporite replacements. Carbon and oxygen stable isotope data indicate progressive vadose diagenesis.

The "Wealden" sequence (Pedroso Gp) also comprises two formations. The Hortigüela Fm consists of distal alluvial clastics and sheet-like oncoidal carbonates deposited on a high-energy lake flat. The Piedrahita de Muno Fm above contains stacked sheetflood and rare channel sands, crevasse splays, distal alluvial mudstones with pedogenic carbonates, and rare lacustrine silts and limestones. These are interpreted as terminal river deposits of a semi-arid inland drainage basin. The Salas Gp (Aptian) chiefly comprises distal alluvial mudstones with pedogenic carbonates, sheet conglomerates and sandstones. This association reflects deposition on a muddy alluvial plain. The Utrillas Fm (U Albian-Lr Cenomanian) consists of quartz-pebbly sandstones, quartzitic conglomerates and rare mottled palaeosols, recording mainly "in-channel" deposition on a high-energy braidplain characterised by multiple internal reworking.

The sedimentology and preservation of Oxfordian, Purbeck, Wealden and Aptian strata were controlled by NW-SE (half-graben bounding extensional normal) faults and NE-SW (wrench or "transfer") faults. The sense of displacement on transfer faults periodically reversed, resulting in temporally-variable subsidence patterns and marked regional stratigraphic heterogeneity. Lacustrine carbonate facies can be used to detect synsedimentary tectonic movements in central basinal areas. The Aptian unconformity is related to Austrian phase rifting in the Bay of Biscay. The M Albian unconformity is a "steer's head" onlap produced by eustatic changes and compactional/thermal subsidence.

Geometry of Oligocene conglomerates is correlated with SW-directed Pyrenean thrusting. NW-SE faults controlled the location of blind thrust and thrust-ramp culminations, while NE-SW transfer faults were reactivated as thrust-compartmental faults influencing the derivation of molasse conglomerates and conditioning the location of lateral ramps and backthrusts. E-W Hercynian thrusts and backthrusts were locally reactivated as Tertiary thrusts. The San Leonardo Fault Zone is a complex stack of imbricate thrusts with décollement on Carniolas and upper Purbeck evaporites. Pliocene, Quaternary and modern drainage patterns continued to reflect the Hercynian structures. The Arlanzon River through Burgos lies on a major transfer structure offsetting the W Cameros and SW Cantabria half-graben systems.

The longevity and repeated (often inverse) reactivation of transfer faults demonstrated here has important implications for the development of foreland basins and collisional orogenic belts.

SEDIMENTOLOGIA Y TECTONICA DEL OESTE DE LA CUENCA DE CAMEROS, PROVINCIA DE BURGOS, NORTE DE ESPAÑA.

Nigel Howard PLATT, B. A..
Worcester College, Oxford.
Department of Earth Sciences, University of Oxford.

RESUMEN.

La cuenca de Cameros constituye la prolongación noroccidental de la Cordillera Ibérica y presenta muchas similitudes geológicas con el suroeste de Cantabria y de Inglaterra.

Las facies y los cambios de potencia del Jurásico reflejan la subsidencia diferencial de bloques limitados por fallas de dirección NE-SW y NW-SE. Las discontinuidades superpuestas del Cimmeriense final se atribuyen a fases sucesivas de extensión. El levantamiento del muro del plano de falla dio lugar a truncaciones locales muy importantes.

Los sedimentos continentales del Jurásico terminal y Cretácico inferior, marcan un periodo de sedimentación rápida. La secuencia "Purbeck" (Grupo Tierra de Lara) comprende dos formaciones: La Formación Señora de Brezales, que consiste en areniscas rojas y conglomerados y que se depositó en una sistema de abanicos aluviales de pequeña escala y llanuras arenosas en clima semi-árido, y la Formación suprayacente Rupelo, que comprende carbonatos de agua dulce (lacustres) de zona marginal y abierta. Las facies lacustres abiertas de esta formación contienen fauna y flora de carofitas, ostrácodos, gasterópodos y bivalvos. Las marginales muestran estructuras pedogenéticas, incluyendo moteado de raíces, desecaciones, brechificaciones, cavidades debido a disoluciones por microkarstificación, rizoconcreciones y costras laminares. Ocasionalmente aparece silex (reemplazamientos de evaporitas). Los datos de isótopos estables de carbono y oxígeno indican una diagénesis vadosa progresiva.

La secuencia "Weald" (Grupo Pedroso) comprende tambien dos formaciones: La Formación Hortigüela consiste en clásticos de origen aluvial distal y carbonatos en cuerpos tabulares con oncolitos depositados en una zona lacustre de alta energía. La Formación Piedrahita de Muñó, situada sobre la anterior, contiene apilamientos de arenas transportadas por corrientes en manto, no canalizadas, y algunos escasos canales, lóbulos de derrame, arcillas aluviales distales con carbonatos de origen pedogenético y ocasionalmente limos y calizas lacustres. Se interpretan como depósitos fluviales distales en una cuenca endorreica en clima semi-árido.

El Grupo Salas (Aptiense) esta formado principalmente por arcillas aluviales distales con carbonatos de origen pedogenético, y conglomerados y areniscas en cuerpos tabulares. Esta asociación refleja la deposición en una llanura aluvial fangosa. La Formación Utrillas (Albiense superior - Cenomaniense inferior) consiste en areniscas con cantos de cuarzo, conglomerados cuarcíticos y palaeosuelos con moteado de raíces, interpretandose su origen como dominio de canales en una llanura fluvial braided de alta energía, caracterizada por un intenso retrabajamiento interno.

La sedimentación y conservación de los materiales oxfordienses, purbeck, weald y aptienses, fueron controlados por fallas de direccín NW-SE (de tipo extensional normal limitando un semi-graben) y fallas NE-SW (de tipo transcurriente o en dirección). El sentido del desplazamiento en las fallas en dirección se invertía de forma periódica, de manera que las pautas de subsidencia variaban con el tiempo y dieron lugar a una marcada heterogeneidad estratigráfica a escala regional. Las facies carbonatados lacustres pueden ser utilizadas para detectar movimientos tectónicos sinsedimentarios en las áreas centrales de la cuenca.

La discontinuidad Aptiense se relaciona con la fase Aústrica durante la cual se produce una etapa de rifting en el Golfo de Vizcaya. La discontinuidad del Albiense medio es un onlap en "cabeza de toro" producido por cambios eustaticos y subsidencia térmica y compactacional.

La geometría de los conglomerados oligocenos se correlaciona con el cabalgamiento pirenaico de dirección SW. Las fallas NW-SE controlaron la localización de las culminaciones de los cabalgamientos ocluidos y frentes de las rampas de cabalgamiento, mientras que las fallas transcurrentes de dirección NE-SW se reactivaron como fracturas de cabalgamientos compartimentados, influenciando la derivación de los conglomerados molásicos y condicionando la localización de las rampas laterales y los ante-cabalgamientos. Los cabalgamientos y ante-cabalgamientos hercínicos E-W se reactivaron localmente en el Terciario. La zona de fractura de San Leonardo es un apilamiento complejo de cabalgamientos imbricados con despegue en las facies "Carniolas" del Infralias y en las evaporitas de la parte superior del Purbeck. Los modelos de drenaje en el Plioceno, en el Cuaternario y en la actualidad, continúan reflejando las estructuras hercínicas. El curso del rio Arlanzón, que atraviesa la provincia de Burgos, sigue la traza de una estructura transcurrente principal, reproduciendo los sistemas de semi-graben del oeste de Cameros y suroeste de Cantabria.

La longevidad de las fallas transcurrentes y sus repetidas reactivaciones (a menudo con movimiento invertido) tiene aqui implicaciones importantes en cuanto al desarollo de cuencas de antepais y cinturones orogénicos de colisión.

Sedimentology and Tectonics of the Western Cameros Basin,
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Nigel Howard Platt, B. A..
Worcester College, Oxford.
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EXTENDED ABSTRACT.

CHAPTER 1. INTRODUCTION.

The Cameros Basin is located approximately 170km NNE of Madrid (fig 1) and occupies an area of 100 x 60km. The Basin forms the NW extension of the Iberian Chains and also exhibits many geological similarities with S Cantabria.

The Cameros Basin contains sediments of Triassic to Eocene age. These sediments rest on a Palaeozoic basement of Cambrian, Ordovician and Westphalian age. Both the orientation of rifting during Triassic and Late Jurassic extension and the direction of thrusting under compression were strongly controlled by the late Hercynian structural grain of the basement. The Late Jurassic - Early Cretaceous continental sediments are particularly well-developed and very thick (up to 5km in the E of the Basin) and mark a period of dramatically-increased subsidence.

The present study has concentrated upon the W portion of the Basin where Late Jurassic - Early Cretaceous continental sedimentation was initiated. The sequences are generally thinner in this area but the influence of tectonics is more clearly evident.

The approach adopted by this study initially involved detailed sedimentological study of the Late Jurassic - Early Cretaceous continental sequence. Mapping of facies distributions and structural studies have been integrated to facilitate improved understanding of the tectonic controls on sedimentation.

This work has involved:

1) Comparison of stratigraphy (fig 2), sedimentology, and structure with other areas in the N Atlantic region, eg W Shelves of the British Isles, Aquitaine, E Iberian Chains and the Atlantic marginal shelves of Portugal and E Canada.

2) The erection of a new and more detailed stratigraphy for the Late Jurassic - Early Cretaceous sediments (fig 3).

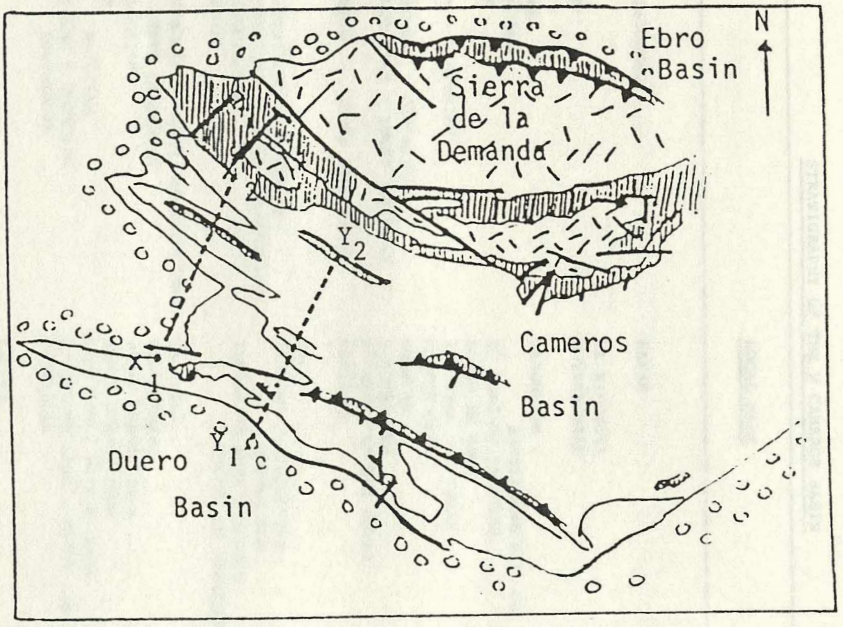
3) Detailed facies modelling of the sediments in this sequence. Particular emphasis has been placed upon the spectacularly developed "Purbeck" facies.

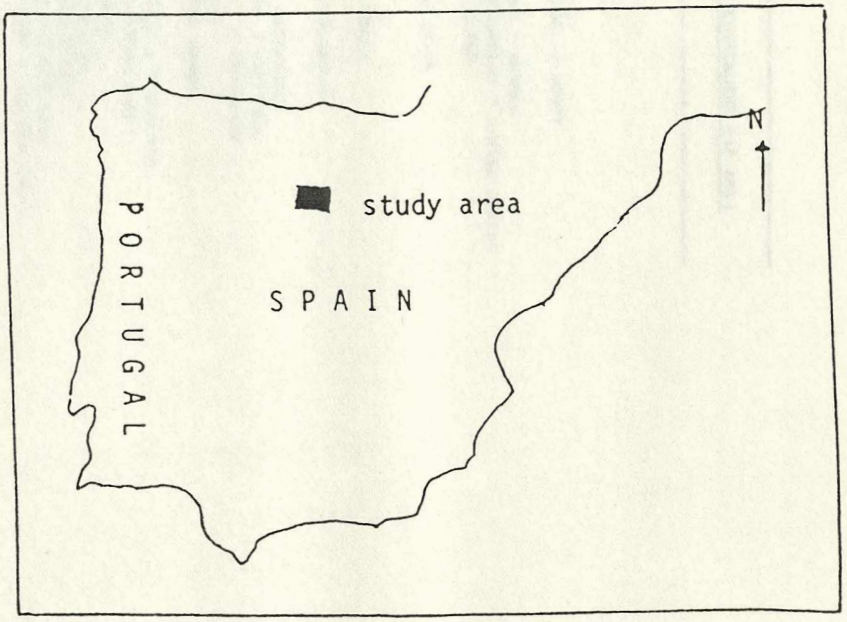
4) Examination of facies distributions and evaluation of the nature and importance of fault control on differential subsidence and sedimentation.

5) Field mapping of a) intra-Mesozoic unconformities
b) Tertiary thrust sheets
c) larger-scale structures
and compilation of structural cross-sections across the Basin. This work has concentrated on areas of particular structural and sedimentological interest.

6) The first integrated account of Mesozoic - Cenozoic basin evolution applying recently-developed concepts about the initiation, development and inversion of basins and the respective influences in sedimentation and deformation of basin-bounding extensional normal faults, cross-graben wrench or "transfer" faults and thrusts.

Geological Map of the Western Cameros Basin showing orientation of sections.





The Iberian Peninsula: Location Map

fig 1

STRATIGRAPHY OF THE W CAMEROS BASIN

PERIOD	STAGE	LOCAL NAME	U. K. EQUIVALENT (IF ANY)
QUATERNARY			
	PLIO-PLEISTOCENE	RANAS	
TERTIARY	MIOCENE	LA HINOJOSA	
	OLIGOCENE	COVARRUBIAS	BEMBRIDGE / BOVEY
	PALAEOCENE	GARUMNIAN / SANTIBANEZ DEL VAL	READING / BARTON
	MAASTRICHTIAN		MAASTRICHTIAN / DANIAN CHALKS
	CAMPANIAN	S^{o} DOMINGO DE SILOS	UPPER CHALK
		BURGO DE OSMA	
	CONIACIAN - CAMPANIAN	HONTORIA DEL PINAR	MIDDLE CHALK
		HORTEZUELOS	
	UPPER TURONIAN - CONIACIAN	MUNECAS	
CRETACEOUS	UPPER CENOMANIAN - LOWER TURONIAN	PICOFRENTES	LOWER CHALK
		S^{a} MARIA DE LAS HOYAS	
	UPPER ALBIAN - LOWER CENOMANIAN	UTRILLAS	UPPER GREENSAND
	APTIAN	SALAS	LOWER GREENSAND
	UPPER VALANGINIAN - BARREMIAN	PEDROSO / PIEDRAHITA / HORTIGUELA	WEALDEN / HASTINGS / FAIRLIGHT
	UPPER BERRIASIAN - LOWER VALANGINIAN		
	? LOWER BERRIASIAN	TIERRA DE LARA / RUPELO / SENORA DE BREZALES	PURBECK
	? PORTLANDIAN		PORTLAND SANDS
	UPPER OXFORDIAN - LOWER KIMMERIDGIAN	TALVEILA	CORALLIAN & KELLOWAYS
	MIDDLE - UPPER CALLOVIAN	VILLAVELAYO	LOWER OXFORD CLAY
JURASSIC	LOWER CALLOVIAN	UPPER ORANGE BEDS	CORNBRASH
	BATHONIAN	LOWER ORANGE BEDS	GREAT OOLITE
	AALENIAN - BAJOCIAN	LIMESTONES WITH SPONGES	INFERIOR OOLITE
	PLIENSBACHIAN - TOARCIAN	LIMESTONE LIAS & MARLY LIAS	MIDDLE & UPPER LIAS
	RHAETIAN - SINEMURIAN	CARNIOLAS	LOWER LIAS & PENARTH GROUP
	KEUPER	KEUPER	KEUPER / MERCIA MUDSTONE
TRIASSIC	MUSCHELKALK	MUSCHELKALK	WATERSTONES
	BUNTER	BUNTER	BUNTER / SHERWOOD SANDSTONE
CARBONIFEROUS	WESTPHALIAN B - C	WESTPHALIAN B - C	COAL MEASURES
ORDOVICIAN	TREMADOC	BARRIOS	
CAMBRIAN	LOWER - MIDDLE	BARBADILLO DEL PEZ	

PLATT N (1986)

fig 2

fig 3

STRATIGRAPHY OF THE "PURBECK" / "WEALDEN" SEQUENCE OF THE W CAMEROS & SW CANTABRIA.

		W CAMEROS					SW CANTABRIA		
PEDROSO GROUP	~40m HORTIGUELA FM	ZURRAMUJERES MB	oncoidal limestones	clastics	TERRAZAS MB	PIEDRAHITA DE MUNO FM up to 2000m+	sandstones & oncoidal lsts	UPPER	CORVIO & EMBALSE DE PISUERGA MBS
				----------black limestones---------					
	~20m	SAN MARTIN MB	marls	clastics	VIZCAINOS MB ~200m				
----------desiccation-cracked surface----------							----------		
	0-20m	RIO CABRERA MEMBER	red ostracod lst chert yellow/vuggy lsts				chert	UPPER-MIDDLE	LOMILLA
		----------					----------		
	0-50m RUPELO FM	MAMBRILLAS DE LARA MB	Dinosaur Bed dark biomicrites & charophyte marls				fossiliferous	MIDDLE	AGUILAR SUP.
TIERRA DE LARA GROUP		----------Torrelara Bed----------					----------		
		--------brecciated surface----------							
	0.5-7m	LADERA MB	white lsts (brecc/pelleted) mottled lsts	red marls	LAS VINAS MB 0-40m		microkarst mottled	LOWER	AGUILAR INF.
	----------laminated limestone----------						----------		
	SENORA DE BREZALES FM 2-7m		crusts polygenetic cgl red sands				crusts polygenetic cgl	BASAL	VILLAESPINOSO
----------karstic surface / Fe crust----------							----------		
	TALVEILA FM 0-75m		lsts with corals ssts/yellow calclithites						MISSING
	----------karst / erosive surface----------								
			MARINE JURASSIC					MARINE JURASSIC	

CHAPTER 2. STRATIGRAPHY.

The Mesozoic stratigraphy is summarised (fig 2). A new and more detailed stratigraphy is presented for the "Purbeck" and the "Wealden" (fig 3). A tentative correlation is proposed between the Purbeck sequences of the W Cameros and SW Cantabria. The similarities in stratigraphic development imply that the two areas were interconnected or intercommunicating during Purbeck times at least.

CHAPTER 3. SEDIMENTOLOGY OF THE "PURBECK" SEQUENCE (TIERRA DE LARA GROUP).

1. Senora de Brezales Formation.

Red-bed sandstone and conglomerate facies were laid down (fig 4) in small-scale alluvial fans and sandflats. An arid to semi-arid climate is inferred from the nature of the facies and from evidence of periodic pedogenesis. The deposits were the erosion products of subjacent marine Jurassic strata.

2. Rupelo Formation.

Carbonate facies were deposited (fig 5) in fresh-water marginal lacustrine and open lacustrine environments. Marginal lacustrine facies display a wide variety of pedogenic features including mottling, desiccation brecciation, microkarst solution cavities, rhizoliths and laminar crusts. Open lacustrine facies exhibit a fresh-water fauna/flora of ostracods, charophytes and gastropods. Sauropod bones and abundant footprints were also found. Lamination is rare in the open lacustrine facies, indicating shallow, oxygenated conditions. The presence of abundant intraclasts suggests significant internal reworking, both of marginal facies washed in from the area surrounding the lake during storms, and of open lacustrine facies along internal "low-stage lake-flat drainage channels" during low lake stands. Open lacustrine intraclasts were locally reworked to form grainstones, which are interpreted as high-energy shoreline deposits.

Thin 2 - 80cm chert horizons are also present. The chert displays pseudomorphs and locally contains length-slow chalcedony, indicative of evaporite replacement.

Diagenesis is discussed and the main features of the early vadose and void-filling fresh-water phreatic cements are described. Late ferroan calcite cements are locally present.

Facies sequences dominantly reflect climatic influences.

CHAPTER 4. SEDIMENTOLOGY OF THE "WEALDEN" SEQUENCE (PEDROSO GROUP).

1. Hortigüela Formation.

Clastic and carbonate sediments were deposited (fig 6) in fluvial and littoral lacustrine environments respectively. The carbonates are sheet-like, containing charophytes and abundant calcified cyanophyte oncoids 1-30cm in diameter. The morphology and palaeoecology of the oncoids are similar to other examples documented from fresh-water settings.

2. Piedrahita de Muno Formation.

This Formation chiefly comprises clastic sediments which are interpreted (fig 7) as fluvial deposits. Channelled, cross-bedded sands are interbedded with overbank muds which contain calcified burrows and display various degrees of pedogenic mottling. Nodular pedogenic carbonates ("cornstones") are occasionally developed. Thin, tabular, bioturbated sand bodies are abundant within the mud units; these sand units are interpreted as crevasse-splay deposits. Greenish

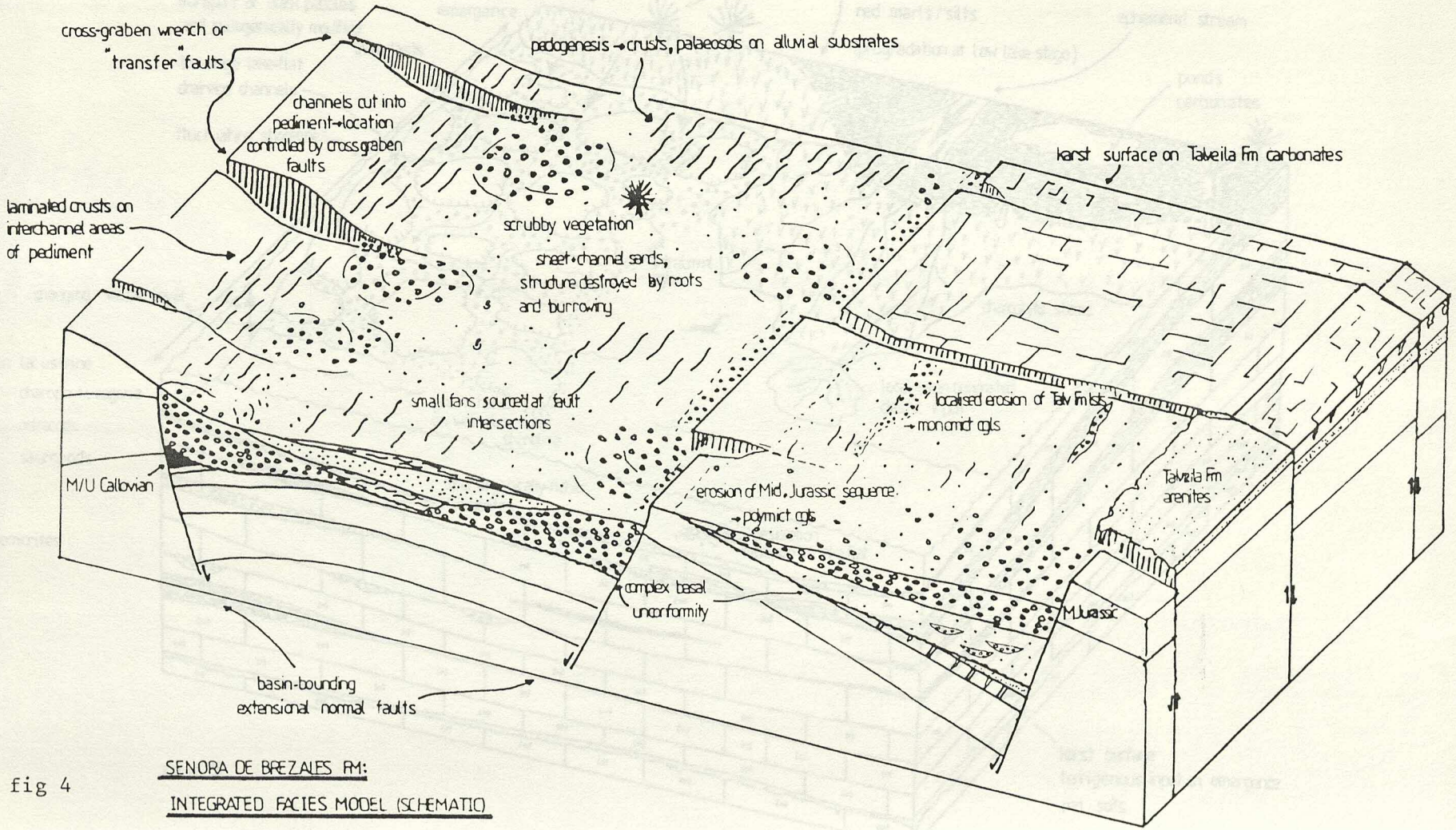


fig 4

SENORA DE BREZALES FM:

INTEGRATED FACIES MODEL (SCHEMATIC)

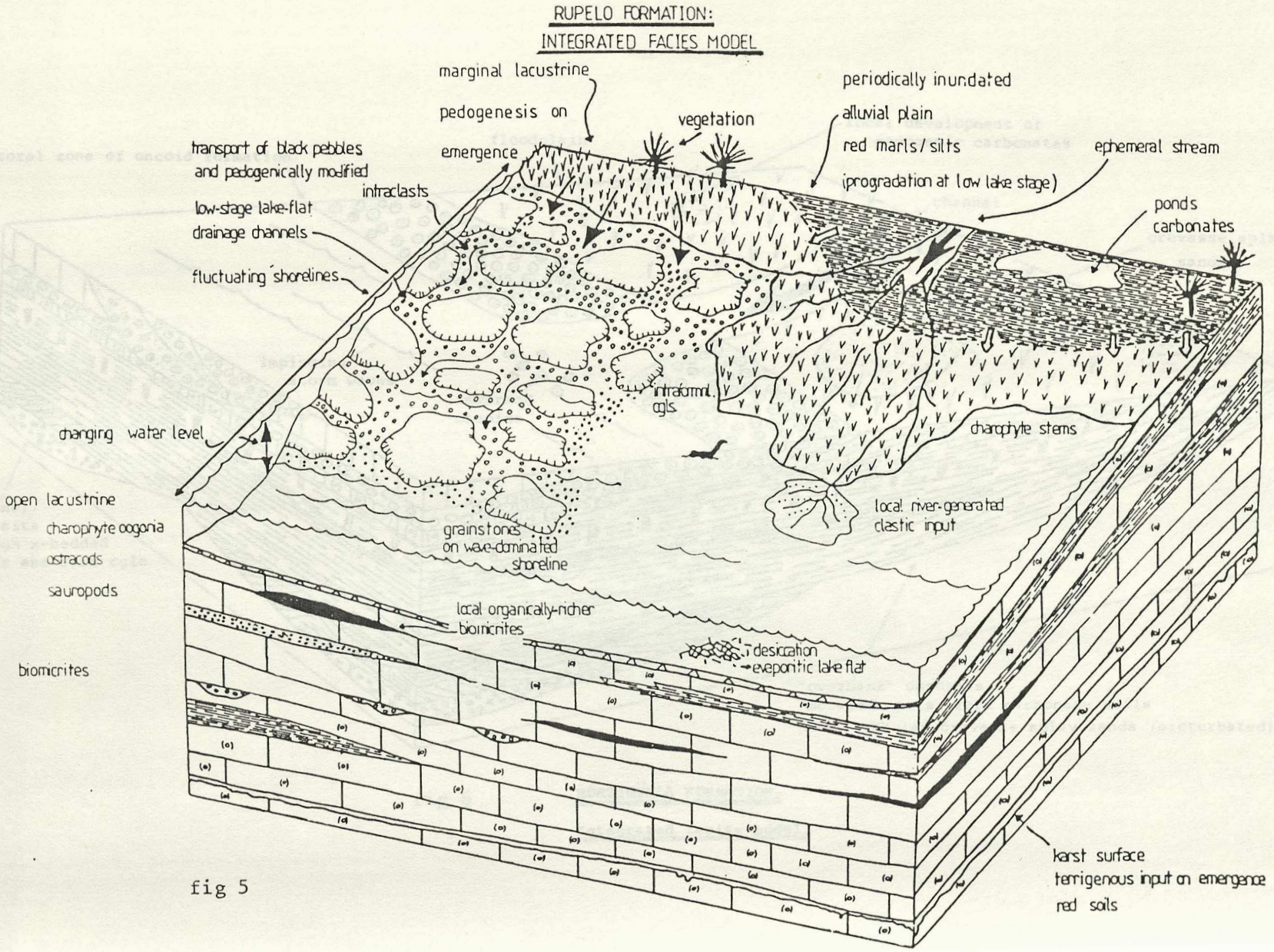

RUPELO FORMATION:
INTEGRATED FACIES MODEL
marginal lacustrine
pedogenesis on
emergence
vegetation
periodically inundated
alluvial plain
red marls / silts
(progradation at low lake stage)
ephemeral stream
ponds
carbonates
transport of black pebbles
and pedogenically modified
intraclasts
low-stage lake-flat
drainage channels
fluctuating shorelines
intraform. ggls
charophyte stems
changing water level
open lacustrine
charophyte oogonia
ostracods
sauropods
grainstones
on wave-dominated
shoreline
local river-generated
clastic input
local organically-richer
biomicrites
desiccation
evaporitic lake flat
biomicrites
karst surface
terrigenous input on emergence
red soils

fig 5

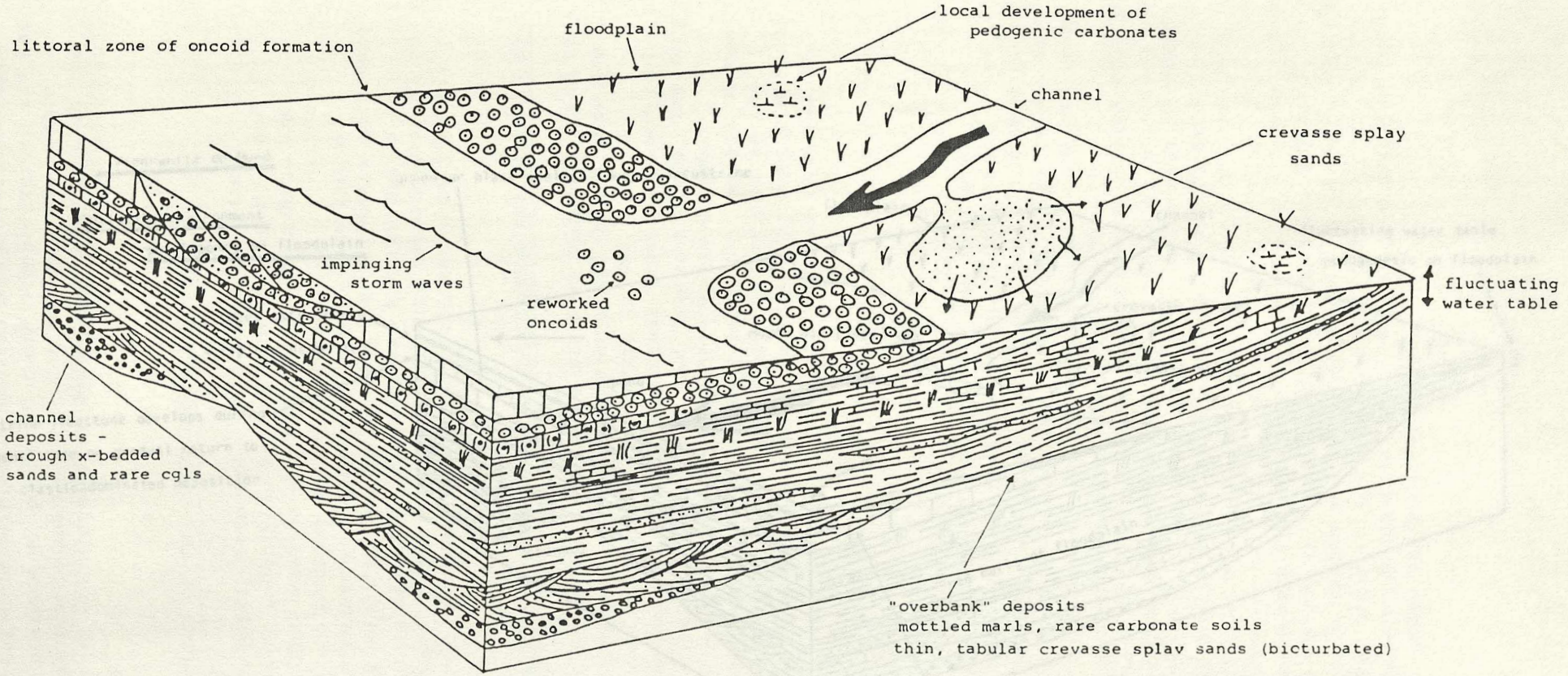


fig 6 HORTIGÜELA FORMATION:

Integrated facies model.

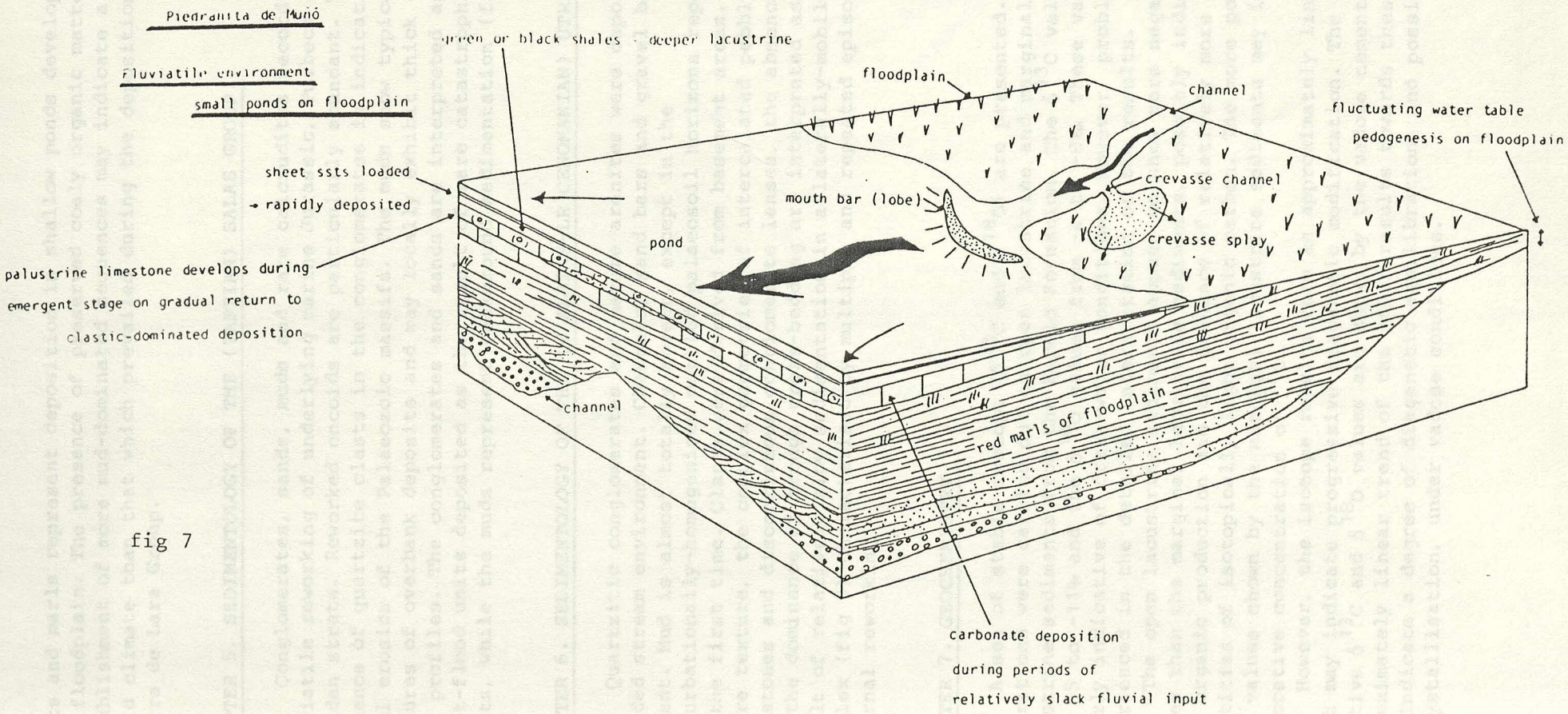

Piedranitta de Muñó
Fluviatile environment
small ponds on floodplain
green or black shales - deeper lacustrine
floodplain
channel
fluctuating water table
pedogenesis on floodplain
sheet ssts loaded
→ rapidly deposited
mouth bar (lobe)
crevasse channel
pond
crevasse splay
palustrine limestone develops during
emergent stage on gradual return to
clastic-dominated deposition
channel
red marls of floodplain
carbonate deposition
during periods of
relatively slack fluvial input

fig 7

silts and marls represent deposition in shallow ponds developed upon the floodplain. The presence of preserved coaly organic matter and the establishment of more mud-dominated sequences may indicate a more humid climate than that which prevailed during the deposition of the Tierra de Lara Group.

CHAPTER 5. SEDIMENTOLOGY OF THE (?APTIAN) SALAS GROUP.

Conglomerates, sands, muds and rare calcrudites record fluviatile reworking of underlying marine Jurassic, Purbeck and Wealden strata. Reworked oncoids are particularly abundant. The presence of quartzite clasts in the conglomerates is indicative of local erosion of the Palaeozoic massifs. The muds show typical features of overbank deposits and may locally exhibit thick carbonate soil profiles. The conglomerates and sands are interpreted as sheet-flood units deposited as the result of rare catastrophic flood events, while the muds represent background sedimentation (fig 8).

CHAPTER 6. SEDIMENTOLOGY OF THE (U ALBIAN-LR CENOMANIAN) UTRILLAS FM.

Quartzitic conglomerates and mature arenites were deposited in a braided stream environment. Channels, sand bars and gravel bars are present. Mud is almost totally absent, except in the bioturbationally-homogenised mottled palaeosoil horizons reported here for the first time. Clasts were derived from basement areas. The very mature texture, the complex geometries of intercalated pebbly sandstones and discontinuous conglomerate lenses, the absence of mud and the dominance of trough cross-bedding are interpreted as the result of relatively slow sedimentation in a laterally-mobile channel complex (fig 9) characterised by multiple and repeated episodes of internal reworking.

CHAPTER 7. GEOCHEMISTRY.

A set of stable isotope ($\delta^{13}C$ and $\delta^{18}O$) are presented. Separations were carried out on open lacustrine and marginal lacustrine sediments from the Rupelo Formation. The $\delta^{13}C$ values range from -5 to -11‰ and the $\delta^{18}O$ values from -4 to -8‰. These values are clearly indicative of fresh-water conditions. However, problems were experienced in the detailed interpretation of the results.

The open lacustrine sediments appear to show more negative $\delta^{13}C$ values than the marginal lacustrine sediments, possibly indicating more organic production and the presence of relatively more elevated quantities of isotopically light organic carbon. The more positive $\delta^{18}O$ values shown by the marginal lacustrine sediments may indicate evaporative concentration of ^{18}O.

However, the isotope results form an approximately linear trend which may indicate progressive diagenetic modification. The most negative $\delta^{13}C$ and $\delta^{18}O$ values are shown by the vadose cements, and the approximately linear trend of the other results towards these values may indicate a degree of diagenetic equilibration, and possibly even recrystallisation, under vadose conditions.

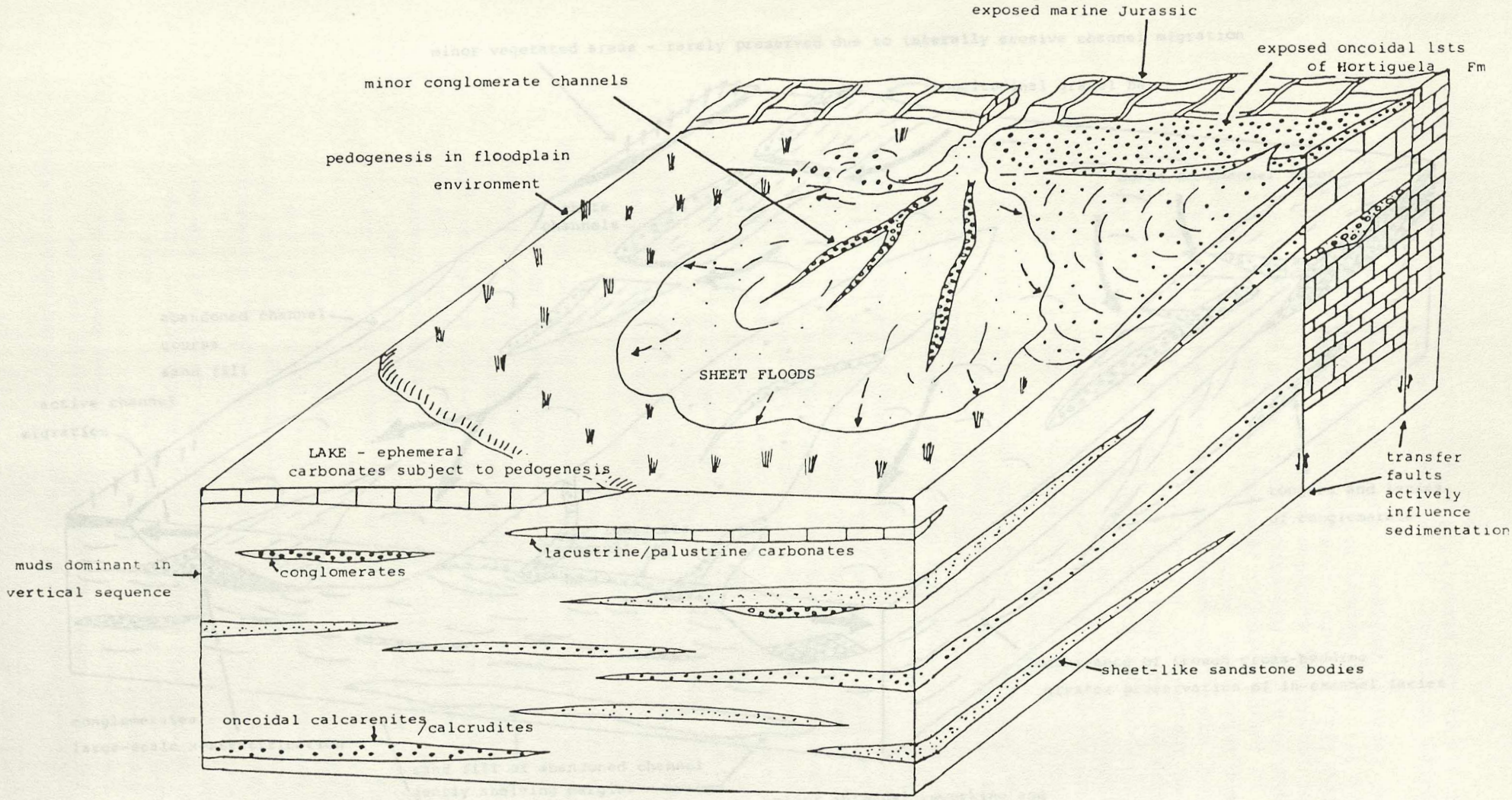


fig 8 SALAS GROUP:
INTEGRATED FACIES MODEL.

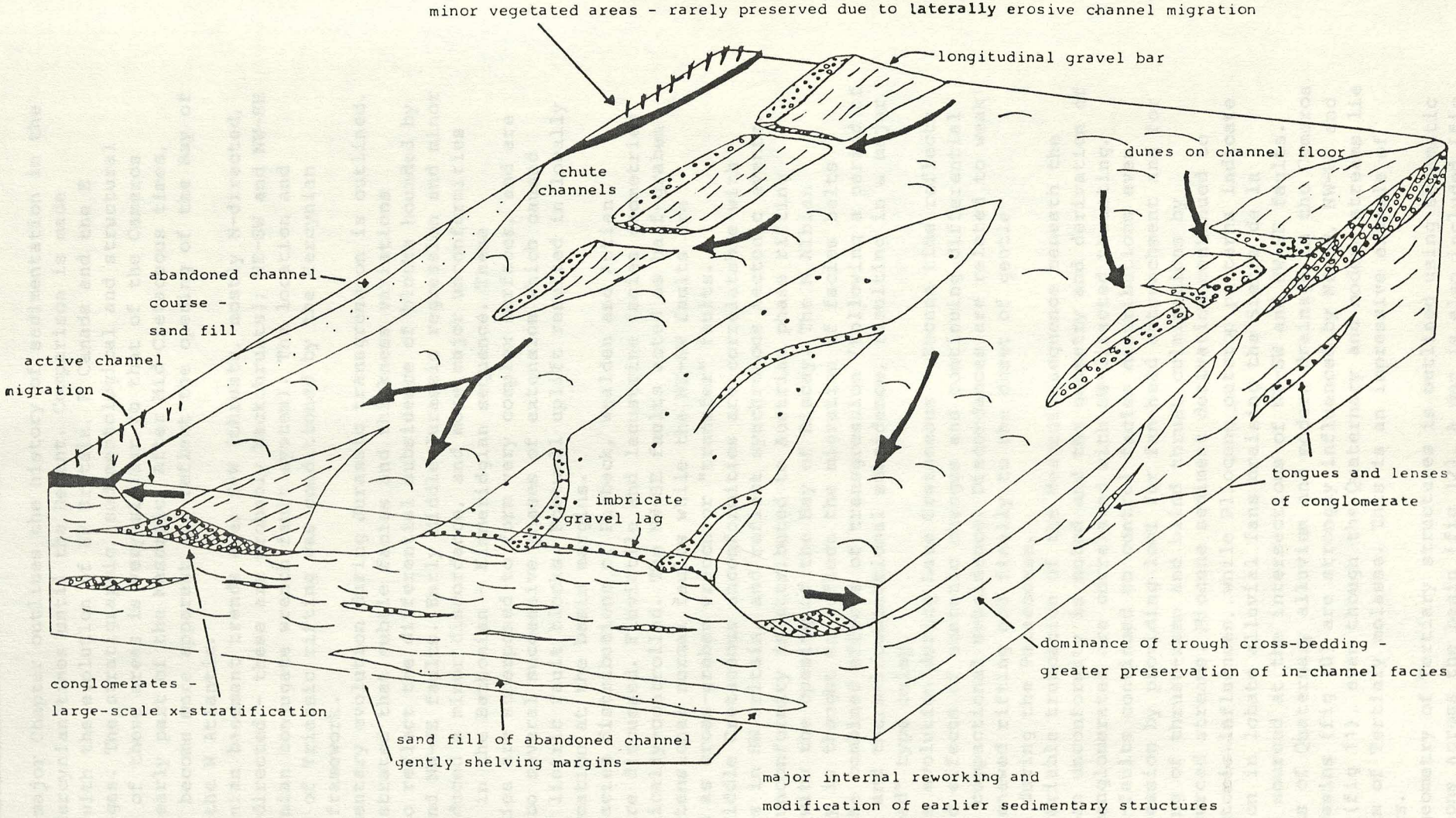


fig 9 UTRILLAS FORMATION:
INTEGRATED FACIES MODEL.

CHAPTER 8. SEDIMENTATION AND TECTONICS.

This major Chapter outlines the history of sedimentation in the area from Hercynian times until the Recent. Comparison is made throughout with the evolution of SW Britain, E Canada and the E Iberian Ranges. The stratigraphic, sedimentological and structural development of these areas is very similar to that of the Cameros during the early part of the Mesozoic. After Mid-Cretaceous times, differences become more apparent and reflect the opening of the Bay of Biscay and the N Atlantic.

Hercynian basement trends are: E-W (thrusts, mostly N-directed, but some S-directed - these are probably backthrusts); NE-SW and NW-SE (late Hercynian conjugate wrench fault system). The location and orientation of Triassic rifting was conditioned by the Hercynian structural framework.

Sedimentary evolution during Jurassic transgression is outlined. It is demonstrated that subtle facies and thickness variations continued to reflect the differential subsidence of blocks bounded by the NE-SW and NW-SE faults. Early Middle Jurassic regression and minor folding produced a minor discordance, and more major unconformities are present in the Bathonian - Kimmeridgian sequence. These unconformities are superposed to form very complex surfaces, and are attributed to several successive phases of extension which caused rotation of listric fault blocks. Footwall uplift resulted in locally severe truncation at the basin margins.

The facies distributions of Purbeck, Wealden and Aptian sediments are discussed. Fluviatile and lacustrine facies geometries were tectonically-controlled. The NW-SE faults acted as half-graben bounding extensional normal faults while the NE-SW faults are interpreted as cross-graben wrench or "transfer" faults.

The Middle Cretaceous unconformities are correlatable with discordances in SW Britain, and reflect synchronous tectonic events. The Aptian unconformity is attributed to Austrian phase rifting associated with the opening of the Bay of Biscay. The M Albian unconformity is thought to reflect the migration of facies belts caused by the combined effects of transgression (following a period of regression) and thermal/compactional subsidence, resulting in a major "steer's-head" type onlap.

Facies evolution during Late Cretaceous - Eocene time reflects the combined effects of eustatic changes and continuing differential thermal and compactional subsidence. Discordances are related to weak phases of renewed rifting and finally to the onset of gentle compression during the Palaeocene.

The variable truncation of the Mesozoic sequence beneath the main Pyrenean unconformity is noted and the geometry and derivation of Oligocene conglomerates are correlated with SW-directed thrusting. NE-SW cross-faults continued to control facies distributions even under compression by providing loci for fan-head entrenchment and for the breaching of thrust-ramp and blind thrust culminations by multiple-sourced streams. Miocene sediment derivation continued to reflect tectonic influences, while Pliocene outcrop patterns indicate sedimentation in lobate alluvial fans draining the Sierra de la Demanda and sourced at the intersections of NE-SW and NW-SE faults. The patterns of Quaternary alluvium and modern drainage in the Cameros and Duero Basins (fig 10) are strongly influenced by NE-SW, NW-SE and E-W trends (fig 11), even though the Quaternary and modern streams lie on up to 4km of Tertiary molasse. This is an impressive example of neotectonics.

The geometry of Tertiary structures is outlined using schematic cross-sections across the basin (fig 12). A map is also included (fig

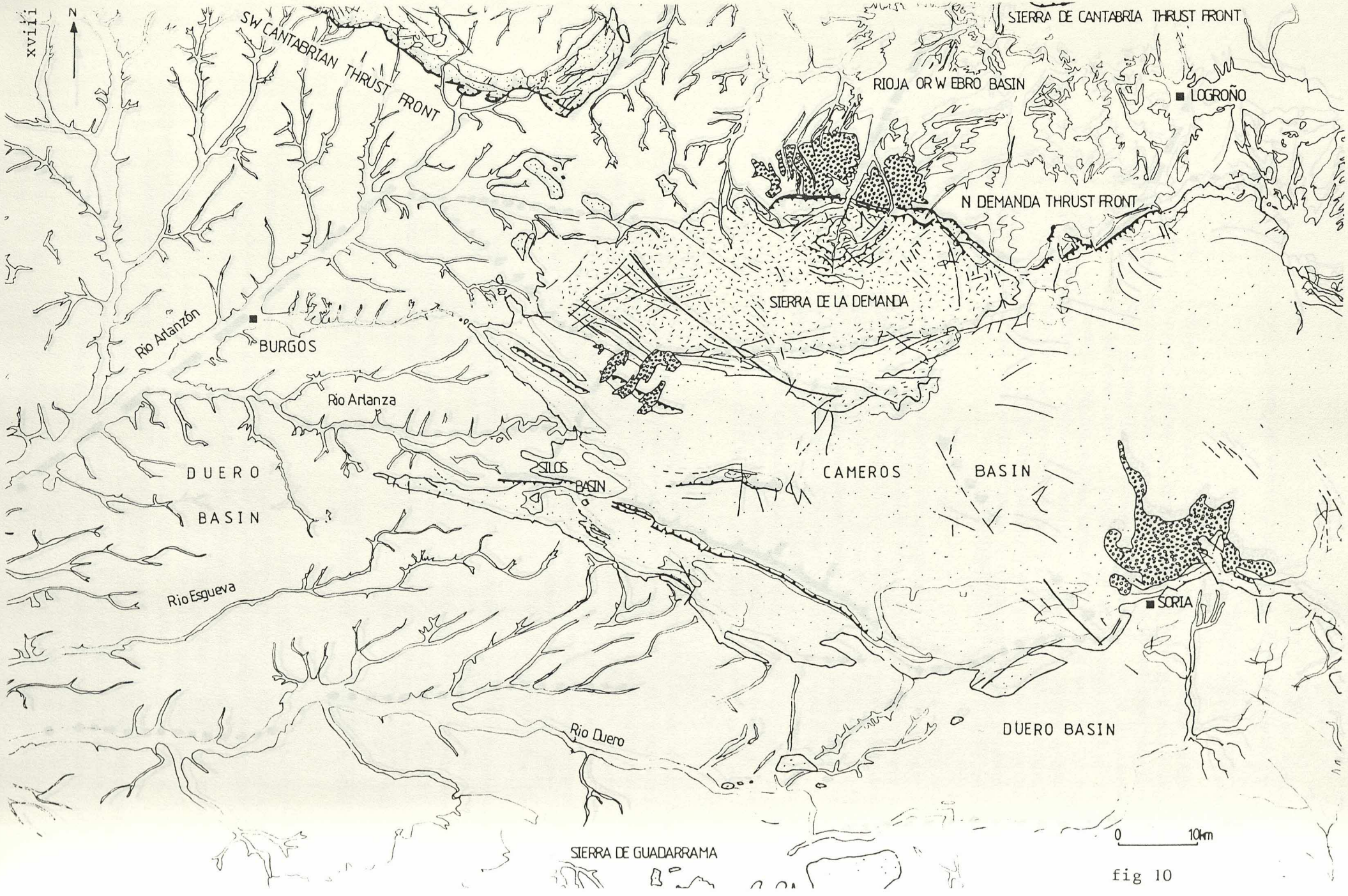

N
SW CANTABRIAN THRUST FRONT
SIERRA DE CANTABRIA THRUST FRONT
RIOJA OR W EBRO BASIN
LOGROÑO
N DEMANDA THRUST FRONT
SIERRA DE LA DEMANDA
Rio Arlanzón
BURGOS
Rio Arlanza
DUERO
BASIN
SILOS
BASIN
CAMEROS
BASIN
Rio Esgueva
SORIA
Rio Duero
DUERO BASIN
SIERRA DE GUADARRAMA
0
10km

fig 10

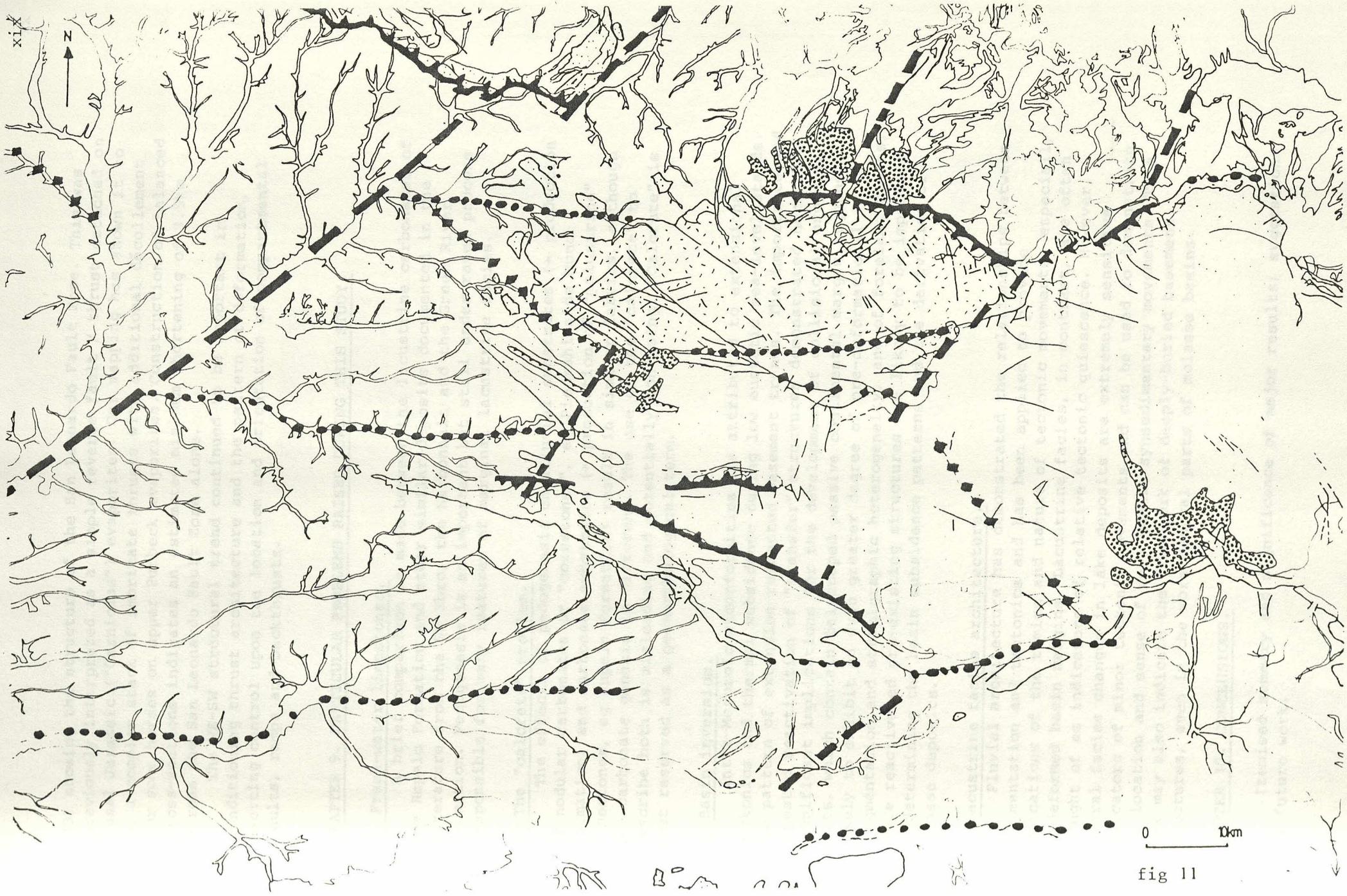


fig 11

13) showing the structure of the San Leonardo Fault Zone. This was previously interpreted as a simple reverse fault / thrust detached on basal Jurassic ("Carniolas") evaporites. Our mapping has shown it to be a complex stack of imbricate thrusts with additional décollement for some horses on upper Purbeck evaporites. Construction of balanced cross-sections indicates an estimated minimum shortening of 3.5km across the San Leonardo Fault Zone alone.

The NE-SW structural trend continued to be important in conditioning thrust architecture and the pattern of deformation, exerting control upon the location and orientation of compartmental faults, ramps and backthrusts.

CHAPTER 9. PARTICULAR PROBLEMS RAISED DURING THIS STUDY.

1. Fresh-water limestones.

A brief comparison is made between the lacustrine carbonates of the Rupelo Formation and other similar deposits documented in the literature from the SE Ebro, the Narbonnais and the Green River Formation. Pedogenesis is an important but still under-rated process responsible for many features of marginal lacustrine facies.

2. The "calcrete" problem.

The effects of pedogenesis on alluvial substrates (→ production of nodular carbonates or "cornstones", eg Piedrahita de Muno Formation) and carbonate substrates (→ production of "palustrine" limestones, eg Rupelo Formation) result in similar fabrics, although the carbonate genesis is different. The use of the "calcrete" to describe both is misleading and potentially confusing. "Calcrete" is best reserved as a geomorphological term.

3. Basin inversion.

Intra-Mesozoic unconformities are attributed to extensional tectonics and thermal subsidence during low eustatic sea-level stands. The pattern of extension reflected basement trends. The importance and repeated reactivation of "transfer" structures demonstrated here have significant implications for the development of collisional orogenic belts, which contain telescoped passive continental margin sequences, likely to exhibit an even greater degree of pre-deformational fragmentation and stratigraphic heterogeneity, and of foreland basins, where reactivated pre-existing structures are likely to be important in determining the basin subsidence patterns and the derivation of molasse deposits.

4. Lacustrine facies architecture.

Fluvial architecture has demonstrated the relationships between sedimentation and tectonics and has been applied to provide indications of the timing and nature of tectonic movements, especially at deformed basin margins. Lacustrine facies, in contrast, are often thought of as indicators of relative tectonic quiescence. However, lateral facies changes in lake deposits are extremely sensitive indicators of minor tectonic movements and can be used to demonstrate the location and sense of very gentle synsedimentary movements. This work may also indicate the framework of deeply-buried basement structures, even in the more central parts of molasse basins.

CHAPTER 10. CONCLUSIONS.

Itemised summary and significance of major results; suggestions for future work.

W CAMEROS BASIN — SCHEMATIC CROSS-SECTIONS

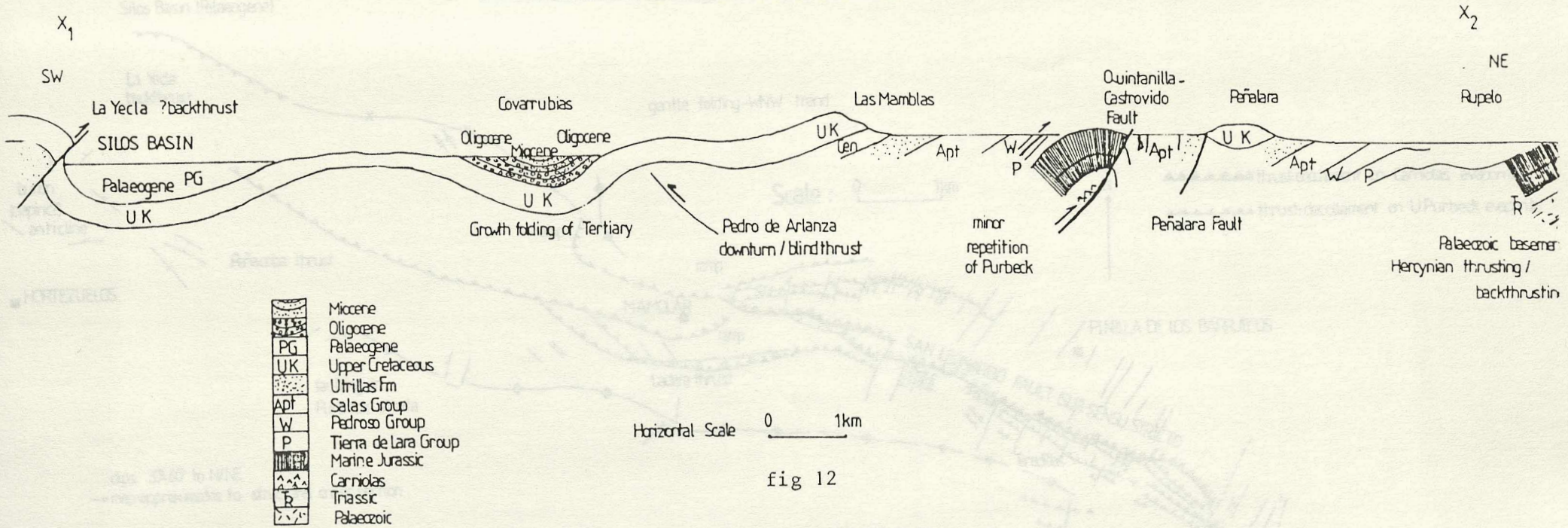
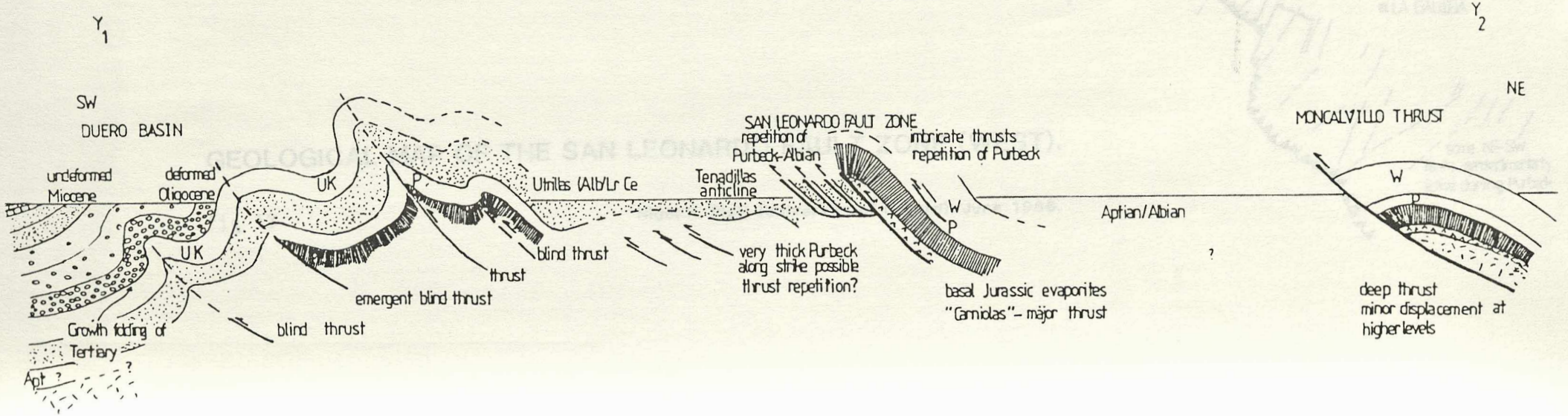
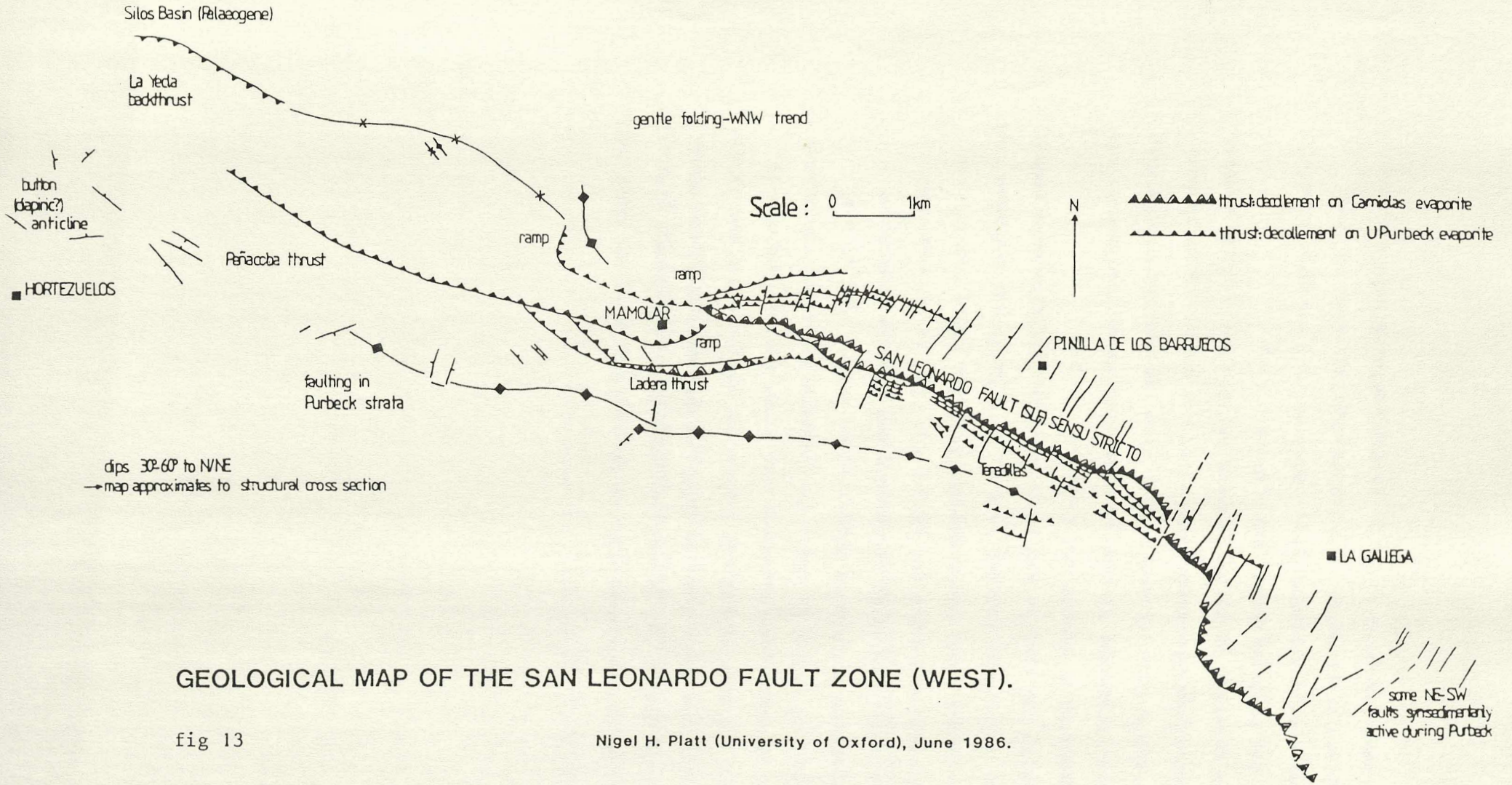


fig 12





GEOLOGICAL MAP OF THE SAN LEONARDO FAULT ZONE (WEST).

fig 13

Nigel H. Platt (University of Oxford), June 1986.

Looking NE from near Villanueva de Carazo towards the Sierra de la Demanda. April, 1983.

ACKNOWLEDGEMENTS.

This project was proposed by Dr H. G. Reading. I thank him for his supervision in the field and in Oxford. Dra Angela Alonso also contributed much to this work both in Spain and during a visit to Oxford. I am particularly grateful to her for encouraging me in my efforts to learn Spanish.

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Nigel PLATT. Department of Earth Sciences, University of Oxford.

SEDIMENTOLOGY AND TECTONICS OF THE W CAMEROS BASIN, N SPAIN.

CHAPTER 1.	INTRODUCTION	1
CHAPTER 2.	STRATIGRAPHY	10
CHAPTER 3.	SEDIMENTOLOGY OF THE "PURBECK" SEQUENCE (TIERRA DE LARA GROUP)	36
CHAPTER 4.	SEDIMENTOLOGY OF THE"WEALDEN" SEQUENCE (PEDROSO GROUP)	89
CHAPTER 5.	SEDIMENTOLOGY OF THE (?APTIAN) SALAS GROUP	108
CHAPTER 6.	SEDIMENTOLOGY OF THE (?U ALBIAN-LR CENOMANIAN) UTRILLAS FORMATION	113
CHAPTER 7.	GEOCHEMISTRY	121
CHAPTER 8.	SEDIMENTATION & TECTONICS	130
CHAPTER 9.	PARTICULAR PROBLEMS RAISED DURING THIS STUDY	191
CHAPTER 10.	CONCLUSIONS	202
REFERENCES.		213

CONTENTS:

Opening Quotations i

ABSTRACT ii

RESUMEN iii

EXTENDED ABSTRACT v

ACKNOWLEDGEMENTS xxiii

CONTENTS xxv

CHAPTER 1: INTRODUCTION. 1

1.1. Location (Geographical) 1
1.2. Setting (Geological) 1
1.3. Aims of the project 2
1.4. Methods 4
1.4.1. Field 4
1.4.2. Laboratory 5
1.5. Previous work in the area. 6

CHAPTER 2: STRATIGRAPHY. 10

2.1. Palaeozoic 10
2.2. Triassic 11
2.3. Triassic-Jurassic "Carniolas" 11
2.4. Lias 12
2.5. Aalenian and Bajocian 12
2.6. Bathonian and Callovian 13
2.7. Talveila Formation (? U Oxfordian - Lr Kimmeridgian) 14
2.8. - 2.11. Continental Sequence (Kimmeridgian - Lr Cenomanian) 16
2.8. Tierra de Lara Group (Group I) 17
2.9. Pedroso Group (Group II) 22
2.10. Salas Group (Group IV) 27
2.11. Utrillas Formation (Group V) 27
2.12. - 2.19. Upper Cretaceous 29
2.12. Santa Maria de las Hoyas Formation (M - U Cenomanian) 30
2.13. Picofrentes Formation (Lr - M Turonian) 31
2.14. Munecas Formation (M-U Turonian - Lr Coniacian) 31
2.15. Hortezuelos Formation (Coniacian sensu lato) 32
2.16. Hontoria del Pinar Formation (Coniacian-Santonian boundary - basal Campanian) 32
2.17. Burgo de Osma Formation (U Santonian - basal Campanian) 33
2.18. Santo Domingo de Silos Formation (Campanian) 33
2.19. Santibanez del Val Formation (Maastrichtian - Palaeogene) 33
2.20 - 2.22. Tertiary 34
2.19. ("Garumnian") 34
2.20. "Oligocene" - syntectonic 34
2.21. "Miocene" - post-tectonic 34
2.22. "Pliocene" 35
2.23. Quaternary 35

CHAPTER 3: THE PETROGRAPHY AND SEDIMENTOLOGY OF THE TIERRA DE LARA GROUP. 36

3.1. Introduction 36
3.2. Senora de Brezales Formation 37
3.2.1. Facies a: Sandstones 37
Interpretation 38
3.2.2. Facies b: Conglomerates 39
Interpretation 40
3.2.3. Facies c: Carbonates 41
Interpretation 43
3.2.4. Senora de Brezales Formation: Synthesis 43
3.3. Rupelo Formation 44
3.3.1. Facies 1: Red marls 44
Interpretation 45
3.3.2. Facies 2: Peloidal micrite 47
Interpretation 48
3.3.3. Facies 3: "Rubbly horizons" 49
3.3.3a. Marly; Prismatic; Reticulate 49
Interpretation 50
3.3.3b. "terra rossa" 51
Interpretation 51
3.3.3c. Rubbly horizons - summary 53
3.3.4. Facies 4: Micrite with open-space structures 53
3.3.4a. Angular-branching and angular sparry cavities 53
3.3.4b. Lined cavities 54
Interpretation 54
3.3.4c. Unlined tubular cavities 55
Interpretation 56
3.3.4d. Cavities with geopetal fills 56
Interpretation 57
3.3.4e. Cavities of uncertain origin 59
3.3.5. Facies 5: Green marls 59
Interpretation 59
3.3.6. Facies 6: Limestone conglomerates and arenites 60
Interpretation 60
3.3.7. Facies 7: Gastropod-ostracod-charophyte-biomicrite and interbedded grey charophyte marls 61
Significance of fauna 64
Interpretation 65
3.3.8. Facies 8: Laminated biomicrites 67
3.3.8a. Diffuse lamination 68
Interpretation 68
3.3.8b. Clastic lamination 68
Interpretation 68
3.3.9. Facies 9: Intraformational conglomerates 69
Interpretation 69
3.3.10. Facies 10: Black pebble conglomerate 70
Interpretation 71
3.3.11. Facies 11: Grainstones and spar-matrix breccias 74
Interpretation 74
3.3.12. Facies 12: Yellow limestones and yellow vuggy limestones 75
Interpretation 76
3.3.13. Facies 13: Cherts 76
Interpretation 78
3.3.14. Charophyte dedolomite 81
Interpretation 81

3.4. Diagenesis. 82
3.4.1. Introduction 82
3.4.2. Micrites 82
3.4.3. Spar 83
3.5. Facies associations 85
3.6. Tierra de Lara Group: Environmental synthesis 85
3.7. Facies sequences 87

CHAPTER 4: THE SEDIMENTOLOGY AND PETROGRAPHY OF THE (?UPPER BERRIASIAN - BARREMIAN) PEDROSO GROUP. 89

4.1. Introduction 89
4.2. Hortigüela Formation 89
4.2.1. Facies 1: Marls / Silts 89
Interpretation 90
4.2.2. Facies 2: Muds / Clays 90
4.2.3. Facies 3: Sandstones 90
4.2.4. Facies 4: Nodular limestones 91
4.2.5. Facies 5: Oncoidal limestones 91
Morphology of oncoids 92
Interpretation and discussion 94
4.2.6. Environmental synthesis for the Hortigüela Formation 96
4.3. Piedrahita de Muno Formation 97
4.3.1. Thick sandstone units 98
Interpretation 99
4.3.2. Thinner sand units 100
Interpretation 101
4.3.3. Mudstones & Siltstones 102
Interpretation 102
4.3.4. Nodular carbonates 102
Interpretation 103
4.3.5. Limestones 103
a) biomicrites 103
Interpretation 104
b) marly micrites 104
Interpretation 104
4.3.6. Facies 6: Green silts / marls 105
Interpretation 105
4.3.7. Environmental synthesis for the Piedrahita de Muno Formation 105

CHAPTER 5: THE SEDIMENTOLOGY OF THE (BARREMIAN - APTIAN) SALAS GROUP. 108

5.1. Introduction 108
5.2. Facies 108
5.2.1. Conglomerates 108
Interpretation 109
5.2.2. Sandstones 109
Interpretation 110
5.2.3. Muds / Silts 110
Interpretation 110
5.2.4. Calcarenites / calcrudites 111
Interpretation 111
5.3. Environmental synthesis 111

CHAPTER 6: THE SEDIMENTOLOGY OF THE "UTRILLAS" FORMATION (UPPER ALBIAN - LOWER CENOMANIAN. 113

6.0. Introduction 113
6.1. Facies 1: Mottled horizons 114
Interpretation 115
6.2. Facies 2: Conglomerates 115
Interpretation 115
6.3. Facies 3: Pebbly sands and conglomarates 116
Interpretation 117
6.4. Vertical and areal distribution of facies 117
6.5. Environmental synthesis 117

CHAPTER 7: GEOCHEMISTRY. 121

7.1. Total Organic Carbon 121
7.2. Stable isotopes 121
7.2.1. Introduction 121
7.2.2 Aims 122
1) Rupelo Formation 122
2) Pedroso Group 123
7.2.3. Techniques 123
7.2.4. Discussion of previous works 124
7.2.5. Interpretation of results presented here 125
1. Rupelo Formation 125
2. Pedroso Group 129

CHAPTER 8: SEDIMENTATION AND TECTONICS. 130

8.1. Introduction 130
8.2. Plate tectonic setting 133
8.3. Hercynian 134
8.4. Triassic 137
8.5. Rhaetic / "Infralias" 139
8.6. Middle and Upper Lias 141
8.7. Aalenian and Bajocian 142
8.8. Bathonian and Callovian 144
8.9. Talveila Formation (U Oxfordian - Lr Kimmeridgian) 147
8.10. Geometry of Late Jurassic Unconformities 150
8.11. Senora de Brezales Formation 154
8.11.1. Basal Contact 154
8.11.1a. Karst surfaces 154
8.11.1b. Conglomerates 154
8.11.2. Tectonic controls on facies in the Senora de Brezales Formation 156
8.12. Rupelo Formation 158
8.12.1. Control of tectonics on facies and thickness 158
8.12.1a. Las Vinas Member 159
8.12.1b. Ladera Member 160
8.12.1c. Mambrillas de Lara Member 160
8.12.1d. Rio Cabrera Member 161
8.12.2. The Rupelo Formation: comparison with other Late Jurassic - Early Cretaceous and "Purbeck" sequences 161
8.13. Pedroso Group 164
8.13.1. Vertical and lateral distribution of Pedroso Group facies 165
8.13.2. Comparison with other "Wealden" sequences 166

8.14. Salas Group 167
8.15. "Utrillas" Formation
(U Albian - Lr Cenomanian) sands 169
8.16. Upper Cretaceous 172
8.17. Oligocene - Miocene 177
8.17.1. Comparison with
the Oligo-Miocene of SW Britain 180
8.18. Plio-Quaternary (Ranas Formation) 181
8.19. Quaternary and Recent 182
8.20. Structure 183

CHAPTER 9: PARTICULAR PROBLEMS RAISED DURING THIS STUDY. 191

9.1. Lacustrine carbonates and pedogenesis 191
9.2 "Calcretes", "caliches", and nomenclature of
carbonate palaeosols. 194
9.3. The influence of tectonics on lacustrine sequences 196
9.4. Structure of the Cameros: the Deformation of a Basin 199

CHAPTER 10: CONCLUSIONS. 202

10.1. Major Results 202
10.1.1. Stratigraphy and Sedimentology 202
a) Talveila Formation 202
b) Tierra de Lara Group 202
i) Senora de Brezales Formation 202
ii) Rupelo Formation 202
c) Pedroso Group 203
i) Hortigüela Formation 203
ii) Piedrahita de Muno Formation 203
d) Salas Group 204
e) Utrillas Formation 204
10.1.2. Basin Evolution; Sedimentation and Tectonics 205
10.1.3. Structure; Regional Significance 208
10.1.4. Implications for Basin Analysis 209
10.2. Suggestions for further work. 210

REFERENCES. 212

Fig 1.0. Looking SW from near Arroyo de Salas, April 1983.

CHAPTER 1. INTRODUCTION.

1.1. Location (Geographical).

The Cameros Basin is located in Northern Spain in the provinces of Burgos, Soria and La Rioja (fig 1.1). The Basin measures approximately 60 km from N to S and 110 km from E to W.

The area studied in this work forms the W portion of the basin and lies mainly within the province of Burgos. The study area (fig 1.2) extends from Hontoria de la Cantera to Barbadillo del Pez in the N and from Tejada to Espejon (Soria province) in the S (sheets 277, 315). Salas de los Infantes, the most important centre of population, lies 56km to the SE of the city of Burgos and approximately 220 km by road NNE of Madrid.

1.2. Setting (Geological).

The Cameros Basin is one of the largest and most important Mesozoic basins in Spain (fig 1.3). Between Triassic and early Tertiary times, a wide variety of sediments was deposited on a Hercynian basement. The basin is bounded to the N by the Palaeozoic massif of the Sierra de la Demanda, while to the W and S lies the Tertiary Duero Basin which contains up to 4km of continental clastic deposits. Another important Tertiary basin, that of the Ebro, lies to the NE.

The Cameros Basin forms the NW extension of the Iberian Chains, a range of mountains running approximately NW-SE and extending from the study area to Maestrazgo in the province of Valencia (fig 1.4). The Iberian Chains are made up primarily of Mesozoic rocks and the range can be broadly described as an elongate trough which was subsident from Triassic time until the onset of uplift in the Oligocene. The trough possessed intermittent connections to the sea both to the N and at its SE end.

To the NW lie the Cantabrian mountains with their important Palaeozoic massifs and the overlying sediments of the Asturias and Vasco Cantabrian Basins. The W extension of the Vasco Cantabrian Basin, the Santander Basin (Pujalte, 1977; Garcia Mondejar, Hines, Pujalte & Reading, 1985), bears many comparisons with the Cameros Basin both in tectonic style and in basin evolution. In general, however, the Cameros Basin shows much more continental influence throughout the Mesozoic and early Tertiary than either the Vasco Cantabrian basin or the E part of the Iberian Chains. This influence reflects proximity to the basement massif (the "Meseta") that forms the core of the Iberian Peninsula, and greater distance from both the developing Bay of Biscay rift to the N and from Tethys to the SE.

The Cameros Basin is especially notable for the very thick continental sequence laid down during Late Jurassic and Early Cretaceous times (Salomon, 1982). A total of approximately 5km of sediment were deposited in the E of the basin (Salomon, 1982b).

The structure of the basin is dominated by two principal tectonic trends. The NE-SW and NW-SE directions are clearly evident on a tectonic map of the Iberian peninsula (fig 1.5), and it has been suggested by several authors, eg Salomon (1982a,b; 1983) and Floquet, Alonso & Melendez (1982), that synsedimentary faults parallel to one or both of these structural trends exerted important control on Mesozoic sedimentation. Apparently, the same directions were also important during Tertiary times (Pol Mendez, pers. comm., 1985) and even today the present drainage pattern reflects the structural grain (see section 8.19).

1.3. Aims of the project.

The project was set up to investigate the interrelationship between tectonics and sedimentation in the Mesozoic of the Cameros Basin. A Spanish research team under Dra A Alonso was to investigate the E part of the basin, leaving the W portion for our study.

The continental sequence of Late Jurassic to Early Cretaceous age was to be given special emphasis. The deposition of great thicknesses of sediment during the comparatively brief interval between the Kimmeridgian and the Lower Cenomanian was noted to be in great contrast to the absence of these sediments of this age from neighbouring areas. This already seemed to point to a very significant tectonic influence on sedimentation.

The first objective was to investigate the SW basin margin. Here, sedimentation was apparently less continuous and it was thought that tectonic influences would be clearly manifested in this region. Initial studies concentrated on the lower part of the continental sequence (ie upon the sediments of Salomon's (1982) Group I - (Kimmeridgian to Lower Berriasian; see figs 2.36 & 2.37). Later, the study area was expanded to include the NW part of the basin where deformation is weaker and where there is excellent exposure along strike. This also enabled the study of the middle part of the continental sequence (Groups IIC, IV and V: Upper Berriasian to Lower Cenomanian), which is largely absent in the SW sector.

As Group I is represented in both areas it was thus further decided to perform a comparative study for this unit. This comparison developed into a basin-wide sedimentological study with the following goals;

i) a detailed description and facies analysis of the sediments.

ii) an investigation of the facies distribution and the factors controlling it.

Groups IIC and IV display patchy and less continuous outcrop. Although facilitating the study of a number of sections in purely sedimentological terms, the nature of the exposure made the recognition of individual horizons extremely difficult. In view of the great thicknesses involved, correlation was thus almost impossible with the available lithostratigraphic criteria alone. It was, therefore, decided to devote rather less detailed study to Groups

IIC, and IV. The upper part of the continental sequence (Group V: Upper Albian-Lower Cenomanian) was studied in minor detail only.

In order to achieve an improved insight into the tectonic control on sedimentation during the Mesozoic, it was first nec essary to unravel something of the structural history of the basin from Hercynian times right up to the recent. By combining the results gained from the present study with material gleaned from the literature, it was hoped to compile a brief history of the basin's overall evolution that would bring together work done in stratigraphy, sedimentology, and tectonics. The aim was to achieve a wider viewpoint unblinkered by the confines of a single stratigraphic system in an effort to attack some of the fundamental problems concerning the formation, development, and subsequent inversion of the basin.

1.4. Methods.

1.4.1. Field.

The philosophy behind work in the field was to gather data which would facilitate detailed sedimentological study and aid the understanding of regional tectonic trends and palaeogeography.

Over twenty logged sedimentological sections were measured at selected localities (fig 1.6) through the Group I sediments in an attempt to achieve basin-wide coverage that would allow the recognition of regional variations. A number of sections were located in close proximity to each other specifically to pick out more local changes (eg across faults). A comparison of these sections provided the basis for lithostratigraphic correlation and also permitted the plotting of areal facies distribution patterns on which the interpretations of palaeogeography were based.

A smaller number of sections were drawn through Group II sediments and these allowed the recognition of the most important facies. One or two short and incomplete sections were also drawn

through parts of Group IV. However, difficulties in correlation made palaeogeographic interpretations correspondingly more difficult.

Comprehensive and detailed mapping of the whole study area was impractical on account of its great size, so instead two main areas were chosen for 1:10 000 work (fig. 1.7; see also enclosures - final maps presented at 1: 20 000 scale). One area selected forms part of an important anticline near Mambrillas de Lara in the relatively less-deformed NW zone, while the second area lies around Mamolar in the tectonically more-deformed SW sector. Over 50km^2 were mapped in each case. Less formal mapping was undertaken in zones of particular interest (eg along the San Leonardo Fault and around Moncalvillo) and helped to answer important tectonic problems. This form of mapping also proved invaluable in initially locating many of the most important logged sections.

Although detailed structural analysis is beyond the scope of this work, a few structural profiles were drawn in areas of more intense deformation and gave a useful insight into structural style.

Over 700 samples were taken during the course of the fieldwork, the sampling generally being carried out at the time of logging. An attempt was made to take samples from most localities and from all of the main facies, although it is recognised that there was an inevitable bias towards the more competent horizons. Fissile units such as shales, silts and marls were more difficult to sample.

1.4.2. Laboratory.

Standard and double-sized petrographic thin sections were prepared for over 300 samples. The double-sized slides were found to be especially suitable for the fabric-oriented approach employed. Most thin sections of carbonates were stained following the technique of Dickson (1965). Although the mineralogies were strongly dominated by non-ferroan calcite, differing intensities of staining resulted from differing grain-size, clay content, and organic content. This had the

effect of emphasizing the fabric and allowed the easy recognition of features not readily apparent in unstained thin sections. Micrites were found to be especially suitable for this technique and many examples which had initially appeared practically homogeneous revealed complex fabrics after staining.

Total Organic Carbon analyses were carried out for a very few samples. The values obtained were relatively low (see section 7.1) and further analyses were not performed.

Petrographic studies were assisted by the use of cathodoluminescence. The machine used was a Technosyn 8200 as modified by S.C Wright and S.M. Baker. Standard techniques were employed (see eg Grover & Read, 1985).

Polished slabs were also made from several samples. These were found to be especially useful for showing larger-scale fabrics of oncoids, laminar horizons and brecciated carbonates.

Stable isotope analyses were done at the British Geological Survey, London. Thirty-five selected ^{13}C and ^{18}O isotopic analyses were performed on samples of Group I and IIC carbonate facies; for details of this work see section 7.2.

1.5. Previous work in the area.

The Cameros Basin has been studied by a number of workers, with most of them having concentrated on one stratigraphic unit or group of units. Early research in N Spain is summarised by Ciry (1939), while a more modern perspective is given by Rat (1982).

The Triassic of the Cameros seems to have received relatively little study. For a general review of N Spanish Triassic sedimentation, the reader is referred to the work of Garrido-Megias & Villena Morales (1977) and Virgili, Sopena, Ramos & Hernando (1977).

Work on the Jurassic is summarised in the Virgili (1972) volume, with a particularly useful contribution by Ramirez del Pozo (1971). There have been a number of excellent German papers on the marine

Jurassic of the area, for example: Dahm (1966); Mensink (1967), and, more recently: Visser (1984); Errenst, Mensink, Mertmann, Schudack & Visser (1984); Mensink & Mertmann (1984). Valladares (1976, 1980) made some of the first sedimentological studies of the marine Jurassic in this area, this work being extended recently by Ureta & Kalin (1985).

The marine Upper Jurassic of the study area has received little study although Diaz-Molina, Goy & Yébenes presented a short communication on this subject (1983), and Errenst (1984) dealt with the marine Malm of the Iberian Chains in general.

The Upper Jurassic - Lower Cretaceous continental sequence was first described by Sanchez-Lozano & Palacios (1884). A later publication by Sanchez-Lozano (1894) introduced the terms "Purbeckiense" and "Wealdense" for the first time, these names taken from comparison with the sequence in S England. This work was superseded by Saenz-Garcia (1932) in which a revised subdivision of the continental sequence was put forward.

The contributions by Beuther (1966) and Tischer (1966), working in collaboration in E and W parts of the basin respectively, led to the erection of a new and independent stratigraphy. Their subdivisions were used in the compilation of a 1:200 000 geological map of the Cameros Basin, this map forming the basis of the subsequent IGME maps of Burgos, Aranda de Duero, Soria, and Logrono (sheets 20, 30, 31, 21) which have been published since 1970. Work by Kneuper-Haak (1966) allowed the correlation of Beuther and Tischer's lithostratigraphy with established biostratigraphy.

Bulard (1972) made a study of the late Jurassic "regression" in the region, while Brenner (1976) and Brenner & Wiedmann (1975) revised Kneuper-Haak's chronostratigraphy and made minor modifications to Beuther and Tischer's dating estimates.

Salomon (1980) proposed a revised stratigraphy after that of Saenz-Garcia (1932), this work being amplified and later modified in detail (1982b,c). Salomon recommended the abandonment of the terms

"Purbeckiense" and "Wealdense" as they had fallen into confusing misuse as descriptions both of facies and stratigraphic age. Salomon's thesis (1982) dealt with both the Cameros Basin and the Vasco Cantabrian Basin. Improved correlation based upon the new stratigraphic scheme now permitted the first discussion of basic sedimentology, and the relationships between sedimentation and tectonics were also discussed for the first time.

Mensink and Schudack (1982) were the first to discuss microfacies in their paper on pedogenesis at the marine - continental transition, this work being based upon Schudack's (1979) Dipl. -Arb. Thesis. The theme of pedogenesis was developed in the recent papers by Salomon (1984) and Platt (1985). Salomon & Drillien (1985) and Platt (op. cit.) described fresh-water carbonate sediments from the E and W of the basin respectively, while Schudack (1984) included a discussion of facies distribution at the base of the continental sequence.

The Upper Cretaceous of the area has been the subject of many publications by Floquet, Alonso and others, eg Floquet (1978, 1979, 1982, 1984), Wiedmann (1979), Alonso & Floquet (1982), and Floquet, Alonso & Melendez A (1982). Pioneering sedimentological studies had been the work of Valladares (1976a,b; also 1985).

Work on the Tertiary of the area in recent years has included that of Pol Mendez (1985). Floquet, Salomon & Vadot (1985) dealt specifically with the sedimentology of the (?Maastrichtian-) Palaeogene basin of Santo Domingo de Silos, this work following on from the original brief description by Floquet (1979).

Tectonic studies of the area have been few. The work of Colchen (1966, 1974) discussed the basement and cover tectonics of the Sierra de la Demanda region, while Morillo-Velarde & Melendez-Hevia (1972) discussed the history of the San Leonardo Fault and postulated on its synsedimentary rôle during the Mesozoic. A more recent paper, by Salomon (1983) deals with extensional tectonics and proposes a rift-basin model for the formation and development of the Cameros.

Recent work on the tectonics of the Iberian Chains has included that of Julivert (1978) and Viallard (1979). The rôle of tectonics in the evolution of Mesozoic and Tertiary basins in N Spain has been discussed by Canerot (in press). Amiot, Floquet et al (1982) and Rat (1983) reviewed the evolution of the N Iberian margin in general terms. Engeser, Reitner, Schwentke & Wiedmann (1984) explored the tectonic evolution of the Vasco Cantabrian basin in particular, emphasizing the rôle of movement along major transcurrent faults during the basin's history. Recently, Hossack (1985, 1986) has studied the tectonic evolution of Pyrenean Cantabria.

Other relevant studies of the Vasco Cantabrian Basin are mostly concerned with stratigraphy and sedimentology. These include Rat (1963), Pujalte (1977, 1979, 1981, 1982), and Garcia-Mondejar, Hines, Pujalte & Reading (1985). Sbeta (1976, 1985) discussed the SW portion around Aguilar del Campoo. This is the zone that is geographically closest to, and which displays the most striking geological similarities with, the Cameros. Wiedmann (1979) suggested that the two areas were intermittently connected during the Mesozoic.

Fig 2.0. Lower Cambrian Barbadillo del Pez quartzites. Rio Pedroso gorge, 2km N of Barbadillo del Pez. September 1985.

CHAPTER 2. STRATIGRAPHY.

A number of workers have outlined the stratigraphy of particular systems. This material is synthesized here in English for the first time and drawn together with additional information which has emerged during the course of this study.

Generalised stratigraphy of the W Cameros Basin is summarised in fig 2.1. Particular emphasis is given to the detailed stratigraphy of the studied formations of Late Jurassic and Early Cretaceous age.

2.1. Palaeozoic.

The basement of the Cameros Basin is of Palaeozoic age. Outcrops of Palaeozoic in the study area are restricted to the S flanks of the Sierra de la Demanda and to an inlier which forms the core of the Moncalvillo - Palacios anticline (fig 1.6). The Duero Basin to the S is also underlain by Palaeozoic rocks. Wells drilled through the thick Tertiary sediments in that area encountered only a thin veneer of Aptian - Albian and Upper Cretaceous deposits resting on the basement, the rest of the Mesozoic being absent (Rios, 1966).

The Palaeozoic of the Sierra de la Demanda consists of Cambrian, Ordovician (Tremadoc) and Carboniferous rocks. For a review of the geology of this area, the reader is referred to Colchen (1974). Fig 2.2 summarises the stratigraphy.

At the N edge of the study area, the Mesozoic lies unconformably upon the Lower Cambrian Barbadillo del Pez quartzites (fig 2.0). Phyllites outcrop further to the N near the road junction between the C-113 and the local road to Pineda de la Sierra and Riocavado de la Sierra (figs 2.3 & 2.4). An agnostid trilobite was recently found at this locality by D C Edwards.

In the Palacios - Moncalvillo anticline (see fig 1.6), further to the SE, the basement is formed of a very distinctive purple meta-arenite. Boulders of this lithology are excellent indicators of

Palaeozoic source areas in Tertiary conglomerates (see sections 2.20 - 2.22; 8.17 - 8.18) and thus document the uplift of the Palaeozoic. Quartzite pebbles in many U Jurassic and Lr Cretaceous sediments are more difficult to trace as some may have been reworked from Ordovician, Carboniferous or Triassic quartzites and conglomerates (which now outcrop chiefly in the NW of the Sierra de la Demanda). Others may be derived from more distant areas (eg Sistema Central).

2.2. Triassic.

The Triassic of N Spain is essentially Germanic (fig 2.5). The Buntsandstein is approximately 80m thick and is developed in its typical clastic red-bed facies. Muschelkalk may occasionally be recognised but the study area was situated approximately at the limit of the Middle Triassic transgression (fig 2.6, see also section 8.4) and these deposits are very thin (max ½m). The 40-50m thick Keuper outcrops only rarely. Further to the SE in the Iberian Ranges, and in the Vasco Cantabrian Basin to the NW, the Keuper is notable for the presence of evaporites. The sparse literature available on the Triassic of the Cameros makes little mention of evaporites, although Gil Serrano & Zubieta (1978) describe the lithologies of the Keuper as "40-50m of red marls with gypsiferous horizons". The significance of Keuper evaporites in determining subsequent sedimentation patterns and structural style is discussed by Hines (1986). Although the influence of halokinetic movements was probably less important in the Cameros, the Keuper may have provided an additional décollement horizon for Tertiary thrusting. This theme is explored further in section 8.20.

2.3. Triassic-Jurassic "Carniolas".

A transition can be observed from the Triassic up into "Carniolas" facies of Rhaetian - Lower Liassic age. Visser (1984) studied the Triassic - Jurassic boundary of the region. The facies developed are in contrast with those of many other regions around the

Fig 2.3-7.0. Marine Jurassic. Classic locality;
Castrovido, ("empty castle").
Storm clouds, April, 1983.

N Mediterranean where organic-rich mudstones are found at this horizon (eg Appenines, Dolomites); in the study area, vuggy dolomites (fig 2.7) and evaporites predominate.

Some of the best exposures of the Carniolas are to be found at Mazuecos in the N of the study area (fig 1.6). The San Leonardo Fault (section 8.5) utilised the Carniolas as a detachment plane. Fig 2.8 depicts a series of laminated carbonates with a teepee structure (Assereto & Kendall, 1977), from Vadillo, approximately 10km SE of San Leonardo.

2.4. Lias.

Valladares (1976) termed this the "tramo margoso". Her palaeogeographic scheme for this time is shown in fig 2.9.

The Sinemurian is termed the "limestone Lias". Approximately 50m thick, it consists of light-coloured 10cm-bedded limestones (fig 2.10) which become more thickly-bedded moving upwards in the sequence. Thin marly horizons are intercalated (Gil Serrano & Zubieta, 1978).

The Upper Liassic (Pliensbachian - Toarcian), or "marly Lias", consists of a series of limestone-marl rhythms (fig 2.11) which are similar to those described by Sbeta (1976) from the Aguilar region.

The Lias is very fossiliferous, yielding a rich open marine fauna of ammonites, belemnites and pectinid bivalves. We may infer that the transgression initiated at the end of Triassic time continued during the earliest Jurassic. The Lias of the study area is thus interpreted as a deposit of quiet epicontinental seas.

2.5. Aalenian and Bajocian.

The Aalenian and Bajocian (corresponding to the "tramo calcareo" of Valladares, 1976) are developed in shallower water facies. A gradual upward change is discernible from marl-dominated rhythms of Toarcian age to bedded limestones with only intercalated marls in the Aalenian (fig 2.12). This shallowing trend continued during Dogger

time and eventually a series of sponge bioherms was developed. At this horizon, an important discontinuity preserves significant relief below the overlying Middle Bajocian beds (figs 2.13 & 2.14). The tectonic significance of this discontinuity is discussed in section 8.7.

2.6. Bathonian and Callovian.

The Bathonian and Callovian are developed in somewhat more sandy carbonate facies. They form a distinctive lithological unit easily recognisable in the field (figs 2.15 - 2.17) by virtue of a rich honey-brown colour - they are the "orange beds" of Callan (unpublished data) and Faulkner (1985). The facies indicate a shallow subtidal environment close to a source of terrigenous input. The presence of quartz might reflect either simple regression or the onset of tectonic instability.

The Lower Callovian contains important discontinuities (Mensink & Mertmann, 1984) perhaps reflecting the marginal nature of the deposits. In the NW of the study area, the Macrocephalus (figs 2.18 & 2.19) ammonite zone is commonly the youngest present below the continental sequence. Fig 2.20 (modified after Mensink & Schudack, 1982) shows the age of the youngest marine Jurassic strata preserved in the NW Iberian Chains, while fig 2.21 shows the palaeogeography during Lower Callovian time as envisaged by Valladares (1976).

In the N of the study area, at Barbadillo del Pez, Villavelayo and at Huerta Arriba, a sequence of dark, bituminous 30cm-bedded limestones is exposed (fig 2.22). These sediments occur above the marine Jurassic sequence and beneath the limestones of the Tierra de Lara Group. Schudack (1979) reported the presence of Quenstedtoceras lamberti, indicative of a Middle or Upper Callovian age. The deposits may record a restricted (and perhaps occasionally anoxic) phase of marine conditions.

2.7. Talveila Formation (? Upper Oxfordian - Lower Kimmeridgian).

In the SW of the study area, a sequence of yellow arenites and conglomerates overlies the marine Jurassic. This was termed the "Jurassique détritique" by Salomon (1982), and is defined here as the Talveila Formation.

Other authors, eg Mensink (1965) and Valladares (1976), made no reference to these deposits, although some of the oolites Valladares described from Dona Santos might now be included in the Talveila Formation. Errenst's (1984) facies distribution maps showed marine calcarenites at Talveila (fig 2.23), but likewise made no reference to the sections further to the NW.

The facies of the Talveila Formation include coarse cross-bedded bioclastic calclithites (figs 2.24 - 2.26), quartz pebbly conglomerates (fig 2.27), and fossiliferous biomicrites (fig 2.28). The biomicrites are generally found near the top of the sequence and form three or four prominent horizons (fig 2.29). The in situ fauna is confined to the carbonates and comprises an apparently restricted assemblage of gastropods and bivalves. Diaz-Molina, Goy & Yébenes (1983) reported charophytes from Talveila, at the extreme SE of the study area.

The style of the cross-bedding, the rapid vertical variation in facies, and the nature of the faunas led Diaz-Molina et al (op. cit.) to propose a barrier island - lagoon model for the sediments. This remains a somewhat controversial interpretation, but, to date, no alternative suggestions have been advanced. A more thorough analysis of palaeocurrents would be useful in the erection of a rigorous sedimentological model.

Regional mapping proved the presence of the formation along an elongate belt to the N of the San Leonardo Fault from Talveila to Pinilla de los Barruecos. Lithologically-similar deposits were also found to the S of the SLF at Ladera and Espejon. The forthcoming IGME 1:50 000 geological sheet 315, "Santo Domingo de Silos", (IGME, in

prep.) also reports Talveila Formation sediments in an anticline near Las Pinarejas, but in fact the sediments at this locality are probably Utrillas (U Albian - Lr Cenomanian) sands and conglomerates which have been tectonically repeated by imbricate thrusts associated with the SLF complex (see fig 8.105; enclosure).

The Talveila Formation was studied in two principal localities - at Hortezuelos and at Pinilla de los Barruecos (fig 1.6). The basal part of the Talveila Formation is only poorly exposed at Hortezuelos. However, nodular, and often mottled, carbonates were observed at the contact with the underlying marine Bathonian carbonates (fig 8.18), perhaps indicating an initial period of subaerial exposure. There are thin (2-3 cm) red horizons throughout the lower part of the formation. These have facilitated the recognition of minor synsedimentary extensional faults (figs 8.19 - 8.22). At Pinilla de los Barruecos, the Talveila Formation sands are apparently in erosional contact with the marine Jurassic, although no angular unconformity was detected.

Where the uppermost sediments of the Talveila Formation preserved below the continental sequence are limestones, karst features can be observed. At Talveila, Hortezuelos and Pinilla de los Barruecos, red sands from the overlying Senora de Brezales Formation have been introduced along fissures and bedding planes in the Talveila Formation limestones (figs 2.30 & 2.31; see also figs 8.30 - 8.39). Where the uppermost sediments preserved are sandstones, then the contact between the Talveila Formation and the overlying continental sequence may either be gradational (as at Ladera) or it may be marked by a ferruginous crust up to 5cm in thickness (as at Aldea del Pinar, fig 2.32). Alonso (pers. comm., 1985) mentioned the presence of a ferruginous crust as a common feature of the Upper Oxfordian sequence of the NE Cameros. The karst features and the ferruginous crust evidently document emergence after the deposition of the Talveila Formation but before the continental sequence was laid down.

Dating of the Talveila Formation is difficult; being developed in apparently marginal facies, fossils, notably ammonites, are absent (Goy, pers.comm., 1985). Diaz-Molina, Goy & Yébenes (1983) attributed the formation to the Oxfordian and ?basal Kimmeridgian, although no biostratigraphic evidence was presented to support this. IGME (in prep.) assign the sediments to the Oxfordian.

Although the karstic surface and ferruginous crust at the top of the Talveila Formation indicate subaerial exposure in the study area after Early Kimmeridgian times, further E in the Iberian Chains, marine conditions persisted, at least intermittently, until the Tithonian (Errenst, 1984).

2.8. - 2.11. Continental Sequence (Kimmeridgian - Lower Cenomanian).

The first subdivision of the continental sequence was by Sanchez Lozano (1894) who recognised five groups (A,B,C,D,E) of which (A,B) were termed "Purbeckiense" and (C,D,E) were "Wealdense". Above this he recognised an "Urgo-Aptian" group and a "pre-Cenomanian" group (fig 2.33).

Saenz Garcia's (1932) stratigraphy proposed five new groups (I - V) of which I was equivalent to (A and B), II and III corresponded to (C,D,E), while IV and V were equivalent to Sanchez-Lozano's (1894) Urgo-Aptian and pre-Cenomanian groups respectively (fig 2.34).

Beuther and Tischer (both 1966) distinguished only between "Wealden" and "Utrillas" - the latter being equivalent to Saenz Garcia's (1932) Group V, while the "Wealden" comprised the remainder of the continental sequence below this (ie Groups I-IV). The Wealden itself was then subdivided lithostratigraphically into five new groups independent of any previously defined units. These groups were named Tera, Oncala, Urbion, Enciso, and Olivan (fig 2.35). Beuther and Tischer's work included a 1:200 000 geological map of the Cameros Basin which formed the basis of the IGME (1970) map.

Kneuper-Haak's biostratigraphic correlation (1966) was based upon ostracods and permitted an age to be assigned to the continental sequence for the first time - exclusively Lower Cretaceous. Brenner (1976) produced new work using charophytes in addition to ostracods and slightly modified some of the dating.

Salomon (1982) proposed a new stratigraphy which followed the spirit of Saenz-Garcia's (1932) work and abandoned the terms "Purbeck" and "Wealden" on account of the confusion that had been created by their inconsistent usage both as stratigraphic and facies terms. The new stratigraphy he proposed is summarised in figs 2.36 & 2.37. Salomon (1983) later slightly revised the nomenclature (fig 2.38) but, in the work presented here, I refer to his original (1982) stratigraphy.

2.8. Tierra de Lara Group (Group I). (See also Chapter 3).

Sediments of both Group I and Group II were formerly included in the Tera Group by Beuther and Tischer (1966), but it seems more prudent to use new nomenclature rather than to resurrect part of their stratigraphic scheme when the remainder of it has now been abandoned.

Group I of Salomon (1982) corresponds to the Grupos Hortezuelos and Castrovido of Salomon (1982b). These were two stratigraphically equivalent groups defined in the S and the N parts of the study area respectively. Salomon was probably influenced in his decision to define two equivalent groups by his belief in the former existence of a mid-basinal high (where Group I was absent) in the area of the Palacios - Moncalvillo anticline. Such a high would have divided the study area into two separate sub-basins. In fact, Group I sediments are present at Moncalvillo (although their thickness is reduced), so clearly the areas to the N and S were interconnected. Thus I prefer to abandon the old terms, and I here define a new unit for the whole of the W Cameros, the Tierra de Lara Group, to comprise the old Hortezuelos and Castrovido Groups.

Comparison of a number of detailed sections (see appendix) has facilitated the definition of two formations within the Tierra de Lara Group (fig 2.39). The lower, Senora de Brezales Formation comprises a dominantly clastic sequence of red sands, conglomerates (often polygenetic) and pedogenic carbonates, while the upper, Rupelo Formation consists mainly of carbonates. The facies within these two formations are discussed in Chapter 3.

Fig 2.40 shows the lithostratigraphy of the Tierra de Lara Group and of the overlying Hortigüela and Piedrahita de Muno Formations.

The base of the Senora de Brezales Formation is defined at the karst surface discussed in section 2.7. Where the limestones of the upper part of the Talveila Formation are absent, the base of the Senora de Brezales Fm may be taken above the ferruginous crust.

The Rupelo Formation is divided into four members (fig 2.40). The lowest of these, the Ladera Member, is developed in marginal lacustrine carbonates and is apparently laterally equivalent in part to the Las Vinas Member, which is made up of red marls interpreted as distal floodplain/playa-lake deposits. The base of these Members is often marked by a prominent laminar crust (see section 3.2.3). The top of the Ladera Member is occasionally marked by a brecciated surface of "calcretised" blocks (fig 2.41).

The base of the overlying Mambrillas de Lara Member is often marked by a limestone conglomerate or sandstone (the Torrelara Bed). The Mambrillas de Lara Member comprises biomicrites and charophyte marls, suggesting an open lacustrine environment. A horizon containing vertebrate remains and traces (the "Dinosaur Bed") is often found at the top of this Member.

The overlying Rio Cabrera Member contains yellow vuggy limestones, interbedded replaced evaporites, and, occasionally, dedolomitised charophyte-ostracod marls and limestones. The deposits are interpreted as fresh-water to hypersaline lacustrine carbonates. The top of this Member is marked by a prominent desiccation-cracked

surface (figs 2.42 & 2.43) which also defines the upper limit of the Rupelo Formation and of the Tierra de Lara Group itself.

Thus correlation is based on lithostratigraphy alone. The Tierra de Lara Group succession is developed exclusively in continental facies so that biostratigraphy is imprecise (as well as being impractical for application in the field).

The formations and members defined are probably diachronous to some extent. Certainly, the sections show evidence of very significant lateral facies changes. Open lacustrine facies are often not developed in localities towards the edges of the study area, particularly in the E. In such cases it is reasonable to assume that the thin carbonates present (developed in Ladera Member type marginal lacustrine facies) may actually date from intervals of higher lake stand when more permanent lakes were established and when open lacustrine facies were being deposited in wide areas within the study area. Thus marginal lacustrine brecciated and peloidal facies developed at the periphery of the study area may actually be equivalent to the open lacustrine biomicrite facies of more central areas.

Despite this problem of diachroneity of facies, some horizons may still be useful time markers. A discussion of the marker horizons is given below.

For example, in view of the wide distribution of Senora de Brezales Fm conglomerates, it could be argued that their appearance was the result of a period of tectonic activity. (The possible timing of such an event is discussed below and in section 8.11).

The change from clastic to carbonate deposition occurring at the boundary of the SdB Fm with the Rupelo Fm (defined at a prominent laminated crust useful as a marker in the NW of the study area) may have been controlled by climatic or tectonic factors occurring at a particular time. If this were true then the base of the Rupelo Fm might be an approximate time plane; as although the first important

carbonates are laterally equivalent to marls (Ladera Member v. Las Vinas Member), both overlie the prominent laminated limestone.

The brecciated surface at the top of the Ladera Member and below the Mambrillas de Lara Member implies a time break between them, while the widespread distribution of the Torrelara Bed may imply a particular sedimentary (or tectonic) event which occurred before the deposition of the Mambrillas de Lara Member. The associated change of facies across this horizon from brecciated and peloidal marginal lacustrine carbonates to fossiliferous open lacustrine biomicrites is such a sharp one that it may have been caused by a major basin-wide change in conditions. Climatic factors or a tectonically-induced alteration of basin configuration might have been responsible for the change in conditions, and either of these may have occurred at a particular time.

The "Dinosaur Bed" also probably dates from a particular interval when conditions favoured colonisation by sauropods and the preservation of their remains and traces. Such vertebrates would have thrived in more permanent open lake conditions and the preservation of their remains would have been facilitated by less frequent exposure and less intense pedogenesis.

The cherts and yellow vuggy limestones are also useful as marker horizons. As these deposits appear to represent the establishment of unusual, possibly hypersaline, conditions throughout the study area, then they are likely to be good time markers and as such they are of great use in correlation.

In contrast, the desiccation-cracked surfaces at the top of the Rupelo Formation represent the final stages of carbonate sedimentation during successive shoreline retreats and are much more likely to have been diachronous.

Thus lithostratigraphic correlation within the Tierra de Lara Group is based upon the recognition of regionally-mappable markers. Some of these horizons date from particular sedimentary events and

thus represent time planes. Others are more likely to reflect lateral facies change and are thus inevitably diachronous surfaces.

Nevertheless, the utility and application of the lithostratigraphy presented here (fig 2.40) is underlined by the similar stratigraphic evolution of facies in the Aguilar del Campoo region, some 120km to the NW. Each of the main units of the Tierra de Lara Group can be recognised there. The presence in SW Cantabria of marker horizons such as cherts and oncoidal limestones in a stratigraphic sequence predictable from the W Cameros succession is particularly striking. Furthermore, the upper part of the Pisuerga Reservoir and Salinas de Pisuerga sections present a succession similar to the Hortigüela Formation of the W Cameros.

To date, the whole carbonate sequence of the Aguilar area has been included in an all-embracing "Aguilar Formation". Fig 2.40 correlates the Aguilar and W Cameros sequences and suggests a possible lithostratigraphy for the Aguilar region. Obviously, much refinement of the lithostratigraphy remains to be done there.

The age of the Tierra de Lara Group has been the subject of considerable debate. Salomon (1982) favours a "Kimmeridgian - Lower Berriasian" age. The youngest marine Jurassic strata is of different age in different parts of the Cameros Basin (see above; fig 2.20). Several authors have mistakenly inferred from this that the continental sequence was itself highly diachronous (see for example fig 2.44, also discussion in section 8.11).

The conglomerates of the Senora de Brezales Formation are sufficiently similar in clast composition and tectonic setting to suggest that they date from a correlatable tectonic pulse (see above). Comparable polygenetic conglomerates have been reported from much of central N Spain and Asturias. They may record a discreet tectonic event, posibly of Kimmeridgian age (see sections 8.10 & 8.11).

The carbonate facies of the Rupelo Formation were probably deposited during the same timespan throughout the basin. Salomon

(1982) favoured a Berriasian age for the carbonates - although he suggested that they did not continue beyond the basal Lower Berriasian. Brenner & Wiedmann (1974) reported the presence of:

Fabanella aff. ornata (STEGH)

Kliana aff. sp. 1 of OERTLI

Darwinula leguminella (FORB),

and also suggested a Berriasian age. Our own charophyte dating on samples from the Mambrillas de Lara section (courtesy of Dr A Strasser) revealed the presence of Globator incrassatus. According to P O Mojon (pers. comm., 1985), this indicates a Berriasian age.

The Aguilar Formation has been assigned a variety of dates. Sbeta (1976) proposed a "Late Jurassic - Berriasian" age. This is in clear agreement with our findings for the Rupelo Formation. In contrast, Pujalte (1979 and pers. comm., 1985) favours an exclusively Upper Jurassic age for the Aguilar Formation. However, in consideration of the closely comparable lithostratigraphy of the two areas, it is highly likely that the carbonates of the Aguilar Formation (excluding the oncoid-rich Embalse de Pisuerga Member, see next section) are also Berriasian in age.

2.9. Pedroso Group (Group II). (See Chapter 4).

Of Salomon's (1982) Group II, only Group IIC occurs within the study area. This corresponds to the Hortigüela and Neila Formations of Salomon (1982b).

The Hortigüela Formation is made up of two lithologically distinct units (figs 2.39 & 2.40): a lower suite of marls, micaceous sandstones and conglomerates (here defined as the San Martin Member) and an upper unit which contains a number of limestones (here termed the Zurramujeres Member).

Correlation within the Hortigüela Formation is made difficult by the absence of readily recognisable marker horizons. Welle (unpublished data) used the first appearance of oncoids in the

sequence as a reference frame. Their last appearance can also be used.

The oncoids may be of local stratigraphic significance since the blue-green algae present in oncoids are dependent on particular conditions of salinity, temperature and water chemistry which may have only occurred during a particular interval. However, such conditions are also controlled by certain environmental parameters such as depth and proximity to shoreline, so that the presence of oncoids may, in reality, be facies- rather than time-dependent. Another problem is that in some localities oncoids are apparently absent altogether. Thus Welle's system has several limitations and may only be used with caution. Nevertheless, it remains the best (and the only) method available for correlation in the field.

Salomon (1982b) referred the thick sequence of Group IIC clastics to the Neila Formation. While the facies developed in the Neila area are broadly similar to those of the W Cameros, the Neila section lies in a faulted sub-basin, the "Villavelayo syncline" (see fig 8.92), so I prefer to define a new formation for the W Cameros. Thus these clastics are assigned here to the Piedrahita de Muno Formation. I have included this new formation with the Hortigüela Formation in the Pedroso Group, a newly-defined group for the W Cameros.

Correlation within the Piedrahita de Muno Formation is also difficult. The only regionally-recognisable marker horizons in a very thick sequence are thin (≃ 1m) fresh-water limestones. Thus a prominent dark limestone present in the Terrazas Road section is tentatively correlated with a lithologically-comparable limestone in the Piedrahita de Muno section and provides an invaluable reference point in a succession of otherwise apparently monotonous facies. The Piedrahita de Muno Formation may be locally subdivided using this marker horizon into the Terrazas Member beneath the dark limestone and the Vizcainos Member above it.

Correlation on the basis of clastic lithologies is hindered by the apparent uniformity of facies and by poor exposure which provides few continuous vertical sections. This is unfortunate as it tends to reduce the application of sedimentological observations. Accurate thickness determinations are not possible and this further hampers the detailed analysis of the tectonic controls on sedimentation and of facies "architecture" for the Piedrahita de Muno Formation.

The stratigraphic relationship between the Hortigüela and Piedrahita de Muno Formations is difficult to elucidate as they do not generally occur together. The Hortigüela Formation is developed in the NW of the study area and may be recognised in a few localities in the SW. The Piedrahita de Muno Formation is restricted to the area E of the Jaramillo-Covarrubias Fault (see Chapter 8 for a discussion of Pedroso Group facies distributions). The two formations may be partial lateral equivalents, in which case the lacustrine limestones of the Piedrahita de Muno Formation may correlate with the Hortigüela Formation oncoidal limestones, although the important differences in the nature of the clastic facies (see Chapter 4) tend to argue against this.

The age of the Hortigüela and Piedrahita de Muno Formations is far from well-constrained. Biostratigraphy is again dependent upon ostracods and charophytes. Salomon (1982) reported the presence of Clypeina parvula and Perimneste horrida. This was interpreted as indicating an Upper Berriasian - Valanginian age.

This interpretation is not generally in dispute and further precision is probably beyond the limits of accuracy obtainable from micropalaeontological correlation. Nevertheless, individual authors have cited Berriasian or Valanginian ages on the basis of comparison with other areas. A problem with this approach is that the sediments of lithologically-comparable successions of other areas are also, by definition, often developed in restricted facies with consequently

similar reduction in the resolution and reliability obtainable from biostratigraphic techniques.

This Chapter has already suggested a possib le correlation between the oncoidal carbonates in the upper part of the Aguilar Formation (proposed Embalse de Pisuerga Member) and oncoidal facies in the Hortigüela Formation (Zurramujeres Member). Pujalte (op.cit.) favours an Upper Jurassic age for (apparently the entirity of) the Aguilar Formation, while Sbeta (1976) proposed a Berriasian age. If we accept Sbeta's dating, then by inference, a Berriasian age for the Hortigüela Formation would be inferred. Salomon (1982b) suggested an Upper Berriasian age for his Group IIC which is in agreement with this. However, to the best of my knowledge, no age determinations have been carried out from the upper, oncoidal, part of the Aguilar Formation (Embalse de Pisuerga Member) which is markedly different from the lower parts, and may, in fact, be rather younger.

Indeed, many other authors favour a Valanginian age for the oncoidal carbonates. A correlation of the Valanginian of the central Iberian Chains with that of the Vasco Cantabrian Basin and of Maestrazgo was proposed by Wiedmann (1979), and Gil Serrano & Zubieta (1978) suggested a Lower Valanginian age for the oncoidal limestones in the study area on the basis of lithostratigraphic correlation with similar facies from Ramales, Santander. Hines (1986) mentions oncoidal carbonates from the coastal section at Cuchia in Santander province (although these may be of Barremian age).

Monty & Mas (1981) described morphologically-similar oncoids of Valanginian age from the SE Iberian Chains in Valencia. The occurrence of these oncoids was correlated with transgression in that area, and this recalls Valladares' (1976) interpretation of the Hortigüela Formation oncoidal limestones as marginal marine.

Although we may not be in agreement with this environmental model (section 4.2.6), a (?Lower) Valanginian age for the Hortigüela

Formation sediments would seem more likely in the light of comparison with the sequences of Ramales and of the SE Iberian Chains.

The Piedrahita de Muno Formation is also very difficult to date accurately. At the time of writing, charophyte-bearing fresh-water limestone samples from the Piedrahita de Muno section were still awaiting detailed biostratigraphic study. However, lithostratigraphy invites comparison with other NW European "Wealden" sequences (tentatively dated as Valanginian - Barremian), but further precision is not yet possible.

The facies of the Pedroso Group are somewhat different from those of the Tierra de Lara Group. Sandstones, especially in the Piedrahita de Muno Formation, tend to be micaceous, and, indeed, clastics dominate over carbonates in the Pedroso Group generally. This points to important changes between the Tierra de Lara and Pedroso Groups, perhaps both in climate and in the intensity of synsedimentary tectonics (see section 8.13).

In NW Germany, the change in facies between the carbonates of the Upper Münder Mergel and the Serpulit to the clastic facies of the Wealden occurred as early as between the Lower and Upper Berriasian. Comparison of this sequence with those of the W Cameros and Aguilar favours a Berriasian age for the oncoidal carbonates.

However, in S England the Lower and Middle Purbeck are thought to be of Berriasian age, while the Upper Purbeck and Hastings Beds (ie the lower part of the Wealden) are generally attributed to the Lower Valanginian. Thus, in S England the switch from carbonate to clastic deposition occurred approximately between the Berriasian and the Lower Valanginian. (However, even there, some authors, eg Hughes (1975), attribute part of the clastic Wealden to the Upper Berriasian).

The palaeogeographic setting of N Spain is likely to have been approximately analagous to that in S England owing to the proximity and similar tectonic evolution of the two areas at this time (see section 8.13.2). Thus a Valanginian age for the Piedrahita de Muno

Formation (and probably also for the Hortigüela Formation) is inferred.

2.10. Salas Group (Group IV). (See Chapter 5).

The "Salas Group" (Group IV of Salomon, 1982) is a very thick (< 1km) unit of continental clastic deposits. Sandstones, silts and conglomerates are the dominant lithologies although there are a few limestone beds, (mostly calcarenites containing reworked oncoids). The important "Urgonian" marine carbonates of Mediterranean regions and of the Vasco Cantabrian Basin to the N are absent here.

Salomon (op. cit.) suggested a "Barremian - Lower Albian" age for the Salas Group. This is consistent with the Hauterivian age proposed by Gil Serrano & Zubieta (1978). The absence of a datable fauna has so far precluded more precise age determination. Thus further discussion of the age range of the Salas Group must remain somewhat speculative.

However, Faulkner (1985) considered there to be strong evidence of synsedimentary tectonic activity at this time. The Salas Group conglomerates include reworked clasts of marine Jurassic limestones, Tierra de Lara Group limestones and Hortigüela Formation oncoids (see Chapter 8). Thus it appears that previously subsident areas were now subject to erosion. An adjustment of the tectonic framework of this type could be termed a "partial basin inversion". Considering the parallel tectonic development of similar basins in the NW Atlantic region, the most favoured time for such a restructuring to have occurred would have been around the Aptian. Thus we tentatively infer an Aptian age for the Salas Group.

2.11. Utrillas Formation (Group V). (See Chapter 6).

A further sequence of continental clastics occurs above the Salas Group. This is often colloquially referred to as the "Utrillas", a name that is generally used to refer to facies rather than as a

formal stratigraphic term. The Utrillas facies remains one of the great enigmas of Iberian geology.

In the study area, these sediments were termed Group V by Salomon (1982), while Floquet, Alonso & Melendez (1982) included them in their stratigraphic scheme for the Upper Cretaceous as the "Tejada sands".

A basal conglomerate and sandy green- or purple-mottled palaeosoils may locally be recognisable (eg in recent road cuts or dry stream scours). Elsewhere the thick and apparently monotonous sequence of pebbly sands and arkoses that forms the Utrillas Fm (see Chapter 6) often overlies the clastic deposits of the Salas Group in areas of poor exposure with little apparent change in facies. As a result, it is often difficult to distinguish between the two in the field. This certainly complicates the analysis of both.

As an aid to the recognition of the two units, Salomon (pers. comm., 1982) pointed out that the Salas Group is generally approximately concordant with the underlying strata, while the Utrillas, in contrast, is apparently unconformable upon the rest of the continental sequence. This criterion is difficult to apply in the field, (our own method for distinguishing the two units is described in section 5.1; see fig 5.1) but, nevertheless, the presence of a significant unconformity is evident from regional mapping. In places, the Utrillas lies directly upon the marine Jurassic or even upon the Palaeozoic (see section 8.15 for a discussion of the origin and regional significance of the unconformity).

The Utrillas contains no datable fauna, so that the age of the Utrillas facies in the study area has been difficult to determine. This problem is complicated by the diachronous nature of the Utrillas facies. Alonso (pers. comm., 1985) suggested that the Utrillas was a "transgressive" suite of continental and littoral facies laid down in advance of the Upper Cretaceous transgression as the seas encroached upon the Meseta (or "Northern Castillian Platform"). The only reliable

Fig 2.12-19.0. Upper Cretaceous. Classic locality; Mazariegos.
Harvest approaching, August, 1982.

dates come from the overlying marine deposits, and these clearly demonstrate that the upper limit of the Utrillas is diachronous (Floquet, 1983). The passage into the marine environments of the Upper Cretaceous sequence occurs progressively later to the S and W:

Upper Ebro.....Lower-Middle Cenomanian

SW Cantabria & N Iberian ranges...Middle-Upper Cenomanian

Sistema Central.....Lower-Middle Turonian.

This thesis does not provide new biostratigraphic data on the age of the deposit. However, my own reconnaissance at Tejada (see Chapter 8) has suggested that the upwards transition from the continental Utrillas into the marine-influenced facies of the Middle to Upper Cenomanian is indeed a gradational one. It is inferred that the upper part of the Utrillas is of Early Cenomanian age.

As mentioned above, mapping has indicated that the Utrillas of the area overlies an unconformity. However, sedimentation was apparently unaffected by faults in the subjacent strata. The Utrillas can be envisaged as a sheet deposit, not itself a product of synsedimentary tectonics but laid down as a sediment blanket draping the palaeotopography and palaeogeology. The post-Aptian onlap (see section 8.15) is elsewhere dated as Middle Albian - Lower Turonian. This is consistent with the Upper Albian - Lower Cenomanian age postulated for the Utrillas of the study area.

2.12. - 2.19. Upper Cretaceous.

The Upper Cretaceous (figs 2.45 & 2.46) has recently been the subject of several new publications by Floquet (1979-1984). The stratigraphy was summarised by Floquet, Alonso & Melendez (1982) and is outlined in fig 2.47, which shows the relationship between the Upper Cretaceous succession of the study area (region 2: Northern Cordillera Ibérica) and neighbouring areas (fig 2.48) including SW Cantabria (region 1), Central Cordillera Ibérica (region 3), N margin of the Sistema Central (region 4).

Figs 2.49 & 2.50 attempt to correlate this stratigraphy with previously defined units (Valladares, 1976; Floquet, 1979). Valladares noted that the lower parts of the sequence (her Units I & II) are present throughout the study area, while the upper parts (her Units III & IV) are found only in the SW. Unit III is only represented between Tejada and Arauzo de Miel, having apparently been removed by erosion preceding the deposition of Oligocene conglomerates in the NW. The area around Santo Domingo de Silos seems to have been particularly subsident during the Late Cretaceous - Valladares reported much greater thicknesses for each of her units there (see section 8.16).

2.12. Santa Maria de las Hoyas Formation (M - U Cenomanian).

Despite problems of poor exposure, the passage upward from the sands of the Utrillas to the marls of the Santa Maria de las Hoyas and Picofrentes Formations is often approximately mappable in the field by the change from white to pink sandy soil (over Utrillas sands) to pale grey muddy soil (over marls). A slope break is also often present (fig 2.51). Faulkner (1985) and Vaughan (1986) noted the presence at the base of the marl sequence of a cemented horizon particularly rich in shell fragments and encrusting oysters. This, the "oyster bed" (fig 2.52) is, in fact, the Santa Maria de las Hoyas Formation of Floquet et al (1982) which is only 5m or so in thickness but forms a marker horizon of regional extent. The oyster bed can be interpreted as a hardground. This apparently developed as a result of slow sedimentation in coastal waters during the transition from continental to marine environments. At Tejada, the Utrillas continental sands pass up into 5-10m of progressively more-cemented and locally multidirectional cross-bedded sands (?indicative of tidal influence).

2.13. Picofrentes Formation (Lr - M Turonian).

Above the cemented surface of the Santa Maria de las Hoyas Formation is a sequence of fossiliferous marls assigned to the Picofrentes Formation. The fauna contains elements reminiscent of those found in the Lower Chalk of S England. Small Micraster-like (Hemiaster) echinoids are abundant, while large gastropods and ammonites (?Acanthoceras) may occasionally be found. The fauna is clearly open marine and reflects continued transgression.

A prominent glauconitic sandstone at the top of the marls was reported by Faulkner (1985), (figs 2.53 & 2.54). The presence of glauconite further testifies to the marine conditions. The sandstone may record the lateral interdigitation in this area of the marly facies of the Picofrentes Fm (defined 100km to the SE) with the more clastic facies of the upper part of the Santa Cruz del Tozo Formation (defined 80km to the NW in the S part of region 1). Alternatively, the input of clastic material at this time could perhaps be correlated with the Turonian clastic influx reported by Alonso (1981) from the Sistema Central.

2.14. Munecas Formation (M - U Turonian - Lr Coniacian).

A varied sequence of limestones together make up the Munecas Formation. The lower part of this formation consists of bioclastic limestones with a diverse marine fauna including corals, bryozoa, echinoderms, brachiopods and foraminifera (Faulkner, 1985), (fig 2.55). Above this is an oolitic limestone consisting of single and compound squashed ooliths, bioclasts and glauconite (fig 2.56). Stromatolites were found at Cuevas de San Clemente and at Penacoba (fig 2.57). We may interpret the lower part of the succession as a high energy shallow-water shelf sequence. Conditions later evolved from shallow marine to restricted marine and occasionally even to supratidal. This sequence demonstrates a general "shoaling upwards" regressive facies evolution. The Late Turonian shallowing observable

here is in contrast to apparently continued high sea levels in Britain and N France at this time (see section 8.16).

2.15. Hortezuelos Formation (Coniacian sensu lato).

Above the oolitic limestones of the Munecas Formation, Faulkner (1985) noted a 90m thick sequence of poorly bioturbated wackstones containing thin broken bivalve shells. This monotonous sequence is interrupted only by horizons rich in encrusting oysters. We may interpret this sequence as the product of continued regression which gave rise to low energy restricted to open shallow marine conditions (as would be found in the internal and more open areas of a large carbonate platform).

2.16. Hontoria del Pinar Formation (Coniacian-Santonian boundary - U Santonian).

This formation corresponds to the "echinoid limestones" and "foraminiferal wackstones" of Valladares (1976). Hontoria del Pinar Formation limestones cap many prominent peaks in the study area, including the Las Mamblas and Penalara ranges in the N and Aguila and Pico in the S. They also form impressive vertical scarps in the gorges to the SW of Cuevas de San Clemente and Hortigüela. According to Floquet et al (1982), the formation consists of a succession of rudist- and foraminiferal-limestones overlain by an alternation of cross-bedded calcarenites and micritic limestones. The calcarenites contain shell débris including echinoderms, foraminifera and rudists, while the micrites contain only bivalves. Faulkner (1985) also reported gastropods. Chert is present in the upper parts of the formation, possibly implying a silica source from sponge spicules. Floquet et al (op. cit.) interpret the environments as external platform, bioclastic bar and protected platform. The observed facies evolution is indicative of post-Coniacian transgression.

2.17. Burgo de Osma Formation (U Santonian - basal Campanian).

This formation consists of rudist limestones, limestones with benthonic foraminifera, and mudstone limestones with algae and fenestrae. Floquet et al (1982) interpreted the environments as "protected lagoon to intertidal and supratidal flat". The succession of facies indicates a general shallowing at this time.

2.18. Santo Domingo de Silos Formation (Campanian).

This corresponds to Unit III of Valladares (1976) and was described by Floquet et al (1982) as the "dolomites, dolomitic marls and limestones of Santo Domingo". The presence of "carniolas" facies and also of horizons rich in benthonic foraminifera, oolitic calcarenites, fenestral structures and rudists indicates a succession of rapidly-changing and generally shallow-water conditions with fluctuating salinities. Floquet et al (op. cit.) interpreted the environments as "internal restricted platform: from subtidal lagoon to supratidal sabkha".

2.19. Santibanez del Val Formation (Maastrichtian - Palaeogene).

These sediments - "the limestones, muds and red sands of Santibanez del Val" - (fig 2.58) are equivalent to Unit IV of Valladares (1976). Floquet et al (1982) reported a series of five detrital shale-sandstone units alternating with four indurated limestones. The facies sequence is outlined in fig 2.59. The presence of cross-bedded sands, red marls, and of micritic limestones with brecciated textures and intraclasts, gastropod- and oncoid-rich horizons, indicates "continental, fluvial to palustrine/lacustrine environments". Vadot & Salomon (1985) demonstrated the strong influence of pedogenesis on these sediments, which provide an interesting comparison with those of the Tierra de Lara Group.

2.20. - 2.22. Tertiary.

The Tertiary of the study area can be divided into four main stratigraphic units:

2.19. "Garumnian".

"Garumnian" is a facies term used informally in the literature to signify continental carbonate successions lying above the Upper Cretaceous but beneath the main Tertiary molasse sequence. These sediments have been described here together with the Upper Cretaceous (see previous section, "Santibanez del Val Formation").

2.20. "Oligocene" - syntectonic.

2.21. "Miocene" - post-tectonic.

The pre-Pliocene molasse deposits of the area are divided on lithological and structural criteria into a lower, generally conglomeratic, and often strongly deformed, succession (fig 2.60), and an upper, generally sandier sequence (figs 2.61 - 2.63), which contains some limestones and shows only weak deformation.

IGME mapped the lower part of the sequence as "Oligocene" and assigned the upper part to the Miocene on consideration of the likely timing of deformation. These are the only grounds available for dating as neither unit has a reliably-datable fauna. Caution must be exercised because a trend of reducing grain-size and decreasing degree of deformation can be recognised moving progressively away from the main faults bounding the Duero Basin, so that the vertical facies evolution is mimicked by the lateral basinward change in facies. Fig 2.64 shows the regional distribution of "Miocene" facies (IGME, 1970), while fig 2.65 shows the OligoMiocene lithostratigraphic units of Pol & Caballeira (1983) and Pol (1985).

2.22. "Pliocene".

A sequence of polygenetic conglomerates and sands approximately 70m thick lies unconformably upon the Mesozoic and Palaeozoic in some places in the N part of the study area. The sequence is given the local name "Ranas Formation" and has been assigned variably to the Pliocene (IGME, 1970); Miocene-Pliocene (Salomon, 1982); or Plio-Quaternary (Gil Serrano & Zubieta, 1978).

The distribution and morphology of the outcrops is consistent with deposition in alluvial fans draining the Sierra de la Demanda, (figs 8.89 & 8.90), see also section 8.18).

Sediments assigned to the Pliocene have also been mapped in the extreme SE of the area (ADARO, 1981).

2.23. Quaternary.

Quaternary gravels outcrop in a number of localities and have been extracted locally for use in construction (eg near Palacios de la Sierra and Dona Santos).

Other Quaternary alluvial sediments are only of minor interest, but the Pliocene, Quaternary and present drainage patterns are very significant because their morphologies reflect important tectonic features (see sections 8.19 & 8.20, also figs 8.91 - 8.92).

Fig 3.0. Tierra de Lara, April 1985. View looking N over arable land on Lower Cretaceous clastics towards Lara de los Infantes, scene of the massacre of the ten "infants of Lara" during the Moorish wars in the tenth century. Perched on Alto de Lara (Lara Height), the hill in the left of the photograph which forms part of the Upper Cretaceous limestone ridge of Penalara (Lara Rock), is a seventh century chapel built by the Visigoths. Mountains in the far distance are the Palaeozoic of the Sierra de la Demanda.

CHAPTER 3. THE PETROGRAPHY AND SEDIMENTOLOGY OF THE TIERRA DE LARA GROUP.

3.1. Introduction.

The description and interpretation of the sediments of the Tierra de Lara Group (Group I of Salomon, 1982) form primary objectives of this thesis. Previous investigations have explored the sedimentology only in general terms, but have nevertheless provided a useful understanding of the main facies and environments. Schudack (1979) gave an illustrated account of the microfacies at and around the contact between the marine Jurassic and the continental sequence. This latter work provided detailed description of the basal metres of the continental sequence in a large number of localities, especially on the S margin of the Sierra de la Demanda. However, because the continental sequence rests upon the marine Jurassic with significant onlapping unconformity (see Chapter 8), the basal metres of the continental sequence, although often developed in similar facies, are in fact of widely differing ages, and in some cases actually belong to Salomon's Group II. Salomon (1982) recognised and distinguished between piedmont, floodplain and lacustrine deposits, and discussed sediments from marginal lacustrine or "palustrine" (=swampy, marshy) settings. Salomon (1984) stressed the rôle of pedogenesis and drew attention to particular root and desiccation structures.

This chapter aims to give a more systematic account of the sediments found within the Tierra de Lara Group. A close examination of facies has facilitated a clearer analysis of the lithostratigraphy (section 2.8) and of facies distribution patterns (sections 8.11-12).

The lithostratigraphy of the Tierra de Lara Group is based upon the recognition of facies, which tend to occur in a consistent sequence. The description of facies which follows is thus organised approximately in stratigraphic order. Some facies occur at several

Fig 3.2.0. Senora de Brezales Formation. Typical sequence of red sands overlain by channelled polymict conglomerates. Snow capping Valdosa, peak of Upper Cretaceous limestone in distance. Near Hortezuelos, March, 1983.

horizons, in which case they are fully described at the first appearance only.

3.2. Senora de Brezales Formation.

The Senora de Brezales Formation forms the basal part of the Tierra de Lara Group (see Chapter 2; fig 2.40). The Senora de Brezales Formation comprises three main facies: sandstones, conglomerates and carbonates. The common facies associations and the environmental model proposed for this formation are shown in figs 3.189 & 3.190.

3.2.1. Facies a: Sandstones.

Red sandstones reach a maximum thickness of 30m and are present: a) forming the base of the Senora de Brezales Formation (SdB Fm) in the SW of the study area, near to the San Leonardo Fault and in the area to the SW of it; and b) above basal SdB Fm limestone conglomerates in the NE of the study area, in the sector to the SW of the Quintanilla-Castrovido Fault and SE of the Jaramillo-Covarrubias Fault.

The sands are generally dark brick red in colour and are of medium grain size. They have subrounded grains, are generally quartzose, and often contain scattered quartz pebbles 1-4mm in diameter. The sands are friable and outcrop poorly.

The sands are often structureless (fig 3.1), although trough-cross bedding, thin pebbly lags and plane lamination are occasionally discernible (figs 3.2 - 3.5). 1-2cm diameter burrows are occasionally present (fig 3.6) and the sands often have a burrow-mottled or bioturbated appearance. At Mamolar and at Hortezuelos, the sands are locally white-red mottled. The morphology and size of the mottles at Mamolar (figs 3.7 - 3.9) suggests that they might be caused by root-bioturbation (or burrowing). Cavities filled with red sand penetrating the post-Talveila Formation karst surface at Hortezuelos may be karstic potholes or large roots later sand-filled.

Round vertical structures found to the N of Huerta del Rey (fig 3.10) and smaller-scale bedding parallel tubes found at Castrovido (fig 3.11) may also represent tree- or bush-roots.

The sands locally display fine angular cracks (figs 3.12 & 3.13). Thin, mottled and brecciated nodular and rubbly limestones are also present (figs 3.14 - 3.17). These display floating quartz grains, angular-branching sparry desiccation cracks and grey carbonate strings 1mm or so in diameter which are interpreted as fine rootlets.

Interpretation.

The red colour of the sands is apparently the product of haematite coating of the grains. The colour of the sands is immediately reminiscent of "red-beds", and their stratigraphic position above a prominent regional unconformity and associated with other continental deposits is consistent with the sedimentological evidence of continental conditions (plant roots, desiccation cracks and brecciated nodular mottled carbonates).

The homogeneous character of the sands is very striking, but the local presence of lags and meso-scale cross-bedding suggests that the sands were water-lain, (or at least reworked by water), probably as weakly-graded units. The absence of mud probably indicates that permanent flows were not established. The common absence of sedimentary structures suggests later homogenisation by strong bioturbation. The nodular and rubbly carbonates are similar to "cornstones" - pedogenic carbonates classically described from the British Old Red Sandstone (figs 3.18 & 3.19). The presence of floating quartz grains indicates the in situ growth of replacive calcite. The brecciation and mottling is strongly indicative of pedogenetic modification.

The locally poorly-defined contact between the red sands and the underlying Talveila Formation sands may suggest partial reworking.

In summary, the red SdB Fm sands are interpreted as the deposits of ephemeral streams which may have partially reworked the Talveila Formation sands. The climate was probably semi-arid. The sporadic nature of the rainfall meant that the streams only flowed rarely, while in the intervening periods the sands were homogenised by burrowing and the action of plant roots. Conditions were sufficiently dry to result in periods of desiccation. Strongly oxidising conditions were responsible for the red colour of the sediments and prevented the preservation of organic material so that the roots are usually preserved only as moulds. Occasional rainfall permitted the growth of hardy scrub vegetation and periodically allowed the transient establishment of high or perched water tables facilitating the development of pedogenic carbonates. Although there is no clear evidence for aeolian deposition, the association with other features indicative of a semi-arid or arid climate suggest that a limited amount of reworking of the sands by wind during dry periods can not be ruled out.

3.2.2. Facies b: Conglomerates.

Conglomerates are a characteristic feature of the Senora de Brezales Fm. In the SW, the conglomerates occur above the red sandstones. In the NE they are found beneath the red sands, occurring in channels which cut down directly into the marine Jurassic pediment (as they do in the NW sector between Hortigüela and Quintanilla de las Vinas, where the red sands are absent). The conglomerates are usually 2-5m in thickness, locally occurring in several pulses. At Espejon, individual conglomerate channels may be recognised (figs 3.20 - 3.23). The channels have strongly erosive bases, with the coarsest conglomerates filling the base and centre of the channel. The conglomerates pass upwards and laterally into red trough cross-bedded sands and are sometimes interbedded with them (as

at Hortezuelos & Castrovido; figs 3.24 - 3.26), suggesting that more organised flow conditions were occasionally set up.

The conglomerates are polymict (figs 3.27 & 3.28), "polygenetique" of the French usage, (except for the basal conglomerates at Hortezuelos, which include clasts only of grey micritic limestone, see Chapter 8 for discussion). The clasts are calcareous, subrounded and generally 2-10cm in size (fig 3.29), although locally reaching 30-100cm near major faults (see below and Chapter 8). Clast lithologies include:

a) fossiliferous yellow-orange packstones containing ½mm qz grains and rhynchonellid shell fragments (Lower Callovian).

b) pink to grey micrites, often partially recrystallised, - possibly weathered clasts from other parts of the marine Jurassic succession (eg Talveila Formation, Middle Jurassic or Lias).

c) pink grainstones / oolites - also apparently from the marine Jurassic succession: perhaps Middle Jurassic or Talveila Formation shallow-water facies.

d) rarer, small (< 1cm) clasts of red sandstone.

The matrix is generally of medium-grained red sandstone, locally with 1-2cm rounded quartz pebbles, these latter being more abundant in the SW.

The conglomerates show few structures and are usually clast-supported. Occasionally, matrix-supported conglomerates are found, as at Senora de Brezales (fig 3.30) but nowhere is the matrix of a sufficiently muddy constitution to suggest that débris flows were an important transport process.

Interpretation.

The presence of channels may reflect the establishment of more permanent drainage courses, but it is still likely that the flows were flashy and intermittent (although powerful and strongly erosional) occurring during sporadic periods of heavy rainfall.

The sands and conglomerates may be laterally equivalent. Thus the conglomerates were perhaps deposited close to the source areas in channels directly overlying the limestone pediment, while the sands were deposited in more distal or interchannel areas. We might imagine a series of wadi-type channels locally cutting down into the bed-rock adjacent to uplifted blocks but spreading out into more sheet-like sand bodies away from the fault scarps.

Alternatively, the presence of conglomerates and sands may be the product of varying strength of flood-events or of periods of active or relatively quiescent tectonics.

The importance of tectonic factors is evident from the location of conglomerate channels, which was conditioned by minor NE-SW synsedimentary faults (see section 8.11.2). The coarsest conglomerates are found at Jaramillo Quemado, very close to the Jaramillo-Covarrubias Fault, a major transfer structure (see section 8.20). Boulders of Lower Callovian limestone up to 1m across occur in the conglomerates at this locality (fig 3.31).

3.2.3. Facies c: Carbonates.

The Senora de Brezales Formation includes pedogenic carbonates within the sandstones (described above). More important carbonates were locally deposited in inter-channel areas.

Laminated limestones or "laminated crusts" are present in the N of the study area, either lying directly on the pediment (as at Castrovido and Mambrillas de Lara), interbedded with or cross-cutting the sandstones and conglomerates (as at Mamolar; figs 3.32 - 3.34) or developed as more extensive sheets at the top of the conglomerates (as at Quintanilla de las Vinas where a 1m-thick laminated or rubbly limestone occurs; fig 3.35).

The laminae in these limestones are irregular and undulating (figs 3.36 & 3.37), and consist of 1mm alternations of light and darker (often black) micrite. Those laminated limestones resting

directly on the pediment (fig 3.38) may be similar to the "soilstone crusts" of Robbin & Stipp (1979), and to laminated "calcrete" described by other carbonate workers from Florida. Laminar crusts have also been described by Multer & Hofmeister (1968), James (1972), and Walkden (1974). They may be distinguished from laminar stromatolites on account of their association with pedogenic features and because of their lack of diagnostic algal structures, such as calcified filaments (Read, 1976).

Many of the laminated limestones not resting on the pediment reveal a fabric of generally horizontal 1mm pale beige carbonate tubes with ½mm spar-filled central cavities. These structures are interpreted as root moulds and the beige micrite coating may either be a replacement of the root tissue itself (in which case the cavity would represent the central root canal) or may be made up of layers of concretionary carbonate precipitated around the root as a result of the local change in sediment chemistry accomplished through root action. A further description of these porous crusts is given in section 3.3.4.

In the SW of the study area (for example at Senora de Brezales and at Espejon) the conglomerate channels are incised into a succession of marly, brecciated and heavily mottled carbonates (fig 3.39). These sediments display a complex reticulate fabric produced by horizontal elements (namely interbedded 2-5cm root crusts and red marly units) and vertical structures (1-10cm diameter root columns penetrating up to 1m). The vertical tubular structures are similar in morphology to occurrences of mangrove roots described by Hofmeister & Multer (1965) (see also section 3.3.4). Fine carbonate strings and spar-filled angular-branching cracks are also a feature of these deposits. The strongly laminar fabric of these root crust / marl alternations (figs 3.40 - 3.44) gave rise to the field term "laminite" for this facies. At Senora de Brezales, the laminites are locally developed in mound-like structures (figs 3.46 & 3.47).

Interpretation.

The presence of important carbonates within the Senora de Brezales Formation is significant as the length of time represented by clastic deposition was probably relatively minor in relation to the intervening quiet periods during which pedogenic carbonates developed. The thickness and abundance of root crusts towards the top of the Formation point to many long periods when marshy conditions were developed thus allowing colonisation by shallow-water plants. These times were punctuated by lengthy periods of emergence when distal alluvial material was occasionally delivered, producing an alternation of laminar root crusts and marly sediment. In the remainder of the time that the deposits were exposed they were subject to pedogenesis causing brecciation and alteration and ultimately producing the rubbly, mottled aspect picked out by weathering today.

The rapidly fluctuating changes of environment suggested by this facies indicate repeated transgressions and regressions of an ephemeral lake margin. Evidently the shorelines fluctuated considerably as a result of seasonal or longer term climatic factors.

Note.

The abundant carbonates towards the top of the SdB Fm succession could arguably be included in the Ladera or Las Vinas Members of the Rupelo Formation. I have described them here with the Senora de Brezales Formation because of their association with conglomerates.

3.2.4. Senora de Brezales Formation: Synthesis.

The facies associations and environments of the Senora de Brezales Formation are summarised in figs 3.189 & 3.190. The integrated facies model (fig 3.190) shows that sedimentation took place in small-scale alluvial fan and sandflat environments in a semi-arid climate. NW-SE basin-bounding (extensional normal) and NW-SE cross-graben (wrench or "transfer") faults were both active influences

Fig 3.3.0. Rupelo Formation. Near Rupelo. Looking NE towards Sierra de la Demanda. Fieldwork, September, 1985.

on facies distribution (see Chapter 8). Minor limestones developed in times of reduced clastic input either as surficial (or near surface) crusts or as nodular pedogenic carbonates within soil profiles.

3.3. Rupelo Formation.

The Rupelo Formation consists of four members and forms the upper part of the Tierra de Lara Group (see section 2.8; fig 2.40). The Formation contains a wide variety of facies. The most frequent facies associations and the environmental model proposed for the Rupelo Formation are shown in figs 3.189 & 3.191.

3.3.1. Facies 1: Red marls.

Red marls are most important in the Las Vinas Member. Outcrop is generally poor, making detailed studies difficult. The best exposures are at Quintanilla de las Vinas, where approximately 40m of poor exposure separates the laminated limestones at the top of the Senora de Brezales Formation from the carbonates representing the upper members of the Rupelo Formation, and W of Torrelara, where 10m of red marl occurs above a rubbly limestone developed on thin impersistent Senora de Brezales Fm conglomerates (fig 3.48). The marls are locally weakly red-green mottled and occasionally faint colour banding is evident on a 5m scale, but otherwise the marls are homogeneous. Structures such as lamination or evidence of bioturbation (eg burrows) are absent. The marl includes lenses of limestone 2-5m in thickness and approximately 30-100m in lateral extent, suggesting that the marls may be laterally equivalent to the limestones of the Ladera Member.

The only fauna noted in the marls were ostracod and charophyte fragments found in one thin section (fig 3.49).

The marls contain a proportion of clay and iron impurities but still have a high carbonate content. A brief atomic absorption study showed that the marls contain little trace of magnesium but are rich

in calcium, suggesting that they consist of low-magnesium calcite and contain little dolomite.

Interpretation.

An analysis of logged sections shows that an important proportion of the total thickness of the Rupelo Formation is represented by poor exposure of apparently marly sediment. Although grey charophyte-bearing marls are common, especially in the Mambrillas de Lara Member, the Ladera and Las Vinas Members contain appreciable thicknesses of red - green mottled marls. It is thus important to understand their sedimentology.

However, in view both of the sparse nature of the fauna found within the marls and of their homogeneous nature, they are difficult to interpret. Thin sections provided little new information as the iron and clay impurities masked other aspects of the fabric.

There are three possible interpretations:

a) that the marls are distal alluvial sediments that were deposited out of suspension in a floodplain environment. In this model, the limestone lenses would be interpreted as the deposits of isolated ponds which locally developed in depressions on the floodplain. These ponds would have been the sites of intermittent biogenic production (and perhaps also of inorganic precipitation) of calcium carbonate.

b) that the marls are deeper lacustrine deposits laid down in water too deep for charophyte growth. In this model, the limestone lenses would be interpreted as isolated mounds or bioherms raised above the general base level and acting as loci of carbonate sedimentation.

c) that the marls are the product of more saline conditions which generally inhibited the activity of carbonate producers.

d) that the marls are the result of subaqueous clastic input which introduced a high proportion of clay and other impurities into

the lake, inhibiting the growth of or eliminating carbonate producers.

Against the deeper lacustrine model (b) we can count the absence of associated open lacustrine facies. The low magnesium content rules out comparison with deeper-water dolomitic marls as described by Link & Osborne (1978) from the Ridge Basin of California.

In favour of the elevated salinity model (c) we can make comparisons with lithologically-similar "playa lake" deposits (eg NW European Keuper Marl / Mercia Mudstone. We may also list the paucity of fauna / flora. The local chertification of "root crusts" (see facies 4 & 13, sections 3.3.4b; 3.3.13) found within the limestone lenses also points to a possible association with evaporites.

In distinguishing between the floodplain (a), playa lake (c), and clastic lake (d) models we face the problem that temporary flooding of a wide distal floodplain or playa lake produces a similar depositional setting to that of a very shallow clastic lake. The processes of sedimentation of fine-grained deposits are effectively identical in these settings so that we must rely on facies association to indicate the environment of deposition. The nature of the interfingering of the red marls with carbonate lenses suggests that they are the lateral equivalents of shallow-water carbonates. Thus we are led to the conclusion that the marls are essentially distal alluvial deposits of a floodplain / playa lake zone fringing an ephemeral carbonate lake.

Freytet & Plaziat (1982) described "sandy mudstones" from the Late Cretaceous - Early Tertiary of SW France. Like the red marls of the Rupelo Formation, these deposits were associated with shallow water marginal lacustrine carbonates and clastics. Freytet & Plaziat also favoured a distal floodplain environment for this facies.

3.3.2. Facies 2: Peloidal micrite.

Brecciated and peloidal carbonates account for much of the Ladera Member succession. They are also found towards the top of the Rupelo Formation in many sections, often occurring immediately below a prominent polygonally-cracked surface (see section 2.8).

The limestones are often mottled from purple to red to grey (figs 3.50 & 3.51). Yellow mottling is also locally developed (as at Ladera). The limestones frequently have a strongly brecciated aspect in the field, especially where they are marly, as in the far NW, (fig 3.52). No fauna is recognisable in the field. Black pebbles from ½ to 5mm (fig 3.53) are sometimes present (see facies 10, section 3.4.10), and ½ to 2mm red clasts are locally abundant. The latter may represent weathered black pebbles or else they may be reworked mottled limestone clasts. Angular branching sparry cracks ¼ to ½mm across are often abundant (fig 3.54).

In thin section, this facies often displays a strongly peloidal fabric, which is especially apparent after staining as the peloids take a darker stain than the micrite matrix. The peloids are generally ½ to 2mm in size and are rounded, occasionally showing weakly-developed internal concentric structure, but more commonly are structureless. The peloids are often partially surrounded by curving sparry cracks (figs 3.55 & 3.56). Smaller (0.1-0.25mm) rounded "pellets" are often abundant. Tubular spar-filled voids 1mm or so in diameter (fig 3.57) are often present, as are other "open space structures" (see facies 4, section 3.3.4). One specimen showed 0.1 x 2mm needles of carbonate developed in a particularly large cavity (figs 3.58 - 3.60).

Fauna is not generally detectable even in thin section, but, rarely, poorly-preserved charophyte stems may occur within the matrix. Intraclasts of darker charophyte-biomicrite (see facies 7, section 3.3.7) may also be found.

Interpretation.

The angular-branching sparry cracks are desiccation shrinkage features and are indicative of subaerial exposure. Skew and craze planes (fig 3.61, Freytet & Plaziat, 1982) are common, although the more classic polygonal cracks are comparatively rare at most localities, generally being confined to the upper part of the Rupelo Formation (Rio Cabrera Member), (figs 2.42 & 2.43).

It is suggested here that polygonal cracks are only likely to be generated by exposure of marly, muddy sediments for relatively short periods rapidly followed by submergence and subsequent burial beneath overlying sediment. Such conditions were favoured in Rio Cabrera Member times when "open lacustrine" carbonates were occasionally exposed and then submerged again apparently without being subjected to prolonged periods of soil formation. More intense pedogenesis causes modification of original fabrics and reworking of the uppermost portion of the soil profile. In such cases polygonal cracks are unlikely to be preserved.

The peloidal structures are similar to the "caliche pisoids" described by Calvet & Julia (1983) and are probably accreted as individual soil grains (peds) as a result of repeated wetting and drying. Shrinkage and rotation of these grains on drying during emergent periods results in the formation of circumgranular cracks, which are thus essentially desiccation features.

Mottling is also a characteristic of sediments subjected to emergence and is thought to be caused by oxidation and redistribution of iron impurities during pedogenesis (Freytet & Plaziat, 1982; see also Buurman, 1980).

The tubular spar-filled cavities are interpreted as root voids and indicate at least periodic colonisation by plants (see section 3.3.4 for a futher discussion of root fabrics).

The needle carbonate crystals are problematic. They might represent leaves or fine, straight root hairs.

The very rare presence of poorly-preserved fresh-water flora in these sediments implies that they are lacustrine carbonates that have been extensively modified and partially recrystallised as a result of pedogenesis. A marginal setting in a shallow ephemeral lake is inferred. The shorelines of such a lake would fluctuate greatly as a result of seasonal or longer-term climatic variations (eg Pleistocene Lake Bonneville, modern Lake Chad) so that periods of submergence and exposure would alternate.

3.3.3. Facies 3: "Rubbly horizons".

This is the field term given to especially brecciated, mottled and often impure carbonates. Two main types can be recognised:

3.3.3a. Marly; Prismatic; Reticulate.

These occur as more poorly-cemented and more strongly-coloured and -mottled horizons (figs 3.62 & 3.63) developed in association with and grading from facies 2 peloidal micrites, as for example at Talveila and El Cangrejo.

The carbonates of this facies are more impure, and ½-2mm red clasts are often abundant, as are angular-branching cracks.

Strongly nodular fabric is locally developed. Fig 3.64 shows 2-5cm carbonate nodules in a red marly matrix. The weak vertical fabric evident in fig 3.64 (prismatic fabric) is often more strongly developed (figs 3.65 & 3.66). This fabric reflects the form of 1-15 x 5-50cm vertical prismatic columns (made of purer carbonate and defined by the style of brecciation - rough vertical fracturing caused by root penetration) set in a mottled, impure carbonate matrix. The fabric is often associated with vertical tubular structures 5-15 cm in diameter and up to 50cm in length. Examples are shown (figs 3.66 - 3.69) from Huerta del Rey and Rupelo. The tubes may show a central darker reddish core approximately 5mm across, with a weak radial fabric occasionally displayed spreading outwards from it.

Horizontal and reticulate fabrics are also occasionally present. These are associated with a combination of horizontal and vertical cylindical roots and sheet-cracks which brecciate and modify original marls or bedded root crust - marl alternations ("laminites", see section 3.3.4b).

Interpretation.

Sediments of this facies show evidence of prolonged and intense pedogenesis associated with strong root- and desiccation-brecciation. Facies 3 (rubbly) sediments thus demonstrate the further effects of the alteration process manifested in the facies 2 (peloidal) carbonates. They also show an increased content of impurities.

Prismatic fabric is a common feature of palaeosoils. Figs 3.70 - 3.72 show similar features developed in pedogenically-modified horizons from Spain (Tertiary) and from Britain (Carboniferous).

The large vertical cylindrical tubes are interpreted as the roots of larger plants. The plants themselves have not been preserved but the size and morphology of the roots invites comparison with mangroves, which today colonise the low-energy brackish marginal marine swampy tidal marshes of tropical zones. Similar structures (fig 3.73) were reported from mangrove-swamps of Florida by Hoffmeister & Multer (1965). Although the environment postulated for the facies 3 deposits is marginal lacustrine rather than marginal marine, the adaptations required by the plants would have been similar. The roots show considerable vertical penetration, and indeed such plants would need to survive dry periods of lower lake level and wetter periods when more swampy conditions prevailed.

Salinities may have fluctuated due to evaporitic concentration of lake or ground waters during dry periods and increased run-off during wet periods. The local presence of strong yellow mottling has been tentatively linked with elevated salinities (Strasser, pers. comm., 1985), although no significant evaporites were recognised in

this facies. A specimen of satin-spar gypsum was collected from one of the rubbly horizons in a section at Huerta del Rey. The exceptionally strong deformation and the rubbly aspect evident at this locality may be partly due to tectonic brecciation during Tertiary thrusting. Whether the gypsum is wholly secondary or was an original feature of the sediments later remobilised during thrusting is not certain, although it is clear that deformation was concentrated along the incompetent marly and rubbly horizons.

3.3.3b. "terra rossa".

At Mambrillas de Lara, the Ladera Member includes several 0.5-1m thick red horizons. These occur separated by 1-2m of facies 2 peloidal micrites (fig 3.74). The red sediments are mottled and impure carbonates containing many 2-10mm subangular red clasts (fig 3.75). Fine 1mm strings of lighter-coloured carbonate are common and some bedding surfaces show bifurcating grey carbonate-filled crack-like features 2mm across and up to 1m in length (fig 3.76).

The red horizons themselves drape irregular basal contacts and locally appear to pass down vertically along joints (fig 3.74). The beds immediately below often have well-developed peloidal fabrics and may contain bedding-parallel pockets of 1mm-peloidal grainstone up to 10cm across and 2-3cm thick. The overlying limestone beds contain rare red clasts 1-5mm in size.

Interpretation.

The red horizons evidently contain a high proportion of terrigenous material in the form of irregular red marl/mudstone clasts. The fine strands of grey carbonate are interpreted as fine root tubes, while the larger horizontal bifurcating stuctures may owe their origin to larger roots extending horizontally or to the filling of larger desiccation cracks with mud-grade carbonate sediment.

The red horizons are interpreted as palaeosoils. They occur above limestones showing strong evidence of pedogenesis. However, they have relatively high contents of terrigenous material and are unlikely to be products of progressive alteration from an original carbonate sediment (in contrast to the mottled marly horizons described above). The irregular basal contacts and the presence of grainstone-filled voids (microkarst cavities; for a full discussion of these see section 3.3.4d - facies 4: micrite with open-space structures) in the underlying limestones suggest that the red horizons are thin soil profiles developed on and irregularly draping / infilling karst surfaces. The clastic material may have been delivered either:

a) by wind, or

b) subaqueously out of suspension from encroaching stream floods as lake fringing alluvial environments migrated towards the basin centre during more prolonged periods of emergence.

The extensive modification of the limestone beneath the red horizons could have occurred: i) prior to the deposition of the red terrigenous material by surface karst proccesses. In this case the rapid input of detrital material would have effectively fossilised the karst features and the mottling, brecciation and root structures of the red horizons would be the products of later pedogenesis; or:

ii) after the deposition of the red material, by subjacent karst processes (fig 3.77). In this model, the potholing and karstic alteration of the underlying limestone surface would have been conditioned by the markedly different porosity-permeability characteristics of the limestone and the red marl/mudstone. The red terrigenous clastic layer would also have suffered some pedogenic modification.

In fact, both surface and subjacent karst processes may have been important.

The red horizons interdigitate with Ladera Member marginal lacustrine carbonates. Red clasts may have been partly derived from

reworking of lake-fringing distal floodplain marls from the Las Vinas Member, which may have been laterally equivalent (see section 2.8).

The red horizons appear to be similar to palaeosoils developed over karst surfaces in Bermuda (Land, MacKenzie & Gould, 1967; Esteban & Klappa, 1983). The red horizons were loosely termed "terra rossa" by analogy with these examples. Recent work by Wright (in prep.) invokes a more rigorous genetic definition of the term requiring illuviation of clays and sediment derivation by aeolian processes. Such definitions, if applied sensu stricto, would exclude the example from the Ladera Member (and possibly also the Bermuda example). However, in the absence of an alternative more appropriate and concise descriptive lithological name for red suprakarstic soils of this type, I have chosen to retain the term, used between quotation marks.

3.3.3c. Rubbly horizons - summary.

The features of the marly, prismatic, reticulate and red "rubbly horizons" are indicative of deposition in more marginal lacustrine environments associated with increases both in the strength of pedogenetic modification and in terrigenous clastic supply.

3.3.4. Facies 4: Micrite with open-space structures.

Limestones from the Tierra de Lara Group (especially in the Ladera Member) frequently possess "open-space structures". This term was coined by Sbeta (1976) to denote millimetre- to centimetre-scale spar-filled cavities. A variety of cavity morphologies is recognised.

3.3.4a. Angular-branching cracks and angular sparry cavities.

Angular branching cracks have already been described in section 3.2.2 and interpreted as desiccation features. Occasionally, angular branching cracks may meet to form larger "angular sparry cavities". The association with angular branching cracks suggests their interpretation as desiccation features.

3.3.4b. Lined cavities.

Numerous 1mm diameter, generally horizontal tubular spar-filled pores are surrounded by a lining of weakly concentric-laminated pale beige to pale brown carbonate. These cavities are locally abundant so that discreet patches (fig 3.78) or bedded units several cm in thickness (fig 3.79) may be entirely made up of a dense fabric of cavities and coatings. Thicknesses of up to 30cm (and occasionally up to 1m) may be attained.

On outcrop scale, the fabric is often crudely laminar. Prominent beds possessing this structure may be traced laterally for over 100m in some cases (as at Senora de Brezales) and may be of local use in the correlation of sections across minor faults or small areas of no exposure. There is often an association with facies 3 rubbly reticulate horizons (as at Talveila, figs 3.44 & 3.45) or with carbonates displaying other pedogenic features such as desiccation brecciation or mottling.

In thin section, the voids and coatings characteristically make up a distinctive lacy and gently undulating fabric of great intricacy (figs 3.80 - 3.83).

Interpretation.

The tubular structures are interpreted as root voids. Similar structures have already been described in sections 3.2.2, 3.2.3, 3.3.2 and 3.3.3. Klappa (1980) discussed the morphology of root structures, or "rhizoliths". Those described here possess associated linings/coatings of carbonate and may thus be classified under Klappa's system (fig 3.84) as "rhizocretions" (sensu stricto).

The tubes are generally parallel to bedding. This morphology could be an adaptation to relatively dry climate - it would have facilitated maximum lateral extension of the root network. Alternatively, it could reflect the presence of firm and perhaps partially lithified subjacent sediments. The bedding-parallel

orientation of the root tubes is ultimately also responsible for the crudely laminar fabric of bedded units, or "root crusts" (a term used to distinguish them from the "laminar crusts" (Multer & Hoffmeister, 1968) or "croûte zonaire" of French usage (Freytet & Plaziat, 1982).

Analogous porous and crudely laminar fabrics have been noted from the Portlandian at Swindon, S England, and from the Lower Carboniferous of S Wales by Dr V P Wright (pers. comm., 1985). Alveolar texture (Esteban, 1976) is often rare in these examples. The laminar horizons from the Mississippian of Kentucky figured by Harrison & Steinen (1978) also contain rounded voids of comparable morphology and may have been formed by a similar process.

Although this porous fabric is very common within the Tierra de Lara Group, the more classic root fabric "alveolar texture" (fig 3.85) is relatively rare. The absence of other root-associated structures might alternatively suggest alternative hypotheses for their origin. Klappa (1981) described biogenic, laminar pedogenic fabrics from Quaternary soil profiles in SE Spain, interpreting them as the product of colonisation by epilithic lichens ("lichen stromatolites"). Krumbein & Geile (1979) recorded the formation of laminated crusts (desert stromatolites) by calcification of coccoid cyanobacterium. Wright (1981) further discussed the formation of laminar horizons, terming them "subaerial stromatolites". Goldbery (1982) described bedded, laminar pedogenic units from lateritic palaeosols in Israel. However, these crusts all display denser, poreless, fabrics of a type exhibited by only one crust in the study area (this crust may be of Recent age; fig 3.86). I prefer to interpret the lined cavities as rhizoliths.

3.3.4c. Unlined tubular cavities.

Tubular cavities 1mm or so in diameter, several millimetres in length and without linings/coatings are present in a number of localities. These cavities are particularly common in fine-grained

micritic limestones which are otherwise remarkably homogeneous and demonstrate only rare manifestations of pedogenesis.

Interpretation.

These cavities are somewhat similar in morphology and size to those of section 3.3.4a (although linings / coatings are absent). They were originally interpreted as root structures (uncoated rhizoliths; Klappa, 1981) but an alternative interpretation was later proposed (Davaud & Strasser, pers. comm., 1985) that these cavities are small burrows or "worm tubes". The formation of these cavities by a different process would explain the difference in morphology from the cavities described in section 3.3.4a.

3.3.4d. Cavities with geopetal fills.

Many cavities in the Tierra de Lara Group show geopetal filling. Cavities of this type are most common in the Ladera Member but are also common in the Mambrillas de Lara Member. The cavities are characteristically elongate (2-20mm in length and 1-5mm in height) and frequently have approximately horizontal planar cavity floors but smoothly rounded roofs (figs 3.87 & 3.88).

The filling of these cavities generally consists of a lower early generation of fine microspar (grain size 10μ) and an upper and later coarse blocky void-filling cement (grain size 100μ - 200μ) (see fig 3.88). Several generations of early filling may be discernible in the field from the presence of weak white to pale pink colour banding in the fine microspar (fig 3.87). The proportion of cavity filling varies greatly. In some cases there may only be a trace of microspar in an otherwise coarsely crystalline sparry cavity. In other instances the whole cavity is filled with microspar with no coarse spar present, reducing the lithological contrast with the "host" micritic carbonate, (eg fig 3.89), making it then exceedingly difficult to recognise the areas of paler carbonate as former cavities.

Rounded peloids ½ to 1mm in diameter may be found "floating" within the cavities (fig 3.90) and there is often a coarsening-upward trend in the grain-size of included clasts within a single cavity. Cavities often link up to form complex networks with perched geopetal fills of internal sediment and abundant "floating" peloids (fig 3.92), and, occasionally, intraclasts of charophyte biomicrite (fig 3.91).

Interpretation.

Although the deposits described here are thought to be entirely non-marine, the sediments of ephemeral bodies of fresh-water are likely to undergo many of the same diagenetic processes on emergence as inter- and supratidal marine sediments during times of lowered sea level. Similar cavities with geopetal microspar fillings have been reported from marine deposits by Dunham (1969a) in a discussion of supratidal diagenesis from the Townsend Mound of New Mexico. Dunham thought that the microspar internal sediment was a product of diagenesis in the vadose zone and termed it "vadose silt". The interpretation of the geopetal cavities of the Rupelo Formation as vadose features is consistent with the abundant evidence for emergence and pedogenesis.

Freytet & Plaziat (1982) described similar occurrences of geopetally-filled cavities from the Thanetian and Eocene of SW France, terming them "pedogenetic pseudomicrokarst" cavities. The mechanism they envisaged for the development of these cavities is explained in fig 3.93. The cavities are thought to be formed initially by root systems and subsequently enlarged by the passage of ground waters and by limited dissolution prior to the deposition of the internal sediment under vadose conditions.

Microkarst cavities, geopetal fills and floating peloids have also been recently noted from pedogenetically-modified lacustrine deposits in the Oligocene Bembridge Limestone of Isle of Wight, S England (Wright; J D Marshall, pers. comm., 1985).

Sbeta (1976) described the occurrence of irregular cavities with internal sediment (and noted coarsening-upward sequences of included peloids within them) from the Aguilar Formation of SW Cantabria (fig 3.94). These examples include both elongate cavities and very spectacular vertical cracks penetrating up to 30cm (fig 3.95). Comas (1981) subsequently interpreted them as microkarst cavities. The equivalent stratigraphic position and similar morphology of the geopetal cavities within the Rupelo Fm implies a similar mechanism for their formation.

Although Freytet & Plaziat (1982) postulated a limited contribution by dissolution to the formation of microkarst cavities, the rounded shape of the cavity rooves (which occasionally show evidence of minor collapse) at least suggests modification of the cavities by solution. Wright (1984) discussed cavities with similar rounded roof outlines from pedogenetically modified Lower Carboniferous sediments of S Wales, suggesting that solutional processes were responsible for this morphology.

Finally, the similarity between the geopetally-filled cavities described here and "stromatactis" cavities (Bathurst, 1959, 1982) was recorded by Schudack (1979) and is worthy of note. The term "stromatactis" is normally restricted to cavities occurring within marine sediments and its use is generally confined to description of Palaeozoic strata. Much debate has centred on these structures and various explanations have been advanced to account for the formation and preservation of cavities with sharp rooves and steep walls. Mechanisms proposes include: trapping of gas bubbles within the sediment; sediment collapse as a result of water escape (Heckel, 1974); or biogenic origin as trace fossils or as cavities left by the decay of sediment binding organisms (eg algae, sponges: Borque & Gignac, 1983).

Is it possible that some recorded instances of "stromatactis" are actually ancient microkarst features which record periods of

exposure hitherto unrecognised from ancient sediments) or are they merely similar structures formed by different processes?

3.3.4e. Cavities of uncertain origin.

1-3mm cavities filled with brown spar have been noted from the Rupelo Formation near Torrelara. These cavities often have angular terminations, but they are not associated with desiccation cracks or other indications of emergence. Strasser (pers. comm., 1985) suggested that they may be "birds-eyes" (Shinn, 1967, 1983) formed as a result of the presence of gas bubbles within the sediment. An alternative explanation is that they represent authigenic minerals (eg sulphates) which were precipitated during early diagenesis of the sediment but were subsequently dissolved.

3.3.5. Facies 5: Green marls.

Porous (root) crusts in the Ladera Member at Mambrillas de Lara are associated with thin (1-2cm) partings of green marl (fig 3.78). These partings either drape irregular top and bottom surfaces of bedded units or occur as small pockets between undulating bedding planes. No fauna was observed.

Interpretation.

The green marls are problematic. Either:

a) the green marls were subaerially deposited, possibly forming within soil horizons,

b) they were deposited subaqueously, perhaps as a result of increased clastic input, or

c) they were formed by reworking of volcanic material.

The green colour might suggest conditions that were not strongly oxidising. Moderately-reducing conditions may occur within relatively organic-rich soil profiles, in buried sediments or under subaqueous hypersaline conditions. Absence of fauna would result in each case.

Wright (pers. comm., 1985) noted the presence of green marls from brackish / fresh-water deposits from the Upper Oxfordian Cabaços Formation of SW Portugal. I have found similar thin green marls associated with pedogenically-modified horizons in the Dirt Beds of the British Purbeck on Portland (fig 3.96), while Strasser (pers. comm., 1984) reported green marls with illite-smectite clay mineralogies from the Purbeckian of Switzerland. Although not volumetrically important, green marls do appear to be common features of Late Jurassic - Early Cretaceous marginal carbonate sequences and merit more detailed investigation.

3.3.6. Facies 6: Limestone conglomerates and arenites.

Clastic deposits are rare within the Rupelo Formation. However, limestone conglomerates up to 1m thick (fig 3.97) and/or thin (2cm) arenites occur at a number of localities in the N tract of the study area between Torrelara and Rupelo. The clastics are assigned to the "Torrelara Bed", which marks the contact between the Ladera and Mambrillas de Lara Members (fig 3.98).

The clastic deposits include conglomerates, grits, and arenites, all rich in carbonate clasts and often displaying grading.

The conglomerates have erosive bases and irregular carbonate clasts ½-20cm in size. Clast lithologies include pale beige micrites and black pebbles. The former exhibit pedogenic features and belong to facies c, 2 and 4, apparently being derived from the erosion of underlying Senora de Brezales Formation and Ladera Member carbonates.

The sandstones have slightly erosional bases and may exhibit trough cross-stratification.

Interpretation.

The conglomerates and arenites of facies 6 are all of local derivation and are confined to the N of the study area. They appear to

have been deposited in discreet events by small-scale streams and may be localised near zones of synsedimentary faulting.

3.3.7. Facies 7: Gastropod-ostracod-charophyte biomicrite and interbedded grey charophyte-marls.

The upper part of the Rupelo Formation contains mid- to dark-grey limestones which show relatively little pedogenic modification. The darker limestones occasionally exhibit diffuse lamination (see section 3.3.8a), and are interbedded with dark- to mid-grey marls (figs 3.99 & 3.100). Bedding is on a 20-40cm scale and is notably better-developed than in the pedogenically-modified facies. Beds are usually tabular but locally show lensoid geometry (fig 3.101).

Mudstones and wackstones occur, the difference depending upon the proportion of ½-30mm intraclasts, which include black pebbles (see facies 10) and clasts of pale micrite (fig 3.102). The pale micrite intraclasts often show peloidal fabrics and internal structures such as angular-branching and circumgranular cracks are common. Detrital quartz grains are extremely rare.

Open-space structures similar to those of facies 4 are rare. Cavities are occasionally found with partial geopetal fills of microspar internal sediment, but desiccation features are absent, except at the very top of the sequence where polygonal cracks are found on bedding planes.

The limestones take a particularly dark red stain when subjected to the technique of Dickson (1965). The intensity of the staining may be the result of either:

a) the fine grain size, which permits rapid reaction with the staining solutions, or:

b) the presence of clay impurities or organic matter.

However, despite the pungent smells released on cutting these limestones, a pilot study revealed low total organic carbon values

(see brief discussion in section 7.1). The odours may reflect the presence of disseminated sulphides.

The marls are often fissile and friable. Samples were not sufficiently consolidated to permit sectioning, but yielded a rich fauna of charophytes (see fig 3.103). The marls often outcrop poorly, but when present they show a uniform mid- to dark-grey or greenish grey colour and appear to be fairly homogeneous. Features indicative of pedogenesis, such as mottling and carbonate nodules, are absent.

The biota of this facies is dominated by charophyte stems (fig 3.105) and by their oogonia, (the female reproductive parts enclosing the egg; fig 3.106). Ostracods and gastropods are common, and thicker shell fragments (possibly broken bivalve shells) also occur (fig 3.105).

Charophyte stems are generally more abundant in this facies than oogonia, which are concentrated in a few darker beds. The structures visible in the stems in thin section depend upon their orientation and the level at which they have been sectioned (figs 3.107 & 3.108). Internode fragments are the most common. These comprise a central tube surrounded by the cortex, which consists of twelve tubes, thus giving the characteristic "telephone dial" appearance in transverse section (fig 3.109). Node sections are much rarer, but can be recognised by the enlargement of six of the cortical cells (fig 3.110). The oogonia also show a variety of structures according to the orientation of sectioning (figs 3.111 & 3.112). They are approximately spherical, about 0.5-1mm in size and have a case consisting of five sausage-shaped cells that spiral to the left. At the end of these enveloping cells, but rarely preserved, is the cornula, a structure consisting of five small cells. Charophyte morphology is summarised in fig 3.110), taken from Sbeta (1976), and a list of charophytes identified in the Rupelo and Aguilar Formations is given in figs 3.103 & 3.104.

Ostracods present range in size from 0.5 - 2mm. Cyprideae appear to be dominant. Both articulated and disarticulated shells occur. The structure of ostracod tests is summarised in fig 3.113. Geopetal fillings of micritic internal sediment may occasionally be found and shelter porosity is preserved beneath some shells (fig 3.114).

Gastropods found range in size from 1 to 3cm (see eg fig 3.108). Specimen SP22 (see fig 3.103) from the Mambrillas de Lara section contained Anisopsis sp., Proauricula sp. (jaccardi?) and Valvata sp.. Many gastropods have geopetal internal micrite sediments with drusy calcite filling the remaining void space.

Large vertebrate bones and fragments were found in marls and biomicrites of this facies at Ladera and Espejón (figs 3.115 - 3.119). The bones have been replaced by a blue phosphate mineral (vivianite). Many bones are too fragmentary for precise identification, but a collar rib and the neural arch region of a vertebra were found. The bones were examined at the Natural History Museum in London and tentatively identified as the remains of dinosaurs from the herbivorous Family Sauropoda. These dinosaurs inhabited fresh-water swamps and many were extremely large (eg commonly 15m for the genus Brontosaurus). The dimensions of those bones that have been anatomically identified confirms that they belonged to animals of quite enormous size.

Dinosaur bones were only collected from two localities, but in many other places bedding planes towards the top of the Mambrillas de Lara Member show 30-50cm depressions that are interpreted as vertebrate footprints (figs 3.120 & 3.121). The shape of the prints suggests formation by feet of rounded outline, possibly with flat soles. Some prints show three protruding toe-marks (3.122).

Significance of fauna.

The fauna/flora of facies 7 biomicrites consists of fresh-water elements (figs 3.123 & 3.124) which together indicate low-salinity, shallow-water environments.

Charophytes belong to a highly-specialised non-marine algal group. Chara ("stonewort", "muskglass", "bassweed") is the modern representative of this group. It is unable to survive marine conditions, but inhabits fresh- and brackish-water. It forms a significant part of the submerged vegetation in streams, ponds and lakes in temperate zones, growing in quiet water with roots buried in the sediment. As a green alga, Chara is restricted to the photic zone, and although known from water up to 20m deep in Finger Lakes, New York and Lake Neuchâtel, Switzerland, it generally favours depths of 1-7m (Wood & Imahori, 1963).

The stems and oogonia are frequently calcified. The stems are especially important contributors to the $CaCO_3$ production in temperate fresh-water, both in modern and ancient hardwater lakes, eg: in Green Lake, Wisconsin (Twenhofel, 1950); in Early Cretaceous Lake Peterson (Glass & Wilkinson, 1980); and in the marl lakes of Michigan (Murphy & Wilkinson, 1980). Charophytes are strongly calcified particularly during the summer months when water temperatures are highest and rates of photosynthesis are elevated. In warm, shallow and highly productive freshwater lakes, calcified charophyte débris makes up most of the sediment. Certainly, in open lacustrine environments at least, biogenic carbonate production often far outweighs generation of calcite by inorganic evaporative precipitation.

The stems are 2-3mm in length and 0.2-1mm in diameter, delicate, and thus not readily transported. The presence in the Mambrillas de Lara Group of almost intact charophyte stems in great abundance thus indicates quiet, shallow-water conditions.

The oogonia, in contrast, are smaller and more readily transported. The removal of oogonia by gentle currents accounts for

their paucity in many beds where stems are abundant and explains their localised presence in a few darker mudstone beds where stems are absent. Where ostracods and oogonia predominate in the absence of stems, slightly deeper (but not anoxic) conditions may be inferred.

The abundance of charophytes (and perhaps of other plants which have not been preserved) and the swampy conditions would have favoured colonisation by sauropods. The presence of vertebrate footprints further implies very shallow water depths, and perhaps even temporary emergence. However, sauropods were so large that some authorities have questioned their ability to support their enormous weight out of water. The prints themselves suggest foot morphology adapted on the same principle as that of the elephant, rhinoceras or hippopotamus, giving a large surface area for supporting great weight without sinking into a soft, muddy substrate.

The gastropods reported are all fresh-water forms, and the predominance of Cyprideae in ostracod assemblages was cited by Anderson (1973) as indicative of brackish- to fresh-water lagoons in subtropical climates similar to those now occurring in the S Mediterranean and N Africa.

Interpretation.

The presence in the Mambrillas de Lara Member of significant thicknesses of well-bedded limestones and marls displaying fresh-water fauna/flora and little evidence of subaerial exposure indicates that more perennial lakes were established later in Tierra de Lara Group times. The predominance of mudstone textures suggests slow sedimentation of biogenically-produced carbonate under generally low energy conditions. The absence of lamination implies that the bottom conditions were oxygenated. Even the diffuse lamination present in the darkest limestones (facies 8, fig 3.125) shows evidence of disturbance by limited bioturbation.

The presence of wackestones (figs 3.127 & 3.128) containing pale micrite (from more littoral environments or from underlying strata, eg in the Ladera Member) and black pebble (see discussion, facies 10) intraclasts indicates reworking from areas around the lake. Intraclasts may have been washed in during storms. According to Freytet & Plaziat (1982; see fig 3.129), an upward increase in the abundance of intraclasts within a single bed indicates progressive lacustrine filling (and hence increasing proximity to littoral zones) while an upward decrease in intraclast abundance may indicate flooding and sudden reworking of brecciated marginal deposits.

The weakly lensoid bed geometry is similar to that present in open lacustrine carbonates in the SE Ebro, interpreted there by Cabrera et al (1985) as the result of current action. Any original differences in limestone bed thicknesses would be emphasised by differential compaction of the interbedded marls.

Although most pedogenic features are absent from this facies, rare cavities do occur with partial geopetal fills of microspar (eg fig 3.126. These cavities are similar to facies 4d open-space structures (microkarst cavities). Sbeta (1976) described similar cavities from unmodified biomicrite facies in the Aguilar Formation and discussed possible hypotheses for their formation. A problem arises because:

i) no other sedimentary structures unequivocally indicative of emergence occur in this facies. This suggests subaqueous formation of the cavities, perhaps by algae or by the development of gas bubbles.

ii) the internal sediment within the cavities is "vadose silt". This implies subaerial diagenesis.

Two possible mechanisms to explain these apparently contradictory features are detailed below.

A. 1) subaqueous deposition of micrite.

2) early formation of cavities by gas bubbles, algae or by burrowing.

3) preservation of the cavities by early subaqueous lithification of the host sediment.

4) later emergence facilitating late-stage vadose diagenesis and the precipitation of vadose silt.

B. 1) subaqueous deposition of micrite.

2) emergence facilitating formation of the cavities as pseudomicrokarst features (but not permitting the onset of intense pedogenesis).

3) vadose diagenesis; deposition of "vadose silt".

4) submergence and resumption of subaqueous micrite deposition.

In view of the shallow depth of the lake implied from faunal evidence (see above) periodic and short-lived emergence would have been likely. Thus the formation of microkarst cavities during brief emergence, as outlined in B, would have been possible.

The formation of microkarst cavities in the absence of other pedogenetic features might seem unusual, but the preservation of polygonal desiccation cracks near the top of the Rupelo Formation (fig 3.130) shows that short periods of emergence did occur without extensive pedogenetic modification. Microkarst cavities can penetrate to considerable depths so that horizons bearing other evidence of exposure might be confined to levels nearer to the surface and may easily be eroded without trace or redeposited as intraclasts. Polygonal cracks themselves would not be preserved if there were no lithological contrast with the overlying beds.

3.3.8. Facies 8: Laminated biomicrites.

Lamination is extremely rare in the Rupelo Formation, being confined to the Mambrillas de Lara Member. Two main types of lamination occur:

3.3.8a. Diffuse lamination.

Darker beds in the biomicrite facies may rarely show poorly-developed, diffuse and slightly undulating, disturbed millimetre-scale alternation of lighter and darker laminae (fig 3.125).

Interpretation.

The dark colour of the limestones suggests elevated contents of organic carbon, although in fact TOC values are still relatively low (<½%). Thus although this type of lamination may reflect the local establishment of slightly anoxic bottom conditions inhibiting bioturbation, the preservation of lamination may merely indicate strong compaction (fig 3.125), or particularly rapid sedimentation and incomplete biological reworking. The dark colour may be due in part to increased clay content or to the presence of disseminated sulphides.

3.3.8b. Clastic lamination.

At Torrelara, greenish grey marls exhibit 1mm graded silt laminae which occur at 2-3mm intervals (fig 3.131). The marls contain ostracod tests and are interbedded with facies 7 biomicrites.

Interpretation.

Detrital quartz is extremely rare in the Mambillas de Lara Member and the presence of clastic laminae may indicate proximity to terrigenous clastic input which was able to penetrate the lake-fringing zone of terrestrial and aquatic plants. This zone normally acted as a trap and filter of clastic sediment. The silt laminae are interpreted as the deposits of lacustrine density currents, these perhaps being generated as a consequence of river input.

Density currents within lakes have been well-documented in the literature, and indeed much early work on "turbidites" was based on

studies of sedimentary processes in modern lakes (eg Lake Geneva). Similar thin graded clastic laminae have been reported from modern E African lakes by Talbot (1984).

The preservation of the lamination may have been facilitated by fairly rapid sedimentation in slightly deeper or more saline conditions. Ostracods are present in the clastic-laminated facies, but it is not clear if these are reworked or if they include forms capable of tolerating more saline conditions.

The rarity of lamination is discussed further in section 3.6.

3.3.9. Facies 9: Intraformational conglomerates.

The Mambrillas de Lara Member is dominated by facies 7 biomicrites, but weathered surfaces locally pick out conglomeratic texture. This was originally thought to be a product of the style of weathering, but is in fact a primary depositional texture only revealed by weathering. The conglomerates are matrix-supported, and contain moderately-sorted, irregular 1-5cm mid grey biomicrite clasts of moderate sphericity plus less abundant 1-3cm black pebbles set in a mid grey marly biomicrite matrix (fig 3.132).

These conglomerates have erosional bases (fig 3.133). 2-3m wide, 30cm deep channels may be observed in the Mambrillas de Lara section. Similar channels also occur in the Aguilar Formation (fig 3.134).

Interpretation.

The conglomerates are difficult to recognise in the field as the clasts and matrix show only weak lithological contrast. The clast assemblages are dominated by facies 7 biomicrites. Marls may have also been important contributors to the sediment load of the channels. The conglomerates indicate localised internal reworking. The only clasts present that are arguably of external derivation are the black pebbles (see facies 10).

The channel fills probably represent powerful single erosion-sedimentation events. Debris flows may have been an important depositional process.

Channelised sediment transport may have occurred on the submerged lake floor, but this would require steep slopes or strong, perhaps storm-induced currents. I prefer to envisage short periods of lower lake-level causing rapid and transient retreat of the shorelines and briefly exposing "open lacustrine" sediments (eg facies 7 biomicrites and marls). Internal drainage streams then locally plied these wide, flat exposed areas much as small dendritic channels drain tidal flats at low tide. The channels may be termed "low-stage lake flat drainage channels". The only material available for stream reworking was the immediately subjacent sediment consisting solely of facies 7 biomicrites and marls (hence the lack of lithological contrast between clasts and matrix). These sediments may have previously undergone rapid early consolidation and partial desiccation brecciation (although there is no evidence of more intense pedogenesis). Clast input into the channels was probably by a combination of direct stream erosion (→ "rip-up" clasts) and minor bank collapse.

3.3.10. Facies 10: Black pebble conglomerates.

Black intraclasts are common at many horizons within the Rupelo Formation (they also occur in the Hortigüela Formation; see Chapter 4). Some occur in "peloidal" (facies 2) carbonates. These may be 2-5 (and rarely up to 30)mm in size, showing a clear association with red clasts and often exhibiting pink-red mottling at their margins. Black intraclasts may also be found in facies 9 intraformational conglomerates (when they are often rounded) and in facies 11 grainstones. However, black pebbles are commonest in facies 7 biomicrites (fig 3.135), where they are more angular, 1-5mm in size and often clustered. Locally the biomicrites are rich in intraclasts

and have wackstone texture (figs 3.136 & 3.137); where black clasts are especially abundant the wackstones are given the field name "black pebble conglomerate". Similar wackstones also occur in the Aguilar Formation (fig 3.138).

In thin section, the black intraclasts occasionally show a diffuse, clotted fabric (fig 3.139), although in other instances there is little lithological contrast with the biomicrite matrix.

Interpretation.

Although common within the Purbeck, black pebbles are a common feature of many ancient and recent marginal marine and fresh-water carbonates (figs 3.140 & 3.141). Ancient examples quoted in the literature date from the Carbonifeous to the Tertiary, while black clasts of Pleistocene and Recent age have been reported from tropical and subtropical zones around the globe.

Two main hypotheses have been advanced to explain the origin of black pebbles:

a) increased content of metallogenic material (eg sulphides, organometallic complexes etc., and

b) elevated organic content.

Klappa (1978) attempted to remove the organic material by reflux. The residue was still dark, and Klappa thus favoured model a). However, the dark colour of the residue may have been due to incomplete removal of the organic matter. Strasser & Davaud (1983), working on black pebbles from the Purbeck of French and Swiss Jura, were able to rule out hypothesis a) as they found that sulphides (eg pyrite) were absent from most black pebbles. Clay minerals are also generally absent. This implies that the dark colour is indeed due to the presence of elevated contents of organic matter (hypothesis b).

Strasser & Davaud's differential thermoanalysis studies indicated that blackened calcretes, black pebbles and blackened root casts owed their dark colour to material derived from higher

plants (possibly after burning and conversion to more chemically stable charcoal), while more diffuse black spots and black limestones are darkened by a combination of algal matter and some pyrite.

Applying this to our sedimentological interpretation, a ready source of higher plant material necessarily implies proximity to shore. The requirement for plant colonisation and for occasional fires suggests a warm and perhaps seasonally semi-arid climate.

Organic matter is adsorbed as molecular layers on the surface of micritic carbonate crystals (Carter, 1978). Adsorption is favoured by small crystal size and may be associated with neomorphism and aragonite → calcite inversions (see section 7.2.5.1). The process is favoured by reducing and slightly alkaline conditions, (eg in fresh-water or hypersaline ponds or around roots in a soil profile).

Once adsorbed, the organic matter could be effectively preserved by early cementation under vadose or phreatic conditions. Preferentially blackened and cemented zones are harder and relatively more resistant than the surrounding host sediment to subsequent chemical changes (eg pedogenesis, further diagenesis or weathering). Thus the black pebbles may easily be concentrated during reworking, as suggested by their observed clustering. The stability of the adsorbed organic matter is such that it will only be removed by bleaching if the carbonate is first dissolved. The high stability may explain incomplete removal of the organic matter by Klappa (1978).

The fixed organic matter will be relatively stable to oxidation even during reworking. However, black pebbles clearly do break down under intense pedogenesis, progressively oxidising to the red pebbles that are often abundant in facies showing evidence of prolonged emergence (ie facies 2, 3 and 4). The red colour may be due to oxidation of trace amounts of pedogenetically-redistributed iron.

This model also explains the size distributions of black and red clasts. Black pebbles in marginal facies are rare but are often large, while in open lacustrine facies they are smaller but more abundant.

Larger and more abundant black clasts would originally have been delivered to marginal lacustrine enironments, which were more closely associated in space and time with the source areas (the forested and burned areas on land) and these black pebbles would have undergone shorter transport and less reworking than those eventually delivered to more "open lacustrine" environments. However, there is a preservational bias in marginal environments towards large black pebbles as: a) they have proportionately smaller surface areas and undergo only slow oxidation on emergence, while smaller black pebbles are often oxidised to red pebbles; and

b) smaller pebbles are transported more easily than larger pebbles and are carried out into open lacustrine environments.

In central lake areas, increasing distance from a shoreline implies greater distances from a source of black pebbles and probably also progressively lower energy. Hence, we would expect a gradual diminution in number and size of black pebbles and also perhaps a greater content of finely disseminated organic matter (especially in the absence of bioturbation). We would also expect red clasts to be largely absent. This is indeed the case.

Thus black clasts are commonest in relatively high energy open lacustrine settings. Conglomerates of clustered black pebbles may be associated with storm or transgressive reworking of blackened horizons into open lacustrine environments (see fig 3.142, from Freytet & Plaziat, 1982; and discussion, section 3.3.7).

Analysis of the size and abundance of black pebbles in vertical sequence can thus provide valuable information on the evolution of facies and palaeogeography, while their very presence has important implications in the reconstruction of palaeoclimate and vegetation patterns.

3.3.11. Facies 11: Grainstones and spar-matrix breccias.

Grainstones are rare in the Rupelo Formation. Small, isolated pockets of grainstone are commonly found within sparry cavities of facies 4 micrites with open-space structures. Greater thicknesses of grainstone occur interbedded with biomicrites and marls near the top of the Mambrillas de Lara Member at Mambrillas and (in the absence of biomicrites and marls) at Las Pinarejas and Huerta del Rey.

The grainstones are made up of abundant 1-30mm carbonate intraclasts with sharp margins set in a spar matrix (fig 3.143). Sorting is extremely variable. The larger clasts are often angular, while the smaller 1-2mm intraclasts are generally more rounded. In the field, poorly sorted grainstones with abundant large angular intraclasts resemble breccias (fig 3.144), while those with well-sorted smaller intraclasts resemble calcarenites.

The commonest intraclast lithology is grey to beige charophyte biomicrite (fig 3.145), although black pebbles (see facies 10) are also present. Charophyte stems are abundant, but oogonia are rare.

Interpretation.

The grainstones of facies 11 contrast with those of facies 4 as they show no evidence of subaerial exposure. Desiccation features are absent and the biota (a shallow-water open lacustrine assemblage of abundant charophyte stems and oogonia) is well preserved, indicating little pedogenetic modification. The intraclasts do not show "jig-saw" fabrics, effectively ruling out in situ formation by mechanical (eg evaporite solution collapse) or tectonic brecciation. The spar (see section 3.4) did not develop by progressive neomorphic replacement of micrite - it is a cement.

The absence of characteristic Mambrillas de Lara Member biomicrite facies in some sections where grainstones are prominent is explained by internal reworking of biomicrites to form grainstones.

Sbeta (1976) described similar "charophyte biosparite" facies from the Aguilar Formation, interpreting them as lake shore or offshore bar deposits. Certainly, relatively high energy reworking is implied. Winnowing could have been performed by traction currents or storm-generated waves in shallow, littoral zones.

The grains may have been ripped up solely by currents, or may have been reworked from biomicrite facies already slightly brecciated either by surface desiccation (see discussion of polygonal cracking, facies 2) or by the growth of microkarst fissures (see facies 7). Intraclasts derived from biomicrites partially-brecciated by these processes would not themselves show pedogenetic fabrics.

Ideally, the absence of pedogenetic features displayed by the clasts should permit the disinction between these "primary" grainstones, reworked under relatively high energy conditions, from "secondary" grainstones (Freytet & Plaziat, 1982) formed in situ from abundant glaebules developed under advanced pedogenesis. However, this distinction may be obscured by mixing of modified and unmodified intraclasts during reworking. Lithological analysis of small intraclasts is difficult in the field, so that in practice it is often only possible to distinguish beds of grainstone (reworked granules) from grainstone pockets (cavity fills / in situ formation).

The two processes of grainstone formation would have occurred in environmental settings in close proximity. In situ grainstone pockets are formed during exposure in marginal lacustrine environments, while the beds of grainstone are laid down subaqueously but close to the shoreline and in shallow water.

3.3.12. Facies 12: Yellow limestones and yellow vuggy limestones.

Massive or vuggy limestones are found in many sections, as nodules/mottles, clasts, or as bedded units. This facies is closely associated with facies 13 cherts (fig 3.146) and most yellow limestones are assigned to the Rio Cabrera Member, the upper division

of the Rupelo Formation (see section 2.8; fig 2.40). Nodules of yellow limestone (fig 3.147) are also locally found just above the top of the Las Vinas Member (compare facies 13).

The yellow carbonates take a red stain and consist of fine microspar or micrite with angular-branching and rare circumgranular cracks (fig 3.148). The vuggy limestones (fig 3.149/3.150) consist of elongate ½ x 2mm calcite crystals arranged in an irregular mesh network. The abundant interstitial angular voids give the rock a highly porous texture.

Interpretation.

The microspar of the massive yellow limestones is the product of diagenetic recrystallisation. The yellow colour may record iron impurities incorporated in the carbonate lattice during dolomitisation and retained during later dedolomitisation.

Freytet & Plaziat (1982) reported dolomites and dedolomites from lacustrine and marginal lacustrine sequences in the Late Cretaceous - Early Tertiary of SW France. The dolomites there are associated with gypsum crystals and both were thought to have been formed by evaporation below littoral sabkhas.

Strasser & Davaud (pers. comm., 1985) encountered similar yellow mottling in the Purbeck of the Swiss and French Jura, noting an association with evaporites. The vuggy texture of the yellow limestones in the Rupelo Formation is consistent with evaporite solution. The assocation with cherts (themselves probably evaporite replacements, see facies 13, section 3.3.13) lends further support to this hypothesis, and points to a complex and probably interlinked diagenetic history.

3.3.13. Facies 13: Cherts.

I initially discovered chert within the Rupelo Formation in September 1985. The best exposure occurs 200m N of Rupelo at GR

(62418358), where a 1m thick chert bed outcrops W of the football pitch (figs 3.151 & 3.152).

Chert was subsequently found near the top of the Rupelo Formation at many localities, either in situ or as loose blocks. The thickness of chert is generally less than at Rupelo, but its resistant nature and distinctive appearance allow easy field identification.

The tracing of chert outcrops in the field is assisted by the frequent stratigraphic association of Rupelo Formation cherts with yellow limestones, yellow vuggy limestones and dark charophyte-rich biomicrites. This association permits the definition of the Rio Cabrera Member (section 2.8). Cherts are also reported here with a similar association and in an equivalent stratigraphic position in the Aguilar Fm near Lomilla, 150km to the NW (figs 3.153 & 3.154).

However, the presence of chert can not be used alone as a stratigraphic marker as thin chert horizons also occur locally at the top of the Las Vinas Member.

The chert is generally white but occasionally may be a translucent grey-blue reminiscent of flint. Blocks often have very smooth surfaces with angular terminations, and a 5-20mm scale "crystalline" fabric (fig 3.156). Square (?cubic) and angular voids 2-5mm in diameter and up to 10cm in length are also common (figs 3.157 - 3.160).

Cut slabs may reveal a fabric of 5-10cm rounded nodules (fig 3.155) which in thin section are seen to be made up of length-fast chalcedony and microquartz (fig 3.161). Angular voids are now mostly filled with length-fast chalcedony (figs 3.162 & 3.165). Minor quantities of length-slow chalcedony (quartzine) may also be present (figs 3.163 & 3.164). Examination of these former voids in plane-polarised light occasionally shows the retention of pale brown colour zoning (fig 3.162).

Patches of polygonal megaquartz also occur within the chert (fig 3.166). Rare lath-shaped micro-inclusions of a birefringent mineral

(?anhydrite) may also be found. The chert often includes 0.2mm diameter dolomite rhombs, commonly clustered (figs 3.162 & 3.163).

Silicification is commonly incomplete and it is locally possible to recognise a silicification front (figs 3.167 & 3.168) where the host sediment (often dark biomicrite) has been progressively silicified by the growth of 0.2mm siliceous spherulites. Relict fabrics preserved by the silicification include peloids (fig 3.168), "root crusts", (figs 3.169 & 3.170; see also facies 2 & 4), geopetally-filled cavities (fig 3.171) and charophyte stems (fig 3.172).

Interpretation.

Many of the cherts possess features indicative of evaporite replacement. These features are discussed below.

The crystal shapes and voids are interpreted as pseudomorphs after evaporite minerals (see Chowns & Elkins, 1974; review in Arbey, 1980). Similar crystal outlines have been illustrated by West (1964), Tucker (1976) and Milliken (1979). This interpretation is supported by the presence of length-slow chalcedony (quartzine), cited by Folk & Pittmann (1971) as clear evidence of sulphate replacement, especially when small inclusions of sulphate minerals such as anhydrite, gypsum or barite/celestite are also present.

Keene (1983) reported the occurrence of length-slow chalcedony from pelagic sediments in the N Pacific and warned that the presence of quartzine need not necessarily have firm implications about the environment. However, in view of the presence of peloidal and nodular fabrics (also reported by Folk & Pittmann (op. cit.) as common features of silicified evaporite sequences) and the tropical / subtropical open lacustrine / marginal lacustrine environmental setting envisaged for the remainder of the Rupelo Formation (see synthesis, section 3.6), the interpretation of many of these cherts as evaporite replacements is reasonable. Anhydrite has been reported from the Aguilar Formation (but only from the subsurface) by Leon

(1980). The original presence of anhydrite in the Rupelo Formation may also be supported by the rare occurrence of secondary selenite gypsum (fig 3.173), often generated by the uplift of anhydrite-bearing sequences (Tucker, 1981).

The relict fabrics show none of the features commonly associated with subaqueous evaporites such as depositional lamination, clear algal structures or evidence of wave or current activity. An origin for the now-replaced evaporites in a sabkha environment is therefore postulated. Formation would have occurred within the sediment pile as a result of the passage of saline ground waters during vadose or shallow-phreatic diagenesis. A change in the net water budget, perhaps brought about by increased aridity may have caused a drop in lake level eventually leading to conditions similar to those found in modern continental and coastal sabkhas (see examples in Kendall, 1984) thus resulting in the diagenetic emplacement of evaporites.

The timing of the silicification was probably relatively early, pre-dating the Pedroso Group deposits, for example. However, the source of the silica remains problematic. Two possibilities exist:

1) that the silica originated from inorganic sources, either from silica-rich ground waters percolating through the carbonate rocks after local dolomitisation or from volcanic glass recrystallised during diagenesis; or

2) that the silica originated from biogenic sources and was subsequently remobilised, concentrated and precipitated.

An association with volcanics is theoretically possible; breakdown of volcanic glass would also yield montmorillonite and chlorite and these may be present within green marl facies. However, of the inorganic sources, groundwaters seem to be a more likely candidate, especially in view of the presence of clustered euhedral dolomite rhombs included within the chert (ie probably contemporaneous with or predating silicification; see Geeslin & Chafetz, 1982).

Knauth (1979) argued that if silicification takes place in a narrow mixing zone and the flux of silica-saturated water is sufficiently large, then almost complete replacement of the limestone in the mixing zone may be achieved without silica input from biogenic sources. The volume of water required is enormous, as under normal pH conditions, the solubility of silica is very low (6-10ppm) relative to that of biogenic opal (120-140ppm). However, under conditions of high pH, the solubility of inorganic silica is increased. Eugster (1969 & in press) reported the occurrence of silica gels associated with magadiite in Lake Magadi (a highly alkaline lake), Kenya, while Peterson & von der Borch (1965) reported the formation of crystobalite in evaporating ephemeral lakes of the Coorong, S Australia after seasonal elevation of the pH because of high rates of photosynthesis (see also Colinvaux & Goodman, 1971; Wheeler & Textoris, 1978; Wells, 1983).

Although the environmental setting envisaged for the Rio Cabrera Member is somewhat similar to the modern Coorong, and despite the association of the cherts with dark biomicrites, a biogenic origin for the silica is still possible. Potential sources include sponge spicules (eg Wilson, 1966; Geeslin & Chafetz, 1982), radiolaria and diatoms, of which only diatoms are found in non-marine waters. None have been discovered within these rocks, although their very delicate structures might inhibit preservation.

Freytet & Plaziat (1982) postulated a structural control on silicification in the Late Cretaceous - Early Tertiary of SW France. They suggested that silicification was confined to anticlines, thus inferring that silicification probably post-dated Pyrenean tectonics. They also remarked on the possible rôle in diagenesis of saline solutions emanating from the Triassic. The thickest cherts in the study area occur at Rupelo, in a button anticline, and at Tenadillas, also in an anticline. Although the form of the anticine at Rupelo might suggest a halokinetic origin, the Tierra de Lara Group deposits

are exceptionally thick there, indicating very rapid subsidence at that time (see sections 8.11 & 8.12). The thicker cherts at Rupelo and at Tenadillas probably simply reflect greater original thickness of Rio Cabrera Member evaporites.

3.3.14. Facies 14: Charophyte dedolomite.

This facies occurs in the Rio Helechal section at Hortezuelos. It is described here with the Tierra de Lara Group facies, although Salomon (1982) included it in Group IIC (~Pedroso Group). The outcrop in question occurs stratigraphically above the Ladera Member carbonates and 15m below the oncoidal limestones of the Hortigüela Formation. The Ladera Member deposits are truncated at this locality by a coarse quartz grit (fig 3.174) which may be equivalent to the Torrelara Bed or may be a basal Pedroso Group sandstone. I have described the facies with the Tierra de Lara Group facies because it occurs well below the first appearance of oncoids and because of its diagenetic history (see below).

The facies consists of an alternation of 1m grey limestones (fig 3.175) and 2-3m blue-grey marly limestones containing 1-5cm iron concretions (fig 3.176). Both the marls and limestones have abundant charophyte stems - see fig 3.177. Charophyte nodes are well-preserved (fig 3.178 and see discussion of charophyte morphology, facies 7 and fig 3.110). The facies is notable for the presence of abundant poorly-staining rhombic crystals nucleated upon the charophyte stems and for the spectacular ferroan calcite cements which fill the central cavities of the stems (fig 3.179).

Interpretation.

The abundance of charophyte stems and the absence of pedogenetic features indicate deposition in quiet, shallow, open lake waters.

The diagenetic history is complex. The primary calcite of the charophytes apparently grew during early diagenesis and was then

partly replaced by dolomite. These now take a weak pink stain (fig 3.179), indicating partial dedolomitisation. The late-stage ferroan calcite cement (blue stain, fig 3.179) may be rich in iron released during the dedolomitisation process.

The iron concretions may also indicate remobilisation of iron and the former presence of sulphates (subsequently reduced to precipitate secondary sulphides).

The dolomitisation/dedolomitisation indicates a similar, perhaps partly contemporaneous diagenetic history to that of facies 12 & 13.

3.4. Diagenesis.

3.4.1. Introduction.

Much of this chapter has been concerned with pedogenetically modified carbonates. The diagenetic aspects have received relatively little attention. This reflects the dominance of micrite in the Rupelo Formation. However, some of the more significant results obtained from studying the cements in those facies containing appreciable quantities of spar are outlined below. The complex diagenetic histories of facies 12-14 have already received preliminary discussion and will not be further dwelt upon here.

3.4.2. Micrites.

Biomicrite facies do not provide a great deal of information about any subsequent diagenetic history when studied using optical petrography and staining techniques on account of their fine grain size and homogeneity. Under cathodoluminescence they show uniform dim luminescence. However, stable isotope studies (section 7.2) suggest that the biomicrites have indeed been affected by vadose diagenesis. The biomicrites show isotope values which indicate equilibrium with flushing meteoric groundwaters and they have probably been

recrystallised or cemented (although the fine grain size precludes optical observation of this).

Pedogenetically-modified facies are rewarding under cathodoluminescence as the constituent elements of composite glaebules and of areas of otherwise apparently homogeneous micrite are revealed, especially in facies 6 limestone conglomerates. These contain pale micrite clasts which may appear homogeneous under the optical microscope, but are revealed under cathodoluminescence as composites of many smaller intraclasts whose margins have been obscured by pedogenic modification.

3.4.3. Spar.

The only facies with appreciable spar are facies 4 (micrite with open-space structures) and facies 11 (grainstones). Optical petrography permits the recognition of five categories of calcite cements:

a) vadose silt (strictly internal sediment rather than cement);

b) early isopachous fringing;

c) pendent cement (vadose);

d) phreatic;

e) ferroan.

Vadose silt can be difficult to recognise in thin section when cavities in the micrite are completely filled by internal sediment and there is no spar present as there is little contrast in grain size between the vadose silt and the host micrite. However, the application of cathodoluminescence (see eg Grover & Read, 1984) allows the easy distinction (figs 3.180 & 3.181) between vadose silt (non-luminescent) and micrite (dim luminescence). According to Fairchild (1983), the intensity of luminescence depends upon the incorporation of transition elements into the carbonate lattice, especially Mn (stimulating) and Fe (quenching).

Pendent cements are very rare, but when present they are excellent indicators of vadose cementation (see Dunham, 1971; Grover & Read (1978); Longman, 1980). As Purser (1969) showed, pendent cements are produced within water droplets held by surface tension beneath larger grains. Fig 3.183 shows "drips" of sparry cement formed beneath glaebules / peloids in a facies 2 micrite. Note the different morphology of the cement from the spar of circumgranular shrinkage / rotation cracks (figs 3.55 & 3.56). The pendent cement of fig 3.183 post-dates the pedogenetic formation of the glaebules and an early isopachous fringing cement. Another example of an isopachous cement can be seen in fig 3.92, (where it clearly post-dates the vadose silt but pre-dates the late-stage ferroan cement).

Cathodoluminescence also facilitates the recognition of an early and a later generation of phreatic cement. Most cavities are filled with non-luminescent phreatic calcite, but the larger cavities were not always completely filled and were subsequently filled by luminescent and zoned fresh-water blocky cement crystals, as shown in a facies 11 grainstone (fig 3.182). Crystal zones correspond to compositional zones (see eg Frank, Carpenter & Oglesby, 1982).

Figs 3.184 - 3.186 show the phreatic cement infil of a large cavity within a facies 2 mottled limestone. Some larger spar crystals within cavities (figs 3.187 & 3.188; see also figs 3.58 - 3.60) have a palisade morphology similar to the speleothem columnar calcites described by Kendall & Broughton (1978). Cathodoluminescence clearly distinguishes the early non-luminescent palisade crystals from the later generation of luminescent, and spectacularly zoned, cement.

Ferroan calcite cements are particularly common in facies 12 & 14 (see above; eg fig 3.179).

Late stage veins (of probable tectonic origin) may be filled with coarse ferroan or non-ferroan calcite cements. Carbonate was dissolved along stylolites, especially in highly deformed areas.

3.5. Facies associations.

In addition to the more obvious examples mentioned in the text (eg of white cherts with yellow and vuggy limestones), there are many other common facies associations. Each association can be assigned to a particular environmental setting, thus permitting a better understanding of the interrelationships between different facies and between different environments. The facies associations are outlined in fig 3.189.

3.6. Tierra de Lara Group: Environmental synthesis.

Analysis of facies sequences and identification of facies associations permits us to build up more accurate integrated environmental models showing the variety and possible arrangement of sub-environments (fig 3.190: Senora de Brezales Formation, see section 3.2.4; fig 3.191: Rupelo Formation). While such models are of necessity somewhat simplified, they still provide informative representations of depositional system geometry.

The Rupelo Formation was deposited in a shallow hardwater lacustrine system. Ruby Lake, Nevada, (Dean & Fouch, 1983) would appear to be a good modern environmental analogue. Distal alluvial red marls and silts were deposited on alluvial plains around the lake. These occasionally encroached over large areas of the basin (eg Las Vinas Member). Localised carbonate sedimentation took place in ponds developed on the floodplain.

The Rupelo Formation carbonates are not generally laminated. Lake waters were unstratified and the sediments thus subject to bioturbation in oxygenated bottom waters. Carbonate was derived dominantly from biogenic sources. Calcified charophyte stems were the dominant contributors to carbonate production, although oogonia, ostracods, gastropods, and bivalves were also abundant. Charophyte stems are delicate and easily fragmented - they were only preserved in quiet, shallow water. Oogonia, in contrast, were easily transported

into deeper water, where bottom waters were more organic-rich and locally preserve weak (partly compactional) lamination. Some organic matter was land-derived, notably in the form of "black pebble" intraclasts of stabilised organic carbon (charcoal) transported (and locally concentrated) by streams from vegetated hinterland areas.

The fresh-water fauna / flora is only preserved in open lacustrine facies (eg M ambrillas de Lara Member). The lake was very shallow and frequent changes in water level led to very wide shoreline fluctuations. Marginal lacustrine facies (eg Ladera Member) were thus frequently subject to subaerial exposure, desiccation, root colonisation and pedogenesis. Pedogenic modification resulted in the loss of original fabrics and the development of mottling, sparry cracks, microkarst cavities, brecciation, surficial laminar crusts, and root crusts.

The marginal lacustrine marshy areas were colonised by abundant shallow-water and land plants which formed a fringing vegetated zone, thus effectively acting as a barrier to terrigenous clastic input. However, rare stronger stream influx resulted in the deposition of thin conglomerate and sandstone (Torrelara Bed). River-generated clastic input was probably also responsible for millimetre-scale fine graded silt laminae which are interpreted as the products of small-scale lacustrine turbidity currents.

Storms and lake transgressions caused reworking of intraclasts from marginal lacustrine into open lacustrine environments. During brief intervals of shoreline retreat, low-stage lake flat drainage channels caused the reworking and resedimentation of briefly-exposed open lacustrine facies. Grainstones of open lacustrine intraclasts were deposited on exposed shorelines.

During the later part of Rupelo Formation time (Rio Cabrera Member), periods of extreme desiccation led to the establishment of hypersaline conditions and the formation of evaporites. Evaporite

precipitation was associated with diagenetic dolomitisation / dedolomitisation and silicification.

3.7. Facies sequences.

The facies of the Tierra de Lara Group occur in a broadly predictable sequence, thus permitting the erection of the detailed lithostratigraphy (see section 2.8; fig 2.40). Fig 3.192 summarises the evolution of environments indicated by the observed sequence of facies. The Rupelo Formation, in particular, shows a very consistent facies evolution, as exemplified by the succession shown in fig 3.193.

In addition to the climatic factors outlined in fig 3.192, tectonic factors also exerted a strong control on the areal (see Chapter 8) and vertical distribution of facies. The relatively higher clastic supply in the Senora de Brezales Formation reflects more active tectonics and the exposure of "basement" lithologies on a pediment with significant fault relief. Clastic supply was much reduced during the deposition of the Rupelo Formation carbonates, reflecting less active subsidence, degradation of the pediment relief and draping of the carbonate facies across the the basin margins.

Facies sequences can also be recognised recording the influence of more minor tectonic and climatic variations. For example, influx of silt/marl into a carbonate-producing environment, as occurs several times in the Ladera Formation at Mambrillas de Lara, reflects the basinward progradation of terrigenous input. This could be the result of a climatically-controlled change in the net water budget and a consequent fall in lake-level, or of tectonically-controlled increase in clastic supply. In this case, the sharp change between limestones and marls suggests that tectonic factors were responsible. However, the abundant and repeated evidence of emergence in the Ladera and Rio Cabrera Members (eg fig 3.194) indicates rapid and frequent oscillations in lake-level and fluctuating shorelines. Such changes

may reflect tectonically-controlled variation in the location and catchment of stream input or climatically-controlled variation in the inflow-evaporation balance. In this instance, the frequency of desiccation points to short-term climatic influences being responsible.

Smaller scale facies sequences can also be observed. The most common recognisable sequences are upward increases in the degree of pedogenic modification, each sequence effectively representing a soil profile, (not necessarily complete). The situation is complicated by the superimposed effects of successive periods of emergence. The style of modification varies according to the lithology of the "host" sediment (fig 3.195).

Other common small-scale sequences include vertical variations observed in the size and abundance of intraclasts in biomicrite facies. The variations are interpreted as the product of storm reworking and/or changes in water depth. Some wackestone beds show irregular erosive bases with an upward decrease in the size and number of pale micrite and black intraclasts (figs 3.132 & 3.133). Similar sequences in the Late Cretaceous - Early Tertiary of SW France were interpreted by Freytet & Plaziat (1982) as evidence of transgressive reworking. However, in some cases (eg fig 3.181) there is no evidence of emergence (although this evidence could conceivably have been removed by erosion) and the large size of the intraclasts (up to 2cm) would seem to imply moderately energetic conditions. Heavy storm rains could facilitate transport of marginal lacustrine facies into more open lake areas. Stream inflow and storm waves could together account for localised subaqueous erosion in shallow water.

Fig 4.0. Pedroso Group. Rio Pedroso lies 10m to left of the road, which provides the best exposure of the Piedrahita de Muno Formation. Upper Cretaceous limestones forming the scarp (to the S of Barbadillo del Mercado) visible in the distance in storm clouds. April 1984.

CHAPTER 4. THE SEDIMENTOLOGY AND PETROGRAPHY OF THE (?UPPER BERRIASIAN - BARREMIAN) PEDROSO GROUP.

4.1. Introduction.

The Pedroso Group (Group IIC of Salomon, 1982) consists of two formations: the limestone-bearing Hortigüela Formation and the clastic-dominated Piedrahita de Muno Formation (see section 2.9; fig 2.40).

4.2. Hortigüela Formation.

The Hortigüela Formation may be compared lithologically with other Lower Cretaceous oncoidal deposits in Cantabria and the SE Iberian Chains (see section 2.9). The Formation outcrops chiefly in the NW of the study area, but there are also a few exposures in the SW, around Hortezuelos and Mamolar. In both sectors, the Hortigüela Formation rests upon the underlying Tierra de Lara Group with mild unconformity. The best outcrops occur in the NW, between Hortigüela and Quintanilla de las Vinas, where several sections through the Formation are available (eg fig 4.1).

Sedimentological study has facilitated the recognition of five main facies - marls, muds / clays, sandstones, nodular limestones and oncoidal limestones. These facies are described and interpreted below and an integrated facies model is presented (fig 4.2).

4.2.1. Facies 1: Marls / Silts.

Marls are the volumetrically most important facies in the Hortigüela Formation. Thick marls occur at the base of the Formation, forming the San Martin Member, and between the oncoidal limestones of the Zurramujeres Member (fig 4.1).

The marls at the base of the Formation are generally brick red, although they are locally green-mottled. The marls occurring between the oncoidal limestones are generally green and often include 5-10mm

oncoids and 1-5mm oncoid fragments. 1-5mm mottled limestone lithoclasts are also present; these exhibit internal structures indicative of emergence and pedogenesis (eg circumgranular cracks). The matrix consists of marly micrite with a few discontinuous fine angular-branching cracks and incipient circumgranular cracks. ½-1mm subangular quartz grains are also present in the matrix.

Interpretation.

The marls are interpreted as the products of fine clastic input in distal alluvial - lacustrine environments. The red marls are lithologically very similar to the red marls of the Rupelo Formation (section 3.3.1). The presence of mottling is indicative of pedogenesis and subaerial exposure, probably in a distal floodplain environment.

In contrast, the green marls are not similar to the green marls of the Rupelo Formation (section 3.3.5) which occur in much thinner beds and in different facies associations. The presence of many oncoid fragments and limestone lithoclasts floating in a marly matrix with included quartz grains and of some desiccation features would indicate deposition under relatively high-energy conditions, with significant terrigenous influx, such as might occur on a lake shoreline.

4.2.2. Facies 2: Muds / Clays.

Approximately 5m of dark grey-blue muds / clays occur above the Rupelo Formation limestones at Tejadas (see fig 1.6). These are interpreted as the fine-grained deposits of clastic-dominated lakes.

4.2.3. Facies 3: Sandstones.

In the NW of the study area, the Hortigüela Formation contains only rare sandstones. Thin, apparently tabular 10-20cm massive medium-grained sands are occasionally intercalated with the marls of the San Martin Member. These sands are interpreted as distal crevasse splay deposits. Thin, fine-grained and micaceous yellow sands also

outcrop locally between the oncoidal limestones of the Zurramujeres Member.

In the S, thicker (up to 5m) trough cross-bedded sands occur in the Hortigüela Formation, eg at Tejadas, near La Gallega, where they locally contain rare 5-20mm wood fragments. These sands are interpreted as in-channel deposits.

4.2.4. Facies 4: Nodular limestones.

Mottled, impure limestones outcrop very occasionally in the Hortigüela Formation. These limestones often exhibit prismatic fabric (figs 4.3 - 4.4). The nodular limestones apparently grade upwards from green marls, but are often truncated by overlying oncoidal carbonates (figs 4.3 & 4.4).

The mottling and prismatic fabric are indicative of subaerial exposure and pedogenesis. The nodular limestones are interpreted as pedogenic carbonates which developed on marl substrates in periods of emergence during low lake-stands.

4.2.5. Facies 5: Oncoidal limestones.

In the NW of the study area, the Hortigüela Formation contains locally as many as six oncoidal limestone units. The oncoidal limestones are from 1 to 4m in thickness. The beds are intercalated with marls / silts and sandstones and are often traceable along strike for several hundreds of metres, although they are discontinuous on a larger scale. Lateral facies changes are particularly evident across the NE-SW faults which dissect the Hortigüela - Cuevas de San Clemente anticline - the thickness of limestone beds and the size of oncoids within the limestones may vary from one side of a fault to another, for example.

At the base of the units, the limestones comprise micritic wackstones with ½-1mm black intraclasts and rare irregular 1-2cm diameter oncoids. These pass upwards into packstones with abundant

2-3cm oncoids and oncoid fragments. Irregular and angular micritic intraclasts up to 4cm across may also occur (fig 4.5). Cross-bedding is occasionally developed (figs 4.6 - 4.10). Larger oncoids (eg figs 4.11 & 4.12) occur locally at the tops of beds, some occasionally reaching 20cm in size. The larger examples are usually more irregular.

Morphology of oncoids.

The terms "pisolite" and "pisoid" have generally been used to describe inorganically accreted coated grains (cf "caliche pisoids", "vadose pisolites") while the terms "oncolite" and "oncoid" are more generally associated with an organic (and frequently an algal) origin. According to Hofmann (1969), "An oncoid is a stromatolite with centrifugal growth vectors and encapsulating laminae formed around an intermittently mobile nucleus".

The coated grains of the Hortigüela Formation have been variously described as "pisolites" and "oncolites". This inconsistency of nomenclature is a common feature of the literature on coated grains in other parts of the world. For example, Jones & Wilkinson (1978) and Murphy & Wilkinson (1980) described "algal pisoids" from marl lakes in Michigan. Other previously used names have included: "algal nodules","algal balls", and "algal-encrusted grains".

The work of Peryt (1983) has done much to clarify this problem by recommending the use of "vadoid" for individual grains formed inorganically as a result of vadose processes (eg "vadose pisolite") and "vadolite" for rocks formed of vadoids; while the terms "oncoid" and "oncolite" are reserved for examples involving an organic origin. Many oncoids formed as a result of the growth of blue-green algae (or: "cyanophyta"; "cyanobacteria"). Riding (1983) recommended the use of the term "cyanoid" for these oncoids. Morphological criteria useful for the distinction of oncoids from vadoids are set out in fig 4.13.

The oncoids in the Hortigüela Formation are generally spheroidal and may be described as SS-type stromatolites under the classification system of Logan et al (1964). Further classification is possible on the basis of internal stacking into SS-I, SS-C and SS-R types (fig 4.14). SS-C oncoids are by far the most common in the Hortigüela Formation, although some oncoids display micro-unconformities in the growth layers, leading to their classification as SS-R type.

Oncoids are generally from 1 to 4cm in size, and are usually spherical or sub-spherical. Larger oncoids from 5 to 20cm in diameter may also be found, particularly towards the top of beds (see section 4.2.5). These larger examples tend to be more irregular (eg fig 4.11).

The internal structure of the oncoids consists of alternating smooth, concentric layers of dense, structureless carbonate and crinkly layers of more porous, spongy micrite characterised by the presence of radial tubiform algal filaments. Alternations of dense layers with porous layers containing this radial fabric result in the laminar concentric component of oncoid structure that is clearly evident in the field (figs 4.15 & 4.16).

The algal filaments include both unbranched single filaments and branched or fan-like clumps (these are especially common in the larger oncoids; figs 4.17 - 4.19). On comparison with the descriptions in Wray (1977) these seem similar to the blue-green algae Girvanella and Cayeuxia respectively (although Monty & Mas described similar algal structures in Lower Cretaceous oncoids from the SE Iberian Chains, which they compared with the modern blue-green alga Phormidium). However, identification of the algae is difficult since algal filaments decay quickly, often soon after the formation of a calcareous microcrust around them. As a result, the porous layers of the oncoids are often only partially calcified. Cavities caused by the decay of algal filaments are later replaced by void-filling spar and early diagenetic sparitisation (as recognised by Nickel, 1983) may also

result in a loss of original fabric. Even in the case of modern oncoids from Lake Constance (Schäfer & Stapf, 1978), oncoid microstructures can only with difficulty be related to the algae which formed them.

The porous layers are 0.3 to 1mm in thickness, while the denser layers may be concentrically laminated on a 10-50μ scale. The layers are irregular in detail, and knobbly "microstromatolitic" (Monty & Mas, 1981) type club-shaped protrusions are particularly common towards the centre of the oncoids (fig 4.20). This morphology may represent a "pioneer growth" - later, as the protrusions grow in size they eventually meet and coalesce to form concentric laminae.

Monty & Mas (1981) interpreted the denser layers as the products of tiny prostrate filaments similar to those of modern Schizothrix. When these algae bloom, they rapidly produce a series of superposed micritic films which may later compact to form a simple micritic lamina.

The oncoids are nucleated on a variety of structures, including green marl clasts, black intraclasts (fig 4.5), oncoid fragments (fig 4.11), silt grains and shell fragments.

Oncoidal micrite is grey-brown to orange in thin section. Organic matter is common in the branching filaments and in some denser layers. The smaller oncoids occasionally contain a little (?trapped) silt or clay.

Interpretation and discussion.

Blue-green algae (cyanophyta) today live in fresh-water environments; they also colonise marine environments but have been partially displaced by the red algae (rhodophyta) from the Eocene onwards (Monty, 1972). Although the spongiostromate oncoids of the Hortigüela Formation are morphologically comparable with modern fresh-water oncoids, the presence of cyanophyta can not be used as a reliable indicator of fresh-water conditions in the ancient.

Nevertheless, algal carbonates do indicate a shallow-water environment (within the photic zone). Algal carbonates are generally restricted to near-shore areas with low clastic input and with relatively few algal grazers (ie often in fresh-water or hypersaline conditions). The association of the oncoidal limestones with continental facies - crevasse splay sands, "overbank" muds and palaeosoils - is suggestive of a fresh-water environment (as, perhaps, do the isotope data; see discussion, section 7.2.5.2). Oncoids have been reported from both lacustrine and fluvial environments (Freytet & Plaziat, 1965, 1982; Weiss, 1969; Sbeta, 1976; Ordonez & Garcia del Cura, 1978; Schäfer & Stapf, 1978, Nickel, 1982; papers in Peryt, 1983).

Opinions differ as to whether oncoids require agitated water conditions for their development. Some authors (including Logan et al, 1964) have argued that the development of concentric structure in SS-C type oncoids must indicate frequent rolling episodes to facilitate more or less symmetrical and equal growth, while more irregular structures in SS-R and SS-I type oncoids are indicative of long periods of stasis. Schäfer & Stapf (1978) noted that modern oncoids in Lake Constance are larger and more abundant nearer to the course of currents induced by stream inflow from the Rhine River, which certainly suggests that agitation favours oncoid growth. However, if the growth of algae is very slow, then perhaps only very occasional disturbance caused by bioturbation, or by storm events, might be sufficient. Jones & Wilkinson (1978) reported colonisation by blue-green algae in situ on the underside of oncoids, so that rolling may not necessarily be a requirement for oncoid development at all.

In the case of the Hortigüela Formation, the association with packstones, the occurrence of very coarse intraclasts, and the presence of cross-bedding, are all strongly indicative of relatively high energy deposition. Water agitation was probably intense, at least periodically, either as a result of storm waves or because of powerful

river-generated currents in times of high stream inflow. The sheet-like nature of the oncoidal limestones in the Hortigüela Formation probably indicates deposition on a lake flat or shoreline.

Although algal growth and photosynthesis involves the precipitation of carbonate, Pentecost (1978) has suggested that the dominant mechanism of oncoid growth is by carbonate accretion as a result of the trapping of detrital carbonate on algal mucilage during agitation. The dense and porous carbonate layers may be partly the result of seasonal fluctuations in biogenic production, the changing conditions reflecting variation in temperature, lake level or salinity. Porous layers are the result of algal growth, which may be restricted to summer or to periods of high lake stand, while the denser layers may record either largely inorganic accretion by precipitation of carbonate during cooler or more evaporative periods, or carbonate trapping during storm events.

An alternative suggestion was offerred by Sbeta (1976) who suggested that algal growth requires minimum water disturbance. Hence sedimentation by encrusting algae occurs only during quiet periods, while under more agitated conditions radial filaments can not form and instead carbonate sedimentation takes place both by growth of algal filaments parallel to the surface and by trapping of detrital carbonate.

4.2.6. Environmental synthesis for the Hortigüela Formation.

Although the exact mechanism of oncoid growth remains somewhat uncertain, it seems likely that the oncoidal limestones do represent fresh- and shallow-water conditions in waters that were at least intermittently agitated. Fig 4.2 shows an integrated facies model for the Hortigüela Formation. Significant reworking of oncoids formed in littoral zones could lead to the deposition of oncoidal wackstones in more open lacustrine conditions and of oncoidal packstones on shorelines and shoals. Particularly large oncoids may have formed

under slightly higher energy conditions, such as those occurring on exposed lake shorelines or near to the course of inflowing water currents. Local current action also led to the occasional development of cross-bedding in the oncoidal units.

The erosive bases of the limestone beds indicate that the onset of carbonate deposition was generally associated with fairly rapid lake transgression in each case. The upward passage from wackstones, with small oncoids, to packstones, with abundant and locally large oncoids, is the opposite of that noted by Nickel (1983), who reported a fining-upward of oncoid size associated with successive deepening cycles in channel-associated oncoidal carbonates from the Eocene Guarga Formation of the S Pyrenees. This suggests that the sequence within each limestone unit in the Hortigüela Fm actually represents a progressive shallowing. This shallowing might reflect gradual shoreline retreat because of increased aridity. Alternatively, the shallowing might record the infilling of any original topographic depression with carbonate sediment to produce a laterally extensive and very shallow lake flat. Such a setting would have been ideal for the growth of large oncoids.

Clastic supply was generally limited during the deposition of the carbonate units, but the relatively thin nature of the oncoidal limestones demonstrates that during times of more intense tectonic activity, of more clastic stream supply or of more humid climate, the lake flats were periodically subject to input of fine terrigenous detrital material from encroaching fluvial systems.

4.3. Piedrahita de Muno Formation.

The clastic-dominated Piedrahita de Muno Formation occurs chiefly in the NE of the study area (between Piedrahita de Muno and Neila and in the Canales de la Sierra - Villavelayo region) and in the SE (between Mamolar and Talveila). The Formation is generally found

where the Hortigüela Formation is absent. The two may be partial lateral equivalents (see section 2.9).

In the NE, the Piedrahita de Muno Formation is extremely thick (1-2km). However, exposure is often poor, especially of the finer-grained facies. Correlation is very difficult on account of the great thickness, poor exposure and the lack of easily recognisable marker horizons.

The most informative outcrops for sedimentological study occur at the following localities (refer to fig 1.6):

a) on the NW side of the road from Piedrahita de Muno towards Vizcainos.

b) on the NE side of the road from Castrovido towards Terrazas.

c) to the E of the village of Moncalvillo.

Analysis of these sections has facilitated the recognition of a number of facies. These are described below and an integrated facies model is presented (fig 4.21).

4.3.1. Thick sandstone units.

The Piedrahita de Muno Formation contains numerous sand units 1-5m in thickness (eg figs 4.22 - 4.23). Although they frequently have mildly erosive bases, when exposure along strike is available (as at Moncalvillo) it is evident that the sandstones are laterally extensive and not significantly channelled.

The sands are generally medium-grained, and consist dominantly of well-sorted subangular cracked quartz grains ½mm or so in size (fig 4.24). Feldspar and white mica are also present, as are variable proportions (up to 1%) of ½-1mm detrital mafic grains.

Trough cross-bedding is locally present (fig 4.25) and larger scale cross-beds also occur (fig 4.26). Between Moncalvillo and Castrillo de la Reina, there is a large outcrop offering three-dimensional exposure of trough cross-bedded sands (figs 4.27 -

4.29). I have included these sands here with the Piedrahita de Muno Fm, although both the thickness of the sands and the large scale of the cross-bedding are rather unusual. In fact, these sands may belong to a higher part of the stratigraphy (eg Salas Group or equivalent).

Sandstones in the Mamolar - La Gallega region (fig 1.6) are particularly rich in quartz pebbles. In the N of the study area, there are a few cross-bedded paraconglomerate beds with abundant rounded 5-20mm quartz pebbles near the base of sandstone units (fig 4.30). Rare solitary and clustered quartz pebbles also occur near the base of sandstones, as do 10-30mm mud "rip-up" clasts and flakes (generally weathered out and visible in outcrop as 1-2cm sub-rounded voids). 5-20mm coaly black organic fragments are also present (and sometimes abundant) at channel bases in a few localities. Finer vegetal débris may drape sets and thus pick out cross-laminae (fig 4.31).

Many of the sandstones are composite, often consisting of numerous stacked and locally plane-laminated tabular beds 20-50cm in thickness (eg fig 4.30). However, one 3m thick sandstone bed at Piedrahita de Muno (fig 4.32) consists of massive coarse and poorly-cemented friable sand. Higher in the same section, there is a 5m thick medium-grained sandstone (figs 4.22 & 4.23), which is generally massive with only minor tabular cross-bedding at the base and an erosive-based reactivation surface with minor pebbly lags.

Interpretation.

The massive coarse-grained sand described above from the Piedrahita de Muno section (fig 4.33) occurs between two lacustrine limestones (see section 4.3.6) and may represent rapid subaqueous clastic deposition in a marginal lacustrine environment - possibly sediment dumping at the mouth of a distributary channel (fig 4.21).

The thick massive sand at the top of the Piedrahita de Muno section (figs 4.22 & 4.23) may represent a channel-fill deposit. Elsewhere, the local presence of cross-bedding indicates that

organised semi-permanent flows were occasionally set up, although there is no evidence for lateral channel migration.

Most of the other sand units are composite, only mildly erosive at their bases, and apparently tabular, which suggests that they do not represent channel deposits. Collinson (1978) suggested that where sand bodies are sheet-like with no observed lateral margins, they may result from major sheet floods or be the result of extensive channel migration. There are no lateral accretion surfaces (epsilon cross-sets), and this effectively rules out channel migration, so that composite sand bodies may, in fact, represent successive rapidly-deposited sheet flood events. Sheet floods carry sediment both as bed load and in suspension.

Plant débris and rip-up mud clasts were picked up by the floods as they passed over the vegetated muddy floodplain.

4.3.2. Thinner sand units.

The thicker sands described above outcrop well, even in areas of limited exposure. When outcrop conditions are good, it is also possible to recognise many thinner sands. These fall into two categories:

1) Lozenge-shaped beds of massive, mid-brown medium sand generally 1m or so thick and wedging out laterally over 5-10m (figs 4.34 & 4.35).

2) 10-50cm beds of medium-fine grained, mid to dark brown sand. Some of these wedge out laterally over 2-3m and are thus discontinuous on the scale of an outcrop (fig 4.36). More commonly they are tabular, with sharp (but not notably erosive) bases, flat tops and locally some plane lamination. A few of the 50cm beds have sharp, undulating bases and are weakly mottled.

Polygonal cracks are locally developed on the tops of sandstone beds (fig 4.37), as are vertebrate prints, which were recorded at one locality 2km NE of Salas de los Infantes (figs 4.38 - 4.40). The

prints are each 20cm in length and are preserved in positive relief. They were made by a large animal (probably a dinosaur such as Iguanodon) with bipedal gait and a stride of 1.7m. The prints are similar to those described from Triassic red-beds of South Wales by Tucker & Burchette (1977). The prints reported here show three pointed toes and a foot outline which contrasts strongly with the prints recorded in the Rupelo Formation (see section 3.3.7; figs 3.120 - 3.122). Whereas the rounded outline of footprints from the Rupelo Formation suggests adaptation to support the weight of a heavy animal living on a soft substrate, the more pointed outline of footprints from the Piedrahita de Muno Formation and the bipedal gait probably indicate adaptation of a much less bulky animal to more rapid progress over firmer substrates on land.

Interpretation.

The tabular sands represent discreet sedimentation events. Deposition took place initially as bed load and later out of suspension. Subsequent bioturbation (chiefly by burrowing) resulted in a loss of primary sedimentary structures (probably mainly plane-lamination) and the introduction of some mud, giving the sands their characteristic dirty brown, locally mottled, appearance at outcrop. The polygonal cracks mark rarely-preserved desiccated top surfaces.

Rapid emplacement of the sands over soft wet muds led to the development of density inversion and the local formation of load structures (fig 4.41).

The tabular sands might be interpreted either as storm-induced distal sheet-floods or as crevasse splay deposits. Their tabular geometry is consistent with either interpretation. The lensoid massive sand bodies are interpreted as minor abandoned channel fills - their morphologies are similar to examples of crevasse channels shown by Stear (1983) and Tyler & Etheridge (1983).

4.3.3. Mudstones & siltstones.

Mudstones and siltstones together probably make up 75% of the vertical section at Piedrahita de Muno. Outcrop is relatively poor, even in road sections. The muds are usually dark brick-red, but are locally green-mottled (fig 4.42). They are frequently structureless, but do contain calcareous tubular structures with irregular surfaces. These are 5-10mm in diameter and up to 3cm in length. Occasionally, the mudstones exhibit fine, white to grey crack-like structures (figs 4.43 & 4.44). These are particularly common (and occasionally abundant) in slightly more indurated muds and siltstones.

Interpretation.

The mudstones and siltstones are interpreted as distal alluvial suspension deposits. The absence of sedimentary structures is the result of intense bioturbation (by root action and burrowing). The more indurated horizons displaying crack-like structures are interpreted as immature hydromorphic soils with weakly calcified desiccation cracks and fine root networks. These features developed during periods of emergence.

4.3.4. Nodular carbonates.

The mudstone facies of the Piedrahita de Muno Formation are characterised by the development of weakly mottled, but strongly nodular, carbonate horizons 5-50cm in thickness (figs 4.45 - 4.47). These have gradational bases and normally show upward increases both in carbonate content and in the degree of brecciation. Prismatic fabric is weakly developed locally (figs 4.48 & 4.49).

In thin section (fig 4.50), the carbonates have a fine microsparry or micritic matrix and display a variety of pedogenic features including development of "peloids" / glaebules, and of circumgranular and angular-branching spar-filled desiccation cracks. The limestones are relatively impure - they have abundant included

0.1-0.5mm quartz grains. These frequently show irregular, embayed margins and often have a thin external coating of carbonate cement.

Interpretation.

The nodular limestones are interpreted as pedogenic carbonates which were precipitated largely inorganically within fine-grained alluvial deposits.

In alluvial settings, pedogenic carbonate is derived in solution both from stream run-off and groundwaters, and is precipitated by evaporation out of fluids moving upwards because of capillary action or from water migrating downwards from the surface by percolation. The process clearly requires the intermittent presence of fluids within the soil profile, and thus precipitation is favoured by fluctuation of the water table. However, water tables need not necessarily be permanently high. Flash-flood events may lead to the temporary establishment of perched water tables (fig 4.51), so that pedogenic carbonates may develop even in semi-arid climatic régimes.

Carbonate is likely to be dominantly replacive (Klappa, 1979). The presence of floating and "exploded" detrital grains was cited by Watts (1978) as evidence for displacive growth of calcite, but, in fact, growth of replacive carbonate preferentially along cracks will tend to produce similar textures.

4.3.5. Limestones.

Limestones are very rare within the Piedrahita de Muno Formation, and thus may be useful for correlation purposes (see section 2.9). There are two main categories:

a) biomicrites.

Approximately 5m of mid- to dark grey biomicrites containing ostracods, gastropods, charophyte stems and oogonia (fig 4.33; fig 4.52) occur at one horizon within the Piedrahita de Muno Formation.

The limestones have mudstone and wackstone textures with a few black intraclasts and some spar-filled cracks and tubes. Outcrop conditions are poor.

Approximately 7m of dark biomicrite occurs in the Terrazas - Castrovido road section. This limestone is lithologically similar, although possibly slightly darker.

Interpretation.

These limestones contain fresh-water fauna/flora and display some evidence of minor emergence. They are interpreted as open lacustrine carbonates which were deposited in shallow, unstratified lakes subject to occasional brief desiccation. Carbonate was dominantly derived from biogenic sources, probably mostly from the break-up of delicate organic structures such as charophyte stems.

b) marly micrites.

One 1.3m thick pale grey, weakly mottled and slightly impure limestone with vertical prismatic fabric occurs in the Piedrahita de Muno section (fig 4.53). The limestone is impure and grades up from green marl.

Fine angular-branching and sparry cracks are locally present.

Interpretation.

This limestone is interpreted as a fresh-water marginal lacustrine carbonate. It was originally deposited in shallow water and was later exposed subaerially and subject to root penetration and desiccation (see discussion of prismatic fabric, section 3.3.3a).

The occurrence of the open lacustrine biomicrite beneath a 5m thick massive sandstone and an overlying marginal lacustrine limestone suggests that a sharp influx of coarse clastic material into the lake rapidly halted carbonate sedimentation, which was

subsequently renewed in shallower water after the dumping of 5m of coarse sand.

4.3.6. Facies 6. Green silts /marls.

A 1.3 m thick bed of green marl occurs immediately beneath the marly micrite bed in the Piedrahita de Muno section. Fissile green silts up to 2m in thickness occur interbedded with tabular (and occasionally loaded) sand units. The silts and marls are apparently unfossiliferous.

Interpretation.

The green marls and silts display no primary sedimentary structures, but, in view of their fine grain size, it is likely that they represent suspension deposits. Their green colour may indicate deposition in conditions that were not strongly oxidising. They are interpreted as the fine-grained deposits of clastic-dominated lakes.

The absence of sedimentary structures in lacustrine deposits suggests that they have been bioturbated, and hence that they were probably deposited in fresh-water (Dean & Fouch, 1982). The green silts and marls were probably deposited either in water too deep for biogenic carbonate production or (more probably) in areas of relatively elevated clastic supply. This interpretation is supported by the association of the green marls with impure marginal lacustrine carbonates (see above; section 4.3.5b).

The sands were deposited rapidly onto soft, wet mudstones. Density inversion led to the development of load structures.

4.3.7. Environmental synthesis for the Piedrahita de Muno Formation.

The mixture of relatively rare cross-bedded and channel deposits with stacked "multi-storey" sheet flood sands found in the Piedrahita de Muno Formation is typical of terminal rivers (Friend, 1978; see also Tunbridge, 1981) or distal/terminal alluvial fan settings

(Mukherji, 1976; Heward, 1978; Parkash, Aswathi & Gohain, 1983), such as those found in basins of internal drainage. Modern analogues include central Australia (Lake Eyre) and central North Africa (Lake Chad). These areas receive only sporadic rainfall, but rains are highly irregular and occasionally torrential. Streams are consequently strongly ephemeral; long periods of drought with no stream flow are punctuated by very heavy rains which lead to catastrophic short-lived flash flood events with very high stream discharge rates.

Sheet flood sands are deposited at the distal end of distributary channels as flows expand laterally, spreading out over vast, often sparsely vegetated, plains of little relief that at other times are characterised by fine-grained distal alluvial and lacustrine deposition. Plane-laminated, and, locally, (see eg Williams, 1971) even trough cross-bedded sands are deposited at high flow stage. Subsequent rapid waning of the flow leads to the deposition of thin layers of mud during the final stages. More proximal areas receive a higher proportion of "in-channel" and discreet sheetflood sands, while in more distal areas sand bodies interfinger with fine-grained suspension deposits (see sections 4.3.2 & 4.3.3).

The presence of limestones and green siltstones/marls in the Piedrahita de Muno Formation indicates that lakes were occasionally developed in more pluvial periods, just as they are in modern internal drainage basins. The input of coarse clastic material during times of elevated clastic supply led to the rapid sedimentation of coarse sand dumped in bars at the mouths of distributary channels (see fig 4.19). When clastic supply was reduced, green silts, marls and grey biomicrites were deposited in shallow, oxygenated fresh-water. Carbonate was derived chiefly by the break-up of calcified charophyte stems, although oogonia, ostracods and molluscan débris also contributed to the carbonate supply. The limestones were later subjected to pedogenesis as a result of emergence during periods of shoreline retreat. This resulted in the modification of original

fabrics. Desiccation features and root penetration fabrics were locally developed.

The interpretations given here for the Piedrahita de Muno Formation differ from the most recent ideas put forward for its lateral equivalent in S England, the "Wealden", by Stewart (1983 and in prep). Stewart reported evidence for lateral migration of channels, for more perennial, stable fluvial systems, and for a relatively more humid climate. Although the presence of channel sandstone bodies in the Piedrahita de Muno Formation demonstrates that networks of semi-permanent distributaries were occasionally set up, the dominance of sheet flood sandstones is more consistent with intermittent sedimentation. Catastrophic, but short-lived, sedimentation events controlled by periodic tectonic activity or storms led to the rapid deposition of stacked sheet sandstone units. Distal alluvial mudstones and pedogenic carbonates were developed in the intervening periods.

Fig 5.0. Soil profiles developed in muds and silts of the Salas Group, overlain and truncated by sharply based channel sands. (See fig 5.3 for closer view). Note vertical and horizontal elements in soil fabric as a result of sheet-like iregular carbonate unit and vertically oriented carbonate nodules. N side of new road between Quintanilla de las Vinas and Lara de los Infantes, approximately 1km SE of Quintanilla de las Vinas.

CHAPTER 5. THE SEDIMENTOLOGY OF THE (?BARREMIAN - APTIAN) SALAS GROUP.

5.1. Introduction.

The Salas Group (Group IV of Salomon, 1982) comprises a thick sequence of conglomerates, sandstones, muds and rare limestones. Its distinction from the Utrillas Formation (Group IV of Salomon, 1982) is facilitated in the field by the criteria set out in fig 5.1.

The Salas Group reaches a maximum thickness of approximately 500m in the vicinity of Salas de los Infantes. At least 300m is present in the NW sector between Hortigüela and Quintanilla de las Vinas. The Salas Group is thin, and locally absent SW of the San Leonardo Fault, because of truncation beneath the overstepping Utrillas Formation (see section 8.14).

Outcrop is often extremely poor. Mapping on a regional scale often relies on soil colour (characteristic rich brown) and isolated conglomerate and calcrudite horizons (fig 5.2). Indeed, although the sequence is probably dominated volumetrically by muds and sands, it is the conglomerates and carbonates that normally outcrop in the field.

5.2. Facies.

As in the case of the Utrillas Formation, (see Chapter 6), detailed correlation is not possible within the Salas Group itself because of the discontinuous nature of the exposure. However, study of the available outcrops has facilitated the recognition and description of four main facies: conglomerates, sandstones, muds and calcarenites / calcrudites. These facies are interpreted below and an integrated facies model is presented (fig 5.3).

5.2.1. Conglomerates.

Conglomerates of the Salas Group are present in 1-2m beds that have mildly erosive bases and are sometimes lenticular on a 5-20m

scale. The conglomerates contain a variety of clast lithologies (fig 5.4) including quartzite pebbles (dominant) and limestone / sandstone intraclasts (less abundant). The subangular and poorly-sorted clasts are set in a coarse, sandy matrix. Cross-bedding is occasionally developed, and is often picked out by strings of larger pebbles.

Interpretation.

The conglomerates record high-energy fluviatile reworking of the subjacent pediment. The presence of clasts derived from the marine Jurassic sequence indicates erosion of the Tierra de Lara and Pedroso Groups beneath. The presence of coarse quartzite clasts suggests erosion of the Palaeozoic massifs. Faulkner (1985) noted that, around Quintanilla de las Vinas, limestone clasts were especially coarse and abundant to the NW of a NE-SW fault that apparently influenced sedimentation. The possible rôle of tectonics at this time is discussed further in section 8.14.

The conglomerates are only rarely cross-bedded or channelled, and generally have only mildly erosive bases. This suggests that they were deposited relatively rapidly, perhaps as sheet-flood deposits on an alluvial plain of only local, fault controlled, relief.

5.2.2. Sandstones.

Sandstones in the Salas Group outcrop only occasionally. They are coarse, locally quartz-pebbly, and occur in 1-2m beds. The sands are often cross-bedded and mildly erosive-based, but may be more massive and sheet-like, commonly truncating mudstones (facies 3; see figs 5.0 & 5.5).

Grains are angular to subangular and moderately to well-sorted, consisting of approximately 85-90% quartz and 10% K-feldspar, with locally up to about 2-5% mafic minerals, these apparently chiefly comprising weathered and altered biotite (or tourmaline?).

Interpretation.

The sandstones record lower energy fluviatile sedimentation. The cross-bedded units indicate the migration of channel floor dunes under more organised flow conditions, while the sheet-like units probably represent more distal or lower energy sheet-floods.

5.2.3. Muds / silts.

These are generally dark red to brown in colour and may be locally green-mottled. Carbonate nodules are often present, especially towards the tops of mud units. The nodules are vertical, subcylindrical and up to 5cm in length (figs 5.0 & 5.5). Occasionally, more tabular carbonate units may occur. These carbonates are impure and are often intercalated with red or green marl. In thin section, circumgranular and angular branching cracks are visible in the micritic matrix.

Interpretation.

The muds/marls are interpreted as suspension deposits from interchannel / overbank areas. The nodular carbonates and mottling are products of pedogenesis - the nodules are probably developed around root systems. The precipitation of carbonate may have been facilitated by changes in sediment chemistry or fluid hydrology around the roots. Carbonate precipitation also probably depends upon upward leaching and downward percolation of carbonate associated with fluctuation of (possibly perched) water tables and repeated desiccation and flooding.

The tabular carbonate units might possibly record minor lacustrine sedimentation. However, the fabrics are strongly indicative of pedogenesis and it is difficult to distinguish with certainty between impure and pedogenically-modified marginal lacustrine carbonates and thicker pedogenic carbonate units (eg "sheet calcrete" of Klappa, 1978) developed within a soil profile.

5.2.4. Calcarenites / calcrudites.

These are commonest at the top of the sequence (figs 5.2 & 5.6) and occur in beds up to 1m thick, occasionally displaying cross-bedding. The grains are generally ½ to 1mm in size and are angular to subangular. Calcrudites contain a high proportion of reworked and fragmented oncoids (fig 5.7 - 5.11) with sizes ranging from 1-3cm (whole oncoids) to 1mm (fragments). Angular to subangular 0.5 to 1mm quartz grains are also often abundant.

Interpretation.

The calcarenites and calcrudites record the localised high-energy reworking of Hortigüela Formation oncoidal limestones. The oncoids are easily transported because of their rounded shape, and they appear to be relatively resistant to weathering on account of their organically-bound structure. Modern erosion sufaces and soil areas between outcrops of Hortigüela Formation limestones are often carpeted by oncoids, which could easily be ripped up as clasts and concentrated by rapid flood events.

5.3. Environmental synthesis.

The Salas Group clearly records continental sedimentation in a fluviatile environment. The high proportion of mud is in great contrast to the overlying Utrillas Formation (see section 6.5). In fact, floodplain muds and marls are probably at least as important volumetrically in the Salas Group sequence as conglomerates, sandstones and calcrudites.

The conglomerates and sands are often lenticular on a large scale, but are not deeply channelled. Certainly, there is little evidence of channel migration. The (local) presence of channels and trough cross-bedded sands, and the elevated proportion of mud indicate deposition towards the frontal zone of large, gently sloping alluvial fans or in terminal rivers (Friend, 1978) spreading out into the

central parts of a continental basin with little relief. The input of coarse clastic material probably occurred sporadically in rapidly-deposited sheet-flood units which spread out onto a muddy and occasionally poorly-drained floodplain during rare storm events or periods of tectonic activity. In a study of a modern catastrophic fan deposit, Tunbridge (1983) reported that conglomerates with boulder-sized clasts can pass laterally into fine sands and muds over as little as 1.5km, so that the lateral extent of conglomerate units onto a wide, flat basin floor might well have been limited. Extensive pedogenesis acted to modify the floodplain sediments during the intervening times. An integrated facies model for the Salas Group is presented in fig 5.3.

As in the case of the Piedrahita de Muno Formation of the Pedroso Group (see Chapter 4), the Salas Group does not closely match the classical "braided" or "meandering" river models. Seasonal climate and sporadic tectonic activity characterised both sequences, and neither of these factors are likely to favour the permanent establishment of stable fluvial systems. There are some similarities between the model shown in fig 5.3 for the Salas Group of the W Cameros and the "muddy braidplain" model developed by P. Allen (1981) for the Wealden of S England, which was envisaged as an example of tectonically-controlled but low-energy alluvial sedimentation in a moderately humid climate.

Fig 6.0. Utrillas Formation sands in road cut, location of sand bars. Typical heather cover in foreground.
Upper Cretaceous sequence forming mountains, right.
N-234 Burgos - Sagunto road, 500m W of turn off to Quintanilla de las Vinas. Looking SE.
Tour of Spain cycle race: Soria - Burgos stage, approximately 30km from finish.
April, 1984

CHAPTER 6. THE SEDIMENTOLOGY OF THE "UTRILLAS" FORMATION (UPPER ALBIAN - LOWER CENOMANIAN).

6.0. Introduction.

The Utrillas Formation (Upper Albian - Lower Cenomanian; see section 2.11) outcrops throughout the Iberian Chains and SW Cantabria. The Formation rests unconformably upon the underlying Lower Cretaceous and Jurassic strata with significant overstep towards the SW basin margin (see section 8.15).

The Utrillas Formation of the study area belongs to the Tejada Sands Member of Alonso et al (1982; fig 2.47). The Member is named after the classic outcrop at the base of Pico de la Sierra, near Tejada (fig 1.6) where over 100m of Utrillas Formation sands, pebbly sands and conglomerates are exposed. Patchy, often discontinuous, outcrops are also present at the foot of other mountains capped by Upper Cretaceous carbonates (eg Penalara, San Carazo). There are excellent outcrops offering three dimensional exposure (but little continuous vertical sequence) to the N of Talveila and near the Puerto Mojon Pardo (see fig 1.6).

Elsewhere, the outcrop of the Utrillas Fm is generally poor. The sands are often poorly consolidated and weather to give sandy soils utilised for wheat cultivation. The base of the formation may be mapped using changes in vegetation cover and soil colour and locally by geomorphological features (fig 6.1). The top of the formation may also be mappable using soil colours and changes in slope (fig 2.51).

In the NW of the study area, the base of the Utrillas Formation is often characterised by a heavily purple-pink mottled fine sand horizon resting above the subjacent Salas Group. This is overlain by a coarse clast-supported conglomerate with abundant poorly sorted, subangular to subrounded 5-20cm white, pink and grey quartzite clasts. The sequence is well exposed beside the new road linking Cubillo del

César and Lara de los Infantes, approximately 500m to the W of Cubiejo. The base of the Formation is marked by a similar regionally-mappable conglomerate in the SW of the study area.

The very top of the Formation generally shows a gradual colour change to yellowish-brown, (this reflecting higher Fe content) and progressively greater cementation, (this probably partly reflecting the passage in groundwaters of dissolved carbonate derived from the Upper Cretaceous above). Multi-directional cross-beds were found in one sandstone block near the top of the Tejada Member - this could record tidal action associated with the deposition either of the top few metres of the Utrillas Formation or of the glauconitic sands (equivalent to the Santa Cruz del Tozo Formation of Alonso et al, 1982) at the base of the Turonian.

Although lithological correlation is possible for the basal and uppermost few metres of the Utrillas Formation, the great thickness (> 200m) and discontinuous exposure preclude accurate correlation of the rest of the sequence. However, there are still some excellent exposures suitable for sedimentological study, notably in road cuts along the N-234 Burgos - Sagunto road between Mambrillas de Lara and Cuevas de San Clemente (fig 1.6). These were excavated during carriageway improvements in 1983-4.

The Utrillas Formation was not studied in great depth and neither detailed sections nor palaeocurrent analyses were made, this reflecting the generally poor and discontinuous nature of the outcrop. Nevertheless, the formation was frequently encountered and briefly studied during the course of reconnaisance and field mapping. This work has facilitated the recognition of three main facies. An integrated facies model for the Utrillas Formation is given in fig 6.2.

6.1. Facies 1: Mottled horizons.

These consist of fine sandstone units generally 1-2 m thick and

exhibiting strong purple-pink to locally greenish diffuse mottling on a 10cm scale. Cross-bedding and quartzite clasts are absent or rare in this facies. The sands weather rapidly in road cuts and often appeared to have a higher proportion of mud or clay.

Interpretation.

The mottled fine sands are interpreted as immature palaeosols developed by the pedogenetic alteration of a sandy substratum. The higher proportion of mud may indicate occasional sedimentation out of suspension - but primary sedimentary structures are absent. This is attributed to bioturbational homogenisation. The mottling may be due to burrowing or to root action.

6.2. Facies 2: Conglomerates.

The conglomerates contain coarse white, pink and grey quartzite clasts typically 2-5cm (and rarely up to 10 or 20 cm) in size. Thicker conglomerate units are clast-supported at the base and often pass gradationally upwards into matrix-supported conglomerates. The matrix consists of coarse white to yellow or pink quartzose or feldspathic sand. Conglomerate units (especially the thicker examples) frequently show strongly erosive bases; sharply incised conglomerate-filled channels may occasionally be recognised (fig 6.3). Thinner conglomerate units are occasionally sheet-like, but are more generally lensoid or otherwise laterally discontinuous. The discontinuous conglomerate units are typically interbedded/intercalated with sandstones or pebbly sandstones on a 5-30cm scale. Imbrication of the conglomerate clasts and large-scale cross-stratification may also be discernible (figs 6.3 - 6.4).

Interpretation.

The large clast-size and steep channel margins indicate transport by high-energy currents. The presence of large-scale

cross-stratification (eg fig 6.3) suggests accretion of large gravel bars within the main channels. The cross-stratification clearly has a dip component towards the channel margins - the bars may be longitudinal bars migrating perpendicular to the channel axis.

The thin conglomerate units intercalated with sands are more difficult to interpret. Some of the more laterally-extensive units may represent sheet floods, but thinner tongues and lenses of conglomerate probably indicate the relatively short-lived establishment of high energy conditions. Such conditions may have been set up periodically where stream flow was strongest, (eg between smaller sand bars migrating within a broad, relatively unconfined channel system) or during conditions of particularly high energy (eg during short-lived minor flood stages). The discontinuous nature of these conglomerate units also points to incomplete preservation of vertical sequences due to repeated erosional modification of the sedimentary structures.

6.3. Facies 3: Pebbly sands and sandstones.

The pebbly sandstones show very complex sediment geometries (eg figs 6.5 - 6.10), owing to intercalation with conglomerate tongues and lenses (see above). However, there are also abundant sandstones with relatively few quartzite pebbles - these tending to be concentrated as imbricate-pebble lags at a few horizons or erosion surfaces. The sandstones with few quartzite pebbles often display excellent trough (or, more rarely, planar) cross-sets (eg figs 6.11 & 6.12), frequently with multiple reactivation surfaces. Sand-filled channels were also recognised, both single (fig 6.13), and stacked (fig 6.14).

The sands are compositionally very mature quartzarenites and feldspathic quartzarenites and are laterally very extensive. The common white colour of the sands is probably partly due to the presence of kaolin produced by weathering of feldspar. There is generally very little or no mud, although there are a few rare thin (1-2cm) pink clay horizons at Tejada. One outcrop near Cubillo del

César contained a single 5mm lignite horizon. Ferruginous crusts are occasionally present and fragments of tree bark (fig 6.15) and whole trunks (fig 6.16) preserved in ferruginous fine sandstone are locally present in the central part of the study area.

Interpretation.

The sands are clearly water-lain. The trough cross-beds indicate the accretion and current migration of dunes. The reactivation surfaces occasionally outline the morphology of larger-scale structures (eg sand bars; fig 6.17) and are indicative of important internal reworking.

6.4. Vertical and areal distribution of facies.

The facies of the Utrillas Formation are laterally very uniform (section 8.15. In vertical sequence, the Utrillas Formation is heavily dominated by apparently monotonous sandstones and pebbly sandstones. However, conglomerates are present at a number of horizons, most notably at the base. Mottled horizons (palaeosols) are also present at the base (NW sector) and at a number of other horizons higher up in the sequence in the Mamolar-Pico area (fig 1.6). It was hoped to use these palaeosols as marker horizons for regional mapping, but the outcrop is poor and the palaeosols also appear to be laterally variable (and probably discontinuous). Both the conglomerates and the mottled horizons are commonest in the lower half of the sequence, but even then probably together account for <5% volumetrically. The upper part of the sequence is characterised by cross-bedded sands and grits with more occasional pebble bands (fig 6.17).

6.5. Environmental synthesis.

The Utrillas Formation is a problematic deposit: it has been recorded and cursorily described by numerous workers, but it has never

received a detailed sedimentological analysis. The most advanced sedimentological treatment is that given by Alonso (1981).

The Utrillas Formation was deposited in a continental, fluvial environment. The compositional maturity reflects multiple internal reworking. The quartzite clasts are apparently derived from the Palaeozoic massifs (eg from Cambrian and Ordovician quartzites; see fig 2.0), and were probably transported many kilometres during a long history of deposition, erosion and subsequent re-deposition.

The presence of: a) thick sequences of coarse sands that are only rarely channelised but are generally very laterally extensive, often trough cross-bedded and intercalated with pebbly sandstones and conglomerate tongues;

b) sheet-like or locally channelised, large-scale coarse cross-stratified conglomerates and pebbly sands;

c) very little mud;

all point to high-energy deposition in a complex depositional system, with many fluctuating sub-environments.

The sedimentary structures immediately suggest a "braided-stream" (Jackson, 1978) environment for the Utrillas Formation, a conclusion reached by a number of authors. However, to take the interpretation beyond this elementary first step is difficult, requiring rigorous and full description. Even though detailed logs and palaeocurrent analysis are not presented here, the observations given above and in the photographs still provide much additional information, allowing us to make a more detailed interpretation of the facies and environments. An integrated facies model for the Utrillas Formation is presented in fig 6.2.

The model shows deposition of the Utrillas Formation on a laterally extensive and unconfined braidplain consisting of many broad and relatively shallow high-energy streams with abundant interchannel

bars and islands. Sediment was transported in the channels dominantly as bed-load. Flow conditions permitted the development of lunate dunes on channel floors, resulting in the abundant trough cross-bedding in vertical section. Larger-scale structures such as longitudinal sand and gravelly bars were developed in the higher energy parts of the channels. Imbricate pebble lags often accumulated on the surfaces of Sand bars as a result of current deflation.

The sedimentary structures were subject to frequent reworking; the intensity of this reworking ranging from minor modification because of variations in the discharge-rate (eg small-scale channels and dunes developed on the backs of sand bars; minor chute channels dissecting gravel bars) to major erosion and even destruction of sedimentary structures as a result of changes in the morphology of the channel network. This led to the development of extremely complex geometries of intercalated sand and sandstone bodies in pebbly sandstone units.

During flood stage, interchannel areas occasionally received finer-grained deposition out of suspension. These mud-grade sediments were subjected to long periods of subaerial exposure. During emergent periods, the "overbank" deposits were subject to pedogenesis and root-bioturbation, which led in turn to the development of mottling and to homogenisation with the underlying sandy sediments.

The features described and the model presented here are broadly similar to those listed in other accounts of "braided-stream" deposition (eg South Saskatchewan River; Cant & Walker, 1978). However, although Cant & Walker's facies model (fig 6.18) shows a wide variety of sub-environments and gives an indication of the complex morphology of high-energy, low-sinuosity systems, careful comparison of the predicted sequences with those of the Utrillas Formation (fig 6.19) reveals the following important differences:

1) The Utrillas Formation contains many more conglomerates and more pebbles in the sandstones than the South Saskatchewan River

sediments. In this respect, the Utrillas Formation sequences show some affinities to the Donjek River Model (Williams & Rust, 1972; Miall, 1977; fig 6.19). The abundance of coarse clasts may indicate a more "proximal" environment than that represented by the South Saskatchewan River; it certainly indicates a source of abundant coarse and resistant quartzite clasts.

2) The Utrillas Formation sequences are apparently extremely mud-poor, containing less mud even than the predicted South Saskatchewan River sequences. This may reflect the compositional maturity of the clastic input into the system (ie erosion only of quarzites would tend to produce a mud-free sequence) but probably also partly reflects the low preservation potential of interchannel deposits in a system characterised by multiple internal reworking.

3) The Utrillas Formation contains mottled palaeosols. The formation of these reflects more active pedogenesis in a warmer climate than that of present-day S Canada. The palaeosoils are rare, but nevertheless, their preservation is something of a problem in view of the virtual absence of mud horizons. In fact, the palaeosoils do appear to have a higher proportion of mud. Primary argillaceous units were mixed with sand as a result of bioturbational homogenisation.

4) The strong dominance of trough cross-bedding over planar cross-bedding is a particularly notable feature of the Utrillas Formation. However, it is clear that the sedimentary structures preserved in a particular section vary according to the position of that section within the depositional system. Analysis of the model (fig 6.2) shows that planar cross-sets dominate in the sand flat environment while trough cross-sets are dominant in the channels themselves. In a system characterised by strong internal recycling, the preservation potential of sand flat sediments is correspondingly reduced, so that the conglomeratic and trough cross-bedded sand elements of the channel environment tend to be anomalously important in vertical sequences.

Fig 7.0. Snow and trees in the extreme NE of the study area, near Laguna Negra. Meteoric waters are depleted in ^{18}O, while soil-derived CO_2 is commonly depleted in ^{13}C, these effects reflecting the importance of evaporative and organic fractionation respectively on isotope geochemistry.

CHAPTER 7. GEOCHEMISTRY.

7.1. Total Organic Carbon.

The recognition of darker open lacustine biomicrite facies suggested that some of the carbonate sediments of the Rupelo Formation might be potential hydrocarbon source rocks. A few of the most promising samples with significant quantities of organic material recognisable in thin section were selected for a pilot study in Total Organic Carbon Analysis. It was hoped that the TOC values obtained might permit the establishment of a "geochemical stratigraphy" by facilitating detailed centimetre- or decimetre-scale correlation of selected intervals (eg within the Mambrillas de Lara Member).

In fact, the dark biomicrites (facies 7; see section 3.3.7) analysed from the Mambrillas de Lara Member of the Rupelo Fm at Senora de Brezales showed disappointingly low TOC values of only around 0.2%. Dark charophyte dedolomite marls (facies 14: see section 3.3.14) from the Rio Helechal section at Hortezuelos contained 1.4% TOC.

Although these results indicate minor to locally moderate source rock potential, these were thought to be likely maximum values and it was evident that values obtained from other horizons would be significantly lower. It was concluded that TOC results would neither

a) provide significant additional information about the Rupelo Formation as a whole; or

b) permit detailed correlation;

and thus further analyses were not performed.

7.2. Stable isotopes.

7.2.1. Introduction.

Stable isotope techniques have great potential for providing additional information about the depositional and diagenetic histories of carbonate sequences. (For a general review of stable isotope

geochemistry, see Arthur et al, 1983). The stable isotope geochemistry of a carbonate sediment will initially reflect the stable isotope composition (as well as the temperature) of the waters from which the carbonate was precipitated. Subsequent modification of the primary isotopic composition is possible due to interaction with later diagenetic fluids.

Stable isotope results are conventionally reported as per mil deviations from the PDB isotopic standard:

$$\delta = \frac{R_{standard} - R_{sample}}{R_{sample}} \times 10^3, \text{ where R is } {}^{13}C/{}^{12}C \text{ or } O^{18}/O^{16}.$$

Cross-plots of stable isotope analyses (eg $\delta^{13}C$ against $\delta^{18}O$) may permit the characterisation of carbonate rocks into particular fields according to their origin (eg fresh-water, marine, hypersaline; see Hudson, 1977; fig 7.1). Stable isotopic studies on carbonate cements (eg vadose, phreatic, late ferroan) provide valuable clues about the rôle of diagenesis (eg Dickson & Coleman, 1980).

7.2.2. Aims.

A brief stable isotope study was undertaken in order to address particular problems not accessible using other petrographic techniques and thus to complement the petrographic approach employed elsewhere in this work.

It was hoped to address the following particular problems:

1) Rupelo Formation.

a) Were the carbonates of the Rupelo Formation deposited from fresh-water? Although the fauna/flora of the Rupelo Formation probably indicates fresh-water, stable-isotope results would provide useful supporting evidence as some of the faunal/floral elements (eg charophytes) might be tolerant of brackish water.

b) Do the pedogenically-modified and marginal lacustrine limestones plot in different fields from the open lacustrine facies? It would be anticipated that the influence of soil-derived CO_2 and

evaporation in marginal lacustrine environments and of higher organic productivity in open lacustrine environments would result in the occurrence of two distinct sample suites with different stable isotopic characteristics.

c) What are the stable isotope values of the black pebbles / dark intraclasts?

d) What are the isotopic characteristics of the vadose cements - is there any evidence of diagenetic modification of primary isotope geochemistry?

2) Pedroso Group.

a) What are the stable isotope characteristics of the nodular carbonates? Do the results support the interpretation of strong pedogenesis? How do the values obtained compare with those for lacustrine carbonates?

b) Do the oncoidal carbonates give values indicative of fresh-water conditions? The presence of blue-green algae (cyanophytes) in oncoids implies an absence of algal grazers. Was this absence the result of hypersaline or hyposaline conditions?

c) Are there significant differences in stable isotopic composition between: i) oncoids - dense layers;

ii) oncoids - bushy (ie algal) layers;

iii) oncoidal limestones - matrix carbonate?

Such differences might indicate the possible influence of organic fractionation processes in oncoid carbonate production.

7.2.3. Techniques.

Stable isotope analyses were done at the British Geological Survey, London. ^{13}C and ^{18}O isotopic analyses were performed on 35 selected samples from carbonates facies of the Rupelo, Hortigüela, and Piedrahita de Muno Formations.

Samples were chosen as far as possible from the same locality (Mambrillas de Lara), with a lesser number of additional samples from

other parts of the basin to complement the range of lithofacies and to act as controls against areal variations.

Approximately 2-4mg of each sample was drilled out of specially-prepared 150μm petrographic sections in finely-powdered form using an engraving tool under the optical microscope. The mounting medium used in the making of the slides was Lakeside. Whole rock extractions were not performed - it was aimed to characterise matrix carbonate without also sampling bioclasts, intraclasts from different environments, or late void-filling cements.

The samples were then evacuated and reacted with phosphoric acid to liberate CO_2 (Mc Crea, 1950), the gas produced being collected using the apparatus shown in fig 7.2 and then analysed for $^{13}C/^{12}C$ and $^{18}O/^{16}O$, using a triple collector Micromass 903 mass spectrometer. The ratios obtained were calibrated against a known laboratory standard, MCS-8, and the results expressed in parts per thousand units in comparison with international standards (Craig, 1957, 1961): PDB (a Cretaceous belemnite) and SMOW (Standard Mean Ocean Water). The results are shown in figs 7.3 - 7.5 & 7.11.

Duplicate analyses were performed on a few selected samples to test the reproducibility of the analytical results and the accuracy of the sampling technique. In general, the $\delta^{13}C$ results were of high statistical validity, while the $\delta^{18}O$ results showed rather less statistical consistency.

7.2.4. Discussion of previous works.

There is little published work on the stable isotope geochemistry of lacustrine carbonates. Two recent studies: Bond, pers. comm., 1985, (Late Cretaceous - Tertiary of the NE Duero Basin, Spain); Oberhänsli & Allen, in press, (Oligocene of the SE Ebro Basin, Spain) have succeeded in defining separate fields for marginal lacustrine and open lacustrine facies (although the latter did not discount the possible rôle of diagenesis). Both have stressed the rôle

of evaporation and organic productivity in controlling isotope dynamics in lacustrine systems (enhanced organic productivity → more negative $\delta^{13}C$; more intense evaporation → more positive $\delta^{13}C$ and $\delta^{18}O$, and correlation of $\delta^{13}C$ and $\delta^{18}O$).

Stable isotope work on pedogenic carbonates and coated grains has included that of Magaritz (1974, 1983), Magaritz et al (1981), Salomons et al (1978), Talma & Netterberg (1983) and Cerling (1984). Interpretations have stressed the importance of evaporation and soil-derived CO_2.

Behrens & Frishman (1971) investigated the stable isotope ratios of modern blue-green algal (cyanophyte) mats and reported an enrichment in the blue-green algae of ^{13}C by 5‰ in comparison with associated marine sediments. In contrast, work on Upper Cretaceous oncoids from NE Spain by Freeman et al (1982) has demonstrated similar values for oncoids and matrix carbonate, so that the occurrence and magnitude of stable isotopic fractionation effects in algal mats and oncoids remain uncertain.

7.2.5. Interpretation of results presented here.

1. Rupelo Formation.

Isotopic values for the Rupelo Formation all lie in the negative $\delta^{13}C$ - negative $\delta^{18}O$ quadrant, with spreads of $\delta^{13}C$ from -7 to -10‰ and of $\delta^{18}O$ from -4 to -7‰. Fig 7.1 shows the characteristic fields of various types of limestone from Hudson (1977). The values reported here (figs 7.3 - 7.5) all lie close to the "average freshwater limestones" values of Keith & Weber (1964), although they are strongly negative even within this field, which suggests that their compositions may not be entirely typical of freshwater limestones. Possible reasons for this are discussed below.

The variables which might affect stable isotope compositions of fresh-water limestones are outlined in fig 7.6. Photosynthetic activity (removal of CO_2 by algae and very rapid carbonate

precipitation → carbonate ^{13}C enrichment) and exchange reactions with atmospheric CO_2 (which has $\delta\ ^{13}C = -7 \pm 2‰$) at times of higher lake level would both enrich the lake waters in ^{13}C. Alternatively, higher organic productivity in open lacustrine waters would result in the production of more organic carbon in lake sediments which might explain the occurrence of highly negative ^{13}C and ^{18}O values - organic C from terrestrial plants has low $\delta\ ^{13}C$ and $\delta\ ^{18}O$.

Higher lake levels might result from heavy rainfalls provoking sheet-wash and thus enrichment of isotopically light soil-derived CO_2 and of detrital vegetal matter from vegetated lake catchment areas (Oberhänsli & Allen, in press): C_3 plants - most vegetation including trees and shrubs - $\delta\ ^{13}C = -26 \pm 4‰$; C_4 plants - mostly arid zone grasses - $\delta\ ^{13}C = \pm 13 - 3‰$.

Soil-derived CO_2 may have $\delta\ ^{13}C$ as low as -25‰ (Talma & Netterberg, 1983; fig 7.7. However, close to the surface of poorly-vegetated, well-aerated soils, the ^{13}C content of the soil CO_2 will approach that of the atmosphere, ie -7‰. Periods with increased precipitation also tend to produce rainfall depleted in ^{18}O. Certainly, as in the case of the Oligocene of the SE Ebro, the open lacustrine facies of the Rupelo Fm do contain a large number of reworked intraclasts, indicating that fresh-water influx was important.

Precipitation of carbonate in lakes may involve CO_2 loss from a carbonate-saturated solution on evaporation. More positive $\delta\ ^{18}O$ and $\delta\ ^{13}C$ values might be the result of increased evaporation in ephemeral lake settings (evaporative fractionation → preferential evaporation of isotopically-light ^{16}O and ^{12}C, see Lloyd, 1966). This effect is further enhanced by fractionation effects in plant transpiration.

However, although the values reported here show a fair spread; there is no clear difference between marginal lacustrine

("palustrine") values and open lacustrine ones. However, there is a strong correlation between $\delta^{13}C$ and $\delta^{18}O$ values. Salomons et al (1978) cited this as evidence in favour of the strong influence of evaporation in carbonate precipitation.

These, then are possible explanations which might be invoked to explain the isotopic variations as primary sedimentary features. However, several authors (eg Hodgson, 1966: Stahl & Jordan, 1969; Tan & Hudson, 1971; Allen et al, 1973; Magaritz, 1975; Walls et al, 1979); have pointed out that stable isotope values may be strongly modified by subsequent diagenesis.

Hudson (1977; fig 7.8) shows several straight-line trends on $\delta^{18}O$ and $\delta^{13}C$ plots which are interpreted as the result of diagenetic modification. Indeed, the linear trend and strongly negative nature of the results presented here are very similar to those of secondary calcites from Bermuda (trend 11, fig 7.8) and could well be significant indicators of diagenesis.

An insight into the reasons for the strongly negative results comes from the work of Allan & Matthews (1977, 1982) which showed that negative $\delta^{13}C$ and $\delta^{18}O$ values from Pleistocene (marine) limestones (such as those reported by Gross, 1964) can result from vadose diagenesis in the presence of isotopically-light, soil-derived CO_2 below exposure surfaces (fig 7.9; see also Beeunas & Knauth, 1985). Depletion of ^{13}C and ^{18}O occurs as a result of the dissolution of unstable carbonate minerals by CO_2-rich groundwaters and subsequent reprecipitation of carbonate as cements and relatively more stable phases. Precipitated carbonate tends to be less ^{13}C depleted and eventually unchanged with increasing depth below an exposure surface. As the fluids percolate downwards and out of contact with soil-derived CO_2, isotopic exchange with the host rock as a result of dissolution and reprecipitation results in progressively less altered values. The final isotopic composition after diagenesis will depend upon:

a) the isotopic composition of the initial carbonate

b) the isotopic composition of the groundwaters

c) the degree of exchange.

In contrast, freshwater phreatic diagenesis will tend to produce more positive δ values (fig 7.10).

Certainly, in this study, the lowest values are those of the vadose cements and the other results show a very linear trend towards those values, suggesting a variable degree of diagenetic equilibration under vadose conditions. Magaritz (1975) noted that $\delta^{18}O$ values readily show diagenetic equilibration with meteoric groundwaters, while $\delta^{13}C$ remains virtually unchanged on equilibration with groundwaters unless there is significant recrystallisation. If recrystallisation does take place as a result of the passage of groundwaters then low negative $\delta^{13}C$ values result. In the case of the Rupelo Formation, where it seems likely that both oxygen and carbon isotope values have been diagenetically modified, we may infer that the vadose diagenesis may not only have involved lithification and cementation - there may have been partial recrystallisation as well.

Black pebbles showed a spread of stable isotope compositions (fig 7.5). This probably reflects original variation in organic carbon content as well as differing degrees of diagenetic modification.

There are clearly a large number of possible variables to consider (see fig 7.6). The complex interplay of primary variations and secondary modification of isotopic values for these lacustrine carbonates makes the isotopic data extremely difficult to interpret. This study was set up with an emphasis on facies and with the aim of assessing any primary variations within facies. A more comprehensive data set concentrating on the variations below particular microkarstic surfaces would provide more information about the isotopic effects of diagenesis. Nevertheless, it still seems clear that the limestones have been extensively modified under vadose diagenesis. The micrites are strongly lithified and cemented (in contrast to the Oligocene

carbonates of the SE Ebro) and may have undergone recrystallisation (and possibly even mineralogical change from high Mg-calcite or aragonite precursors), under vadose diagenesis. A study of trace element (eg Sr) contents in the carbonates might provide useful supporting evidence for such mineralogical changes.

2. Pedroso Group.

The nodular pedogenic carbonates show $\delta^{13}C$ values of -7 to -9‰, and $\delta^{18}O$ of -3 to -7‰ (fig 7.11), although there is considerable variation and the number of samples was insufficient to characterise this lithology adequately. Still, these results are within the range of (if rather negative for) "calcrete" values as reported by Talma & Netterberg (1981; fig 7.12). Once again, there is a possibility of correlation between $\delta^{13}C$ and $\delta^{18}O$ values.

Stable isotope determinations on Hortigüela Fm oncoidal limestones (fig 7.11) showed no significant difference in composition between oncoids and oncoidal limestone matrix. In addition, similar values of $\delta^{13}C$ = -7 to -9‰ and $\delta^{18}O$ = -4 to -7‰ were obtained for dense and porous - ?non-algal and algal layers within the oncoids themselves. These values and the lack of differentiation between oncoids and matrix are similar to those reported by Freeman et al (1982, fig 7.13). The highly negative values obtained and the apparent absence of organic fractionation effects observed in both cases are probably at least partially attributable to vadose diagenesis.

Fig 8.0. Panoramic view of the San Leonardo Fault Zone, Mamolar. Ridge running centre to left consists of Marine Jurassic carbonates and overlying Rupelo Formation fresh-water limestones thrust SE-wards over Utrillas sands. Looking S. Stratigraphic younging towards camera. Steep dips to NE, hence view effectively appears as an inverted structural section. Décollement at far left is along "Infralias" "Carniolas" evaporites, but locality is at end of SLF zone. Lateral ramp visible to left of village where thrust cuts up section, detaching on Rupelo Formation and then cutting Utrillas sands. Ridges behind are parallel thrusts (eg Ladera Thrust; fig 8.112) lower in the structural pile and thrusting Rupelo Formation over Utrillas.

CHAPTER 8. SEDIMENTATION AND TECTONICS.

8.1. Introduction.

Although several workers have considered the rôle of tectonics during particular intervals, to date there have been no integrated studies of the structural and sedimentological evolution of the Cameros Basin.

The chief works on the structure of the area are those of Colchen (1966, 1974). Colchen considered the geology of the Sierra de la Demanda and the immediately surrounding area. Colchen recognised two "Hercynian" phases of tectonic activity and commented briefly on "Pyreneo-Alpine" tectonics and the coincidence of structural trends from this episode with earlier structures, suggesting that Hercynian structures in the basement exerted strong control on the Tertiary deformation of the cover sequence. This set the scene for the IGME memoirs of the 1:50 000 and 1:200 000 geological maps of the area.

Many of these follow Colchen in not recognising tectonic activity during the Mesozoic (so-called "Cimmerian" events, see Ziegler, 1982). However, in their memoir to the IGME tectonic map of the Iberian Peninsula, Julivert et al (1980) mention that "fracture tectonics" was important in the Iberian Chains during Late Jurassic and Early Cretaceous sedimentation, and several other workers have inferred from sedimentological studies that synsedimentary faulting occurred at particular times during the Mesozoic and Cenozoic.

Salomon (1982) discussed the influence of tectonics on the facies and facies distribution of the Upper Jurassic - Lower Cretaceous continental sequence, and produced maps showing important synsedimentarily active tectonic lines. Salomon also postulated intra-Mesozoic folding, suggesting that the Moncalvillo - Palacios de la Sierra anticline (fig 1.6) formed a mid-basin high separating two individualised sub-basins during Berriasian time (see section 2.8). Morillo-Velarde & Melendez-Hevia (1972) discussed the rôle of the San

Leonardo Fault, suggesting that it defined the SW basin margin during the Late Jurassic - Early Cretaceous.

Floquet, Alonso & Melendez A (1982) suggested that Upper Cretaceous facies distribution patterns reflected differential subsidence which was associated with "gentle tectonic movements over deep axes". This theme was subsequently developed by Alonso et al (in press) and by Floquet (1984), which was especially significant in attempting to demonstrate and date the effects of tectonic and eustatic influences and to compare the Upper Cretaceous sedimentary and stratigraphic evolution with other areas in NW Europe.

Studies relating the structural evolution and sedimentary development of the area during Tertiary times include the work of Pol Mendez (1985, and pers. comm., 1986) which considered the importance of fault control on Tertiary facies.

The Mesozoic geological evolution of N Spain was considered by Amiot, Floquet et al (1982). They interpreted the area as a NE-facing passive continental margin made up of a series of NW-SE half grabens which were controlled by major normal faults downthrowing to the NE, and pointed to the dissection of the margin by important NE-SW wrench faults. Salomon (1982, 1983) discussed the initiation and development of the Cameros Basin as a NE-SW oriented half-graben, "La fosse de Soria". Salomon noted the apparent progressive NE-ward migration of the basin depocentre. This was interpreted as the product of dextral strike-slip on NE-SW faults, a hypothesis expanded on and refined by Seguret & Guirand (1985). However, recent structural work in N Spain by Hossack, (1985, 1986, & pers. comm., 1986) has applied modern structural concepts to the study of the Cantabrian Pyrenees.

This chapter aims to apply a new approach to the study of sedimentation and tectonics within the Cameros Basin. Previous workers have concentrated on the documentation of small parts of the stratigraphy and have rarely considered the significance of the results within the wider context of basin analysis. This chapter

suggests a different emphasis by adopting a multidisciplinary perspective embracing concepts of stratigraphy, sedimentology and tectonics. Data obtained from the study of sedimentary facies and their spatial and temporal distributions throughout the Mesozoic and Cenozoic are considered in the light of information derived from structural mapping. This material is then interpreted, applying recent ideas about the initiation, development and deformation of basins, aiming to relate the first integrated account of basin evolution.

This chapter also adopts a wider regional viewpoint. A problem has been that although strata from particular stages and systems have been correlated (and sometimes compared) with equivalents from other areas of Spain, there have been few comparisons with areas outside Iberia. Furthermore, the importance of the Tethys in the Mesozoic and Cenozoic geological history of Spain and Portugal has been such that while parallels have been drawn between the Vasco Cantabrian and Aquitaine basins, the literature on the Iberian Chains has rarely looked further N across Biscay for inspiration. Thus while the striking contrasts with the Betic Cordillera are well-known, the considerable similarities in the structural evolution of S Cantabria and the Iberian Ranges with that of offshore S Ireland and S Britain have been consistently overlooked. This seems particularly surprising in view of the frequent references to the Mesozoic stratigraphy of Wessex!

The recent advances in the understanding of basins were themselves facilitated by hydrocarbon exploration on the NW European continental shelf. It is appropriate, then, that this chapter aims to employ its new conceptual and palaeogeographic emphasis by outlining the dynamic sedimentological and structural evolution of the Cameros within the context of the N Atlantic margins and shelves.

8.2. Plate tectonic setting.

The Cameros Basin forms part of the Iberian plate. Palaeomagnetic and structural evidence suggest that, at the end of the Palaeozoic, Iberia, Armorica and Cornubia formed an arcuate Hercynian cordillera (fig 8.1). Neither the Atlantic nor the Bay of Biscay were in existence at this time.

Rifting was initiated in the area during the Late Permian and Triassic associated with tensional stresses caused by the opening of the Tethys. Late Triassic volcanism occurred in Cantabria. Subsequent expansion of the Tethys, and the onset of sea-floor spreading in the central and N Atlantic to the W of Iberia, caused more extensional episodes during the Jurassic and Early Cretaceous.

The Iberian plate experienced somewhat similar development during this time to other areas of NW Europe, but sea-floor spreading in the Bay of Biscay commenced in the Middle Cretaceous causing further extensional (?and compressional) stresses and, as Biscay opened, Iberia was rotated anticlockwise and rafted to the SE. After this time, the geological history of Iberia was increasingly different from NW Europe and the N Iberian margin displays sedimentological and structural similarities from then on only with the extreme SW of the opposing continental margin - the Aquitaine Basin and the "Western Shelves" (Ziegler, 1982) off SW Britain, namely the Western Approaches, the Hampshire and Channel Basins, the Celtic Sea and the Bristol Channel. The Porcupine Trough displays a broadly similar evolution to the NW Iberian margin off Galicia (for description of the latter see Montadert, 1981). Fig 8.2 shows the position of Iberia during Middle Cretaceous time and shows the relationship between the W Shelves and N Iberia. Early Mesozoic sequences of the E Canadian Basins also show remarkable similarities to the N Iberian successions and this reflects the proximity and parallel evolution of the E Canadian and Iberian margins prior to and in the early stages of N Atlantic opening.

Fig 8.3.0. Hercynian. Calizas de Montana (Mountain Limestone), Carboniferous.
Picos de Europa, Cantabrian Mountains.
August, 1983.

Sea-floor spreading had ceased in the Tethys during the Middle Cretaceous and the first orogenic phases in the Alpine Domain occurred during the Upper Cretaceous as Africa began to move N towards Europe. In N Iberia, compressive effects were felt approximately contemporaneous with the end of sea-floor spreading in the Bay of Biscay during the Senonian. From the Campanian onwards, Iberia began to move N towards France. Further mild compression coincided with Laramide tectonic events during the Palaeocene, and the collision of Spain with France approximately at the start of Oligocene time was manifested in Iberia by the Pyrenean orogeny which produced extensive uplift and deformation in the N of Spain. Previously downfaulted blocks were elevated, there was important thrusting and rapid sedimentation in molasse basins. Hossack (1985) reported 120km of Tertiary shortening in the Pyrenees and Cantabria. Continued slow N-ward movement of Iberia towards France resulted in mild folding and faulting during the Miocene and Pliocene to produce the structural pattern we see today.

8.3. Hercynian.

The W part of the Iberian plate is made up of an ancient microcraton, the Iberian massif or "meseta". The meseta consists of Palaeozoic and older rocks and has been an important positive area since Hercynian times. The Cameros Basin lies near the NE margin of the meseta, and it is useful to appreciate the structure of the Hercynian basement on which the Mesozoic sediments were deposited.

Fig 8.3 shows the Hercynian structural units of the Iberian peninsula as given by Julivert (1980). The map shows the morphology and general NW-SE trend of the orogenic belt in Iberia. An "internal" zone of intense deformation and high-grade metamorphism is flanked on either side by more "external" zones characterised by lower grade metamorphism and thin-skinned tectonics. The two outer zones display opposed senses of dominant vergence so that, while thrusting was to

the SW in the SW of the orogen, the areas to the NE of the Central Iberian Zone (see fig 8.3) show NE-ward Hercynian thrusting.

Figs 8.1 & 8.4 show the morphology of the Hercynian mountain belt in Europe. The orogenic belt apparently described a tight arc in Cantabria so that the continuation of the S Portuguese Hercynian structures may be traced through the SW parts of Ireland, Wales and England and further E into the Rheno-Hercynian and Moravo-Hercynian zones. The Cameros Basin lies on the opposite side of the Central Iberian Zone and forms part of the W Asturian - Leonese Zone which has its NE extension in France, eg the S extremities of the Massif Central, (Montagne Noire), and its SE extension in the Palaeozoic outcrops of the Iberian Chains.

The Hercynian basement to the Cameros Basin outcrops in the Sierra de la Demanda, which lies to the N of the study area, and in the Moncalvillo - Palacios anticline. The intensity of Hercynian deformation was markedly less than in the fold-and-thrust belts of the Cantabrian mountains, although metamorphism was more important.

Colchen (1974) recognised two main Palaeozoic tectonic phases in the Sierra de la Demanda: a) post-Tremadocian - pre-Westphalian B and b) post-Westphalian - pre-Triassic. Julivert (1981) and Colchen (op. cit.) remarked that the effects of the Caledonian orogeny have not been recognised in Iberia, and thus inferred that any Palaeozoic deformation must have been caused by Hercynian disturbances. Colchen (1974) thought that the early phase was the result of three events, two of which occurred during the early Carboniferous and a third which caused rejuvenation of relief prior to the deposition of Westphalian B-C conglomerates. Julivert (1981) suggested that deformation may have begun as early as the Late Devonian.

The early tectonic phase was essentially one of folding (Colchen, 1974). The intensity of deformation was variable and reflected the lithological inhomogeneity of the Lower Palaeozoic sequence. Further compression caused large-scale thrusting with

décollement on the San Anton dolomite, which forms the top of the Lower Cambrian (fig 2.2). The thrust planes are parallel to approximately E-W fold axes developed during the early stages of deformation. Thrusting directions are dominantly to the N, although thrusting was to the S in the portion of the Demanda immediately to the N of the study area, the Sierra de San Millan. The opposed sense of vergence shown by these structures may be the result of local S-directed Hercynian (?back) thrusting in this area.

The post-Westphalian tectonic phase postdates greenschist facies metamorphism and is manifested by the mappable discordance between the Westphalian and Triassic sediments. This late-Hercynian episode was characterised by major faulting (Vegas, 1975; Arthaud & Matte, 1975, 1977).

The thrust planes are cut by abundant NE-SW and NW-SE faults. Fig 8.5 shows the pattern of Stephanian - Autunian faults in NW Europe. These fault trends are important throughout much of the meseta (fig 8.6) and include several faults traceable for hundreds of kilometres, eg the Hispano - Portuguese Fault (NE-SW) and the Aviles Fault (NE-SW). It is immediately evident that the Cameros Basin lies at the intersection of several major late-Hercynian faults.

During the final stages of the Hercynian orogeny, the nature of the faulting apparently recorded a change in structural style. The earlier N-S compression was followed by wrench and dip-slip movements on the NE-SW and NW-SE faults, which formed a conjugate set to the N-directed thrusts.

Heward & Reading (1980) suggested that late-Hercynian wrench faults actively controlled sedimentation as early as the Westphalian. The occurrence of isolated Westphalian basins along the major faults in the Sierra de la Demanda would tend to support this view (although other authors generally favour Stephanian (- Permian?) fault movement).

Although Permian strata are apparently absent in the study area, Sopena et al (1977) report that considerable thicknesses were deposited in other parts of the Iberian Chains (up to 400m), Cantabria (maximum 300m) and the Pyrenees (locally as much as 2 000m). The N Spanish Permian is represented by continental facies similar to those of W Europe. Sedimentation took place in intramontane half-grabens. The Autunian is characterised by black or grey fluvial/lacustrine deposits and includes abundant volcanics and rhyolitic volcanics suggestive of crustal instability.

The onset of red-bed deposition in Saxonian times probably records further tectonic movements during the Saalian phase (Sopena et al, 1977).

8.4. Triassic.

Literature on the Triassic of the Cameros is extremely sparse, being confined to the lithological descriptions of IGME memoirs (IGME, 1970, 1971; Gil Serrano & Zubieta, 1978). However, the Virgili (1977) volume includes several works available on the Triassic of other areas of the Iberian Ranges.

The Triassic of N Spain is developed chiefly in continental facies and is essentially "Germanic" showing the classic tripartite divisions. As in other areas, the Buntsandstein records fluviatile deposition and continued erosional denudation of the Hercynian mountains in arid and semi-arid conditions. A brief marine incursion led to the deposition of Muschelkalk facies carbonates and then fluvial/lacustrine conditions took over as the Keuper facies were deposited.

Although similar to the Trias of much of NW Europe, the N Spanish Trias is especially comparable in its tectonic setting and lithostratigraphic succession to the Trias of the W Shelves of Britain and Ireland described by Ziegler (1982) and Naylor & Shannon (1982).

Gandin et al (1982) classified the Triassic sequences of N Spain into three types - Hisperian, Iberian and Mediterranean - showing progressively greater marine influence and increasing thickness of Middle Triassic carbonates to the E. The Middle Triassic transgression brought the temporary installation of marine conditions as sea-levels rose and shorelines progressively encroached onto the Iberian massif. The sea advanced from the Tethys to the E and perhaps also from the incipient Bay of Biscay rift to the N.

The division between the areas of Iberian and Hisperian Triassic is defined by the disappearance of Muschelkalk facies. The Triassic of the study area (see section 2.3) has a reduced thickness of Muschelkalk (½-2m), (which is reminiscent of SW Britain) and hence just qualifies as "Iberian". Sporadic and short-lived marine incursions during later Triassic times were responsible for the deposition of Keuper evaporites in SW Britain and Cantabria, but in the study area the proximity to the meseta resulted in the deposition of more marginal facies so that evaporites are less important and siliciclastics more dominant than further to the N or E.

As outlined in section 2.3, halokinesis of Keuper evaporites (Brinkmann & Lögters, 1968) was a significant factor in the control of Mesozoic subsidence and facies in Cantabria (Garcia-Mondejar, Hines, Pujalte & Reading, 1985), and the Keuper forms a décollement horizon for Tertiary thrusting there. In contrast, in the study area halokinesis does not appear to have been more than locally significant around button anticlines (eg at Hortezuelos, Tejada and Rupelo; see fig 1.6, and detachment during Tertiary thrusting mostly occurred on other slip-planes within the "Carniolas" and "Purbeck" (fig 8.81).

The Buntsandstein isopach map (fig 8.7) of Garrido-Megias & Villena-Morales (1977) shows the configuration of Triassic basins in Iberia. The Iberian Chains already formed an identifiable trough at this time between the meseta and the Ateca-Calatayud Ridge.

In the study area, the basal Buntsandstein conglomerates rest unconformably upon the Cambrian; further to the N in the Demanda they lie on Tremadocian and Westphalian strata. According to Virgili, Via et al, (1977), this unconformity is the product of late Hercynian movements which resulted in the establishment of moderate to high palaeoreliefs, especially in the W, which they suggested had been accentuated at the start of the Mesozoic. I prefer to explain the unconformity as the result of sedimentary onlap onto a Palaeozoic pediment with relief produced by a combination of earlier faulting during the Late Carboniferous or Permian (see section 8.4) and (?limited) Triassic reactivation of the same faults.

Elsewhere in W Europe, the Triassic was an important time of rift development. The isopach map (fig 8.8) gives an indication of the subsidence pattern. Areas bordering the Bay of Biscay rift system received thick sequences of Triassic sediments as differential subsidence initiated during the Late Permian became accelerated, especially during the Late Triassic. Extension is further documented by basaltic volcanism in Aquitaine and Cantabria, and, more locally, in the N of the Sierra de la Demanda, (Triassic "ophites").

8.5. Rhaetic/"Infralias".

The onset of Carniolas facies deposition records a time of rising sea levels. Sbeta (1976) reported that the Lower Carniolas (~?Rhaetic) is dolomitic with gypsum and anhydrite pseudomorphs; the Middle Carniolas (~?Hettangian) has abundant evaporite pseudomorphs and is extensively brecciated; while the Upper Carniolas (~?Lower Sinemurian) is calcarenitic and oolitic with algal lamination, pellets and an increasingly diverse marine fauna (see fig 8.9). This sequence is approximately consistent with the Exxon Global Coastal Onlap Curve (fig 8.10) which shows latest Triassic transgression, a regressive phase in the early Liassic and renewed transgression afterwards.

This, then, appears to be a eustatically-controlled sequence. Although Garrido-Elorza (1969) reported an unconformity at the base of the Carniolas in the extreme NW of the study area near Los Ausines, in other areas the succession is apparently conformable and, as in SW Britain, there is little evidence that the Rhaetic tectonic pulse ("Early Cimmerian Phase" of Ziegler, 1982) was as important as in Central and Eastern Europe.

The Rhaetic was a time of general transgression over NW Europe as the Tethyan seas advanced from the Burgundy Trough and through the Bay of Biscay rift (Ziegler, 1982). In SW Britain, the latest Triassic - earliest Jurassic sequence is represented by Rhaetic marine shales (Westbury Beds) and marginal evaporitic carbonates (Penarth Group: stromatolitic Cotham Member) overlain by calcareous marine shales with rare calcarenites over basement highs (eg Mendips & Glamorgan). In the W Shelves, shallow marine carbonates were deposited, but brackish and fresh-water microfaunas were reported from the base of the succession in the Fastnet Basin by Naylor & Shannon (1982).

Although these sequences are demonstrably shallow- and marginal-marine, and record a parallel facies evolution to that of the W Cameros, the closest analogue to the N Spanish sequence is found in the Aquitaine Basin where transgressive Hettangian carbonates are overlain by up to 900m of cyclical evaporites. Important evaporites of this age also occur in E Canadian basins (Naylor & Shannon, 1982).

Very thick Lower Jurassic sequences occur on all of the W Shelves (W Approaches: 2 000m; Celtic Sea: 1 500m, and Bristol Channel: 1 000m. In contrast, Sbeta (1985) recorded a mere 250m Carniolas sequence at Aguilar and this is further reduced to only 60-80m in the study area (IGME, 1970; Gil Serrano & Zubieta, 1978; Quintero Amador et al, 1981) although as the Carniolas is apparently utilised as a décollement horizon in the study area, values of thickness may well be underestimated.

The W Cameros Carniolas sequence demonstrates very shallow environments and limited subsidence, both of these factors reflecting proximity to the Iberian massif. Climate may have beem more arid and circulation more restricted than in SW Britain but similar to the Aquitaine Basin.

8.6. Middle and Upper Lias.

The sequence of limestone-marl alternations which characterises the Middle and Upper Lias of the W Cameros reflects sea levels that were higher than during the deposition of the Carniolas facies.

Valladares (1976, 1980) reported that by the Upper Sinemurian uniform marine conditions had been established over the study area, with the exception of Valparaiso where the occurrence of cross-bedded oolites indicates somewhat higher energy, shallower water conditions.

The change from the "Limestone Lias" of the Upper Sinemurian to the "Marly Lias" facies of the Pliensbachian - Toarcian is the result of continued transgression, but wackestones at Valparaiso (fig 1.6) indicate that conditions continued to be somewhat shallower there. Moncalvillo may also have experienced slightly shallower conditions than the rest of the area during the Upper Lias.

Valladares' (1976) isopach map (fig 8.11) indicates thicknesses increasing towards the NE (away from the meseta). The site of the Tertiary Ebro Basin may have been a positive block during Jurassic times, but although thicknesses decrease in that direction NE of the study area, this "Ebro Massif" was too distant to influence sedimentation in the W Cameros.

Although in broad terms the Liassic Cameros may be envisaged as a stable and gently subsiding NE-facing shelf, with facies changes chiefly controlled by eustacy, the shallow conditions at Moncalvillo and Valparaiso point to local tectonically-induced facies variations. Both these localities lie close to major NW-SE faults and Valparaiso also lies very near to the postulated NE-SW Jaramillo-Covarrubias

Fault. Faults may well have influenced sedimentation as differentially-reduced subsidence resulted in shallower and higher-energy conditions over highs, or "swells", while deeper and lower-energy conditions prevailed in the lows, or "basins".

The Liassic facies of the W Cameros are essentially similar in character to those described from the Lias of S Britain. Analogous limestone-marl rhythms in these areas have recently been interpreted as the product of orbitally-forced climatic cycles (Weedon, 1986). The Liassic of the study area may reflect the influence of similar large-scale factors, although their effects were superimposed on a sequence recording generally somewhat shallower-water conditions.

The vertical facies evolution demonstrates some parallels with that of S England. For example, it has been reported that the Lower Toarcian is absent in the W Cameros; in Dorset this interval is condensed within the "Junction Bed". Jenkyns & Senior (1977) suggested that extension was renewed during the Toarcian in S England. Certainly, the tectonic controls on sedimentation in the two areas appear to have been comparable (see Sellwood & Jenkyns, 1975; Whittaker, 1975; Hallam & Sellwood, 1976; also Kent, 1978, 1980), although the presence of significant quantities of clastics (eg Downcliff, Thorncombe and Bridport Sands) in the S England Lias may reflect more active faulting and lesser dominance of shelf carbonate production there. Fig 8.12 shows a highly simplified model for the production of basins and swells as a result of basement faulting.

8.7. Aalenian and Bajocian.

The Toarcian and Aalenian are thin (~30m) and the presence of hardgrounds in the Lr - M Jurassic sequence (eg at Tejada) indicates that sedimentation was not continuous. IGME (1971) reported that the base of the Toarcian is marked by a hiatus of regional extent, although this does not seem to have been recognised by other authors.

The apparent continuation of the Aalenian in limestone-marl alternations indicates broadly similar conditions to those of the Lias, although an increase in the proportion of limestones and the presence of wackstones might indicate a degree of shallowing and the more restricted fauna may point to less open marine conditions (Faulkner, 1985).

Our work has demonstrated an important discordance within the Aalenian - Bajocian sequence in the NW of the study area. The discordance is an important marker horizon useful in mapping (Faulkner, 1985). A sequence of biopelmicrites passes up into sponge bioherms which demonstrate significant relief (fig 8.13), and are overlain by calcarenites made up of structureless angular grains of micrite deposited onto the undulating bioherm surface.

The coarse clastic grainstones then gradually pass up into mud/wack stones containing thin bivalve and gastropod shells and rare black (?algal) clasts (fig 8.14) and stromatolites.

A similar sequence was reported by IGME (1971) from the SE of the study area including "synsedimentary breccias, oomicrites and black pebbles".

Ureta & Kalin (1985) discussed the early Middle Jurassic of the Cameros Basin and noted the presence of discontinuities and reductions in thickness in the area bordering the Sierra de la Demanda. Fig 8.15 shows their inferred sedimentary evolution for the basin in the U Toarcian - Lr Bajocian. They suggest that facies evolution in the NE was probably controlled by tectonics while sedimentation in the "Central Zone" (Canales de la Sierra area) and in the SW (the study area) was primarily influenced by short-term changes in sea-level.

Gomez (1979) described analogous early Middle Jurassic discontinuities in the SE of the Iberian Chains. Fig 8.16 shows his model for the evolution of these discontinuities in that area.

Discordances and unconformities are a common feature of the Early - Middle Jurassic transition of NW Europe. Regressive facies are

developed as far afield as N Germany (Ziegler, 1982) and E Canada (Naylor & Shannon, 1982). The regression coincides with the onset of sea-floor spreading in the Central Atlantic and the W Tethys.

This major rifting pulse is termed the "Mid-Cimmerian" tectonic phase and was expressed on the NW European continental shelf by intensified extension in the N Sea and in the Porcupine Trough. Early Jurassic strata are often truncated in these areas and deltaic deposits accumulated in the N Sea while continental conditions prevailed in the Porcupine area.

Gomez (1979) reports the occurrence of early Middle Jurassic volcanics further to the E in the Iberian Chains along the NW-SE trending Caudiel and Alcublas Faults. These volcanics were extruded at the Aalenian - Bajocian boundary (previous minor phases of volcanism had occurred along the Caudiel Fault during the Middle Pliensbachian and Early Toarcian).

It is easier to appreciate the causes of the regression and tectonic movements evident in the W Cameros when they are viewed in this wider context. Within the study area, it is likely that differential subsidence was controlled by the same faults as those active in the early Jurassic, just as in Britain the Mendip and Market Weighton swells continued to be active during the Middle Jurassic (Sellwood & Jenkyns; Kent, op. cit.). Indeed, the Aalenian - Bajocian facies and the geometry of their discordances are comparable with those of the thin condensed sequences in the laterally-equivalent Inferior Oolite of S England.

8.8. Bathonian and Callovian.

The Bathonian - Callovian of the study area is a marine limestone sequence which shows a gradual upward change from relatively low-energy wackstones and marls at the base through high-energy shallow-water packstones and grainstones to bituminous limestones and marls. The lower part of the sequence contains only thin-shelled

bivalves and gastropods, but the middle part displays an open marine fauna which also includes ammonites (Macrocephalus) and brachiopods (Faulkner, 1985). The upper part is present only in the NE of the study area near Barbadillo del Pez and in the vicinity of Canales de la Sierra, although the facies present (dark, well-bedded organic-rich limestones and marls) appear to be similar to those described by Bulard (1972) from the NE Iberian Chains (Lr to basal U Callovian) and contain the same ammonites (Quenstedtoceras lamberti SOW.).

The facies evolution is indicative of transgression from restricted internal platform (Bathonian) through agitated open shelf (Early Callovian) to organic-rich muddy marine (up until earliest Late Callovian). This interpretation is consistent with the conclusions of Sbeta (1976) who noted that the Bathonian of the Aguilar del Campoo region contained few ammonites and suggested that the Lower Callovian limestones and marls of that area represented deeper-water conditions. However, previous studies in the Cameros have often failed to record the organic-rich facies and have erroneously interpreted the sequence as regressive.

The presence of a lower, fine grained and restricted unit overlain by bioclastic and fossiliferous (possibly condensed) and organic-rich facies is again reminiscent of the successions of SW England and N France (eg "Great Oolite Series" - Bathonian; "Cornbrash" and "Kellaways Beds" - Lower Callovian; Lower Oxford Clay - M and U Callovian). This demonstrates the continued parallel evolution of N Spain and S Britain.

The facies sequences largely record global sea level rise at this time. However, on a local scale, facies distributions probably continued to reflect tectonic controls. The Lower Callovian near Valparaiso (fig 1.6) is an especially coarse bioclastic limestone which may indicate locally higher energy conditions there (possibly shallower water or exposed platform edge). Certainly, in S England, facies variations coincided with major (Palaeozoic) fault lines: for

example the Bristol Channel - St Valery line marked the approximate edge of the Bathonian carbonate shelf (Sellwood, Scott et al, 1985).

Thus, although it is not possible to produce facies distribution or isopach maps for the Callovian (because it is absent from the S of the study area), it is likely that differential subsidence continued along the same faults as in the earlier parts of the Jurassic.

The local absence of parts of the Bathonian - Kimmeridgian succession is a common feature in many parts of NW Europe owing to the presence of a number of complex superposed unconformities (see next section and compare Beach, 1984; Badley, Egeberg and Nipen, 1984).

Gomez (1979) reported that the Lowermost Callovian is often absent from the SE Iberian Chains. Mensink & Mertmann (1984) reported two major disconformities in the Lower Callovian of the NW Iberian Ranges (including the study area), and attributed them to regressions. The presence of quartz clasts in the Lower Callovian may imply denudation and possibly tectonic uplift of basement massifs.

The frequent absence of part or all of the U Callovian and of the Lr Oxfordian in the study area and in many other sectors of the Iberian Ranges is also particularly striking. Bulard (1972) recorded rare cases of U Callovian and Lr Oxfordian faunas in condensed ferruginous horizons in the SE Iberian Chains, although elsewhere in Spain and Portugal sediments of this age are frequently missing. Bulard (op. cit.) favoured a progressive E-ward retreat of the seas from the end of U Bathonian times and during the Callovian (fig 8.17). However, the bituminous marine facies of the M and U Callovian are indicative of uniform low-energy conditions and probably of relatively elevated sea levels (see above) so that it is reasonable to infer that the dark limestones were probably deposited over the whole area.

In many other parts of NW Europe, organic-rich "Oxford Clay" facies continue into the Oxfordian without major break until the rapid change into "Corallian" facies (see section 8.9). However, Lower Oxfordian bituminous facies are absent from the W Cameros, and indeed

faunas of this age are very rare throughout the Iberian Chains (Gomez, 1979).

There are two possible explanations for this. Either:

1) The Lower Oxfordian was not deposited. This would imply anomalously early and tectonically-controlled regression.

2) The Lower Oxfordian was deposited, probably in similar facies to the U Callovian, but was subsequently removed along with all or part of the Callovian as a result of erosion associated with regression immediately prior to the deposition of the Talveila Formation.

The second explanation is the most likely. The rapid change of facies observed in S England between the Oxford Clay and the Corallian clearly documents regression. This regression, the "Lusitanian Crisis" of Mouterde (1972), coincides with the onset of sea-floor spreading in the S Atlantic to the S of Walvis Ridge during the Middle Oxfordian and was probably tectonically controlled. The tectonic effects of this event appear to have been particularly pronounced in Iberia as witnessed by the extensive truncation of pre-Upper Oxfordian strata.

The Lusitanian Basin of S Portugal began to experience more rapid subsidence at this time as Cabaços and Vale Verde Formation lagoonal to brackish carbonates were deposited during the Upper Oxfordian (Wright & Wilson, 1985). The Lusitanian Basin is bounded by the NE-SW Hispano-Portuguese Fault and the subsidence may have been controlled by faulting on that trend. If this were the case then it would be reasonable to suppose that the NE-SW faults transecting the study area (and the rest of N Spain) were active during this phase.

8.9. Talveila Formation (Upper Oxfordian - Lower Kimmeridgian).

The Talveila Formation is developed only in the S of the study area, where it lies unconformably upon older marine Jurassic strata. At Hortezuelos, the Talveila Fm rests on a reddened and karstfied surface of Bathonian limestones (fig 8.18); the Callovian is absent there. In the San Leonardo - Talveila area, the presence beneath the

Talveila Fm of more ferruginous limestone facies reported by Quintero Amador (1983) suggests that Lr Callovian strata are locally present.

The cross-bedded arenites which dominate the lower part of the sequence probably indicate shallow marine deposition. The sands are notable for the presence of synsedimentary extensional normal faults (figs 8.19 - 8.21).

Carbonates are particularly abundant towards the top of the succession and contain a variety of faunal elements (Diaz-Molina, Yébenes & Goy, 1983). Restricted charophyte-ostracod faunas dominate, but corals are present near the top, indicating more marine influence later.

Benke (1981) dated the coralliferous beds at Talveila as Lower Kimmeridgian in age. Corals are particularly widespread in the Iberian Chains during this interval (Errenst, 1984). Quintero Amador et al (1981) quote a "Sequanian" (uppermost Oxfordian - Lr Kimmeridgian) age based on the microfauna of the limestones. The sandy beds at the base of the Talveila Formation are more difficult to date as zonal fossils are absent (section 2.7). However, a M - U Oxfordian age is likely.

Quintero Amador et al (1981) commented on the presence of "secuencias positivas" (shallowing/fining sequences) from coarse sands to biosparudites, sandy sparites and micrites. Diaz-Molina et al (op. cit.) interpreted the succession as a barrier-island - lagoon sequence, suggesting that the vertical arrangement of facies (fig 8.22) resulted from the migration of facies belts in a direction perpendicular to the shoreline. In fact, the age and variety of these facies and the presence of corals and shoaling-upward cycles invites immediate comparison with the Upper Oxfordian - ?basal Kimmeridgian "Corallian" beds of S England. Talbot (1973) described well-defined "shoaling upward" cyclic sequences (fig 8.23) and interpreted them as the result of a series of rapid eustatic sea-level rises followed in each case by periods of coastal progradation. Alternatively, the

cycles could be explained by uniform sea-level rise with superimposed tectonically-induced pulses of sediment input.

The detailed stratigraphy put forward by Talbot for the Corallian (recently revised in detail by Wright, 1986) is shown in fig 8.24. Talbot noted the wide occurrence of the four cycles over Britain and France. I tentatively suggest that the at least some of these cycles might be recognisable in the W Cameros, in which case the Talveila Fm succession may be correlatable with the stratigraphy of the Corallian. The two sequences also demonstrate comparable vertical facies evolution, and certain horizons in the Corallian have approximate facies equivalents in the Talveila Fm (eg oolites; Ringstead coral bed), suggesting that the environments represented were comparable. Fig 8.25 shows the depositional model Talbot proposed for the Corallian. The chief difference between the S England and W Cameros facies is the presence of more clay in the Corallian (eg in Dorset). This may be a reflection of climatic differences between N Spain and S England (or of less active tectonism in S England).

Discussion of facies distributions within the Talveila Formation is hampered by its absence from the N of the study area and by its generally poor outcrop in the S. However, important differences are immediately apparent in thickness as even at Pinilla de los Barruecos in the S, the Talveila Fm is only 10-15m thick in comparison with 30m at Hortezuelos and ~50m at Talveila. The facies also appear to show lateral variation. For example, the succession at Hortezuelos is dominated by calcareous arenites with only relatively minor carbonates at the top, while at Talveila carbonates are abundant, especially towards the top of the sequence. Rapid lateral facies and thickness changes would be expected in a marginal marine deposit of this type. Indeed, such variations are a prominent feature of the S England Corallian (Wilson, 1968) where they may be related to the influence of the London-Brabant landmass which may have been rejuvenated at this

time. Similarly, the clastic supply in the Talveila Fm may have been controlled by periodic tectonic uplift of the Iberian meseta.

Our mapping has shown that the S part of the study area is transected by numerous NE-SW faults. In the Talveila area, the carbonates outcrop only on the fault blocks to the SE of (and within 1km) of the village (fig 8.26). The intuitive explanation would be that synsedimentary fault movement restricted carbonate deposition to particular fault blocks. However, consideration of thicknesses suggests the more likely explanation that the faulting largely post-dated the deposition of the Talveila Formation and allowed preservation of the carbonate-dominated upper part there but resulted in its removal elsewhere. Thus the apparent lateral change in facies may be largely explained by fault-controlled variable truncation of a sequence exhibiting vertical facies heterogeneity.

In much of NW Europe, the onset of the Kimmeridgian is marked by transgression and return to the deposition of organic-rich muds (Kimmeridge Clay facies). Gomez (1979) reported the presence of marly facies (Sot de Chera Marls) in the lowermost Kimmeridgian of the SE Iberian Chains. In the study area, shallow water conditions prevailed in the Lower Kimmeridgian and deeper water facies were either not deposited or were removed prior to the deposition of the coral beds.

Important changes began to occur in Iberia at this time. The Kimmeridgian - Tithonian saw a major regression in the Iberian Chains and in NW Europe generally. Tectonic movements occurred in the Lusitanian Basin of Portugal as subsidence was further accelerated and the lacustrine-lagoonal carbonates of the Cabaços and Vale Verde Formations were superceded by the rapid influx of Kimmeridgian clastics (Wilson, 1979). NE-SW faults were probably active once more.

8.10. Geometry of Late Jurassic Unconformities.

As we have seen in the preceding sections, the study of Late Jurassic facies distributions is complicated by the presence of

important unconformities. Fig 8.27 summarises the Middle and Upper Jurassic marine successions from four sectors of the study area.

In the S, marine Callovian is absent and U Oxfordian Talveila Formation clastics lie unconformably upon the marine Bathonian. The Talveila Formation sands are in turn overlain by continental clastic deposits of the Senora de Brezales Formation. The limestone facies of the upper part of the Talveila Formation (Lower Kimmeridgian) are usually absent because of truncation beneath this unconformity.

In the N sector, the youngest marine Jurassic strata are of Lower Callovian age ("orange beds" facies) to the W, and of Middle - Upper Callovian age ("bituminous limestones facies") in the E. There are minor discontinuities in the Lower Callovian, but the Talveila Formation is completely absent in this zone so that Senora de Brezales Formation clastics lie directly on the Callovian.

Thus there are two significant unconformities - one pre- Senora de Brezales Formation (probably Middle Oxfordian) and a later one - Pre- SdB Fm (-?post- Lr Kimmeridgian). The superposition of these two unconformities causes problems in unravelling the depositional history of the study area. The absence in the S and NW of all or part of the Callovian and the absence of the U Oxfordian - Lr Kimmeridgian from the N are two cases in point. Although it is reasonable to assume that the trangressive M and U Callovian facies were originally present over the whole area (see sections 8.8 & 8.10), this is less certain for the marginal facies of the Talveila Fm. U Oxfordian and Lr Kimmeridgian deposits are present in the NE and SE of the Iberian Chains and might have had a more widespread distribution in the study area prior to removal before the deposition of Senora de Brezales Fm clastics.

The apparent geometry of the unconformities and a model for their evolution is shown in fig 8.28. The reality is undoubtedly more complicated than illustrated by this very simplified diagram. Effects of local faulting are likely to have been significant in the localised preservation of strata beneath unconformities (see Badley, Egeberg &

Nipen, 1984; Beach, 1984) and thus it is significant that although the outcrops of Middle and Upper Jurassic marine strata are widely-spaced across the basin, these outcrops are all on present-day structural highs, with the exception of those bordering the Sierra de la Demanda. It is at these latter localities in the NE that M and U Callovian strata are preserved. Sediments of this age might thus be much more widely present in structural lows than is evident at first appraisal of the surface data. This has important implications for hydrocarbon exploration as the bituminous limestones and black marls of the M and U Callovian are good potential source rocks, (and source rocks are scarce in N Iberia). In depositional "basins" the facies, if present, are likely to be even more anoxic and organic-rich. It is notable that the few wells drilled in the Cameros Basin have been sited on the structural highs of the SW and SE margins (presumably in search of anticlinal and/or stratigraphic traps) and have thus been sited exactly where these strata are most likely to have been absent.

In the study area, the unconformity between the Senora de Brezales Formation and the underlying strata is the most impressive of the intra-Mesozoic unconformities. It marks a major change in basin configuration and a change from the relatively uniform gentle subsidence and generally shallow marine conditions of the Rhaetic - Lower Kimmeridgian to the onset of the more rapid, more strongly localised subsidence and continental sedimentation which characterised the latest Jurassic and Early Cretaceous.

Many authors have recognised a Late Jurassic - Early Cretaceous unconformity in NW Europe. This "Late Cimmerian" unconformity is variably dated as earliest Cretaceous (eg Ziegler, 1982) or Volgian [Tithonian] (Beach, 1984). In fact, in NW Europe, as in W Cameros, accurate dating of this surface is complicated by the superposition of several unconformities and by problems of stratigraphic resolution in the marginal/restricted Upper Jurassic facies.

In the study area, the basal Senora de Brezales Formation unconformity can only be dated as post-Lower Kimmeridgian and pre- (?Lower) Berriasian. In NW Europe, tectonic movements probably occurred during the Kimmeridgian - Badley, Egeberg & Nipen (1984) noted that the Kimmeridge Clay of the North Sea is often unconformable on the underlying Heather Formation (?Oxfordian) and suggested that this was the result of tectonic activity. This idea supports the inference of Hallam & Sellwood (1976), who invoked intra-Kimmeridgian tectonics to explain the apparent rapid increase in subsidence implied by thick Kimmeridge Clay sequences and the presence of Lower Kimmeridgian sands in Normandy. Andrews, (1985) has also recently postulated intra-Kimmeridgian tectonic activity in S England to explain shortening and small-scale thrusting observed in Kimmeridge Clay dolostones of Dorset, although these may be diagenetic features.

The widespread truncation of earlier strata has often been interpreted as evidence of compression (viz. the "Cimmerian Orogeny"). A more modern explanation (eg Beach, 1984) would interpret the complex association of superimposed unconformities at this time as the product of variable rates of extension. Variable extension rates result in episodically-accelerated subsidence as the active tectonically-controlled subsidence of active rifting phases is added to the more passive, thermally-controlled / compactional subsidence from earlier phases. The effect of superimposing eustatic variations onto such complicated subsidence histories is to produce exceedingly complex sedimentary onlaps during sea level rises and truncation during times of lower base level, particularly on the footwall crestal axes of rotated fault blocks, (where Jackson & McKenzie, 1983 estimated that footwall uplift may amount to 10% of the hanging wall subsidence).

Fig 8.29 shows hypothetical reconstructions of the geometry of the unconformities predicted after accounting for probable effects of local faulting. Consideration of these unconformities is of great importance in the study of the Senora de Brezales Formation.

8.11. Senora de Brezales Formation.

8.11.1. Basal Contact.

The Senora de Brezales Formation (SdB Fm) post-dates a period of emergence which began some time after the deposition of the Talveila Formation carbonates (Lower Kimmeridgian). The combined effects of tectonic activity and erosion during this period resulted in the formation of a pediment with significant relief so that the SdB Fm rests unconformably on a variety of different strata. In places, the pediment was incised by basal SdB Fm channelised conglomerates; in others erosion was less powerful and karst surfaces predominate.

8.11.1a. Karst surfaces.

The SdB Fm lies on karstified Lr Kimmeridgian limestones at Talveila (fig 8.30) and on karstified thin lower Talveila Formation limestones at Pinilla de los Barruecos (figs 8.31 - 8.34) and Hortezuelos (fig 8.35). The passing of red SdB Fm sands into cracks and along joints is very evident. The Hortezuelos outcrop is apparently disturbed by a fault, and scarp or karst conglomerates drape the karst surface (fig 8.36). Nearer to La Yecla (fig 8.37), and probably at Ladera, the SdB Fm overlies lower Talveila Formation calcareous sands and the palaeokarst is not so impressive.

In the N, the SdB Fm overlies marine Callovian strata (fig 8.38; see previous section) and in interchannel areas the marine Jurassic strata are heavily altered and brecciated, locally forming palaeosoils (figs 8.39 & 8.40). In both localities, laminated limestone crusts (section 3.2.3, facies 2) are developed in association with these palaeosoils.

8.11.1b. Conglomerates.

The SdB Fm conglomerates are often extensively channelled into the pediment (figs 8.41 - 8.43), particularly in the N, where the

Talveila Fm and the basal red sands of the SdB Fm are absent. The siting of these channels shows control by NE-SW faults (fig 8.44).

The composition of the conglomerates strongly reflects the underlying palaeogeology. For example, the first conglomerates at Hortezuelos contain clasts only of grey micritic limestone. These clasts appear to have been extensively derived from laminated crusts deposited in interchannel areas, and perhaps also from upper Talveila Formation lagoonal carbonates. The red sands above these first conglomerates consist mainly of quartz, probably including a component reworked from the lower Talveila Formation sands. The overlying polymict conglomerates may date the exposure of the underlying marine Jurassic sequence, since the abundance of reddened oolitic limestone clasts indicates erosion and reworking of Jurassic regressive facies (ie Bajocian, Bathonian, Upper Oxfordian ??and perhaps Callovian). The clast derivation is discussed in fig 8.28.

In the NW of the study area, polymict conglomerates lie directly on fossiliferous Lower Callovian limestones. The presence in these conglomerates of abundant and extremely coarse clasts of Lower Callovian indicates that the clasts were derived directly and locally from the pediment.

In the NE, polymict conglomerates overlie the Marine Jurassic directly, and at Castrovido the clast lithologies include Lower Callovian limestones and packstones/grainstones (perhaps of Upper Oxfordian or Middle Jurassic age). These conglomerates are in turn overlain by red sands which may indicate continued local erosion of Talveila Formation sands, or alternatively which may reflect sand derivation from different sources (eg Palaeozoic or Triassic).

Thus study of clast lithologies in SdB Fm conglomerates reveals the strong control of palaeogeology on conglomerate composition. Vertical changes in clast assemblages may record the progressive erosion history of the source areas (which in most areas were the pediments of uplifted adjacent fault blocks).

The polymict conglomerates of the Senora de Brezales Formation are lithologically similar to the basal conglomerates described by Sbeta (1976) from the Aguilar Fm of SW Cantabria. Comparable environmental and tectonic conditions were also operative there at this time. The palaeogeology of the pediment in the two areas would have consisted of marine Jurassic limestones developed in similar facies. This permitted the deposition of locally-derived fault-controlled conglomerates of comparable composition.

8.11.2. Tectonic controls on facies in the Senora de Brezales Formation.

A schematic NW-SE stratigraphic section for the SdB Fm in the N of the study area (fig 8.45) demonstrates the tectonic control on facies which is evident if we take the base of the overlying Rupelo Formation carbonates as a reference datum. On a large scale, variations in facies can immediately be attributed to tectonics. For example, the red sands are restricted to the area E of the Jaramillo - Covarrubias Fault, which is a major NE-SW trending fracture. The coarsest basal conglomerates (clasts 1m in size) are found just to the E of this fault (fig 3.31). Immediately to the W of the fault, at Valparaiso and at Mambrillas de Lara, basal conglomerates are generally absent, karst surfaces are present and the SdB Fm is thinned or absent. Moving to the W, conglomerates are present firstly on isolated fault blocks and eventually in fault-controlled channels. The picture emerges that the conglomerate channels followed the structural lows. On intervening highs, karst erosion dominated, acting to equalise the pediment relief. The fault blocks can be envisaged as a series of depressed or uplifted piano keys, the relative subsidence of which controlled thicknesses and facies. Complications to this simple pattern arise because, in some instances, the thickest sequences occur on fault blocks (piano keys) which are now uplifted. This is interpreted as the result of subsequent reversal of the sense of fault

movement. This reversal might have occurred relatively early in some cases since not all of the reversed faults appear to cut the overlying Rupelo Formation strata (see fig 8.40; also figs 8.46 -8.48).

An analogous situation occurs in the S of the study area. This sector is also cut by numerous NE-SW faults, which locally control the presence of SdB Fm conglomerates. Movement on the faults resulted in variable preservation of the Talveila Formation (see sections 8.9 & 8.10). These faults were also subsequently reactivated, (sometimes with an opposed sense of displacement), as evident from offset of the overlying Rupelo Formation carbonates.

NE-SW faults also cut the section at Moncalvillo and apparently controlled the deposition of SdB Fm conglomerates there (although the faults cut the overlying strata and thus clearly have been reactivated after SdB Fm times).

Changes in facies and thickness also occur across strike. However, the exposure is not continuous in this direction and the basin structure only allows comparison between outcrops on NW-SE trending anticlines often several kilometres apart.

Despite the problem of discontinuous exposure, differences are immediately apparent in the S sector of the study area between generally thick sequences to the S of the San Leonardo Fault and thinner sequences to the N (fig 8.49), (although this simple trend is complicated by variations along strike). Similarly, in the N sector we can compare the considerable SdBFm thicknesses S of the Quintanilla-Castrovido Fault with the much reduced thicknesses immediately to the N of it.

The red sands show particularly impressive thickness variations: from up to 40m thick in areas S of the San Leonardo Fault and 20-50m thick S of the Quintanilla-Castrovido Fault (between Jaramillo Quemado and Castrovido) to very thin or absent in most other areas.

From this pattern we could infer "down to the S" displacement across the NW-SE faults at this time (fig 8.50). However, comparison

of sections directly across faults is often not possible so that apparent variations observed across strike may be partially attributable to along-strike variations. In any case, the exposure is not complete enough to rule out antithetic fault movements of the type shown in fig 8.51. However, NW-SE faults probably were active during deposition of the SdB Fm, apparently bounding large domains of differing subsidence and with different detailed stratigraphies.

Of the NE-SW faults, a few were major features (eg Jaramillo-Castrovido Fault) but many others were of lesser importance and exerted only local control on facies or thickness.

8.12. Rupelo Formation.

8.12.1. Control of tectonics on facies and thickness.

The Rupelo Formation exhibits a variety of facies (section 3.3). Facies sequences show a generally consistent evolution from marginal/ephemeral lacustrine (Las Vinas Mb, Ladera Mb), to more perennial lacustrine (Mambrillas de Lara Mb), ultimately to desiccating and locally evaporitic lacustrine (Rio Cabrera Mb). As discussed in section 3.7, this sequence probably reflects climatic factors.

However, the Rupelo Formation does show very marked lateral changes in thickness, and individual members (particularly the upper ones) are often absent. These changes are interpreted as the result of two parameters:

1) continuing differential subsidence
2) post-depositional truncation.

Figs 8.52 - 8.54 indicate the tectonic controls during deposition of the Rupelo Formation. Important trends are immediately apparent. The Rupelo Formation is thickest in the NW and SW, but very thin in the NE, in the Moncalvillo anticline and in the SE (except at Talveila). The upper members are absent in the extreme SW and in the

SE, but are generally present in the NW and especially thick at Rupelo.

The maps can be viewed as mosaics made up of areas with differing Rupelo Formation sequences and thicknesses bounded by important NE-SW and NW-SE faults. Very rapid changes can be observed directly across these faults (eg between thick and fully-developed sequences at Rupelo and Mambrillas to thin and partial sequences at Vizcainos and Cubiejo).

The distinction between originally thick, but later severely truncated, sequences and sequences that were originally thin can be made using stratigraphic criteria. For example, at Espejon the thin Rupelo Fm includes the Ladera and Mambrillas de Lara Members as well as several metres of overlying strata assigned to the Rio Cabrera Member so that the thin sequence there reflects an originally small thickness with only minor erosional thinning. In contrast, at Hortezuelos, thin sequences include only the Ladera Member and the reduced thickness is apparently due to truncation.

The presence of obvious stratigrahic gaps above the Rupelo Formation may also indicate truncation. An example of this occurs in the SW of the study area where the (? Lower Berriasian) Rupelo Fm is often directly overlain by, and truncated by, Utrillas facies sands (Upper Albian - Lower Cenomanian).

8.12.1a. Las Vinas Member.

The Las Vinas Member is present only in the NW sector, where it interdigitates with, and locally underlies, the Ladera Member, (to which it forms a partial lateral equivalent). The sandy marl/mudstone facies probably indicate distal alluvial sedimentation in a marginal lacustrine / floodplain fringe setting during a period of raised fine clastic supply (see section 3.3.1). Thickening to the NW suggests clastic influx from that direction.

8.12.1b. Ladera Member.

The Ladera Member consists of marginal lacustrine carbonate facies. In thin sequences it is the only member recognised. All measured sections included some marginal lacustrine carbonate facies and so the Ladera Member has an apparently wide distribution reflecting the maximum extent of Rupelo Formation sedimentation.

In fact, the marginal facies of thin sequences may be laterally equivalent in more fully-developed sequences to more open lacustrine facies (eg Mambrillas de Lara Mb) which coincide with the highest lake stands (see section 2.8). Thus diachroneity of facies poses problems in comparing depositional thicknesses.

However, the Ladera Mb does show important thickness changes across the NW-SE San Leonardo and Quintanilla - Castrovido Faults (tendency to much thicker sequences in the S and to more intense pedogenetic modification in the N), and across the NE-SW Jaramillo-Covarrubias Fault (thick sequences of up to 80m in the NW and very thin sequences of often only 1-2m in the SE). The Ladera Mb is also thick on the N edge of the study area between Rupelo and Torrelara, exhibiting a tendency to marlier sequences towards Torrelara and to thicker sequences (with evidence of very strong pedogenesis in the basal part of the Member) towards Rupelo.

The inferred fault control of thickness is shown in fig 8.52. Tectonic controls on subsidence were gentle, although very localised coarse intraformational conglomerates 2km W of Mambrillas de Lara (fig 8.55) may indicate erosion from the crests of NE-SW fault scarps.

8.12.1c. Mambrillas de Lara Member.

Fig 8.53 shows the thicknesses measured or estimated for the Mambrillas de Lara Member. Absence due to truncation is indicated for sections in the SW sector. The thickest sequences are those at Tenadillas (S of the San Leonardo Fault), at Mambrillas de Lara (to the S of the Quintanilla-Castrovido Fault), and between Rupelo and

Torrelara (on the N margin of the study area). This suggests a similar subsidence pattern to that inferred above for the Ladera Mb. The local presence of a conglomerate horizon between the Ladera and Mambrillas Members (the Torrelara Bed; see section 2.8) may indicate a phase of mild tectonic activity between the deposition of the two units.

8.12.1d. Rio Cabrera Member.

Fig 8.54 shows the thicknesses measured or estimated for the Rio Cabrera Mb, with chert thicknesses also indicated. Significant chert thicknesses are confined to the areas between Rupelo and Torrelara in the N and between Pinilla de los Barruecos and Tenadillas in the S. The widespread absence of the Member in the other areas is largely due to truncation and hampers reconstruction of the subsidence pattern. Significantly, though, the Rio Cabrera Member is present both N and S of the San Leonardo Fault. This suggests either a modified subsidence framework or reduced importance of local tectonic controls.

8.12.2. The Rupelo Formation: comparison with other Late Jurassic - Early Cretaceous and "Purbeck" sequences.

The study of the Late Jurassic and the Rupelo Formation of the W Cameros Basin has revealed several points important in the context of the Late Jurassic - Early Cretaceous history of NW Europe:

1) The Kimmeridge Clay and Portlandian facies are absent in the study area although well-developed in other parts of NW Europe.

This is notable as the facies evolution of the W Cameros was closely parallel with other areas of NW Europe during the Lower and Middle Jurassic, as it was again during the Lower Cretaceous. The absence of the Kimmeridge Clay and Portlandian facies in the study area could have been a product of non-deposition or of subsequent truncation as a result of particularly rapid "Late Cimmerian" extension in N Spain. Alternatively, these facies might not have been deposited, because of regression and consequent retreat of the seas

from this generally less-subsident and more "continental" area. Comparison of Lr - M Jurassic facies and thicknesses in the W Cameros with those of other areas shows that the study area was hardly a "basin", but was indeed a relatively marginal area during this time.

2) The Rupelo Formation succession is very similar to the Aguilar Fm sequence, (although somewhat thinner).

The stratigraphic and sedimentological similarities suggest that the two basins were connected or intermittently communicating.

3) The Rupelo Fm and Aguilar Fm (fig 8.56) "Purbeck" facies may broadly be correlated both sedimentologically and stratigraphically with the Purbeck of parts of the E Iberian Ranges, S England, the Paris Basin and the French and Swiss Jura.

The extensive distribution of Purbeck facies indicates the widespread establishment of comparable conditions. The seas retreated from large areas of the Pangaean continental shelf. This led to the onset of restricted shallow-water carbonate sedimentation in many areas previously characterised by earlier-Jurassic shelf carbonates. Regression occurred in association with important modifications in the pattern and rate of regional subsidence.

4) Although broadly comparable with other Purbeck sequences, the Rupelo Formation is unique for a number of reasons:

a) The apparent absence of marine influence and the dominance of lacustrine environments in the W Cameros - the other "Purbeck" sequences generally show some marine influences, as exemplified by the sequence in the E Iberian Chains (Canerot, 1982). Even the Aguilar Fm of S Cantabria may show transition into brackish/lagoonal environments, (eg Aroco and Arco Somero Fms; Garcia-Mondejar & Pujalte, 1981).

b) Although pedogenesis is a common feature of other Purbeck sequences, the degree and frequency of pedogenic modification is much greater in the Rupelo Formation even than in the Aguilar Formation.

c) The nature of the evaporitic facies within the Rupelo Formation and the Aguilar Formation (yellow vuggy limestones and white cherts) are very different from the Purbeck evaporitic facies of S England, although possibly more similar to those of the Jura.

d) Stromatolites (sensu stricto) are not developed although they are present in some other Purbeck sequences. In contrast:

e) Root action appears to have been sedimentologically much more important in the Rupelo Formation than in other Purbeck sequences.

These differences may largely be attributed to the more "continental" location of the Cameros Basin (apparently cut off from marine influences), which meant that:

i) Tides and fair-weather waves were effectively absent.

This would have resulted in reduced littoral reworking of the marginal facies. Peri-lacustrine or "palustrine" facies would thus have had a higher initial preservation potential.

ii) The response to climatic factors was very rapid.

Large-scale water level variations can occur very rapidly in lakes, especially in shallow lakes which desiccate or expand their areas quickly in response to changes in the net water budget. Lagoonal/marginal marine sequences are more dependent on much longer-term eustatic sea-level changes.

iii) Shoreline energy was low and the waters fresh. This would favour much more widespread colonisation by land- and shallow-water plants.

This would have permitted the more frequent preservation of root traces and of relatively delicate structures (eg charophyte stems).

iv) The basin was isolated from the sea.

In a semi-arid climate, this would have favoured the desiccation of the basin. However, the precipitation of evaporite minerals would have required proportionally much more intense evaporitic concentration of the available "fresh" water than would be necessary in the case of marine or brackish water. The salts precipitated might have been markedly different in composition from those of desiccating

marginal marine basins. Mineral suites produced in lakes closely reflect lake chemistries, which are highly variable and dependent upon the palaeogeology of the drainage area.

5) Alternatively, the intense pedogenetic modification manifested in the Cameros Basin could have been the result of particularly intermittent and slower subsidence.

8.13. Pedroso Group.

Of Groups II and III as defined by Salomon (1982) only Group IIC (≈the Pedroso Group) is present in the study area, having been subdivided here into the limestone-bearing Hortigüela Fm and the clastic Piedrahita de Muno Fm (see section 2.9). Thick sequences of sediments assigned to Groups IIA, IIB, IID and III are present in the E Cameros Basin but absent in the study area, suggesting more intermittent sedimentation and less rapid subsidence in the W.

The absence of Groups IIA and IIB in the study area implies that there was a hiatus between the deposition of the Tierra de Lara and Pedroso Groups. However, it is not generally possible to recognise an unconformity, except in the Hortezuelos area, where Ladera Mb limestones are truncated by sands, marls and limestones (possibly belonging to the Hortigüela Formation). However, reworking of sauropod bones from the "Dinosaur Bed" of the Mambrillas de Lara Mb to the N of Mambrillas indicates an erosive discordance in that area at least.

The Hortigüela Fm includes sands with mica and quartz and the Piedrahita de Muno Fm is almost entirely made up of clastics with locally abundant white mica and some mafic minerals. The underlying Rupelo Formation consists of limestones with very little clastic material. The sharp increase in clastic supply associated with the Group II sediments in the study area indicates erosion of basement massifs, probably as a result of tectonically-induced uplift.

8.13.1. Vertical and lateral distribution of Pedroso Group facies.

The frequent absence of Hortigüela Formation sediments from the SW of the study area renders lateral comparison of facies difficult. However, the presence of oncoidal limestones at Ladera and Hortezuelos suggests that the Hortigüela Formation may have also originally been also present over much of the S sector.

In the NW, the Hortigüela Formation is best developed to the W of the Jaramillo-Covarrubias Fault (although extending slightly across this line to Jaramillo Quemado). Local facies changes occur across more minor NE-SW faults (eg N of Mambrillas de Lara), and the limestones apparently thin and eventually disappear towards the NW. Although the sequence probably did display NW-SE lateral changes in depositional thickness, the observed thinning is probably largely the result of erosion and reworking during the deposition of the overlying Salas Group.

The Piedrahita de Muno Fm clastics are present, but thin, in the S sector to the N of the San Leonardo Fault. They are thick in the NE of the study area (to the E of the Jaramillo-Covarrubias Fault) and in the area of Canales de la Sierra. Thin limestones, often very dark in colour, are also occasionally present.

Lateral changes in facies are difficult to assess because of:

a) the poor quality of exposure;

b) the extremely thick sequences; and

c) problems in corelation (see section 2.9).

Nevertheless, I have attempted to reconstruct the Pedroso Group facies distribution (summarised in fig 8.57). The nature and geometry ("architecture") of the facies was evidently strongly controlled by tectonics.

The restriction of Groups IIA, IIB, IID and Group III sediments to the E of the basin is probably related to the influence of renewed activity on NE-SW faults, perhaps with a wrench component of displacement (Guirand & Seguret, 1985). Although clastics dominate the

Group II and Group III successions, lacustrine carbonates (Salomon & Drillien, 1985) and evaporites (fig 8.58, see Salomon, 1983) are present at several horizons in the E of the basin (Salomon, 1982b, his figs 7.16-7.20).

Sandstones are less common and the sequence is more mud-dominated towards the top in the area to the S of Castrovido.

The change from the carbonate-dominated facies of the Rupelo Fm to the clastic-dominated facies of the Pedroso Group is equivalent to the transition recognised throughout NW Europe from "Purbeck" to "Weald" facies (see discussion in sections 2.8 - 2.9). In the W Cameros, this boundary is a very sharp one, as it is in S England (Kennedy, pers. comm., 1986). Although climatic changes (eg increase in humidity) probably occurred at this time (Hallam, 1983) as the Atlantic began to open, the sharp increase in clastic supply and the acceleration of subsidence rates can only realistically be attributed to important tectonic movements.

The changes in facies and subsidence rates are analogous to those which affected the Lusitanian Basin somewhat earlier, during the Kimmeridgian tectonic pulse (Wilson, 1975), but in this case document the widespread Early Cretaceous renewal of rapid subsidence in NW Europe.

8.13.2. Comparison with other "Wealden" sequences.

In the E Cameros Basin there are up to 4km of U Berriasian - Barremian sediments. Similar very thick continental - "paralic" sequences were deposited in the Channel and Hampshire Basins (800m), W Approaches (1500m), the Celtic Sea (2 000m), Asturias, Cantabria and E Canada (all < 1 000m). These sequences all show evidence of important synsedimentary fault control, related by Ziegler (1982) to the onset of renewed rifting/wrench movements in the Bay of Biscay area. The clastic sequences were apparently associated with erosion of the ancient massifs - the Iberian Meseta, Armorican Massif, Cornish

Platform and London landmass. Extension was locally associated with volcanic activity - for example the Wolf Rock phonolite off Cornwall (Valanginian) and Alexis Fm lavas of Labrador (Early Cretaceous).

The Early Cretaceous sediments in the SE Iberian Ranges also show strong tectonic control and evidence of renewed rifting (fig 8.59, taken from Vilas et al, 1982). Although that area was closer to Tethyan influences, with the result that marginal marine facies were also developed, in areas more remote from this influence continental "Wealden" facies are present (Mas, Alonso & Melendez N, 1982).

Despite their similarity with other oncoidal carbonates in N Spain, the carbonate facies of the Hortigüela Formation have more in common with the marginal marine deposits of the Paris Basin than with the S England Wealden. In contrast, the clastic Piedrahita de Muno Fm readily finds sedimentological equivalents in the fluvio-lacustrine "paralic" clastics of both Cantabria and S England as described by Pujalte (1981) and P Allen (1975) respectively. The tectonic controls on sedimentation seem to have been comparable (see previous section).

The Pedroso Group represents only part of the Wealden succession of the Cameros. The thicknesses in the E are very much greater, indicating subsidence that was even more rapid than in other parts of NW Europe. The presence of lacustrine carbonates and evaporites in the Cameros is indicative of an internally-draining more "continental" setting. Climatic conditions possibly more arid than those developed in other parts of NW Europe facilitated the development of isolated lacustrine and sabkha systems in the central sediment-starved parts of the basin during periods of relatively low clastic input.

8.14. Salas Group.

The Salas Group (Group IV of Salomon, 1982) is a thick unit of conglomerates, sands, muds and rare calcarenites (Chapter 5) developed only in the W of the Basin. The absence of Groups IID and III in the W implies that there was a hiatus before the deposition of the Salas

Group. This hiatus and the switch of more rapid deposition from the W to the E were attributed by Salomon (op. cit.) to tectonic movements ("une restructuration du bassin").

Salomon also noted the abundance in the Salas Group of quartzose conglomerates and feldspathic sandstones, suggesting that this reflected tectonic movements and erosion of the granitic basement of the meseta. Our own studies have shown that the Salas Group is particularly rich in yellow-brownish quartzite pebbles with additional important contributions from Mesozoic strata, eg Bathonian-Callovian, ?Rupelo Fm and Hortigüela Fm limestones. Reworked oncoids are particularly common at some localities and indicate erosion of the immediately subjacent Hortigüela Fm (see section 5.2). The presence of Jurassic limestone clasts is perhaps more significant as it may imply local removal of the entire underlying continental sequence.

Salomon (1983) mentions NE-SW thickness variations with "condensed" sequences in the SW - I interpret this apparent thinning to be the result of greater truncation below the Utrillas facies U Albian - Lower Cenomanian sands. Although Salomon (op. cit.) suggested that facies changes were attributable to NW-SE faulting, our own work (Faulkner, 1985) has indicated that synsedimentary movements on NE-SW faults were responsible for the deposition of immature sands and conglomerates containing Mesozoic limestone clasts in some areas and more mature quartz arenites in others.

As mentioned in section 2.10, the age of the Salas Group is not certain, although a Barremian - Aptian age is likely. In many other areas of N Spain, S Europe and the Gulf Coast of the US, the Aptian is notable for acceleration of subsidence, transgression and the establishment of "Urgonian" carbonate platforms. These facies are absent from the Cameros which was situated well inland of the coast and was subject only to continental influences. Garcia-Mondejar (1982) described continental clastics which form the lateral equivalents of the Urgonian facies in S Cantabria. Some of these sediments

(Cilleruelo de Bezano, Las Rozas, Quintanilla de An conglomerate, La Canal Fms; fig 8.60) may be partial lateral equivalents of the Salas Group.

Pascal (1983) and Garcia-Mondejar (1979, 1985) discussed the importance of tectonic controls on Urgonian sedimentation, stating that E-W, NW-SE and NE-SW faults were active in Cantabria. Canerot (1982) suggested that NW-SE faults were also active in the E Iberian Ranges at this time, while Vilas et al (1982) recognised discreet episodes of tectonic activity during the Hauterivian, Barremian and Aptian.

Tectonic movements thus affected N Spain at repeated intervals during this period. Garcia-Mondejar (1979) attributed these events to the early stages of rifting in the Bay of Biscay. Movements probably also affected SW Britain. The Rockall Trough was formed at this time and volcanic activity is documented by the extensive development of montmorillonitic "Fuller's Earths".

However, despite the tectonic similarities with Cantabria, in SW Britain there are no Urgonian carbonates, these being confined to Tethyan margins. As transgression linked S Europe and Texas, for the first time the the S Biscay margin experienced development that had more in common with Tethyan than NW European areas.

8.15. "Utrillas" Formation (U Albian - Lr Cenomanian) sands.

The Utrillas facies are difficult to date. The hiatus between the Salas Group and Utrillas facies is of unknown length. In the NW of the study area, Utrillas sands lie unconformably upon the Salas Group while in the SW they truncate older strata. Outside the study area, in SW Cantabria and the Sistema Central, the Utrillas oversteps onto Palaeozoic basement. The basal Utrillas conglomerates above the unconformity are rich in white and liver-coloured quartzite pebbles which are presumably reworked from basement massifs, perhaps via Lower Palaeozoic or Triassic conglomerates. The sands above are quartzose or

arkose generally with little clay. Deposition was probably relatively slow with repetitive internal reworking.

The sub-Utrillas unconformity (≃ supra-Urgonian unconformity of Cantabria) is one of the most impressive of the intra-Mesozoic unconformities in the study area and marks the end of rapid differential subsidence. The Utrillas facies drape pre-existing faults with little locally-discernible change in facies or thickness, and during the Upper Cretaceous (see section 8.16), the area experienced relatively gentle and uniform subsidence.

The geometry of the unconformity is outlined and alternative models for its evolution are presented in fig 8.61). The unconformity is of potential economic significance in some localities as thin lignite horizons may be found in the uppermost Salas Group beds if preserved.

The Utrillas facies clearly post-date a tectonic event, probably of Late Aptian - Early Albian age. The sense of displacement on some faults active earlier in the Mesozoic was partially reversed before Utrillas deposition. In SW Britain, a Middle Cretaceous phase of folding (the Ringstead or "Austrian" phase) has long been recognised (fig 8.62). The progressive W-ward overstep there of the M Albian Gault over older strata (fig 8.63) is well-known. Owen (1971) demonstrated that block faulting of the London Platform occurred prior to deposition of the Upper Cretaceous sediments, and Kemerling (1979) discussed a phase of Aptian block faulting and wrench-induced basin "inversion" in the Bristol Channel. The broad, open "Austrian" structures are now better-documented, forming important hydrocarbon exploration targets in S England. In E Canada, too, important tectonic changes affected the Newfoundland Shelf. In that area, Early Cretaceous (Mississauga Fm) clastics pass up into more clastics (Logan Canyon Fm), which, like the Utrillas, continues into the Cenomanian (Naylor & Shannon, 1982). Effects were apparently more minor in the N Sea where Beach (1984) recognised the last major rifting as

Hauterivian - Barremian, interpreting subsequent unconformities as the result of eustatic changes and thermal subsidence.

Certainly, in N Spain, the pre-Utrillas or Supra-Urgonian unconformity could be regarded as marking the end of "rifting" and the onset of "drifting". The Utrillas is a highly diachronous deposit pre-dating the Upper Cretaceous transgression (see section 2.11). One of the reasons for the impressive nature of the unconformity is the major overstep onto and erosional reworking of basement massifs.

The overstep could be explained by simple thermal subsidence analogous to that described by Watts (1982), ("steer's head onlap", fig 8.64). However, this seems to be an oversimplification, as although in the N the passage from the Salas Group to the Utrillas is marked only by a change in lithology with no unconformity, in the S the Salas Group deposits have clearly been erosionally truncated. This might imply renewed extension. Further rotation of fault blocks and footwall uplift (Jackson & Mackenzie, 1983; fig 8.65) at a time of low sea-level stand would have caused significant erosion particularly at the crestal axes of rotated fault blocks.

The reversal of the sense of displacement of some faults may have been caused by modified stress trajectories. Although large-scale wrench movements may ultimately have been responsible, on a local scale the fault reversals can be explained by changes in the relative importance and sense of dip-slip and wrench displacement on the basin-bounding and cross-graben "transfer" faults (Gibbs, 1984). However, Hossack (pers. comm., 1986) interpreted the reversal of fault displacement as the result of local inverse reactivation of normal faults as thrusts, this apparently being coincident with a Middle Cretaceous phase of shortening in the Pyrenees (Hossack, 1985).

To summarise, then, the Middle Cretaceous (Austrian) tectonic phase in the study area encompassed both pre/intra-Salas Group movements (major extension and rifting) and pre-Utrillas (major overstep, local compression and/or extension).

8.16. Upper Cretaceous.

During the Upper Cretaceous, the study area formed part of the N Castillian Platform, a large area of very shallow, dominantly carbonate-producing seas open to the Atlantic in the N and to the Tethys in the S. Work on Upper Cretaceous sequences has often stressed the control of eustacy on vertical facies evolution (eg Floquet, 1979). However, the influence of tectonics is now receiving more emphasis (eg Floquet, 1982, 1984; Alonso et al, in press). The account given here draws heavily on these sources. Fig 8.66 gives the lithological succession of the Upper Cretaceous, indicating the approximate timing of inferred tectonic and eustatic events.

The deposition of Utrillas sands probably continued into the Cenomanian. The M - U Cenomanian transgression is marked by the presence of marginal marine facies. Transgression continued during the Early Turonian, although there is evidence for a break in sedimentation at the Cenomanian - Turonian boundary with glauconitised surfaces, hardgrounds and local clastic influx (Floquet, 1982, 1984).

Alonso et al (in press) report that the uppermost Turonian and lowermost Coniacian are frequently missing. The U Turonian includes shallow water rudist and stromatolitic facies in the Sistema Central (Alonso, 1981). Karstification, brecciation and palaeosoil formation are also reported (Floquet, 1979b). Alonso et al (op. cit.) interpret this sequence as the result of a hiatus in subsidence with vertical sediment accretion leading to a progressive shallowing. However, the discordance may be due to tectonic activity (see fig 8.67). Ziegler (1982) reported the occurrence of a Turonian unconformity in the Porcupine Trough (post-Turonian sediments onlap onto older strata). This unconformity was attributed to a final extensional pulse.

The Coniacian initially saw renewed transgression. After this, a brief regression around the Coniacian - Santonian boundary was followed by a further phase of transgression in the Lower Santonian. These sea-level variations were interpreted by Floquet (1982) as

recording the combined effects of global eustatic sea-level rise (Hancock & Kaufmann, 1979; Vail et al, 1977) and changing relative subsidence/sedimentation rates.

The Upper Santonian - Lower Campanian includes regressive sequences: (rudist limestones → inter-supra tidal deposits with blue-green algae and evidence of desiccation; Floquet, 1982). Coastal sabkha facies dominate in the central parts of the platform (eg the study area) although evidence of transgression may be found on the platform margins (Alonso et al, in prep).

The Upper Campanian - Maastrichtian and Palaeogene consist of marginal facies and Floquet (1982) argued that the control of eustacy on facies evolution would have been reduced with increasing continental influence. Alonso et al (in prep) suggest that the climate may have been more humid by the Maastrichtian since evaporites are rare and fluviatile and lacustrine facies are developed. The observed succession of facies is consistent with the global trend of Late Cretaceous regression.

The Upper Cretaceous sequence of the study area is markedly thinner than that of N Cantabria, which is up to 1300m in thickness (fig 8.68); the increased subsidence there reflecting greater proximity to the Biscay margin. In the study area, the sediments are developed in generally shallower water facies than in other Upper Cretaceous sequences (eg N Cantabria, S England, the N Sea and E Canada). Limestone-marl rhythms predominate in N Cantabria (Hines, 1986) while the Chalk sequences of NW Europe are well-known.

The Chalk sequence represents much deeper water conditions than those recorded by the N Castillian Upper Cretaceous sequence. Transgression was smaller in magnitude and occurred later in N Spain than in S Britain. There is an approximate diachronous parallelism of facies (fig 8.69) between the Cameros and other areas. Thus we may compare the Albian Gault and the Cenomanian Lower Chalk/Chalk Marl of S England with the U Cenomanian - Lower Turonian muds/ marls and the U

Turonian limestone-marl alternations of the study area. However, at least in the lower part of the Upper Cretaceous sequence, many of the facies developed in the study area find approximate stratigraphic facies and analogues on and around the basement massifs of the British Isles (eg W Dorset, Yorkshire, N Ireland. Thus the "Glauconitic Marl", "Glauconitic Sandstone", "Grey Chalk" and "Chalk Marl" facies are all broadly recognisable in the study area, if still perhaps in slightly shallower water equivalents. Furthermore, the hardground at the base of the Glauconitic Marl (~M Cenomanian) may well be equivalent to the condensed facies of the "oyster bed", while the U Cenomanian - Lr Turonian marls of the study area (Santa Maria de las Hoyas and Picofrentes Fms) may partly correlate with the Plenus Marls of S England and NW Europe and with organic-rich facies commonly developed in many parts of the world at about that time (Cenomanian - Turonian Oceanic Anoxic Event; Schlanger & Jenkyns, 1976). Likewise, the Late Turonian hardground marking the boundary between the Middle and Upper Chalk in S England (Chalk Rock) may correlate with the pre-Cenomanian karstification and emergence reported by Floquet (1979b). The post-Turonian strata in S England (Upper Chalk) locally onlap onto basement massifs, much as they do in N Spain and the Porcupine (see above).

The silty shales of the N Sea and the silts, shales, and turbidites of E Canada do not immediately invite comparison with the facies present in the Cameros. The Canadian sequences may reflect more active tectonics in that area associated with the separation of Greenland from Labrador. However, the presence of glauconitic sandstones at the base of the sequence in E Canada, of some glauconitic sands and silts (Dawson Canyon Fm), evidence of shallower conditions in the Early Tertiary (Wyandot Fm Chalks) and of regional uplift in the Palaeocene (Banquereau Fm clastics), do at least point to a similar timing of tectonic and regressive/transgressive events.

In the plate tectonic context, the Late Cretaceous saw continued opening of the Bay of Biscay and translation of Spain towards the SE until the uppermost Santonian - Lower Campanian. Further rifting also occurred off Rockall.

This interval may be termed the "drifting" period for the N Spanish margin. Post-rift subsidence is likely to have been thermally-controlled (Alonso et al, in press), or compactional, as the greater thicknesses of Mesozoic sediments in basinal areas compacted much more than the crystalline or metamorphic basement massifs. However, the regressions, which Floquet (1982) interpreted as the result of "positive movements", may have been due to final weak extensional pulses. Alonso et al (op. cit.) reported that N120-N130 listric faults were active (fig 8.70) and that N60 faults were also locally important, especially in the S (eg controlling the Ateca High).

These fault trends are the same as those conditioning graben geometry during the Late Jurassic - Early Cretaceous rifting period and controlling differential subsidence during other relatively less tectonically-active periods (eg Lower and Middle Jurassic; Purbeck).

However, the uniformity of sedimentation over the N Castillian Platform is, nevertheless, very striking. Upper Cretaceous facies and thickness changes in the study area were much less marked than during the Late Jurassic - Early Cretaceous.

Late Cretaceous (Subhercynian) and Late Palaeocene (Laramide) tectonic movements caused both local folding (Gale, 1980) and widespread basin inversion in E and N Europe (eg Sole Pit Trough; Glennie et al, 1981) but were hardly manifested in N Spain, or indeed in most places away from the influence of the mobile NE-trending Tornquist-parallel lineaments (eg to the W of the Pays de Bray Fault, SW Britain; Cartwright, pers. comm., 1986). Inversion in N Spain and the W Shelves was dominantly of Oligocene and Miocene age. However, the Lower Santonian, Campanian and Maastrichtian phases of shallowing

inferred from the facies sequences may have been at least partially controlled by tectonic movements (Floquet, 1984).

Floquet (1982) mentioned latest Cretaceous - Palaeocene folding and Alonso et al (in press) discussed the Maastrichtian development of small isolated basins, suggesting that these movements were the result of early compression. The Silos Basin is just such an isolated basin and the progressive discordances within it indicate that sedimentation and deformation were partly contemporaneous. Floquet, Salomon & Vadot (1985) postulated mild folding before the deposition of the Maastrichtian - Palaeocene sequence (fig 8.71).

However, although small individual depressions may have formed, the present isolation of the Silos Basin with its thick Maastrichtian - Palaeocene sequence may be partly due to reduced truncation there. The basal Oligocene unconformity has demonstrable relief and the absence elsewhere of the "Garumnian" could be due to greater erosion.

The Silos Basin is in an unusual tectonic setting, lying to the N of the La Yecla backthrust (fig 1.6). This situation may have been responsible for isolated local subsidence (see section 2.12-2.19). Alternatively, the loading caused by the development of a particularly thick continental sequence may have acted against uplift and truncation under compression, or finally, and more likely, the local development of a backthrust there may have conditioned the preservation of Maastrichtian - Palaeocene strata. Equivalent strata may have originally have been present in other areas, either in other similar small intermontane basins or perhaps over much of the region. "Garumnian" facies are widespread in Spain, occurring in many areas of the Iberian Chains and S Pyrenees. The lateral persistence of the facies could argue against their deposition in only a few isolated basins. Moreover, these sediments correlate with similar facies of Late Cretaceous - Early Tertiary age which occur on the opposite side of the Pyrenees (eg Freytet & Plaziat, 1982) and which were deposited in very extensive basins.

Even further afield, the age of the youngest-preserved Upper Cretaceous (Chalk) strata in S England also shows considerable variation (see Anderton et al, 1979; fig 8.69): from Maastrichtian - Danian in weakly-deformed areas (eg Central North Sea) and Upper Campanian in the central parts of inverted basins (eg Dorset/Isle of Wight) to Lower Campanian or even Santonian in strongly-inverted basin marginal areas (eg Yorkshire, Sole Pit, Wiltshire and the Chilterns). Senonian facies in these areas are demonstrably pelagic, so that their frequent absence must be the result of truncation.

The Palaeocene - Eocene sequences of the Paris Basin and S Britain consist of terrigenous clastic deposits with thick sands, clays and lignites. Soils are developed, notably in the Reading Beds which contain mottled clays interpreted as vertisols by Buurmann (1980). These sediments may be preserved in small basins (eg Isle of Wight - Hampshire Basin) but comparable facies also extend over large basins (eg London - S North Sea Basin). Again opinions differ whether sedimentation was confined to or concentrated in these basins (Plint, 1982) or fairly uniform over the whole region (King, 1981).

Despite these similarities in the style of basin evolution, the Tertiary facies of S Britain show more marine influence and evidence of a more humid and possibly cooler climate (fig 8.72) reflecting their deposition considerably further to the N. The Bay of Biscay had opened during the Upper Cretaceous and although the high sea-levels of the Late Cretaceous permitted a degree of communication between the N and S Biscay margins, Early Tertiary regression led to their separate development now over 1 000km apart and in different climatic zones.

8.17. Oligocene - Miocene.

The sediments of this age represent the marginal facies of the Duero Basin. Basin margin structures consist of emerging blind thrusts / thrust-ramp culminations. Sedimentation can be broadly

correlated with progressive deformation of the Mesozoic (and locally of the Palaeozoic) under compression.

"OligoMiocene" conglomerates are spectacularly developed in the Arlanza Gorge between Hortigüela and Covarrubias. Progressive disconformities observable in the gorge indicate that deformation and sedimentation were partly contemporaneous. The presence of large-scale cross-bedding suggests the large-scale accretion of alluvial fans. The Oligocene is clearly a syntectonic "molasse" deposit.

The conglomerates are extremely variable. Some have essentially monomict clast assemblages derived locally from the subjacent Upper Cretaceous limestones. Clast sizes may reach 1m (figs 8.73 & 8.74), but are generally around 5-30cm. These conglomerates are locally channelled (fig 8.75) and are generally strongly carbonate-cemented (fig 8.76).

Polymict conglomerates (figs 8.77 - 8.79) are also developed. These display a wide variety of clast lithologies including: purple meta-arenites (Cambrian); white quartzitic conglomerate (?Tremadoc or Utrillas); brown sandstone (?Pedroso Group); yellow limestone (marine Jurassic); grey limestone (Rupelo Formation); polymict conglomerate (Senora de Brezales Fm or Salas Group). Thus the clast assemblages in the polymict conglomerates record the uplift and erosion of the whole Mesozoic and Palaeozoic sequence. The Sierra de la Demanda was probably now uplifted for the first time. The nearest outcrops of Palaeozoic strata are over 15km away and hence some clasts were transported a considerable distance. Polymict conglomerates appear to incise the Upper Cretaceous scarp near the monastery of San Pedro de Arlanza just as the modern River Arlanza does today. Indeed, the clast assemblages present in these polymict Oligocene conglomerates are similar to those of the Quaternary and recent Arlanza river gravels.

Thus OligoMiocene drainage patterns followed the present course of the Arlanza Gorge, itself located where the Jaramillo-Covarrubias Fault (JCF) cuts the Las Mamblas scarp of Upper Cretaceous limestone.

The JCF was a major cross-graben transfer structure during the Late Jurassic - Early Cretaceous (see earlier in this Chapter), and was evidently reactivated during the Tertiary.

Polymict "Oligocene"conglomerates are also found in the SW of the study area near Tejada where they occur N of a fault which is apparently the continuation of the La Yecla backthrust (fig 1.6).

The geometry of the Oligocene conglomerates may be related to the location and style of early thrusting (fig 8.81). However, the rôle of thrust ramp culminations appears to have been local and other factors (eg the size and palaeogeology of the source area) were also important in determining conglomerate facies and composition,as recently discussed in the S Pyrenees by Jolley (pers. comm., 1985).

In the study area, the localised development of polymict conglomerates in the Arlanza Gorge was conditioned by a major fault perpendicular to the strike of the thrusts which effectively increased the size of the fan catchment area (fig 8.80). The intersection of an emergent blind thrust / ramp culmination (Las Mamblas scarp) and a fault perpendicular to it (JCF) acted as a point fan source and conditioned the locus of fan-head entrenchment. Homewood & Allen (1985) have recently demonstrated the importance of faults perpendicular to thrust fronts in determining fan location in the Swiss Molasse Basin (fig 8.82). Faults cross-cutting thrust fronts were also important in alluvial fan sedimentation in the Siwaliks of Pakistan (Parkash, Sharma & Roy, 1980; fig 8.83). In the W Cameros, "cross-faults" active in the Tertiary clearly were reactivated Mesozoic faults.

The rôle of pre-existing structures in controlling sedimentation patterns of "foreland basins" is now becoming more widely appreciated (eg Farrell, 1986). The location of entrenched point-source fans and palaeovalleys (again often utilised by modern drainage) mentioned in the Southern Pyrenees by Jolley (1986), Nichols (1986) and Turner (1986) could similarly have been determined by pre-existing N-S faults

perpendicular to the thrust front (see sketch, fig 8.84). Farrell (pers. comm., 1986) has demonstrated that "cross-faults" in the footwall were active during foreland basin sedimentation.

The sediments further away from the Duero Basin margin are less amenable to field structural-sedimentological analysis owing to their generally poor outcrop. Studies of the distribution of fresh-water limestone facies may aid reconstruction of subsidence patterns in this context (see section 9.3 for discussion).

However, the influence of faulting on thickness and facies in the more central parts of the Tertiary Basins can readily be detected from geophysical evidence. Unreleased seismic sections of the Duero Basin margin further to the SE indicate strong basin asymmetry with greater thicknesses, as would be expected, near the Cameros margin. However, on seismic-generated isopach maps, very rapid thickness changes are also evident in a NW-SE direction. These latter variations are probably the result of movement on NE-SW faults parallel to the JCF and to those bounding the SE Cameros Basin and the Ateca High.

8.17.1. Comparison with the Oligo-Miocene of SW Britain.

In offshore areas of SW Britain (eg the Porcupine Trough, N Sea) the OligoMiocene is developed in marine facies, as in N Cantabria.

However, continental OligoMiocene deposits are also well-known from S England (Bovey Tracey and Petrockstow Basins; see Edwards, 1976; Holloway & Chadwick, 1986) where sedimentation can be related to (?renewed) movement on NW-SE faults. These faults (fig 8.85) cross-cut the Mesozoic basins of the area (Bristol Channel / Celtic Sea and Channel / W Approaches Basins). Working from onshore outcrop data, Gayer & Underhill (1985) did not recognise Mesozoic facies changes across the NW-SE faults, but analogy with the Cameros would suggest that they are likely to have been active as transfer faults during extension. Thus Mesozoic facies changes are likely to be present, and may be evident from offshore data, particularly where the Late

Jurassic - Early Cretaceous is better developed (this part of the succession is largely absent from the Bristol Channel).

The deposits of the Bovey Tracey and Petrockstow Basins display similarities to the OligoMiocene of the study area (great thickness, coarse conglomeratic facies and ?control of sedimentation by reactivated Mesozoic transfer structures) but are much more localised. Larger areas of Oligocene sedimentation were the Hampshire - Isle of Wight and Lough Neagh Basins. However, the facies developed there are significantly different from those of the study area, the differences reflecting more marine influence, weaker uplift of source areas and more humid climate in that area (see fig 8.86).

8.18. Plio-Quaternary (Ranas Formation).

Polymict Ranas Formation conglomerates lie unconformably upon the Mesozoic sequence in the N of the study area (figs 8.87 & 8.88). Ranas Fm conglomerates include clasts of: Jurassic limestone; grey (Rupelo Fm) limestone; and red sandstone (?Triassic, Senora de Brezales Fm, Pedroso Group or Salas Group). The clasts were apparently chiefly derived from the erosion of the pre-Upper Cretaceous part of the Mesozoic sequence. This derivation explains the compositional similarity with the Senora de Brezales Formation conglomerates.

The map outcrop of the Ranas Fm (fig 8.89) outlines sediment bodies with elongate, approximately triangular lobate geometry. This geometry is very clear on aerial photographs (eg fig 8.90), and immediately suggests deposition in NW-SE trending alluvial fans draining the Sierra de la Demanda. The depositional apices lie close to major NW-SE faults bounding the SW margin of the massif. The location of the fans was thus clearly fault-controlled. The fans were best developed along the Jaramillo-Covarrubias Fault and along nearby parallel NE-SW faults offsetting the Quintanilla-Castrovido anticline.

The Ranas Fm conglomerates are generally assigned to the Plio-Pleistocene (section 2.22), although Pol (1985) suggested that they formed the proximal equivalent of conglomerates mapped in the Covarrubias area as Miocene in age. The apparent absence of Upper Cretaceous limestone clasts from the Ranas Fm conglomerates in the N sector indicates pre-Pliocene removal of the Upper Cretaceous cover of the Sierra de la Demanda. This is consistent with the uplift history already deduced from the presence of Palaeozoic clasts in Oligocene and Miocene conglomerate clast assemblages.

Outside the study area, fault movements occurred over much of the Iberian Chains, where Ranas Fm conglomerates are widespread. Tectonic movements also affected other areas of NW Europe at this time - Ziegler (1982) reported wrench activity in Armorica and truncation of Neogene sediments in the W Channel. As in the study area, these movements were probably associated with renewed movement on reactivated Mesozoic transfer structures.

8.19. Quaternary and Recent.

The present drainage pattern and the Quaternary network of alluvium (fig 8.91) show evidence of structural control. Breaks in Upper Cretaceous limestone scarps (eg at Huerta del Rey and Espejon) are probably coincident with cross-cutting NE-SW faults, as is the break in the U Jurassic - Lr Cretaceous ridge at Talveila. There are numerous other examples. My mapping has indicated that many of the faults were active during the deposition of the Tierra de Lara Group.

The Arlanza Gorge follows the Jaramillo-Covarrubias Fault, itself active in controlling earlier Tertiary drainage (see section 8.17). The upper Arlanza and its tributaries run in NW-SE, NE-SW and E-W segments, following the main structural trends. The course of the lower Pedroso river is apparently controlled by the Jaramillo-Covarrubias Fault, as is its confluence with the Arlanza. Many other

Fig 8.20.0. Structure. Vertical Upper Cretaceous strata, vergence to N. Looking NE from near Hortezuelos to La Yecla.
Dusk, October, 1982.

small streams are parallel to the Jaramillo-Covarrubias Fault and probably follow more minor faults parallel to it.

Quaternary and recent drainage in the Duero Basin (fig 8.91; see also interpretations: figs 8.92 - 8.94) shows dominant NE-SW trends, as well as important NW-SE and E-W trends. The NE-SW trend clearly influences the course of the Arlanzon river through Burgos itself and may indicate the line of a major Mesozoic - Tertiary transfer fault originally initiated in the late Hercynian (see next section) and responsible for the apparent offset of the Aguilar and W Cameros Purbeck/Weald half-graben segments (fig 8.93) in a manner similar to that described by Bally (1982) and Lister, Etheridge & Symonds (1986) and shown in figs 8.95 & 8.96. The NW-SE trend parallels a conjugate set of late Hercynian faults (again also active during Mesozoic extension and Tertiary compression) while the E-W trend parallels the strike of the Hercynian thrusts, which were also locally reactivated during Tertiary thrusting/backthrusting (see section 8.17). The Mesozoic and Palaeozoic structure underlying the thick Tertiary sequence of the Duero and Rioja Basins is poorly-known and only discernible from seismic sections, but this study of drainage patterns has at least indicated the likely continuation of the major structural trends observable in the Palaeozoic and Mesozoic strata at the basin margins. A study of this type could prove a useful preliminary to seismic investigation and might predict areas of particular interest at the intersections of major structures.

8.20. Structure.

At first sight, the structure of the W Cameros appears simple to the visitor, with the dominating regularly-bedded scarps of Las Mamblas apparently indicating flat-lying layer-cake stratigraphy.

In fact, the W Cameros is characterised by extensive tracts of relatively gently dipping strata with intervening small areas of intense deformation. Fold axes are tight and subparallel. The

structural style is reminiscent of S England (eg Stonely, 1982), although the intensity of the folding is greater in the Cameros.

A geological map of the study area, (fig 1.6) shows the dominance of NW-SE trending structures. I have already discussed the possible rôle of these in controlling Mesozoic facies. Major NE-SW faults are not immediately apparent from the geological maps, but are demonstrated by studies of facies distributions (eg figs 8.49 & 8.52 - 8.54). E-W faults (Hercynian thrusts, locally reactivated in the Tertiary) are evident in the Palaeozoic massifs and at the Tertiary Basin margins (fig 8.92 shows S-directed thrusts at the N margin of the Rioja Basin and N-ward thrusting at its S margin).

The Iberian Chains have recently been described as a "mountain belt with double vergence" (Rosell, 1985). In the study area, the S sector shows dominantly SW-verging folds and thrusts apparently developed in a S-directed imbricate fan (Boyer & Elliot, 1982) similar to the examples recently given by Morley (1986); see figs 8.97 - 8.98). The N sector of the study area displays NE vergence. However, the structure is further complicated by the presence of some N-verging structures developed in U Cretaceous and "Garumnian" strata in the S (eg between La Yecla and Tejada). Although Banks & Warburton (1986) have recently described instances of inverse thrust polarity in the upper part of the structural pile ("passive roof duplex", fig 8.99). I prefer to interpret these structures as simple backthrusts. The dominance of opposite vergence in the N sector is what is predicted from theoretical and experimental models for parts of a mountain belt behind the main thrust front and to the rear of the deforming thrust wedge (Davis et al, 1983, fig 8.100 ; Platt J, in press). Occurrences of backthrusts and mechanisms for their formation have been discussed by Butler, (1982), and Fischer (1984), (see fig 8.101).

In the study area, it seems that the location and occurrence of backthrusts may be restricted to particular segments of the thrust belt defined by cross-faults (see below, also Hossack, 1985 and pers.

comm.,1986), and may locally be related to the development of hanging-wall ramps. Backthrusting at La Yecla is apparently also favoured by the intersection of a major NW-SE (and perhaps also NE-SW) structure with an E-W fault. The W-ward continuation of this fault (fig 1.6) bounds the E-W trending basin margin to the N of Tejada and is mappable persisting into the Duero Basin from study of the Quaternary drainage pattern (fig 8.94).

Other areas of particularly intense deformation (eg around Espejon in the S of the study area and along the Mataviejas Gorge E of Santo Domingo de Silos) may also be controlled by the intersection of NW-SE, NE-SW and E-W structures (see figs 8.93 & 8.94 & 8.102 - 8.104). The complexities may be broadly interpreted as structural interference patterns. This is a subject worthy of a more-disciplined structural analysis!

Regional mapping has facilitated the compilation of schematic cross-sections across the basin. Detailed mapping (see enclosure) was carried out in the area of the San Leonardo Fault (SLF). This area was selected for special attention as it affords good exposure along strike in an area of structural interest which is also characterised by important truncational unconformities and sedimentological complexities.

The SLF has previously been interpreted as a simple reverse fault or thrust with décollement along Carniolas evaporites. Our mapping (summarised by fig 8.105) has shown the presence of a stack of up to at least five imbricate thrust slices , the additional detachment occurring on the Rio Cabrera Mb ("Purbeck") evaporites. The steep dip of the SLF is interpreted as the result of back-rotation caused by the emplacement of later thrusts (fig 8.106).

The whole thrust system is cross-cut by numerous NE-SW faults which could be interpreted as thrust-compartmental or "hanging wall drop faults" (sensu Butler, 1982). Hossack (1985, 1986) indicated that deformation style and vergence in the Pyrenees and Cantabria changes

abruptly across major transfer zones (eg the Pamplona, Toulouse and Rio Miera lineaments, see fig 8.107).

In the study area, the segmentation by cross-faults is analogous, but on a different scale, to that described by Hossack for the whole Pyrenean - Cantabrian mountain belt, and there is clear evidence to suggest that the thrust-compartmental faults were actually pre-existing structures reactivated under compression.

Important NE-SW faults offsetting the SLF at Pinilla de los Barruecos, La Gallega, Espejon Road, Vadillo, Casarejos and Talveila (fig 8.105), see sections 8.9 - 8.11) have clearly influenced Mesozoic sedimentation (Senora de Brezales Fm conglomerates) and truncation histories (preservation of Talveila Fm, see fig 8.26; SdB Fm and Rupelo Fm). The NE-SW faults in the study area (eg Jaramillo-Covarrubias Fault) have been interpreted earlier in this Chapter as transfer faults conditioning structural and sedimentological development during Mesozoic extension. Intra-Mesozoic reversal of the sense of fault displacement (see section 8.11.2) suggests that the faults may have had important wrench components. Similar reversals of transfer fault displacement were recently reported from S Australia by Etheridge et al, (1985) and Etheridge (1986).

These NE-SW faults now delimit areas of differing thrust vergence and geometry, bound zones with thrusting dominantly above the main SLF or below it, and also separate areas with different numbers of imbricate slices in the stack complex (ie they define the edges of duplexes), and may control the geometry and location of lateral ramps. This is outlined in fig 8.108.

To my knowledge this is the first recorded example of synsedimentary transfer faults being reactivated as thrust-compartmental faults, but I would expect there to be many more examples in other thrust belts, and these may be recognisable by integrated structural/sedimentological studies. Tricart (1984) and Tricart & Lemoine (1986) have recently suggested that deformation

style in the Western Alps was influenced by pre-existing "Tethyan" normal and strike-slip faults. Similar structural controls may have been operative in the Swiss Molasse Basin and in the Jura Mountains where Laubscher (1972) reported Late Miocene - Pliocene basement-involving wrench faulting (fig 8.109) along fractures probably established before folding. These views remain somewhat controversial. (Further discussion of the structure of this area is given in Allen *et al* 1985). Their palaeogeographic reconstruction shows clear facies control by cross-faults (fig 8.82).

The local development of complex fold patterns (figs 8.110 & 8.111) at the W termination of the SLF was interpreted by Vaughan (1986; fig 8.112) as the result of accommodation above lateral ramps. Similar structures have been described by Coward (1984) and Coward & Potts (1983), (see also fig 8.113).

The lateral change in thrust geometry along strike may also reflect the differing response to compression of sequences varying laterally in stratigraphy, thickness and rheology. Potts (in prep) has suggested that the varying geometry along strike of the Tariskavaig thrust on Skye, NW Scotland was conditioned by the presence or absence of the Torridonian. In the study area, lateral variations in stratigraphy, thickness and facies (and, of particular significance, in the presence and thickness of "Purbeck" evaporites and "Wealden" clays) manifestly do occur along strike. These variations would have influenced the behaviour of the sequence under deformation and would have conditioned the development and horizon of décollement and the style of folding/faulting. Significantly, these changes in the characteristics and rheology of the sequence occur very sharply across pre-compressional cross-faults, so that sharp change in thrust architecture across these faults would be expected.

Shortening estimates across the W Cameros are problematical. The area-balancing technique for the construction of restored cross-sections (Hossack, 1979) is difficult to apply because of

probable fault-controlled thickness changes. The outcrop also provides poor depth control. However, an attempted line-balance on the base of the Salas Group / Utrillas across the San Leonardo Fault Zone gives a useful first approximation to the minimum shortening of ≈3.5km (figs 8.114 & 8.115). Total shortening across the entire W Cameros Basin may be estimated at twice to three times this value, say 7-10km.

Chadwick (1985) outlines a method for calculating shortening from estimates of the amount of isostatic uplift resulting from compression. This technique gives a value of 7-9km (fig 8.116). Hossack (1985) suggested that Tertiary shortening in the Pyrenees was around 75km, and thus shortening in the Cameros is an order of magnitude less than in the Pyrenees, but significantly greater than the 1.9km estimated for S England by Chadwick (1985).

The significance of this considerable Tertiary shortening is that Mesozoic palaeogeographic maps may require palinspastic correction. Apparently rapid facies changes across NW-SE faults might partly be a product of the tectonic juxtaposition of individual sequences actually deposited a considerable distance apart, rather than wholly being the result of synsedimentary fault movement on the NW-SE faults. When lateral facies changes were very gentle, as they were during the Upper Cretaceous, for example, then the localised shortening of strata in particularly deformed areas (eg SW Cameros Basin) might give a partly erroneous impression of rapid lateral facies change.

The deformation itself is arguably thin- or thick-skinned. Traditional basement-fault reactivation models would favour thick-skinned movements, yet the magnitude of the shortening of the cover sequence argues strongly against direct basement control. Thrust movements calculated at several km put into question the assumption that areas of intense deformation occur above basement faults when the cover is apparently largely allochthonous.

However, both extension and compression of basins probably involves elements of thick- and thin-skinned tectonics (see eg Gibbs, 1983). Thus although some deformation, eg in the area of the San Leonardo and other parallel faults, is thin-skinned, the involvement of Palaeozoic strata in some thrust slices (eg at Moncalvillo) indicates that ramping down to the basement must have occurred (fig 8.81). Indeed, the Palaeozoic massif of the Siera de la Demanda is itself allochthonous and has been thrust N-wards over Tertiary sediments on its N margin. Hossack (pers. comm., 1986) provided a schematic model of thrusting in the Demanda involving detachment within the Keuper/Carniolas and within the basement (fig 8.117).

It is also conceivable that thrusting of the cover sequence might be the surface expression of movement on deeper faults. The observed stacking of thrust slices might partly record the development of local thrust splays as a consequence of inversion of listric normal faults. If this were the case, then regional shortening could well be much less than indicated by consideration of surface thrust geometry.

Hossack (1984) discussed the mechanics of extension in the Hornelen Basin of Norway and suggested that "roll-back" on pre-existing (Caledonian) thrusts may have been important (fig 8.118). A similar mechanism has been proposed to explain subsidence of offshore basins of NW Scotland (eg the Minch Basin) by Mesozoic (and probably Devonian) roll-back of Caledonian thrusts (eg Outer Isles Thrust).

It would be possible to explain the development of the Mesozoic basins of S England and N Spain (eg Hampshire-Channel, W Approaches, Bristol Channel, Celtic Sea etc; Cameros Basin, Vasco Cantabrian Basin, Asturian Basin) by extensional roll-back on Hercynian thrusts. Chadwick et al (1983) argue that reactivation of Hercynian thrusts striking E-W controlled the orientation and position of the major extensional faults in the Wessex Basin (figs 8.119 - 8.121). The subsequent deformation of these basins could be achieved by compressional reactivation of the thrusts (figs 8.122 - 8.124).

Fig 8.20.1. Penacoba Thrust. This previously unmapped thrust extends for over 6km WNW of Mamolar and displaces mottled (?Rupelo Formation) limestones (purple slope) over Utrillas Formation sands
Looking N. 1km W of Penacoba. September, 1985.

Cross-sections through the Wessex Basin inversion structures (figs 8.125 & 8.126) clearly show development of thrust splays in the hangingwall and footwall to the reversed Mesozoic extensional faults.

The structural evolution of some areas of N Spain may have involved the reactivation of E-W trending Hercynian thrusts during Tertiary compression (eg the N margin of the Rioja Basin and in NW Santander; figs 8.92 & 8.127) where the Mesozoic basin extension was essentially N-S and perpendicular to the strike of the Hercynian thrusts). However, in the W Cameros and SW Cantabria, the dominant Mesozoic extension was NE-SW, oblique to the Hercynian thrusts and instead utilised the NW-SE and NE-SW conjugate set of late-Hercynian wrench faults as graben-bounding and cross-graben transfer faults respectively.

The Mesozoic cover of the Cameros may also be thicker than in NW Santander or in the Sierra de Cantabria, so that the influence of basement thrusts under Tertiary compression might have been further reduced. Thus deformation resulted in the preferential compressional and wrench reactivation of the NW-SE and NE-SW Mesozoic extensional and transfer faults respectively to give dominantly SW- and NE-verging structures (fig 1.6) offset and bounded by NE-SW compartmental faults. There was only minor N- or S-directed thrusting.

In contrast, the geometry of the small (and less subsident) faulted sub-basin of the "Villavelayo syncline" (fig 8.94) apparently was controlled by Hercynian thrust orientation, and in the Duero Basin, where the Mesozoic cover sequence is probably thinnest, the E-W Hercynian trend is clearly picked out by the Quaternary drainage pattern.

Fig 9.0. Lacustrine environment. Laguna Negra, near Neila. September, 1983.

CHAPTER 9. PARTICULAR PROBLEMS RAISED DURING THIS STUDY.

9.1. Lacustrine carbonates and pedogenesis.

Lacustrine carbonates have received relatively little attention in the N American and British literature. This reflects the relative rarity of ancient lacustrine carbonates in the USA and the UK. In contrast, lacustrine carbonate sequences are well-documented in the European literature. Some of the most spectacular examples are to be found in the Tertiary basins of Spain and France.

For many years, work on lacustrine sediments has been dominated by models derived from studies of the Green River Fm (Picard & High, 1972; Surdam & Wolfbauer, 1975; Hardie et al, 1978). The Green River Formation contains some carbonates (eg Smoot, 1978) but these seem to reflect deep fresh-water or saline chemical sedimentation in stratified lakes.

Work on modern carbonate lakes has included studies in Alpine regions (eg Müller & Wagner, 1978), W and E Africa (Talbot, 1976, 1984; Stoffers & Hecky, 1978; papers in Reading, 1986), and in N America (Glass & Wilkinson, 1980; Murphy & Wilkinson, 1980; Treece & Wilkinson, 1980; Dean, 1981; Last, 1982). This work has concentrated on temperate and equatorial zone carbonate lakes. Semi-arid zone carbonate lakes (eg of N Africa, N India) have received relatively little study. Although biological carbonate production was stressed by Kelts & Hsu (1978), Glass & Wilkinson (1980), and Murphy & Wilkinson (1980), most work on deep, stratified or saline lakes has tended to emphasise chemical and hydrological aspects of carbonate production (eg Müller, Irion and Förstner, 1972).

The classical accounts and reviews of lacustrine carbonate sedimentation (eg Picard & High, 1972, 1981; Lerman, 1978; Collinson, 1978; Dean & Fouch, 1983; Allen & Collinson, 1986) are often inappropriate in the context of the S European Tertiary (and of the

Rupelo Fm of the Cameros) as they do not emphasize the many examples of fresh-water lake carbonates which provide clear evidence of shallow water conditions, high rates of biogenic carbonate production and frequent emergence and pedogenesis.

Most modern English language reviews have omitted the outstanding work of Freytet, Plaziat, and Freytet & Plaziat (1965, 1973, 1982, 1984, and numerous others: see bibliography). This work on Late Cretaceous - Early Tertiary continental sedimentation in SW France has provided the basis for an expanding European literature on fresh-water ephemeral lake deposits (eg Nickel, 1982; Arribas, 1982, 1984; Cabrera et al, 1985; Floquet et al, 1985; Platt, 1985 - see enclosures).

Freytet & Plaziat identified the importance of pedogenesis in lacustrine carbonate sequences. Since then, and especially since 1973, workers on the Continent have made great progress towards documenting and evaluating the effects in continental carbonate successions of pedogenesis, a process barely mentioned in even the latest N American and British reviews. This omission has meant that pedogenic fabrics in continental carbonates have often been overlooked (with a few notable exceptions, eg Wells, 1983) and remain generally poorly understood in the English-speaking world. For example, the Bembridge Limestone of the Isle of Wight, the best-documented post-Palaeozoic "lacustrine" carbonate deposit in the UK, possesses many clear manifestations of pedogenesis that are only now being investigated, even though it outcrops in one of the most frequently visited sedimentological sections in the country.

The superposition of pedogenic features on lake carbonates is thus a common, but still under-rated, feature of many continental sequences (eg Late Jurassic - Early Cretaceous and Late Cretaceous - Tertiary of S Europe). Marginal lacustrine carbonates frequently bear comparison with peritidal carbonates in that they exhibit many features indicative of subaerial exposure and pedogenesis. Freytet &

Plaziat (1982) used the term "palustrine" for lake carbonates showing pedogenic fabrics. This term has advantages over "perilacustrine" both in brevity and in its reference to "paludal" (≈marshy / swampy) environments.

Klappa (1978) pointed out the important problem of distinguishing lacustrine carbonates and "calcretes" in view of the similar fabrics documented from both in the literature. Klappa noted that reliable petrographic criteria for the distinction between lacustrine carbonates and "calcretes" were difficult to find. Klappa's descriptions of calcrete fabrics included reference to: crusts (chiefly in areas of 300-400ml annual precipitation); vertical roots; horizontal roots; root mats; root rock ("rhizolite"); solutional cavities; glaebular conglomerates; black pebbles (especially in coastal areas); sheet calcrete; soils; and brecciated textures. All of these features are present in the Rupelo Formation (see section 3.3), and in many other "lacustrine" carbonate sequences. Klappa suggested that some carbonate rocks classically described as "lacustrine" might actually be "calcretes", although he accepted the alternative as a possibility, namely that lacustrine carbonates might provide "embarrassments for students of calcretes". Klappa ultimately had to "leave the problem of distinguishing between lacustrine carbonates and calcrete deposits with unsatisfactory resolution".

In fact, it is indeed likely that some of the rocks previously described as "calcretes" are actually pedogenically-modified lacustrine carbonates.

I consider the superposition of pedogenic fabrics on lacustrine carbonates simply to be the necessary consequence of frequent emergence caused by rapid shoreline fluctuations in enclosed lake basins, especially where the climate is semi-arid. Thus "calcretes" and marginal lacustrine ("palustrine") carbonates have similar fabrics because they both record fresh-water carbonate production in

continental settings and because both display the fabric expression of pedogenesis.

The problem of defining lacustrine carbonates and calcretes is thus resolved, any remaining ambiguity arising because of insufficiently precise use of the term "calcrete". In its present usage (see section 9.2), it is commonly employed to denote almost any carbonate exhibiting evidence of pedogenesis.

9.2. "Calcretes", "caliches", and nomenclature of carbonate palaeosols.

Clastic sedimentologists commonly use the terms "calcrete" and "caliche" when describing pedogenic carbonates developed on (or in) alluvial substrates (eg Allen, 1974b; Leeder, 1975; Steel, 1977; McPherson, 1979; Tandon & Narayan, 1981)[1]. Workers on calcretes (eg Siesser, 1973; Klappa, 1978; Watts, 1978; Hay & Reeder, 1978; Netterberg, 1980) use the term to refer to any carbonate palaeosol whether developed on clastic, carbonate or even crystalline substrates[2] - Klappa (1978) reported that calcretes could develop on any substrate provided that a source of $CaCO_3$ existed, either allochthonous or autochthonous. Carbonate sedimentologists use the term rather differently, ie for horizons in carbonate sequences which show pedogenic modification (and, commonly, vadose cementation)[3]. Examples include Dunham (1969), James (1972), Walkden (1974), Walls, Harris & Nunan (1975), Harrison & Steinen (1977), Somerville (1979), Hay & Wiggins (1980), Adams (1980), Braithwaite (1983).

Clearly, the term "calcrete" is used by different workers to describe pedogenic carbonates in a variety of different environments. Although all the usages imply the operation of pedogenetic processes, there are obviously fundamental genetic differences between:

(1) pedogenic carbonate developed on alluvial substrates soon after their deposition; carbonate derived essentially by precipitation within soil profile on fluctuation of water table (example: nodular carbonate horizons within Piedrahita de Muno Member of Pedroso Group);

(2) carbonate crusts or soil profiles developed on any exposed ancient substrate during emergence; carbonate precipitated on pediment with or without modification (example: Senora de Brezales Formation crusts/palaeosols developed on Lower Callovian pediment, as at Castrovido and Mambrillas de Lara).

(3) pedogenic modification of carbonate on emergence soon after its deposition, often without precipitation of additional carbonate. (example: Ladera Member of Tierra de Lara Group).

Freytet & Plaziat (1982) noted this problem and recommended the dropping of the term "calcrete", at least for pedogenically modified alluvial substrates (hydromorphic and calcimorphic soils, depending on carbonate content). Freytet & Plaziat also shied away from using "calcrete" for pedogenically modified lacustrine carbonates (hypercalcimorph soils), preferring the term "palustrine limestone". The term "palustrine" has freshwater connotations and hence can not be applied to soils developed by modification of marine carbonates. We are lost here between the necessity to reduce ambiguity on one hand and the requirement for a general term on the other.

The term "calcrete", although it may be useful in a geomorphological context to describe duricrusts (eg Goudie, 1973), is not really sufficiently precise for detailed work. The use of the term to describe pedogenic carbonates formed by different processes in a variety of different environments may be confusing because of the possible genetic implications. Certainly, I have avoided the term in this work wherever possible, and have not found it to be a great

disadvantage. An alternative is to use soil terminology; this has the advantage of technical accuracy but tends to be highly obscure and intimidating to the non-specialist. However, more widespread and consistent use of basic soil terminology such as "calcisol", "vertisol", "rendzina", etc., might well be possible. In fact, I have found that a more descriptive approach such as the one employed here necessitates more rigorous, and hence better, observations. These in turn make for more informative descriptions.

9.3. The influence of tectonics on lacustrine sequences.

The control of tectonics on continental sediments is now becoming more widely appreciated. Studies to date have concentrated on the potential of coarse conglomeratic alluvial sequences for indicating synsedimentary fault movement. This approach has recently been extended to sandy alluvial deposits and it has been recognised that the geometry of such sequences and their facies, or "fluvial architecture" can provide valuable information about the patterns of subsidence and the influence of tectonic controls.

Section 8.12 applied a similar technique to the Rupelo Formation, a lacustrine carbonate sequence. Tectonic control of lacustrine facies has previously been implied by Tucker (1977) and Truc (1978), but still the classical view of lacustrine facies distributions comprises a concentric arrangement of facies belts, as shown in Picard & High (1981; fig 9.1). In fact, all of the examples given show a marked asymmetry in facies development. It is widely accepted that most major lake basins are ultimately controlled by tectonics, and so it is reasonable to suppose that tectonics may also have been important in controlling facies distribution and geometry within these lacustrine systems.

The deposition of thick lacustrine carbonates requires a virtual absence of clastic supply and might be taken to indicate relative tectonic quiescence. In the case of the Rupelo Formation, although

tectonic control over subsidence was less dramatic than for the underlying alluvial sediments of the Senora de Brezales Formation, the facies distribution maps (figs 8.52 - 8.54) clearly show that differential subsidence was extremely important in controlling thickness changes within the Rupelo Formation. Thus sedimentation was still influenced by tectonics. The absence of clastic supply may have been partly due to the absence of available clastic source areas.

Most major lacustrine carbonate deposits are underlain by thick marine carbonate sequences. For example, the Early Cretaceous Peterson Limestone (and, in part, the Eocene Green River Formation) of the mid-western United States was deposited in an area drained by streams passing over (and transporting) abundant clasts of limestone and dolomite in Palaeozoic Formations; the Late Cretaceous - Early Tertiary lacustrine carbonates of S France and N Spain lie on thick Upper Cretaceous limestone sequences and the Cabaços and Vale Verde Formations of Portugal and the Rupelo Formation of the Cameros overlie thick Jurassic carbonates.

If the underlying and surrounding strata are exclusively carbonates, then clastic supply is likely to be limited and development of lacustrine carbonates is favoured. When basement or other clastic lithologies (eg sands or conglomerates draping the carbonate substratum) are also exposed at the surface (eg Miocene of the Duero Basin, Tertiary of the Hampshire Basin, Triassic of S Wales), then lacustrine carbonates are much more likely to be intermittent in vertical sequence and localised in areal distribution. In such cases, they may be restricted to areas (such as the Isle of Wight or the NE margin of the Duero Basin) where erosion from the carbonate substratum (Upper Cretaceous of the Duero Basin, Chalk of the Hampshire Basin, Carboniferous Limestone of S Wales) is locally dominant. Lacustrine carbonates also occur chiefly during those intervals when the carbonate substrata are exposed by erosion or when

erosion of basement lithologies is reduced (eg ?Lower Oligocene in the Hampshire Basin, Miocene of the Duero Basin).

Obviously, climate will also affect the rate of chemical weathering of the substratum relative to the rate of mechanical erosion. This will influence the rate of dissolved carbonate supply to a lake system, and of fresh-water carbonate production within it.

It is also clear that if tectonic activity is intense and the climate is arid or semi-arid, then lacustrine limestones will form only in areas away from basin marginal alluvial fans (as was the case in the Tertiary of the SE Ebro Basin (Cabrera et al, 1985). Location of the fans themselves, and thus of lacustrine carbonates in inter-fan areas, will reflect the pattern of normal faults, thrusts and transfer structures. However, even where tectonic movements are gentle, lateral changes in thickness of lacustrine sequences are likely to reflect the regional tectonic framework.

Tectonics will also govern lateral changes of facies. Alluvial-lacustrine transitions of the type observed in the Tertiary of the SE Ebro Basin are likely to reflect tectonic controls (see above; also Allen & Mange-Rajetzky, 1982), but even within lacustrine systems facies changes are likely to be influenced by tectonics. For example, deeper water facies are likely to dominate in particularly subsident areas, while pedogenesis is likely to be much more prevalent in areas of reduced subsidence. Impure carbonates are more likely to develop towards the basin margin and nearer to distant sources of clastic supply while purer carbonates would be expected in the basin centre.

Climate is likely to be the prime control over the hydrologically "open" or "closed" nature of lacustrine systems and may thus determine the deposition of organic-rich or evaporitic facies. However, fault movements are also likely to be important in this respect, as controls on individual fault block movement will condition whether distribution of these facies is localised or widespread.

This work has suggested that the architecture of lacustrine systems and the geometry of their facies distributions, or "lacustrine facies architecture" may also be a useful tool in basin analysis. Studies of this type can provide useful information about basin evolution and subsidence patterns during intervals when continental clastics are sparse and during intervals often previously regarded as tectonically quiescent.

9.4. Structure of the Cameros: the Deformation of a Basin.

The structural evolution of the Cameros provides an insight into the deformation of intra-cratonic rifts. Structural studies can provide valuable lessons for analysis of more weakly-deformed areas. For example, the style of Mesozoic basin development in SW Britain (Ziegler, 1982) appears to have been very similar to that of N Spain (for detailed analysis of the Wessex Basin see Lake, Dewey et al, 1985; Whittaker, 1985; Chadwick, 1986; Holloway & Chadwick, 1986; for the Weald Basin see Lake, 1975). Basement thrusts, cross faults and extensional normal faults were active in both areas throughout their Mesozoic and Cenozoic histories (eg see figs 8.49; 8.52 - 8.54; 8.57; 8.91 - 8.94; 8.119 - 8.124).

The timing of compression was the same ("Pyrenean" ie Late Eocene - Miocene; see Masson & Parson, 1983) although the degree of Tertiary deformation or "basin inversion" was very much more spectacular in N Spain. Continued or more intense compression in S England might have produced structures similar to those of the Cameros.

Conversely, seismic data available from SW Britain, an area which has suffered less Tertiary deformation, can provide valuable clues to the possible importance in the Cameros of basement faulting versus thin-skinned detachment under extension. The more deformed Cameros, on the other hand, can provide evidence on the likely importance of thick- or thin-skinned deformation in SW Britain.

The San Leonardo Fault provides an excellent small-scale model of a thrust system - the exposure is comparable with classically mapped areas of the Moine Thrust Zone in NW Scotland, and the structures are not dissimilar (compare fig 8.113, taken from Coward, 1983, with figs 8.110 - 8.113).

The rôle of transfer faults may also be compared. Stonely (1982) noted that the monoclines of S England (?blind thrusts, see Thompson, 1981; Morley, 1986) were offset. The fold axes in the Weald Basin (Lake, 1975) are also discontinuous in detail. Work from the Cameros would suggest that these offsets probably occurred across basement wrench faults in a manner similar to that shown in figures 8.119 - 8.121 taken from Chadwick (1985). In the Wessex Basin, the offsets may occur across faults parallel to the Pays de Bray fault, one of the major structural lineaments of NW Europe (fig 8.125). The wrench faults are subperpendicular to the Hercynian thrusts and were possibly originally of Caledonian age. They are clearly long-lived structures which show polyphase reactivation.

However, an important difference between the W Cameros and SW Cantabria on one hand and the Hampshire-Channel, NW Santander and Central Vasco Cantabrian Basins on the other lies in the relative importance of reactivated Hercynian thrusts and the conjugate wrench faults during extension and compression (see section 8.20). Whereas Hercynian thrusts showed only minor reactivation in the W Cameros and SW Cantabria, they were significantly reactivated under compression in SW Britain, NW Santander and the Central Vasco Cantabrian Basin and probably were also active under extension in NW Santander. Tertiary thrusting occurred along late-Hercynian wrench faults in the W Cameros and SW Cantabria (eg San Leonardo Fault, W Cameros; Aviles/Ubierna Fault, SW Cantabria) but was of minimal importance in the other areas, although local W-directed thrusting (Jerra thrusts, see fig 8.127) did occur in NW Santander (Spriggs, 1984) along a major NNE-SSW trending transfer fault / "flower structure" (Harding, 1985) near San Vicente

de la Barquera. However, in all the areas, extremely rapid extensional subsidence, sharp lateral facies and thickness changes, intense compressional deformation, and, locally, halokinesis, occured characteristically at the intersection of the major fault lines (see for example Sellwood, Scott et al, 1985). The lines of deeply-buried basement faults or thrusts might thus be predicted from sedimentological and structural studies.

These studies also have important implications for the investigation of more complexly-deformed zones. The sedimentological and stratigraphic complexities evident in the marginal areas of basins such as the Cameros demonstrate that it is dangerous to hold too much store by shortening values obtained in restoring mountain belts to flat-lying sheets of layer-cake stratigraphy.

The assumptions of relatively simple stratigraphy and basin geometry inherent in the construction of balanced cross-sections may be justified in the case of sedimentary sequences from stable cratonic shelf settings (as in the thrust belts where the techniques were first developed, eg Mesozoic of Central North America / Rocky Mountains and Cambrian of NW Scotland / Assynt). However, the complexity in timing and orientation of extensional faulting; the stratigraphic, sedimentological and rheological heterogeneities and thickness variations of the sediment fill reported here from an essentially intra-cratonic rift basin makes the concepts of two-dimensional cross-section balancing in orogenic belts seem highly simplistic. The cover sequences of ocean-marginal basins incorporated into collisional orogenic belts are likely to show even more complicated pre-deformational geometries.

Nevertheless, the lesser complexity of the W Cameros structures makes their interpretation more accessible and could provide useful models for the early stages of deformation in orogenic belts, which can themselves be (simplistically) viewed as complexes of highly deformed basins.

Fig 10.0. NNE fault trend utilised by road heading out of Pinilla de los Barruecos. Looking N from Jurassic scarp N of San Leonardo Fault to steeply-dipping Pedroso Group ("Wealden") conglomerates showing slight offsets on faults perpendicular to strike. Wide plain beyond of Salas Group and Utrillas Formation clastics stretching towards Salas de los Infantes (right distance) and dominating Upper Cretaceous mountain of San Carazo (left). Palaeozoic of Sierra de la Demanda in the far distance, some 30km to the N.
August, 1985.

CHAPTER 10. CONCLUSIONS.

10.1. Major Results.

10.1.1 Stratigraphy and Sedimentology.

1. The Oxfordian - Cenomanian sequence of the W Cameros Basin can be subdivided into:

a) Talveila Formation (Upper Oxfordian - lowermost Kimmeridgian) comprising cross-bedded calclithites as well as charophyte and coralline limestones. The Talveila Formation was deposited in a marginal marine carbonate inner shelf - clastic shoreline - lagoonal complex. Cycles in the Talveila Formation may be correlatable with similar cycles recognised in Corallian facies of S England and N France. The cycles may be eustatically controlled, or alternatively may reflect the onset of late Cimmerian tectonism;

b) Tierra de Lara Group ("Purbeck"), comprising two Formations:

i) Senora de Brezales Formation (?Kimmeridgian - ?Portlandian) consisting of continental red-bed sandstones and conglomerates deposited as small-scale alluvial fans and sandflats in an arid to semi-arid environment. The location of conglomerate channels was strongly influenced by NE-SW faults;

ii) Rupelo Formation (?Berriasian) dominantly containing carbonates deposited in a shallow hardwater lake, which was briefly hypersaline in its closing stages. Red marls (Las Vinas Member) were deposited on floodplains around the lake. Lenses of carbonate were laid down in small ponds developed on the floodplains. Marginal

lacustrine facies (eg Ladera Member) were subject to frequent subaerial exposure and display a variety of pedogenic features including mottling, desiccation brecciation, microkarst solution cavities, rhizoliths and laminar crusts. Open lacustrine facies (eg Mambrillas de Lara Member) exhibit a fresh-water fauna and flora of charophytes, ostracods, gastropods and bivalves. Sauropod bones and abundant footprints are also present. Lamination is rare in the open lacustrine facies, indicating shallow, oxygenated conditions. The presence of abundant intraclasts suggests significant internal reworking both of marginal facies washed in as a result of lake transgressions or during storms and of open lacustrine facies reworked along internal drainage channels during low lake stands. Intraclasts of open lacustrine facies were locally reworked to form poorly-sorted grainstones, which are thought to represent high-energy shoreline deposits. Cherts also occur (Rio Cabrera Member); these are interpreted as evaporite replacements;

c) Pedroso Group ("Wealden") also comprising two Formations:

i) Hortigüela Formation (?Upper Berriasian - Valanginian) containing distal alluvial clastics and sheet-like oncoidal carbonates of a high-energy lake flat / shoreline;

ii) Piedrahita de Muno Formation (?Valanginian - Barremian) consisting of: dominantly stacked sheetflood sands; local cross-bedded channel sands; crevasse splays; mudstones displaying local mottling and the occasional development of nodular pedogenic carbonates; rare lacustrine deposits including green silts, marls, marly limestones and charophyte-ostracod biomicrites. These are interpreted as the deposits of terminal rivers and lakes in an internal drainage basin with arid to semi-arid climate. Sedimentation was mainly from episodic sheetfloods during catastrophic storms;

d) Salas Group (Aptian) which chiefly comprises distal alluvial mudstones with pedogenic carbonates. Relatively thick sheetflood sands, polymict conglomerates, and calcarenites with reworked oncoids are also present. The Salas Gp sediments are interpreted as deposits of a muddy alluvial plain subject to periodic incursions of large-scale sheetflood sands and conglomerates. The presence of Jurassic and Hortigüela Fm limestone lithoclasts is indicative of tectonic activity and the unroofing and erosion of the subjacent Mesozoic strata;

e) Utrillas Formation (Upper Albian - Lower Cenomanian) consisting of quartz-pebbly sandstones and intercalated quartzitic conglomerates, with occasional mottled, bioturbated and slightly mud-rich hydromorphic palaeosoils. This association records high-energy sedimentation in a fluvial system characterised by multiple internal reworking and preferential preservation of "in-channel" sediments.

2. Isotope data for the Rupelo, Hortigüela and Piedrahita de Muno Formations indicate strong influence of meteoric groundwaters and progressive vadose diagenesis.

3. The Rupelo Formation (although similar to the Aguilar Formation of SW Cantabria) is somewhat unique amongst NW European Purbeck sequences. The dominance of fresh-water, continental environments and the spectacularly-developed pedogenic features reflect stronger cratonic influence and possibly a more slowly subsident history.

4. The Pedroso Group is broadly similar lithologically to the "Wealden" of NW Europe. However, channel fills are relatively rare in the Pedroso Group and the dominance of sheetflood sands is in contrast to the S England Wealden which displays good evidence of lateral channel migration.

5. The lacustrine carbonates of the Rupelo Formation are similar to those of the Late Cretaceous - Early Tertiary of S France (Narbonnais, Le Synclinal de l'Arc, Mormoiron, Barrème syncline) and N Spain (SE Ebro, NE Duero & NE Tajo Basins). Shallow-water biogenic lacustrine carbonates are important sediments in semi-arid continental environments, but they are relatively poorly-documented in the N American and British literature. Pedogenesis is an important (but often overlooked) process responsible for many features of marginal lacustrine facies.

6. The production of nodular pedogenic carbonates within alluvial substrates involves dominantly inorganic precipitation within the soil profile of carbonate derived largely from groundwaters. This mechanism is significantly different from the production of pedogenically-modified or laminar horizons on pre-existing carbonates. The use of the term "calcrete" for both types of soil profile is confusing and potentially misleading. "Calcrete" is best reserved for use as a geomorphological term.

10.1.2. Basin Evolution; Sedimentation and Tectonics.

7. Hercynian basement trends are: a) E-W trending thrusts with transport mostly to the N; those with transport to the S are interpreted as backthrusts, and

b) NE-SW and NW-SE faults. These faults form the late Hercynian conjugate wrench fault system.

The location and orientation of Mesozoic tectonic movements were conditioned by the Hercynian structural framework.

8. An early phase of Triassic rifting was followed by transgression in the Jurassic. It is demonstrated that both subtle facies changes and

thickness variations in the Jurassic reflect differential subsidence of blocks bounded by NW-SE and NE-SW faults.

9. Early Middle Jurassic regression and minor folding produced a minor discordance.

10. The Bathonian - Upper Callovian is a transgressive carbonate shelf sequence, passing from lagoonal micrites to high-energy sandy packstones and eventually to organic-rich limestones with intercalated marls. However, the sequence is often imperfectly preserved because of major truncation below late Cimmerian unconformities.

11. There are several major unconformities present in the Bathonian - Kimmeridgian sequence. These unconformities are superposed to produce very complex surfaces. Each unconformity reflects a discreet phase of extension which caused rotation of listric fault blocks. Footwall uplift resulted in locally severe truncation, especially at the SW basin margin.

12. The fluvial and lacustrine facies of the Purbeck, Wealden and Aptian show strong tectonic control. The NW-SE faults acted as half-graben bounding extensional normal faults, while the NE-SW faults were active as cross-graben wrench or "transfer" structures. Complex, mosaic-like facies distributions were produced. Repeated reversal of the sense of displacement on transfer faults resulted in temporally-variable subsidence patterns. Differences in depositional and erosional histories led to marked regional variation in stratigraphic and sedimentological development.

13. The Aptian unconformity is attributed to the Austrian phase of rifting associated with the opening of the Bay of Biscay. The mid-Albian unconformity reflects the combined effects of gentle

(compactional or thermal) subsidence during regression, followed by transgression and consequent migration of facies belts, finally resulting in a major "steer's head" type onlap.

14. Facies evolution in Late Cretaceous to Eocene times is interpreted as the result of eustatic change and continuing differential compactional or thermal subsidence. Discordances are correlatable with hiatuses in the NW European Chalk sequence and are related to renewed rifting and finally to the onset of gentle compression during the Palaeocene.

15. The Mesozoic - Eocene sequence suffered variable truncation beneath the main Pyrenean unconformity. The degree of truncation was structurally controlled and dependent on location with respect to Tertiary thrusts, backthrusts and transfer structures, and to basin margins.

16. The geometry and derivation of Oligocene alluvial fan conglomerates can be correlated with Pyrenean thrusting (with tectonic transport dominantly to the SW).

17. NE-SW faults (perpendicular to thrust planes) continued to control facies distributions even under compression by conditioning the location of fan-head entrenchment and for the breaching by streams of thrust-ramp and blind thrust culminations. Multiple-sourced fans with polymict conglomerates (clast lithologies include Palaeozoic) were located at the intersections of NE-SW transfer faults with NW-SE thrust culminations. Monomict conglomerates were derived from local sources by erosion of Upper Cretaceous strata and were deposited in front of blind thrust culminations and weakly-emergent thrust fronts in zones between major transfer faults.

18. Pliocene outcrop patterns indicate sedimentation in lobate alluvial fans draining the Sierra de la Demanda. Fan apices were located at the intersection of NE-SW and NW-SE faults.

19. The patterns of Quaternary alluvium and modern drainage show strong influence of NE-SW, NW-SE and E-W basement fault trends, even though the basement was covered by up to 4km of Tertiary molasse. The Arlanzon River through the city of Burgos is thought to lie on a major NE-SW transfer structure offsetting the W Cameros and SW Cantabria half-graben systems.

10.1.3. Structure; Regional Significance.

20. The San Leonardo Fault Zone (previously interpreted as a simple reverse fault or thrust detached on basal Jurassic "Carniolas" evaporites) comprises a complex stack of imbricate thrusts with additional décollement for some horses on Rio Cabrera Member (upper Purbeck) evaporites.

21. The NE-SW structural trend continued to be important in conditioning Pyrenean thrust architecture and deformation history, exerting control upon the location and orientation of thrust-compartmental faults, lateral ramps and backthrusts. NW-SE faults previously active as normal faults under extension may have been reactivated as contractional faults under compression. At the least, they are likely to have influenced the location of frontal ramps and culminations.

22. The structural, stratigraphic and sedimentological evolution of the W Cameros Basin is parallel to many other N Atlantic-marginal basins, eg: S Cantabria; SE Iberian Chains; W Approaches; Celtic Sea; Porcupine; Lusitanian and Wessex Basins; E Canadian Basins. Ancient

(Hercynian and/or Caledonian) structures were reactivated in all these areas. Basin development reflected extensional tectonics associated with the opening of the N Atlantic and the Bay of Biscay. Compressional tectonics related to Pyrenean and Alpine collision resulted in variable degrees of structural inversion in European Atlantic-marginal basins. The Cameros Basin is particularly notable among these for its relatively "cratonic" setting, the extremely rapid Late Jurassic - Early Cretaceous subsidence, and the very strong Pyrenean tectonic inversion.

23. Basement is involved in Tertiary thrusting in the W Cameros Basin. Hercynian thrusts were locally reactivated (eg at Moncalvillo). Deformation was probably at least partially thick-skinned. Thin-skinned thrust splays developed off basement-involving faults in the upper part of the sedimentary pile in areas of major shortening where convenient décollement horizons were well-developed (eg in the W part of the San Leonardo Fault Zone).

10.1.4. Implications for Basin Analysis.

24. The stratigraphic development of the W Cameros is highly variable both along and across structural strike and is strongly controlled by both normal and transfer faults. Stratigraphic heterogeneity in rift basins of this type strongly affects their response to compression. This has important implications for more-deformed collisional orogenic belts, many of which contain telescoped passive continental margin sequences likely to have exhibited even greater pre-compressional heterogeneity and structural segmentation. In these circumstances, some of the assumptions inherent in thin-skinned tectonics (eg relatively simple stratigraphic geometry, little lateral variation along strike, unfragmented extensive slip horizons at the onset of deformation) are probably invalid to a lesser or greater degree.

25. Transfer faults active under extension may be reactivated as thrust-compartmental faults under compression.

26. Transfer faults are critical elements in rift basins which condition the location and style of basin extension. Inverse reactivation of transfer structures under compression is demonstrated here. This has important implications for the evolution of foreland basins and orogenic belts, where the rôle of pre-existing structures is important and currently under-rated.

27. Lacustrine facies are extremely sensitive indicators of minor tectonic movements and might be used to demonstrate the location and sense of very gentle movements in currently poorly-understood central basinal areas. In the case of the S European Tertiary, this technique could be applied in the study of foreland basins to evaluate the importance of possible extensional movements early in basin development. Areal facies distributions in Mesozoic and Tertiary lacustrine systems may be applied to reveal deeply-buried basement structures. Vertical facies sequences will record the timing of tectonic events and climatic change.

10.2. Suggestions for further work.

1) More rigorous structural analysis, particularly of the SW basin margin. The Iberian Chains have been much neglected by structural geologists.

2) Detailed sedimentological study of the Talveila Formation, involving microfacies analysis and palaeocurrent studies. Aim: lithostratigraphic comparison and correlation with other Late Oxfordian marginal marine facies on the N Tethyan margin. A study of this type should provide valuable insights

into the respective influence of tectonics and eustacy on facies evolution at the onset of the late Cimmerian phase.

3) Detailed sedimentology of organic-rich Middle - Upper Callovian strata. A suitable approach might involve: microfacies analysis; visual kerogen descriptions; vitrainite reflectance; geochemical analyses including total organic carbon determination and pyrolysis yield.
Aim: Characterisation of hydrocarbon source type and maturity / hydrocarbon generation potential.

4) Detailed testing of the applicability of "Lacustrine Facies Architecture" as a tool in basin analysis. There are important potential economic applications - a better understanding of the tectonic controls on lacustrine sedimentation would facilitate more accurate prediction of the location of organic-rich lacustrine source rocks, and perhaps even of subsurface structure.

Approaching Puerto de Palombrera (1260m).
SW Cantabria, March, 1985.

Looking W towards Salas de los Infantes.
Dusk, September, 1983.

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100

HORTEZUELOS II

HORTIGÜELA
FORMATION

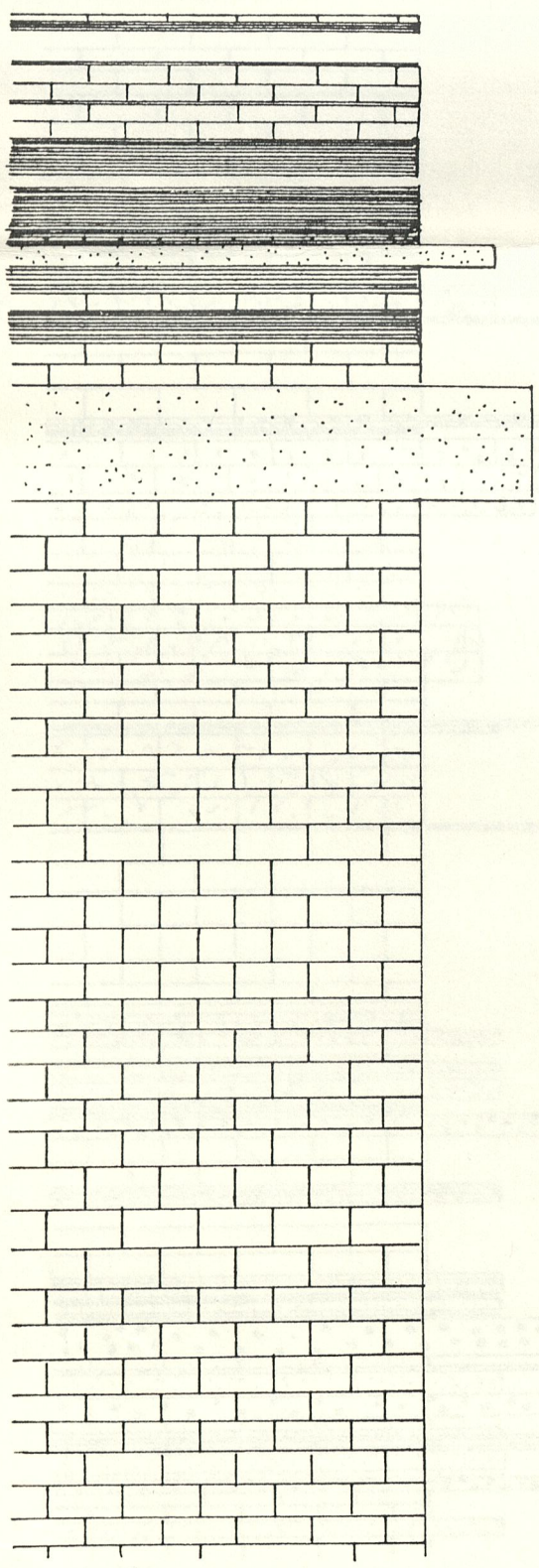
90

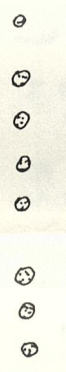
80

70

60

50





? Rio Cabrera Member

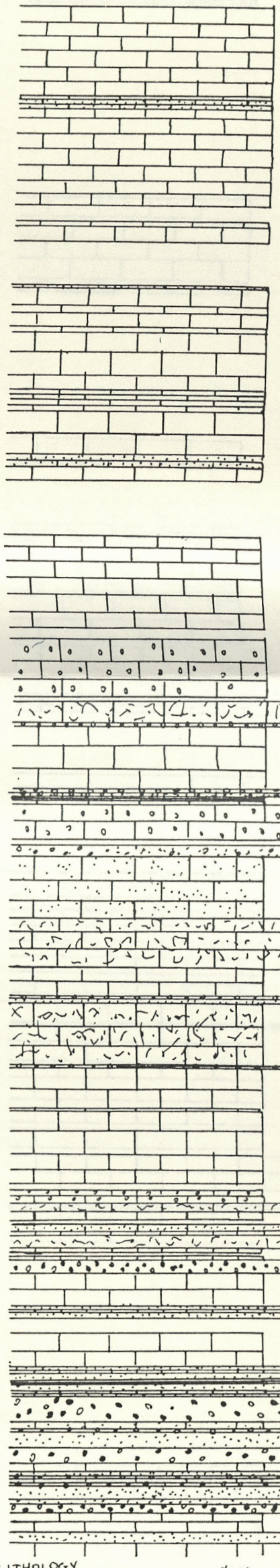
Ladera Member

RUPELO
FORMATION

mudstones - pedogenic modification
black pebbles, eggs, geopi, abcs, etc.

metres LITHOLOGY clay+ silt f m c grit gl gravel

RUPELO I



grey — mudstone/packstone □

slightly M mudstone

mid grey — mud/packstone

midgrey — slightly M

midgrey — M X □ geops abc

mid grey — M X

pink

M
M
M

wackestone

M

M

mudst

X □ |E ■ abc

|E □ nodular M

rubbly

M □ ■

mud/wackestone nodular □

X

|E □ abc

|E abc □ ■ prismatic

red/purple

rubbly M

|E abc □ prismatic/reticulate

polymict

□ wack/packstone

□ M ■

X X

grey/red — nodular rubbly/wackestone

X M X

pink/grey — brecciated □

X

X X □

red/grey

M

red/grey — X X

X X

polymict

pink red — polymict

grey grey — X

polymict M

X

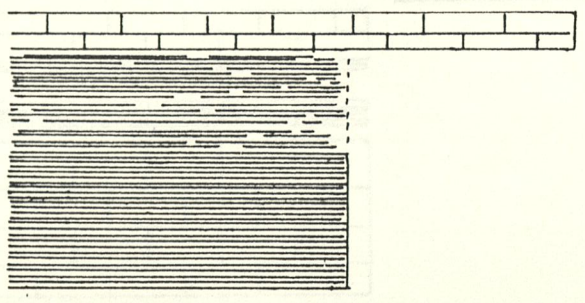
yellow yellow — calclithite/calcarenite packstone

Ladera Member

Señora de Brezales Formation

Talveila Formation
Marine Jurassic (? Lower Callovian)

RUPELO II



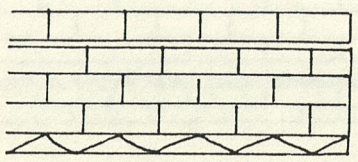
140

grey-brown wackestone / gl

purple-grey M

130

(Football pitch)



diffuse ?

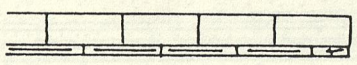
dark grey

orange-brown packstone

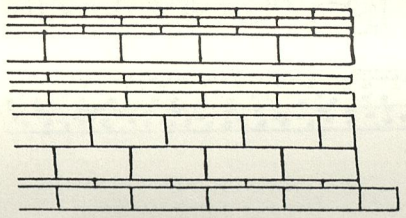
yellow ruggy limestone

120

Rio Cabrera Member



mid grey wackestone



110

grey wackestone

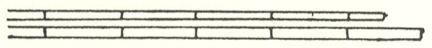
dark grey

mid grey wackestone

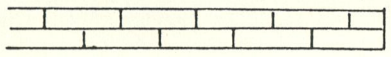
grey-brown

grey-brown

dark grey-orange yellow-grey wackestone packstone



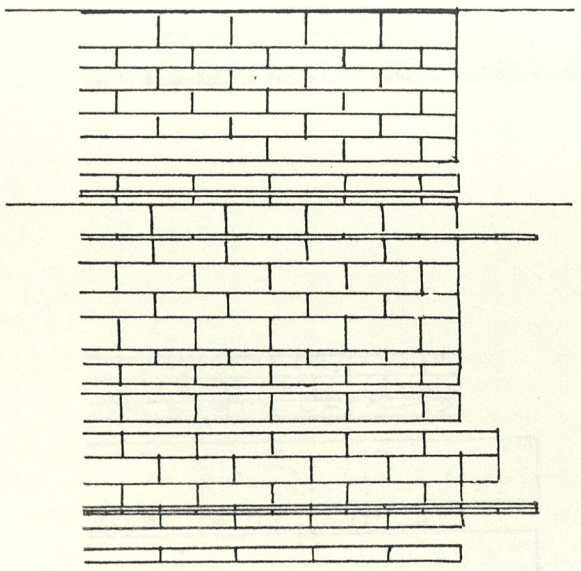
dark grey packstone



grey-brown mudstone

100

Mambrillas de Lara Member



90

mudstone geops abc

mudstone / wackestone

geops

White wackestone / abc packstone

massive

grey-pink M wackestone

pinkish wackestone

packstone abc

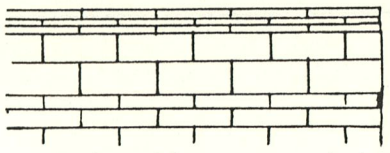
dark grey Bioturbated

wackestone

80

Ladera Member

grey marly soil



70 metres

beige mudstone wackestone geops

mudstones / wackestones geops M abc

white mudstone

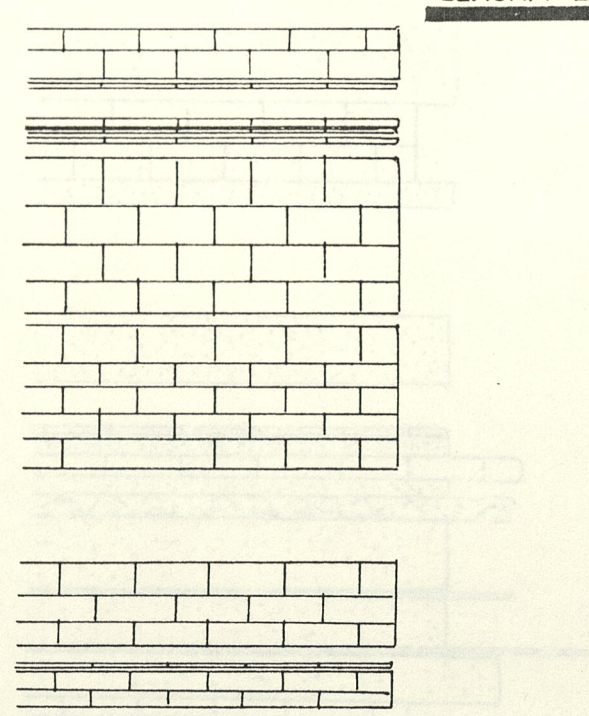
dark grey mudstone

LITHOLOGY clay+ silt f m c grit gl gravel

SEÑORA DE BREZALES

RUPELO FORMATION

Ladera Member



X

M

M

M

XX

XX

X

conglomerates

SEÑORA DE BREZALES FORMATION

polymict limestone conglomerate

red sands

M

Red

M

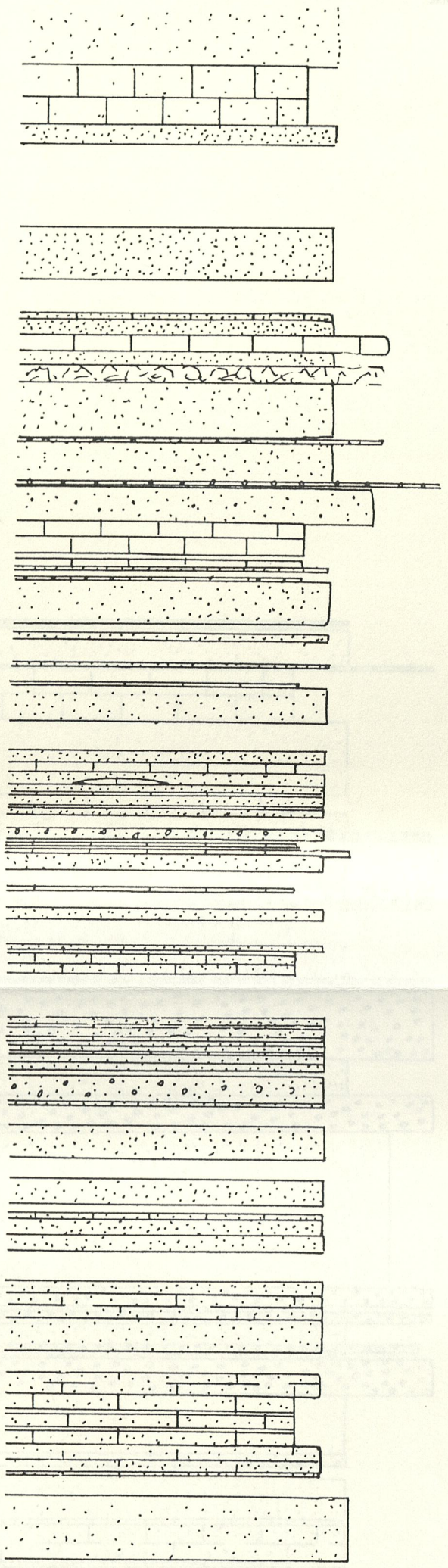
Red

Red

Red

Red

CASTROVIDO I

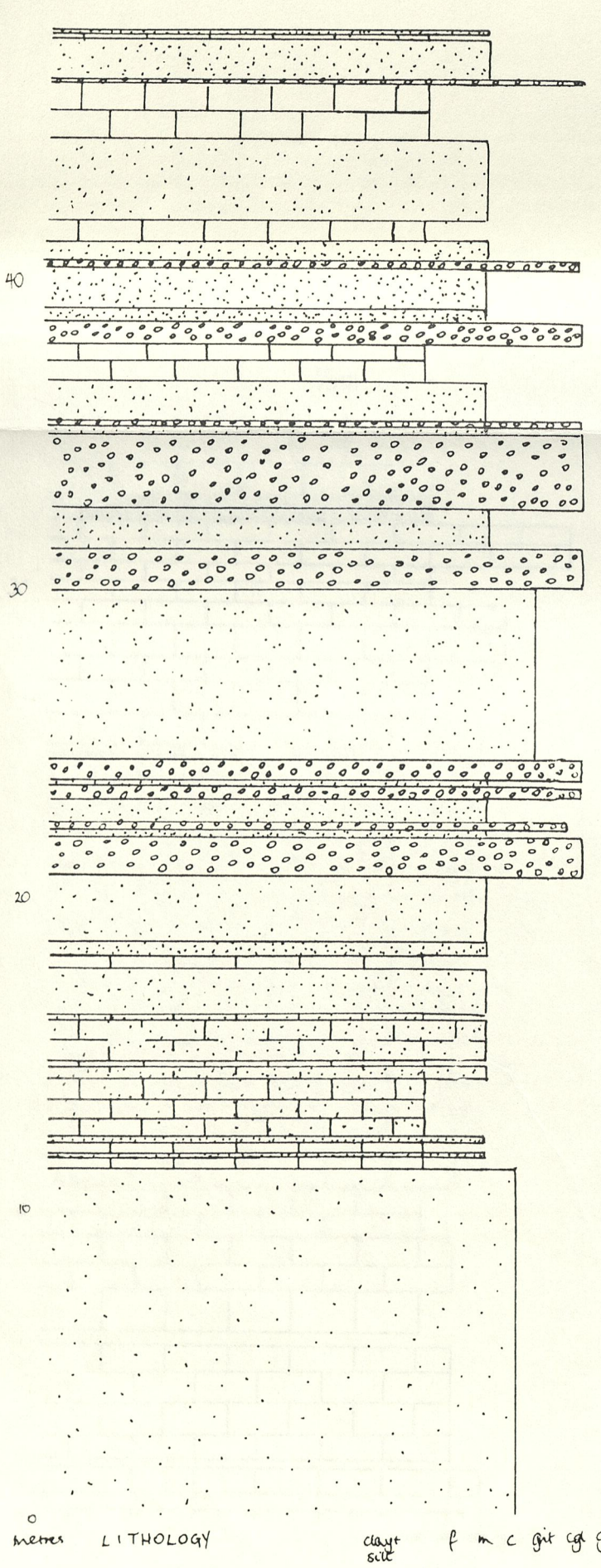


?red
grey
red
red
red
rubbly
red
rubbly palaeosoil
sandstone brecciated
red
grey
red grey
grey
red
red
red
black
red
grey
red
red
red
red
red
red
matrix-supported
matrix-supported
clast-supported
clast-supported polymict

Señora de Brezales Formation

Marine Jurassic
Lr / ?M Callovian

HORTEZUELOS



red

red

SEÑORA DE BREZALES FORMATION

yellow /beige

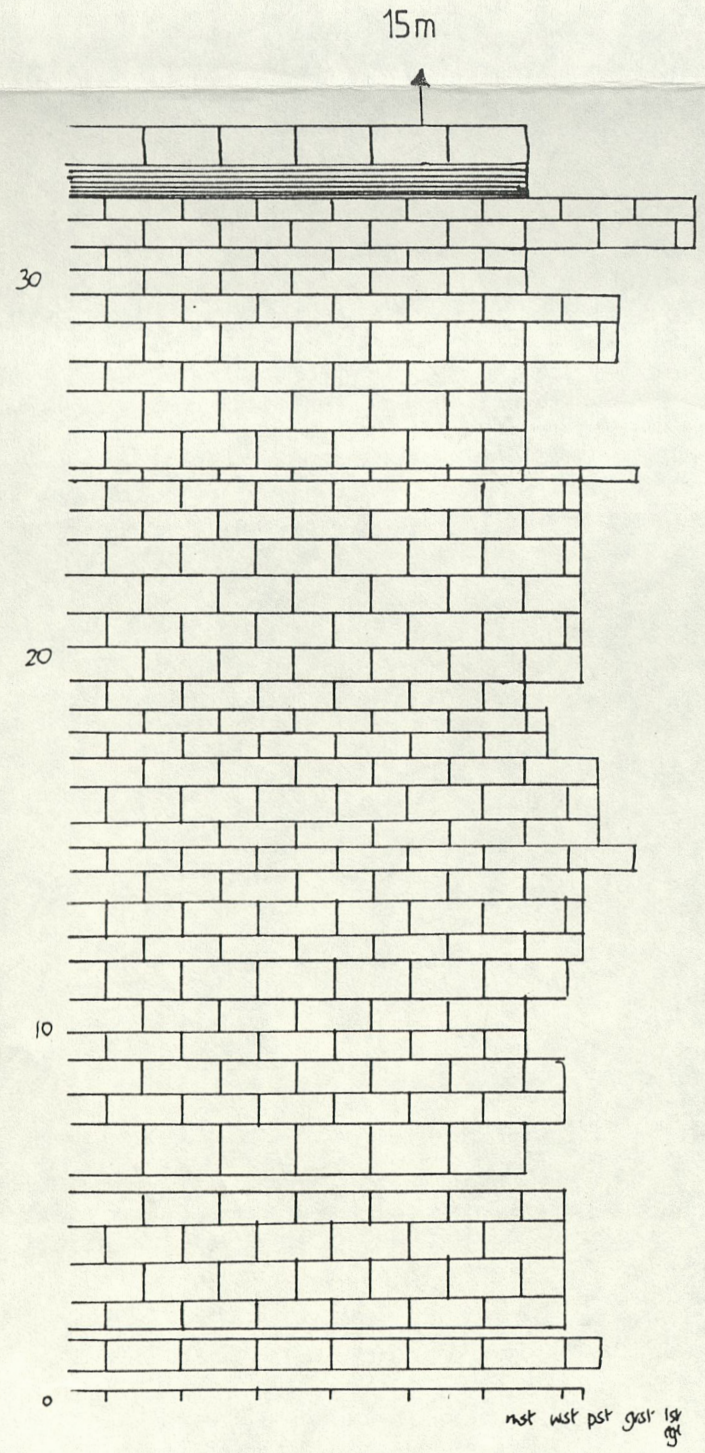
yellow /beige

TALVEILA FORMATION

yellow

sandstones + grits

VALPARAISO



dark grey — Rupelo Formation

dark grey

dark grey

mid grey

grey → grey-brown — Mambrillas de Lara Member

dark grey

mid grey

mid grey

mid-dk gy

mid grey — Rupelo Formation

mid grey

grey - pink

pale grey

pale grey

pale grey

pale grey — Ladera Member

pale grey

pale grey

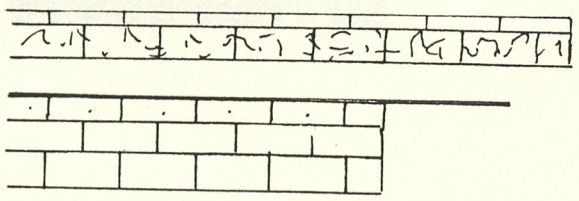
pink-grey

pale grey

210

200

RUPELO III

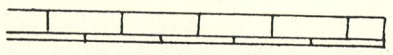


rubbly

rubbly top

grey

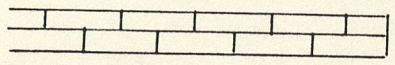
190



mid grey

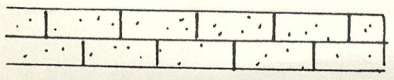
packstone abc

180



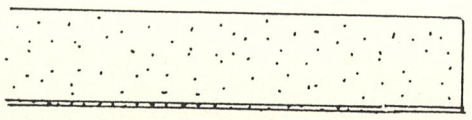
abc

170

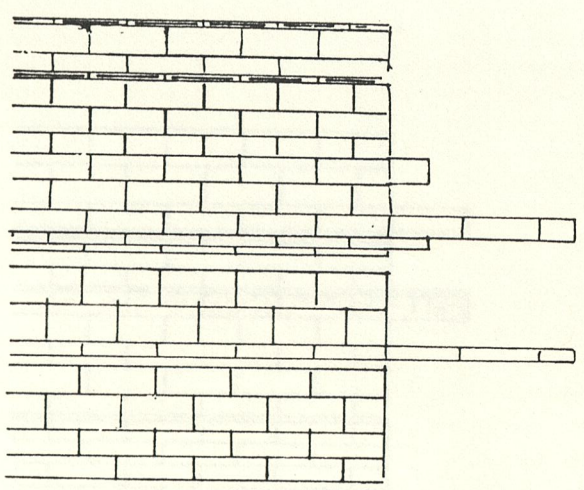


m

Rio Cabrera Member



calcareous



160

calcareous

grey

wackestone
packstone

m

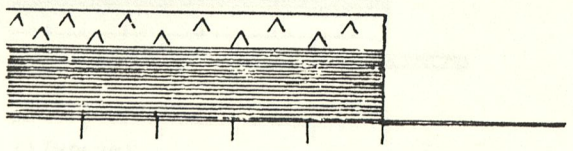
packstone

conglomeratic wackestone

brecciated packstone

wackestone

150



white

chert

grey

conglomeratic top

140 metres

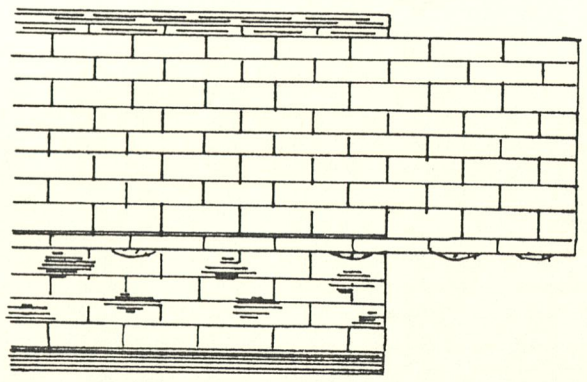
LITHOLOGY

clay+ silt

f m c grit cgl gravel

RUPELO IV

270



marly wackestone
brecciated lst/
grainstone
lst cg bs/
wackestone
dark grey
green-grey grainstone channelised
mid grey
green
mudstone

260

250

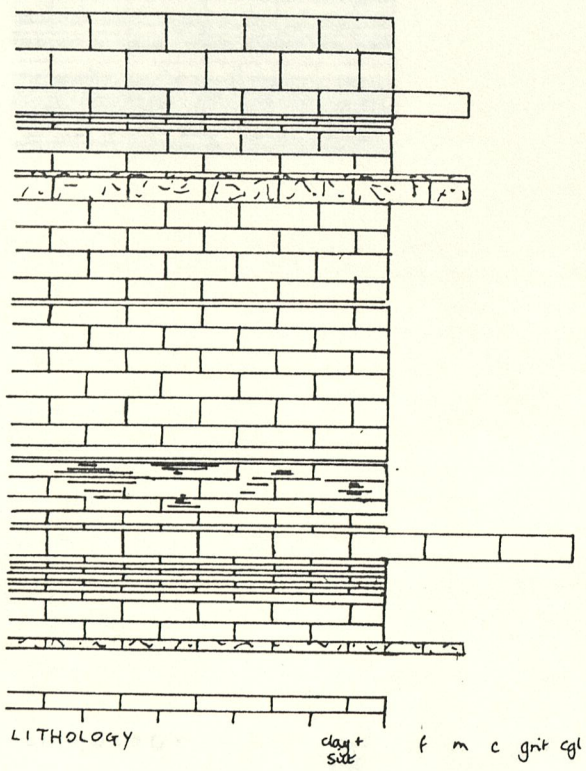
?red marl / green marl
? tabular
fine sandstones

240

230

Rio Cabrera Member

220



brecciated
brecciated
grey
wackestone abc
rubbly, conglomeratic
geops
packstone
wackestone-f
packstone
yellow patches
wackestone
yellow
mudst
geops
purple
packst/grainstone
packst/grainstone
grey
grey
mudstone/wackestone
mudstone/wackestone
off white
mid grey
mudstone/wackestone
mudstone

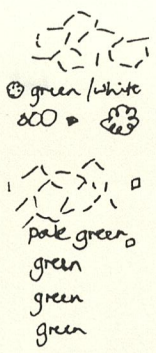
210

200 metres

LITHOLOGY

clay + silt f m c grit cgl gravel

RUPELO V

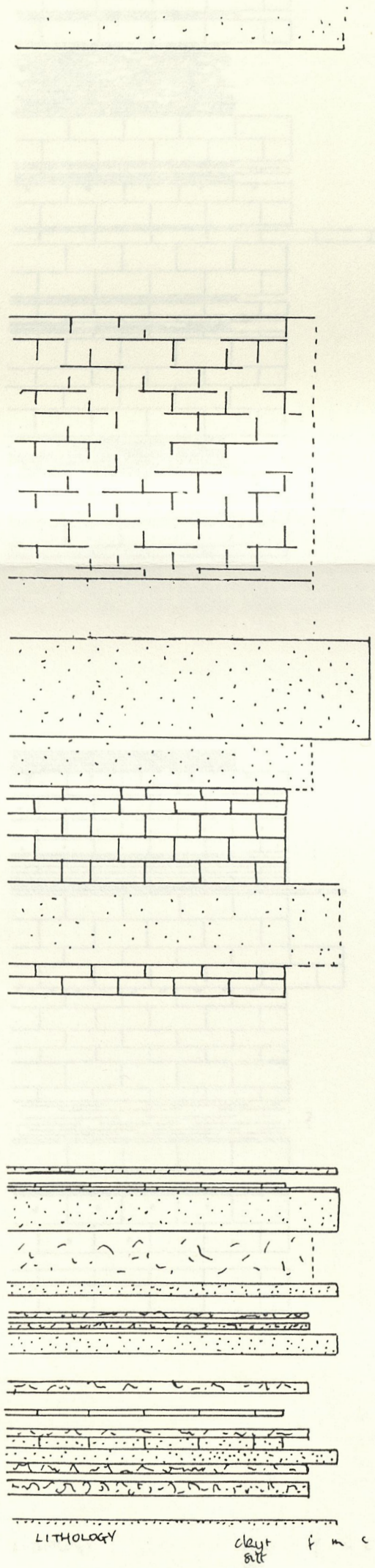


Rio Cabrera Member

brown marl?

metres LITHOLOGY clay+ silt f m c grit cgl gravel

CASTROVIDO II



Piedrahita de Muño Formation

yellow wack/packstone
grey

?red

? Rio Cabrera Member

Ladera Member

RUPELO FORMATION

red
grey
white
pink

rubbly palaeosol ?

pink

rubbly palaeosol
rubbly

pink

rubbly

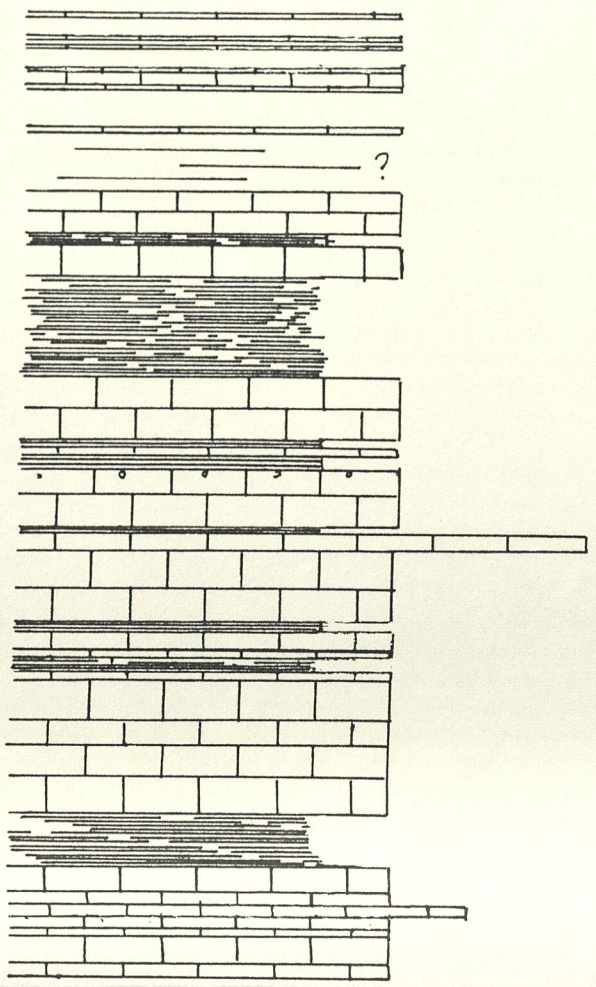
palaeosol

pink

rubbly palaeosol
rubbly palaeosol

Señora de Brezales Formation

MAMBRILLAS DE LARA



grey muddy
grey muddy pale grey

mid to darker grey
mud/wackestone
wackestone

green-khaki

wackestone
grey
wackestone
cgl top

green-grey
green marl parting
wackestone
some grainstones

grey
grey
wackestone
wackestone

grey/green
grey/green
karst top

grey/green
green
marly lst
brecciated wackestone
brecciated top

grey
wackestone

breccia? pedogenic

brecciated top wackestone

green marl wackestone

darker grey
grey marl?
darker grey
grey

grey
laminar top

RUPELO FORMATION

Rio Cabrera Member

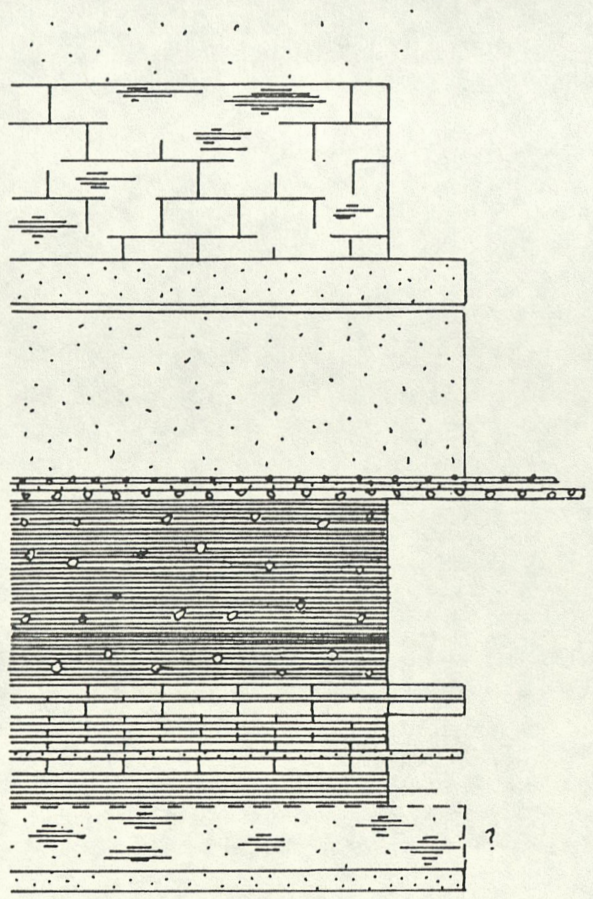
Dinosaur Bed

Mambrillas de Lara Member

Ladera Member

MONCALVILLO

30



channel sands tabular
overbank muds/ silts.

Piedrahita de Muño Formation

Hortigüela Formation ? / Zurramujeres Member

Hortigüela Formation?/ San Martín Member

nodular (paleosol?)

Rupelo Formation / Ladera Member

nodular (paleosol?)

grey
dark grey red
red

Señora de Brezales Formation

grey
yellow
grey
yellow

polymict

packstones

samples, structures

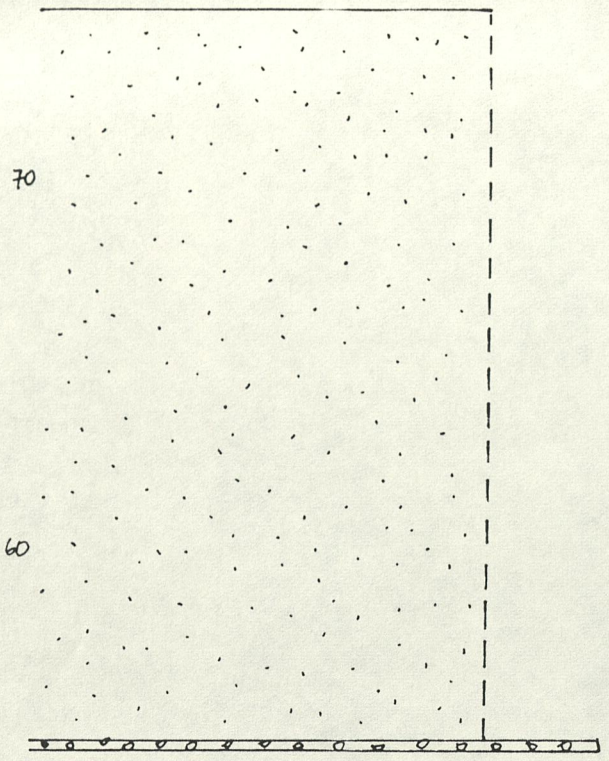
Marine Jurassic / Lower Callovian ?

TALVEILA

80

Utrillas Formation

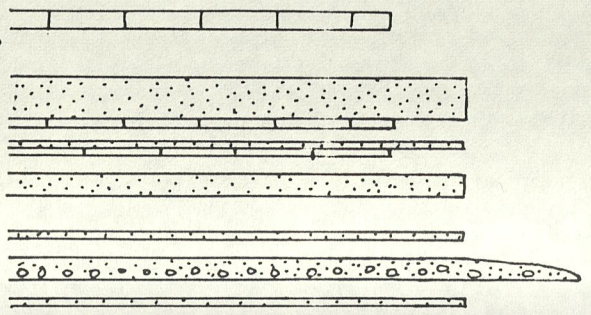
70



60

Salas Group

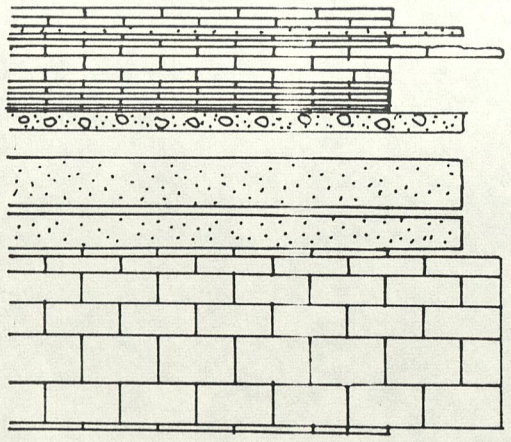
50



Pedroso Group

40

30



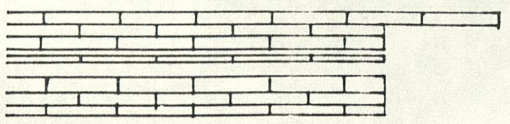
Rupelo Formation

(Ladera Member)

20

Señora de Brezales Formation

10



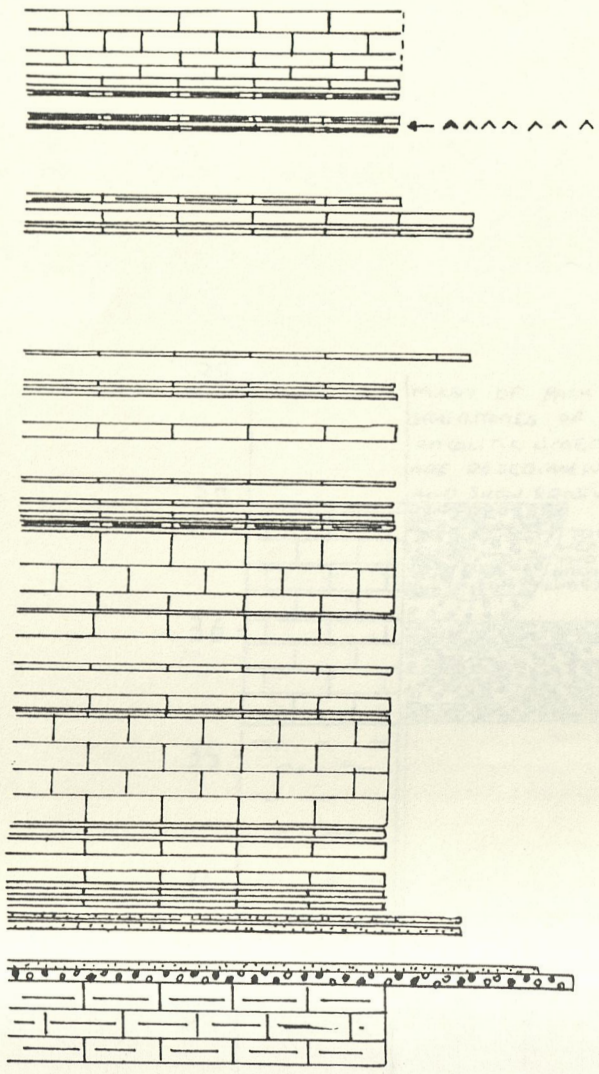
Talveila Formation

metres

0 LITHOLOGY clay+ silt f m c grit cgl gravel colour samples, structures remarks.

SEÑORA DE BREZALES II

TORRELARA



chert partig?

Rio Cabrera Member

yellow

packstone

packstone

wackstones

Mambrillas de Lara Member

wackstones

white to grey

cg

interbedded khaki marls
– diffuse clastic lamination

Torrelara Bed

polymict

RUPELO FORMATION

white – grey

marly mudstones

Ladera Member

grey-green

very marly

red

Las Viñas Member

M

M

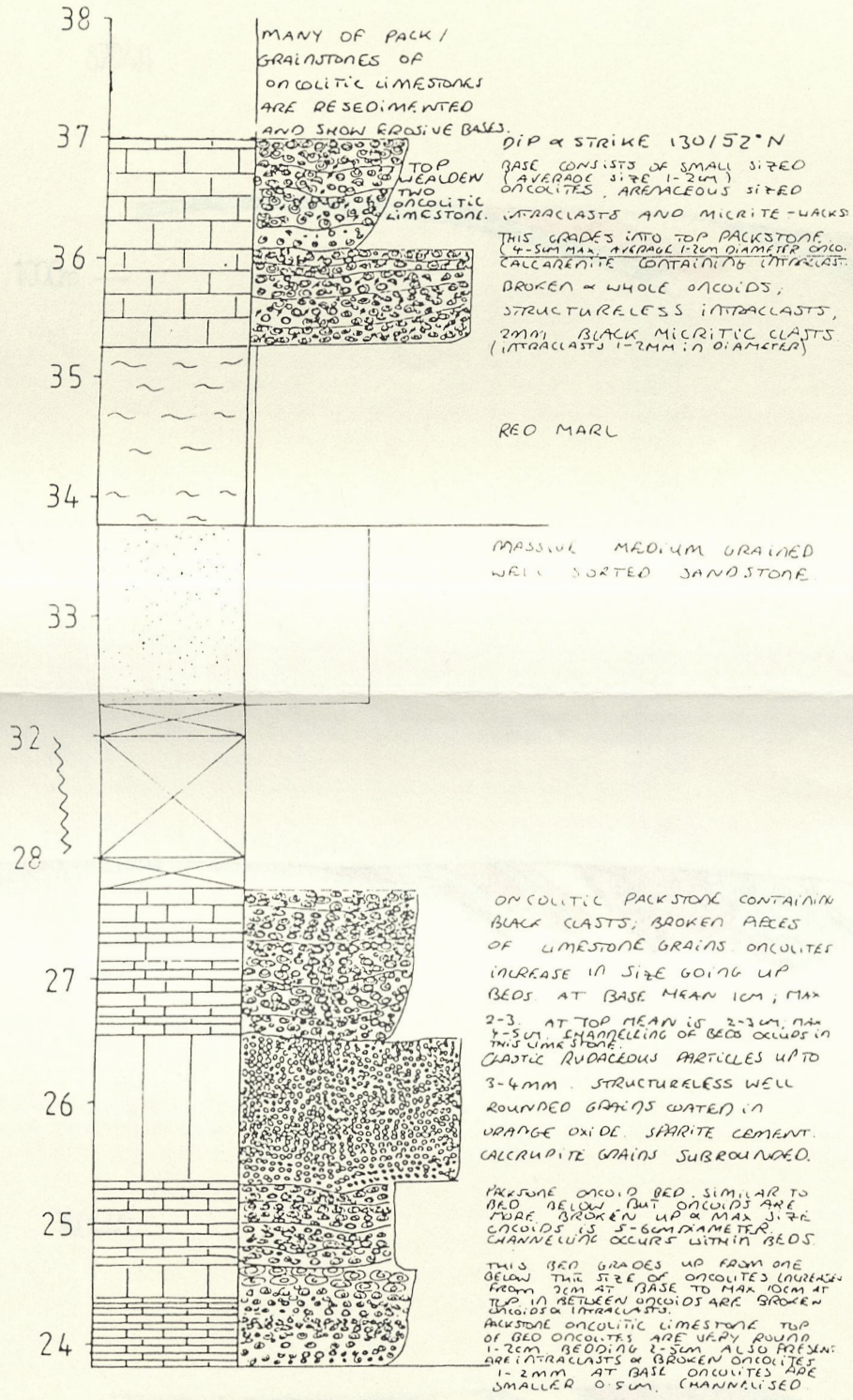
Marine Jurassic
(?Lower Callovian)

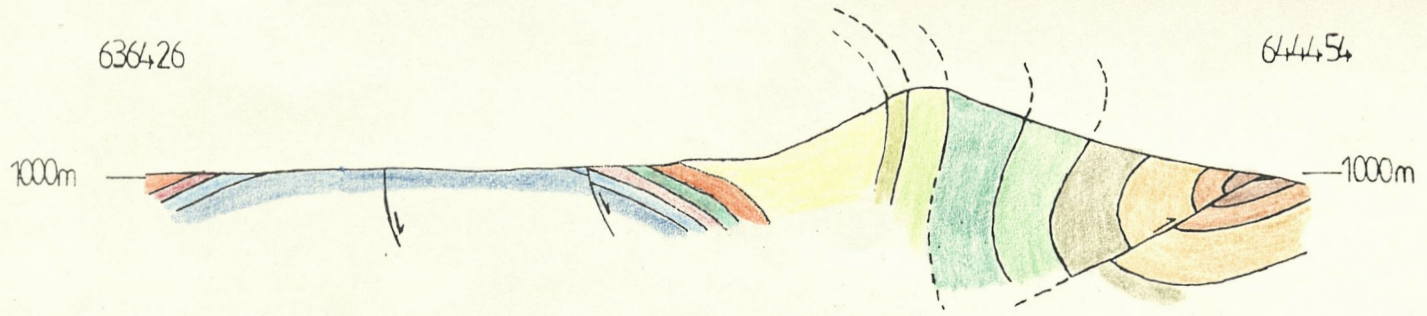
MAMBRILLAS DE LARA II (NORTH)

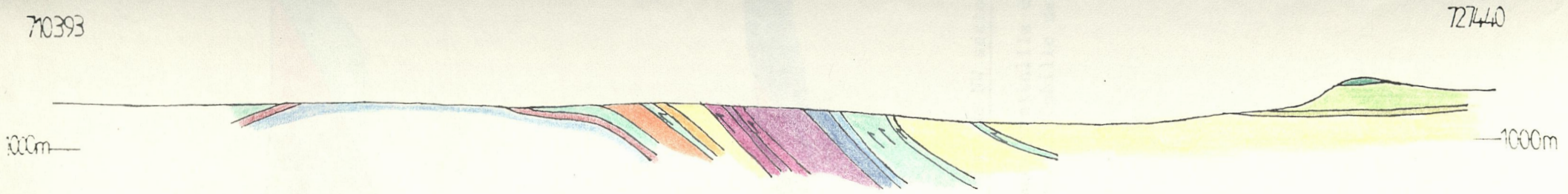
Salas Group

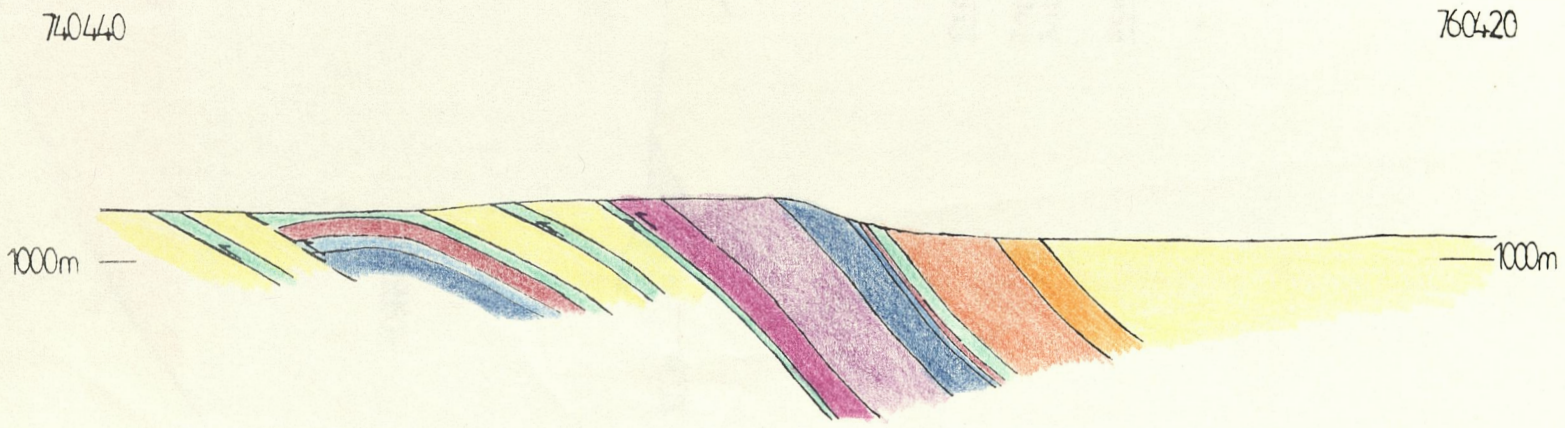
Hortigüela Formation

(Zurramujeres Member)



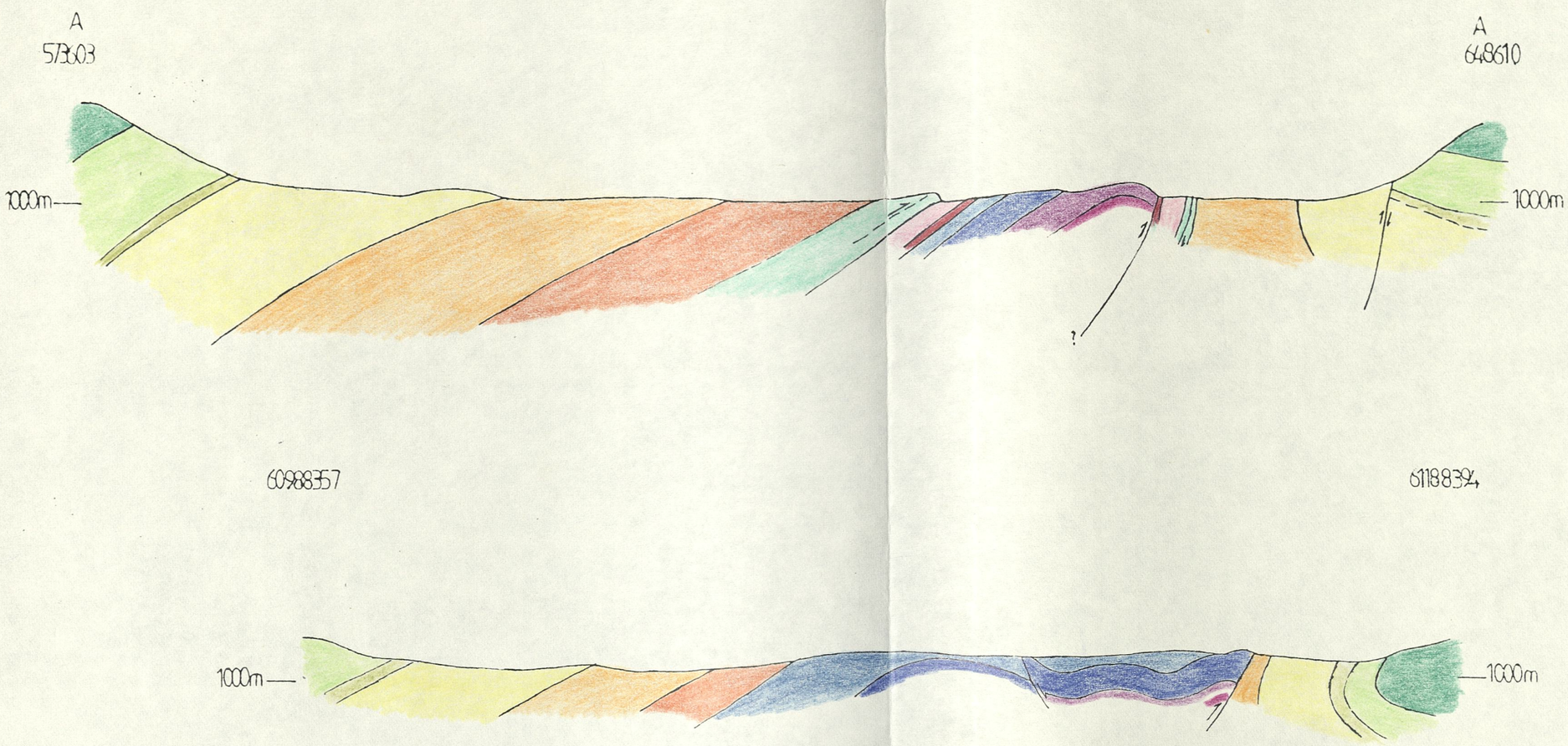






Cross-sections across the San Leonardo Fault Zone (West).

Scale 1: 20 000.



Cross-sections across the NW sector of the study area.

Top: Section through Quintanilla de las Vinas. Area mapped by Faulkner (1985).
Bottom: Section through Cubillo del Cesar. Area mapped by Welle (1982).

Scale 1: 20 000.

PIEDRAHITA DE MUNO: PALAEOCURRENT DATA.

	BED		CROSS-BED			
NO.	DIP (°)	STRIKE (°)	SET THICKNESS & TYPE	X-BED THICKNESS & TYPE	DIP (°)	STRIKE (°)
1)	86	310N	1-2cm planar	30cm planar	54	308N
2)	86	310N	1cm planar	43cm trough	68	306N
3)	60	314N	3cm planar	25cm planar	65	308N
4)	60	314N	½cm planar	19cm trough	70	308N
5)	63	312N	<½cm planar	15cm trough	68	306N
6)	63	316N	<¼cm trough	18cm trough	74	308N
7)	40	274N	<¼cm planar	40cm planar	45	274N
8)	12	324N	½-1cm planar	35cm planar	25	302N
9)	04	318N	<1cm planar	24cm planar	23	312S

PLATT N (1986).
with thanks to Tom Faulkner.

Sedimentology and Tectonics of the Western Cameros Basin,
Province of Burgos, Northern Spain.

Nigel Howard Platt, B. A..
Worcester College, Oxford.
Department of Earth Sciences, University of Oxford.

Thesis submitted for the degree of Doctor of Philosophy,
Trinity 1986.

LIST OF ENCLOSURES.

1: 20 000 Maps: San Leonardo Fault Zone
Quintanilla de las Vinas - Mambrillas de Lara
Quintanilla de las Vinas - Cuevas de San Clemente

Cross-sections: San Leonardo Fault Zone (4)
Quintanilla de las Vinas - Mambrillas de Lara (1)
Quintanilla de las Vinas - Cuevas de San Clemente (1)

Stratigraphic Column: Key for maps and cross-sections.

Published paper: PLATT, N. H., (1985). Fresh water limestone facies from the Cameros Basin, N. Spain. Int. Assoc. Sediment. 6th Reg. Mtg., Lérida, Spain, 1985, (abstracts), 363-366.

APPENDIX.

Selected sedimentological logs:

1: 200 scale: Mambrillas de Lara (upper part of Rupelo Fm)
Moncalvillo
Talveila
Valparaiso
Rupelo
Castrovido
Senora de Brezales (3km N of Espejon)
Hortezuelos (Rio Helechal section)
Torrelara

1:580 scale: Mambrillas de Lara North (Hortigüela Formation)

1: 20 scale: Quintanilla de las Vinas (Senora de Brezales Fm)
Quintanilla de las Vinas (Rupelo Fm: Ladera Mb)
Cubillo - Cubiejo
Mambrillas de Lara (basal part of Ladera Member)
Piedrahita de Muno (basal part of Pedroso Group)
Cubillo del César (part of Salas Group)

Key: to sedimentological logs

Palaeocurrent data: Piedrahita de Muno section.

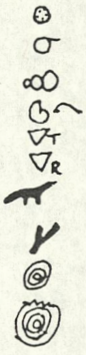
KEY TO SEDIMENTOLOGICAL LOGS.

Lithologies

Conglomerate
Conglomerate (polymict)
Grit
Sandstone
Sandstone (calcarous)
Silt/marl
Marly limestone
Limestone
Brecciated limestone
/palaeosol
Limestone (peloidal/pelleted)
Limestone (sandy/impure)
Chert
Vuggy limestone
No exposure

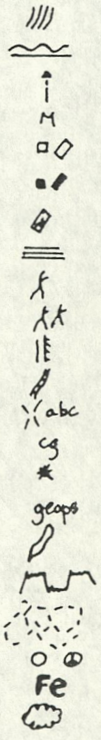
Bioclasts

Charophytes
Ostracods
Gastropods
Bivalves
Brachiopods (Terebratulids)
Brachiopods (Rhynchonnelids)
Sauropods
Vegetal débris
Oncoids
Fragmented oncoids



Sedimentary Structures

Cross-bedding
Ripples
Normal grading
Mottling
Limestone intraclast
Dark intraclast / black pebble
Marl/silt intraclast
Lamination
Roots
Root Crust
Large rhizoliths
Sparry cracks
Angular-branching cracks
Circumgranular cracks
Crystallaria
Geopetal Cavities
Bioturbation / burrows
Stylolite
Polygonal desiccation cracks
Nodules
Iron concretions
Smell (?bituminous)



IAS
International Association Of Sedimentologists
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Universitat Autònoma De Barcelona

ABSTRACTS

6th EUROPEAN REGIONAL MEETING

1985 • LLEIDA • SPAIN

pp363 -366.

FRESH WATER LIMESTONE FACIES FROM THE CAMEROS BASIN, N. SPAIN

N.H. PLATT,
Department of Earth Sciences,
University of Oxford,
Parks Road,
Oxford. OX1 3PR
England.

INTRODUCTION

A thick sequence of continental sediments was deposited in the Cameros Basin, approximately 170km NNE of Madrid, between Kimmeridgian and Albian times (Salomon, 1982). The present study concerns carbonate facies from the western part of the basin which belong to Group I (Kimmeridgian to Lower Berriasian).

The fresh water limestones can be assigned to the following facies associations:

1) open lacustrine

2) palustrine

3) alluvial plain.

The lacustrine and palustrine carbonates reach maximum thicknesses of 50-100m but are locally absent or dramatically thinned in positive areas adjacent to previously active synsedimentary normal faults. The palustrine/lacustrine sequence also thins to the NE where a transition can be observed into an alluvial plain facies association which includes pedogenetic carbonates but shows evidence of only minor lacustrine incursions.

1) Open lacustrine facies association

This consists of micritic limestones characterised by fresh water fauna and flora (ostracods, gastropods, charophyte stems and oogonia) and by the absence of pedogenetic features. Lamination is rare in these sediments, although it has often been cited as a typical feature of lacustrine deposits (Picard and

High, 1972). The absence of lamination suggests oxygenated conditions. The association with almost intact charophyte stems suggests a quiet, shallow water environment as charophytes today live in shallow water and their stems are relatively delicate. Where charophyte stems are absent and ostracods and charophyte oogonia predominate, slightly deeper (but not anoxic) conditions may be inferred. Rarely, 1mm graded silt laminae are found. These are the result of terrigenous clastic input into the lake, perhaps associated with river-generated density currents. Intraclasts from more marginal environments are locally common and indicate transport from the lake margin.

2) Palustrine facies association

A spectrum can be recognised from partially altered lacustrine micrites to palaeosols, thus demonstrating the formation of palustrine limestones by the progressive pedogenesis of lacustrine carbonate substrates at fluctuating lake margins.

Periodic exposure has resulted in a variety of pedogenetic features. Desiccation cracks and circumgranular shrinkage cracks are common, the latter often developed around glaebules ("caliche pisoids"). The presence of abundant rhizoliths (Salomon, 1984) indicates widespread bioturbation by roots, while solutional and microkarst cavities, partially filled by geopetal crystal silt, document the onset of karstification on emergence. Intensely mottled and brecciated palaeosols punctuate the palustrine sequence, pointing to occasional longer emergence. The palaeosols locally include red silt rip-up clasts, probably derived from alluvial plain areas.

The presence of laminar crusts indicates that some carbonate was added to the sequence within the palustrine zone. Planar laminar fabric is displayed at several horizons within the

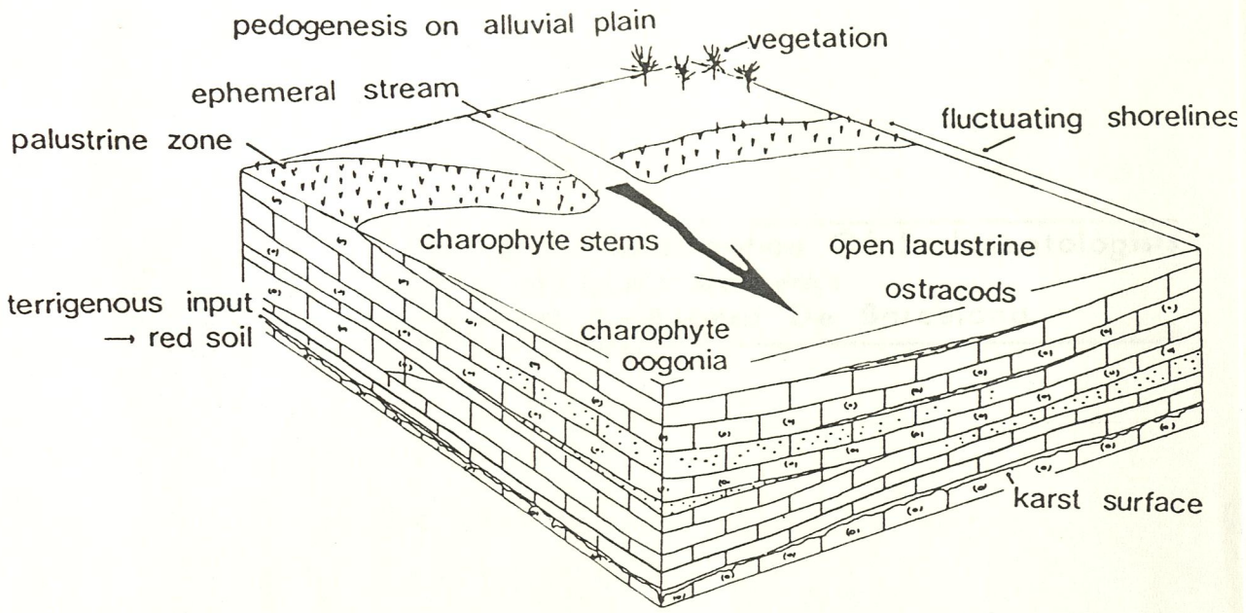
palustrine sequence, as well as in a 1 metre thick limestone bed which lies either directly on the marine Jurassic pediment (Mensink and Schudack, 1982) or above the channelised basal conglomerate that cuts down into the pediment. The planar laminar fabric consists of millimetre-scale wavy laminations of alternating light and darker micrite and is associated with pedogenetic features and rhizoliths. The encrustation of irregular and apparently lithified surfaces suggests a subaerial origin. Laminar crusts (croûtes zonaires) have been described by many workers, but their origin is controversial, having been attributed to precipitation of carbonate, either at the surface or within a soil profile (Multer and Hofmeister, 1968; James, 1972; Robbin and Stipp, 1979; Freytet and Plaziat, 1982) or to colonisation by endolithic and epilithic lichens (Klappa, 1979). Either interpretation is consistent with the apparent absence of algal filaments from these laminar crusts, and the observed variation in their detailed morphology might reflect several mechanisms of formation.

3) Alluvial plain facies association

This consists of an alternation of bioturbated siltstones with massive sandstones and probably represents rapid sediment deposition associated with terminal rivers (Friend, 1978) or an alluvial fan toe (Heward, 1978). The vertical sequence observed is consistent with either interpretation.

The fine grained sediments display pedogenetic features. Slightly calcimorph soils, showing root fabrics, can be recognised and nodular limestones (calcimorph soils) locally exhibit floating quartz grains and displacive carbonate fabrics. Carbonate was derived largely by inorganic precipitation within the soil profile.

Fig. 1. Simplified facies model (modified after Freytet & Plaziat, 1982).



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Cubillo del César (3), September 16th, 1983.

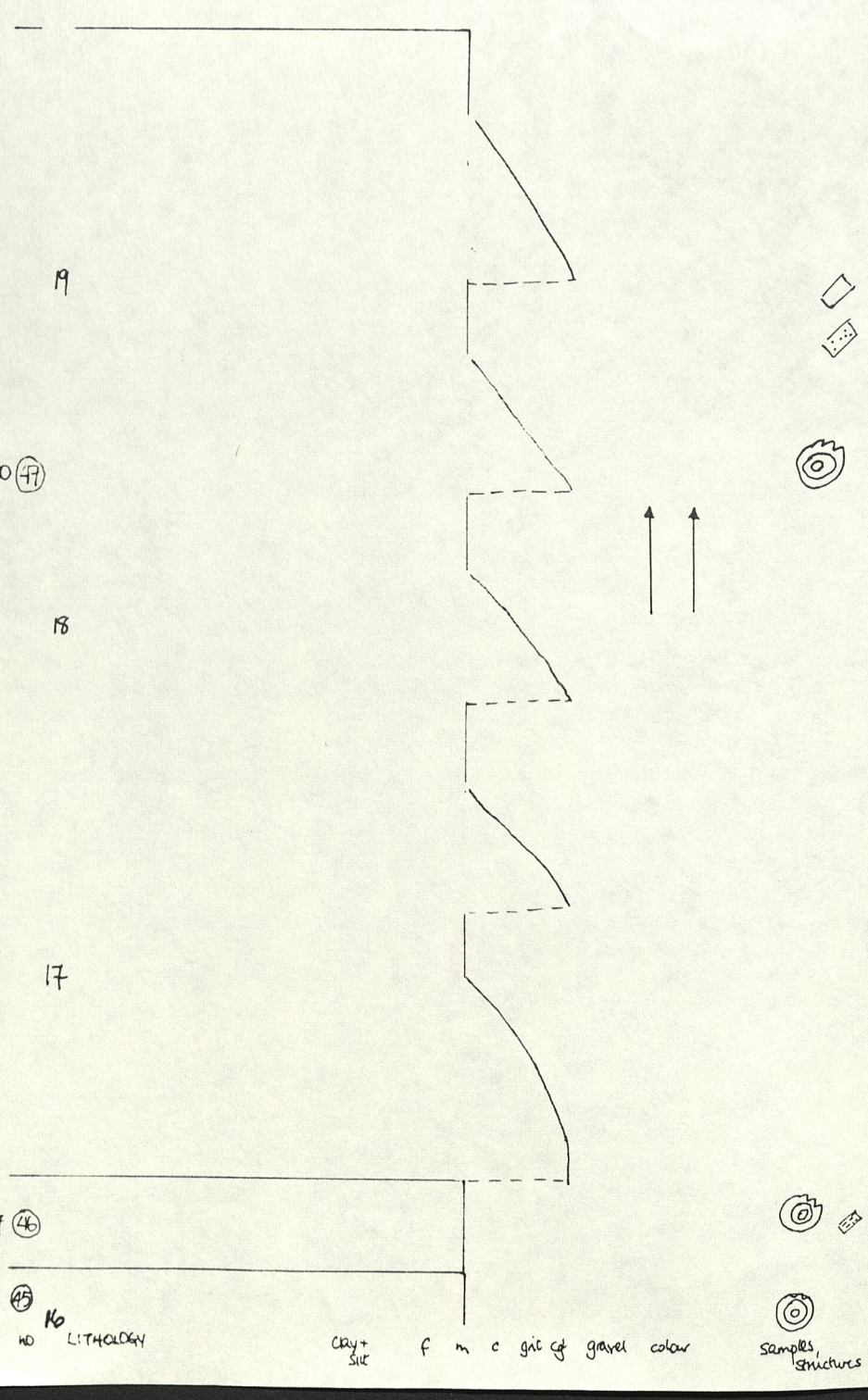
24 Part of Salas Group

23

22

21

20



continuing alternation of ssts as below with sandy, pebbly conglomerate.

clasts : white micritic limestone, grey weathering ?5%, up to 5cm. commonly 2cm.
: med-fine sst/silt ?5%,
: orange carbonate
: qtz clasts subround or cracked subangular
-white, light brown, orange, grey, rarely red.

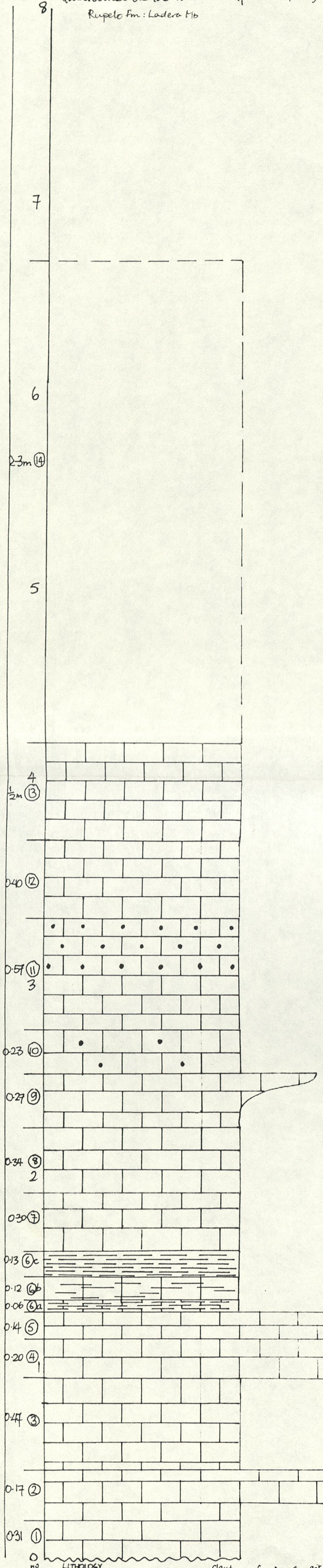
complex organis": 30-50cm fining up "flaws" – matrix supported with reworked oncoids.

Poorly cemented void rich sst., pebble strings 3-4cm separation – clasts include qtz, (subrounded, angular, cracked) rich pink shale, yellow orange carbonate and oncoids

cemented matrix, clasts qtz (rounded, cracked 1-2cm). Voids + pebbles in strings as before.
light brown-orange oncoids 1cm, larger pale brown to ochre gold carbonate clasts.

Quintanilla de las Viñas April 27th, May 16th, 1984.

Rupelo Fm : Ladera Mb



horizontal exposure - pavement.
thickness very estimated.

estimated thickness

top of exposure

dull grey [498]

massive

beige brown grey ● (●) (■) [496] [496A] [497]

pelvidal micrite. excellent 1-2mm pelads and circumgranular spar cracks and rare 1mm dark clasts

darker brownish grey ((◆)) (b) [495]

local pelvidal fabric only.

mid grey [493] [494]

massive. top is a prominent surface - lithology change?
top is a limestone breccia, 2-3mm angular pale beige micrite clasts and dark micritic matrix.

pale white to beige [492] (●)

very rare ½mm dark points, somewhat diffuse.
generally massive.

dark grey to mid grey (weathered

weak surface at top
1mm black clasts.
2-3mm pale white micrite intraclasts.

grey brown ?red

few limestone fragments. ants nest bringing brick red soil to surface.

mid grey (yellow-white weathering) dark grey [491]

?muddy. black flecks scattered, irregular 1mm - 3mm.

marly, apparently very muddy impure micrite "underneath". dark material ½mm flecks.
top is major ledge surface

white b

Similar to (4). light coloured micrite intraclasts, rounded 1-2mm. abundant 2-3mm dark clasts.
1mm brown micrite flecks

weathers - giving
white marly limestone matrix, rounded "clasts" up to 1cm of pure grey micrite, dark clasts up to 5mm.

White b

larger micrite clasts include dark flecks, some brown micrite.

pale grey-beige

more massive. diffuse mid grey intraclasts as (1)
black clasts 2-3mm. common, irregular.

light grey - pale beige

weathers into 1-2cm nodules. some yellow green staining ½cm diffuse patches
2mm - 10mm dark grey intraclasts. dark clasts, 2mm - 4cm. resistant intraclasts weathering out of clasts themselves. some clasts up to 4-5mm

mid grey

intraclasts mid grey, 1cm rounded set in a paler beige-grey matrix.
dark clasts - dark grey micrite ½-2cm. rare? ostracod shells.

remarks

Cubillo - Cubiejo (1) August 10, 1704

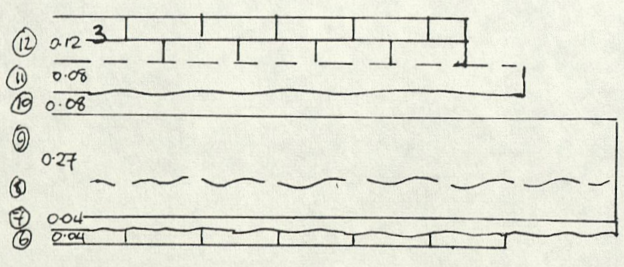
8 Tierra de Lava Group - entire thickness of 3m

7

6

5

4



white

pink - brown

yellow beige locally pink

white, pink rich red

pink to red mottling, 1mm qz pebbles, 1mm qz grains subangular. floating

calcareously cemented · qz pebbles 1mm, rare, floating

no exposure

clasts generally 1-3mm, rarely up to 1cm golden brown or pink-stained micrite. clasts rounded qz floating in very impure calcareous matrix. fine sand grade.

8cm clast golden brown microspar limestone clast, rare ½-1mm qz, mottled, spary. poorly sorted.

clasts 1-5mm, mostly orange white + pink subrounded qz, also rarer golden brown 8mm lst, rare 2-4mm grey micrite micrite matrix, 1mm pink spar, recrystallised spary limestone, orange-brown weathering.

2

5 1·04

T

*

no exposure,

loose golden brown limestone fragments as (3).

local equivalent heavily mottled pink orange, impure limestone, micritic, including terebratulids 5cm and ½mm spar crystals in matrix.

some purple mottles 1-2mm, rich purple, well-defined.

4 0·27

golden brown

T

mottled, locally pinkish, impure, 1mm qz pebbles.

2cm terebratulid.

1

0·10?

R

no exposure, blocks of orange-brown limestone as (1), (2),

fossils include rhynconellid shell fragments and whole same, up to 5cm, strongly ribbed.

rock probably similar to (1), (2).

3 0·89

2 0·05

1 0·18

0

purple-brown

golden brown

*

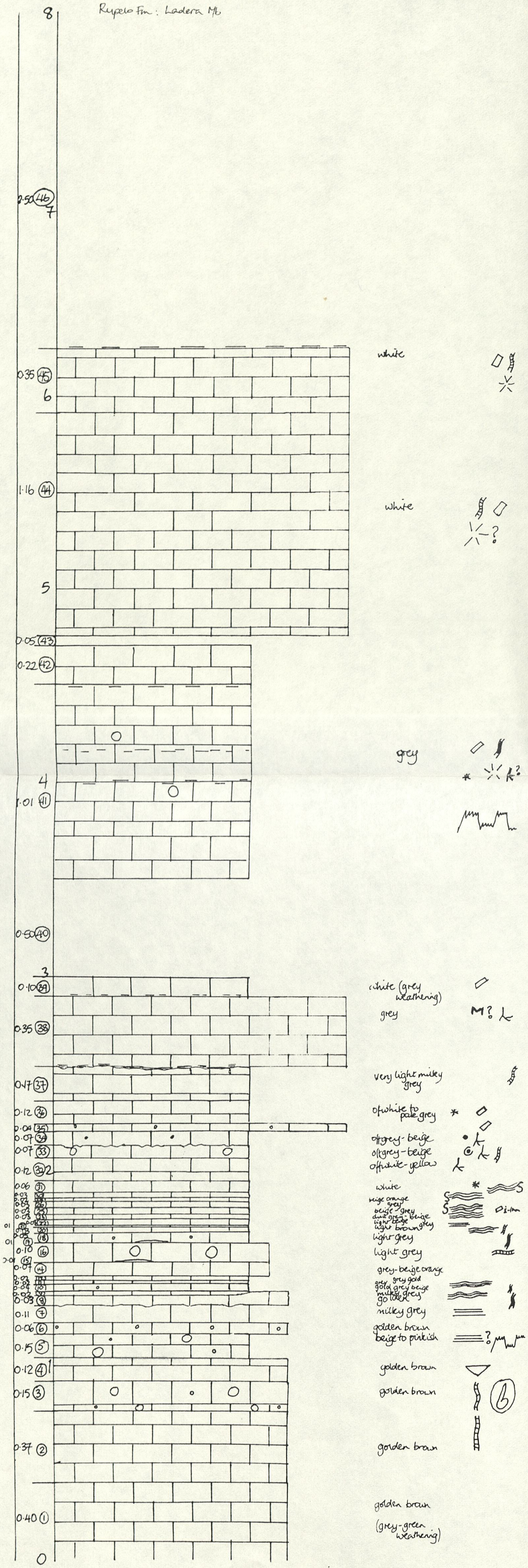
mottled, impure, gradational base.

local purple staining. Irregular (? diagenetic) base to bed.

no metres | LITHOLOGY | clay + silt | f | m | c | grit | cgl | gravel | colour | samples, structures | remarks

Mambrillas de Lara (1), April 19th, 1984.

Rupelo Fm: Ladera Mb



no exposure

patchy pink purple mottling on top surface.
breccia. subweathered surface rarely shows 1-2cm angular clasts of white micrite set in a network of spary cracks.

1st breccia. white micrite "clasts" 1-2cm very irregular, broken up by ½-1mm spary veins numerous complex patterns — desiccation?

no exposure.
top surface locally yellow-pink mottled, poorly brecciated.

nodules locally, 1-2cm
spar filled cracks ½-2mm, complexly bifurcating
locally heavy pink mottling at 50 and 68cm, 98cm

? charophytes or crystallaria.

no exposure.

marly fills, local purple staining on depressions.

clasts 2mm rounded - 50mm subangular, sometimes red tint. clasts rarely up to 100mm.
all clasts white micrite. some SST clasts on surface, red with grey carbonate ? Microcodium.
depressions in surface filled with marly mauve-purple lst.
cgl patchy - some areas appear massive.

patchy pink to purple mottling, in marly more impure areas with 1-2mm spar filled cracks

crystallaria ½×3-4mm wispy, discontinuous + associated with poor orange staining. cgl texture-rare 2mm "clasts" of bu/t micrite.
qz, rounded 5mm-20mm angular irregular lst clasts. local pink mottling - karst cgl ?

some bifurcate downwards, but no clear vertical orientation.

rarely orange hiting. crystallaria 4mm.

(27) ½cm light grey off white micrite (28) mottled (29) 1cm wavy laminated milky beige micrite
(24) 2mm crack, horizontal (a). micrite (b) sparite fills
(22) ½cm nodules/clasts of dull brown micrite
(19) ½cm grey micrite
(17) 1cm grey micrite lens, discontinuous nodules 1-2cm, subangular
(15) 1cm lens grey micrite continuous over 15cm
abundant 1 to 5mm thin and discontinuous lenses grey micrite
locally purplish mottled (12) 1cm lens as (12).

mottled - alternating bands of grey then purple mottled micrite

mottled, orange stylolites

shell fragments

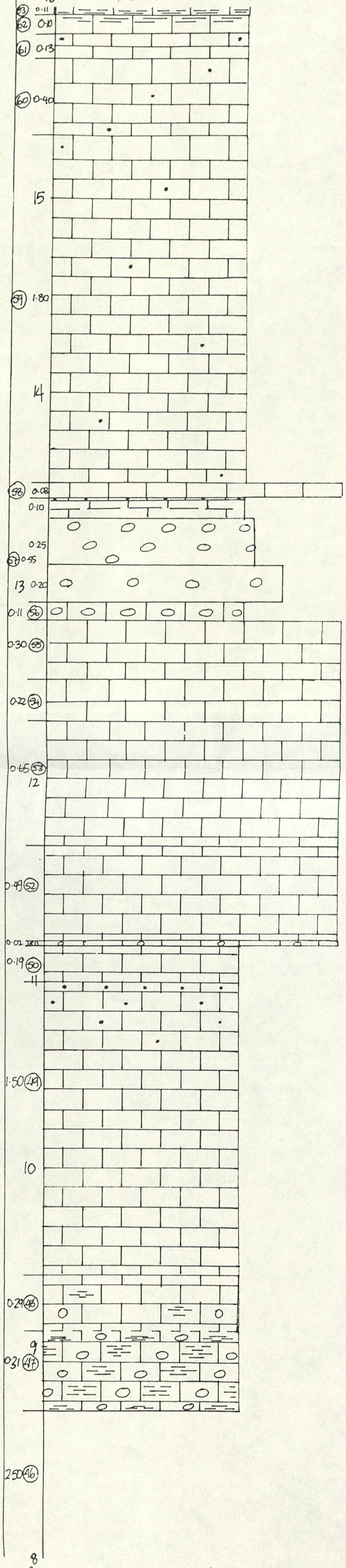
individual rhombic crystals visible ½mm.

no metres	LITHOLOGY	clay+ silt	f	m	c	gr	gk	gravel	colour	samples, structures.	remarks

Mambrillas de Lara (2), April 19th, 1984.

16

Rupelo Fm : Hodrea Mb

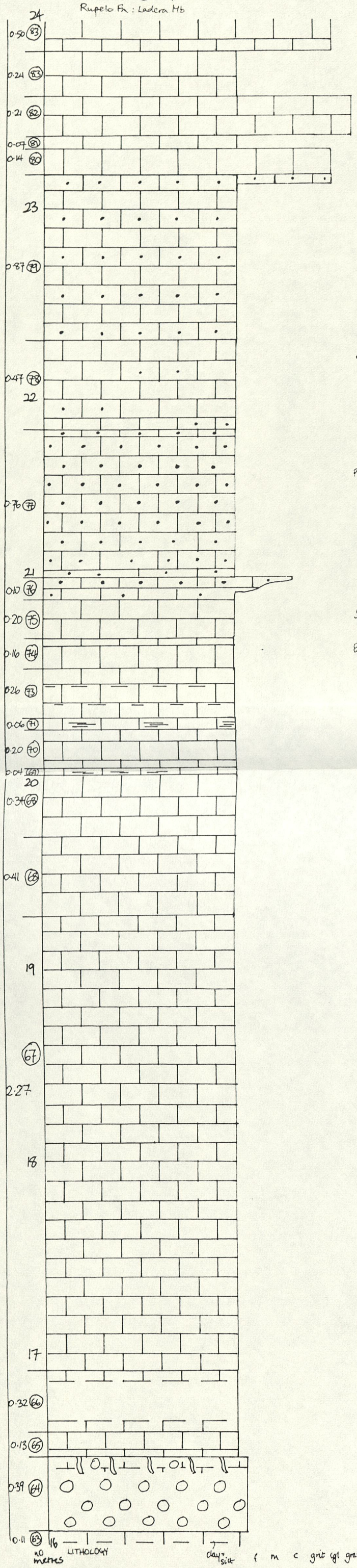


red pink-grey grey host, red staining		continuation of (62). mottled, brecciated 0-6cm mottled. strongly brecciated 4-10cm. } grey angular clasts 1-2cm set in red marly matrix
white		as (59)
white		as (59). rare mottling.
white		massive. occasional yellow weathering no vertical change.
grey		3-4cm irregular clasts of grey micritic lst, locally showing pelvidal fabric and 1-2mm darker clasts carbonate becoming more and more abundant. nodules more abundant up to 2cm, some angular - may be reworked very nodular fabric, broken up into 1-2cm nodules of dark red silt sand with carbonate rim surrounding clasts horizontally elongate void.
red	?M	nodules rare, 1cm.
grey		transition zone - purple mottled. 1-3mm rounded red to white/pink nodules in a complex network of ½mm bifurcating cracks filled with fine spar.
grey		weathered surfaces appear brecciated, internal surfaces less clear.
white		similar to (53)
white		"clasts" bigger - up to 10cm, otherwise similar to (52)
white	?	massive. weathered surfaces appear broken up into irregular 1-3cm "clasts" of white micrite separated by brecciated zones of angular small white micrite intraclasts, intervening complexly bifurcating 1mm wide spar-filled cracks.
purple pink		mottled, brecciated, discontinuous.
milky grey		mottled strongly. some dark 1mm red rounded grains. sequential fills in cracks white micrite coating to purple 1cm subangular "clasts"
		top 10cm one 3mm pelaid and circumgranular crack
grey weathering white		mottling, pelvidal structure + spar-filled cracks more abundant to top
		local strong mottling 1-2cm patches of purple red-brown.
		local partial brecciation.
milky grey		more massive, a few 1cm rounded clasts at base otherwise slightly mottled micrite, 1cm irregular orange stained areas and more diffuse purple-stained patches.
pink purple		yellow greenish mottled. friable. ? palaeosol. irregular clasts/nodules micritic, powdery chalky marly matrix.
		no exposure

no metres	LITHOLOGY	clay+ silt	f	m	c	gr	gl	grain colour	samples, structures	remarks

Mambrillas de Lara (3), April 19th, April 20th, 1984

Rupelo Fm : Ladera Mb



pink		mottled, red silty clasts and poorly conglomeratic texture.
grey		
white		rubbly lst. local only, spar matrix 1-8cm angular clasts. otherwise white micrite, associated desiccation cracks + mottling.
white	** ?	local pink mottling patches. abundant very fine spar crystals + ½mm areas, associated with fine spar-filled cracks micritic but locally brecciated and conglomeratic appearance. mottled pink
grey weathering white		top surface conglomeratic appearance – poorly developed exposure/soil surface ? red clasts 1-2cm down to 2mm, more abundant to top, subrounded ?reworked. some-red silt
		2 reworked clasts clasts dark grey micrite ovoid, subrounded at t5cm. others - mottled lst-?
greyweathering		numerous smaller ones 1-2mm
		large spar veins rare - numerous small cracks associated with peloids
massive		alternation of 10cm massive lst, 10cm peloidal, fine cracks – desiccation?
white	*	rare ½cm dark clasts. top 2cm laminar.
		locally appears mottled, brecciated, 1-2cm pure white micrite in spary matrix with rounded peloids 2mm.
		some spar cracks show weak orange staining
pink/white		darker pink clasts/grains 1-2cm set in a white micrite matrix.
		peloidal, very clear circumgranular cracks.
	*	top 2cm peloidal fabric - sparry, peloids 2-3mm irregular 4-8cm abundant 1-2mm spar areas set in pure white micrite, <½mm rare peloids. weakly laminar at 7-8cm
grey pink		as (74)
grey-pink		mottled. abundant spar filled cracks 1mm.
	?	coarser spar-filled cracks subhorizontal, complexly bifurcating and rarely joining up with circumgranular cracks. peloids at margin + rarely included within cracks themselves. black clasts 1-4mm - various shades of grey.
red-grey	M?	complex mottling. dark clasts – rarely up to 4cm. other dark grey clasts smaller, irregular, possibly resorbed/dissolved margins.
pink purple to buff beige		strongly mottled – buff beige colour often following 1mm crack-like paths.
pink red, grey		mottled - grey matrix. grey clasts 1cm rounded. rare discrete bands 1cm richer in red silty material.
		poorly exposed.
grey		top 30cm becoming increasingly mottled towards top. rare ½cm dark red clasts within central parts of red mottled zones (2-5cm). also rare ½cm dark grey clasts.
		conglomeratic weathered appearance – white "clasts", varying proportions of spar cracks.
	(1m/4)	
grey		
		red clasts present in basal 30cm in decreasing proportions. silt gradational base and top from grey micrite some? concentric structure of silt clasts – weathered rims?
red		
pink/grey	412	mottled. angular ½-2mm dark red grains – silt from below. rare black clasts up to 1mm. fewer nodules, more matrix carbonate, often piped down into silt on subvertical tubes. karst or roots?
red		carbonate clasts 1-3cm subrounded, irregular, horizontally elongated.

no metres | LITHOLOGY | clay+silt | f | m | c | grit | gl | gravel | colour | samples, structures | remarks

Mambrillas de Lara W, April 20

32 Rupelo Fm: Ladera Mb

31

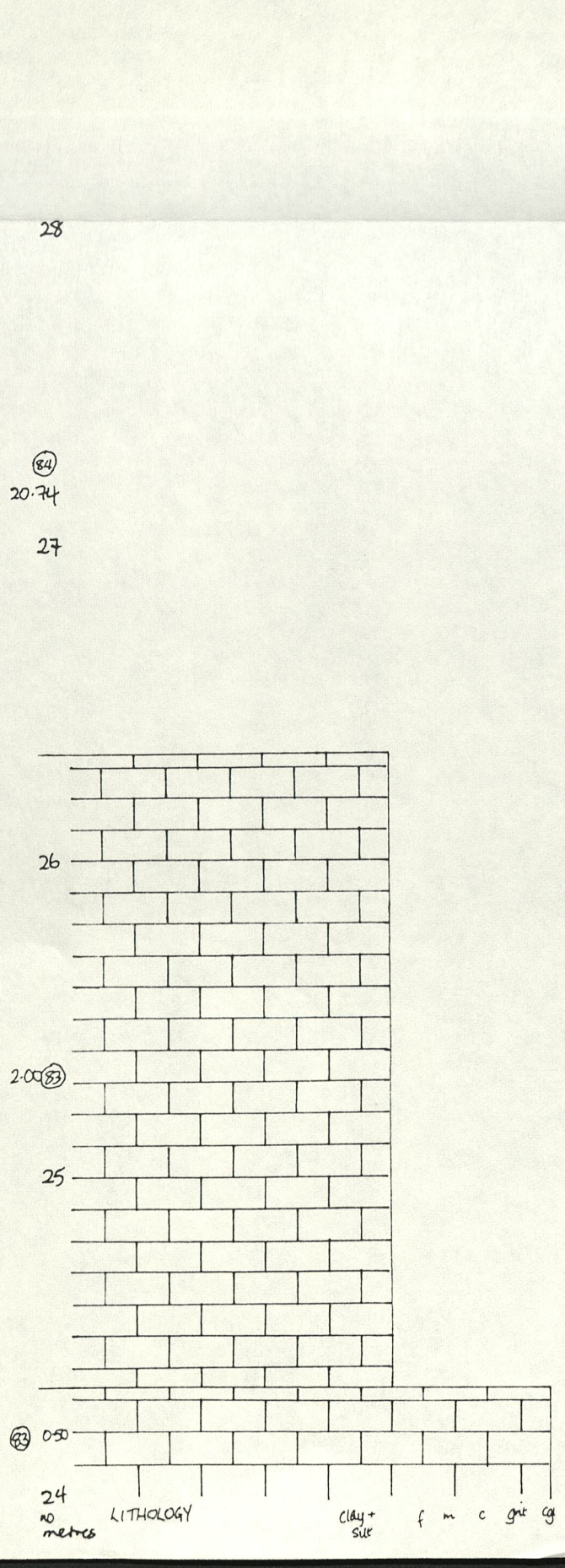
30

29

28

(84)
20.74

27



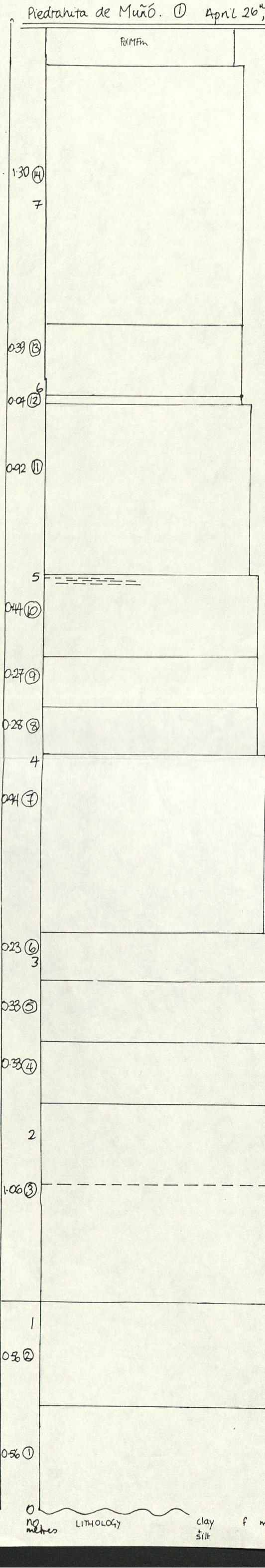
no exposure - red/brown soil, possibly red purple marl underneath

grey — massive. pink mottling very rare, locally present at top.

pink — mottled, red silty clasts, mottling, conglomeratic texture, matling + brecciation greatest at top.

no metres	LITHOLOGY	Clay + silt	f	m	c	grit	cgl	gravel	colour	samples, structures	remarks

Piedrahita de Muñó. ① April 26th, 1983.



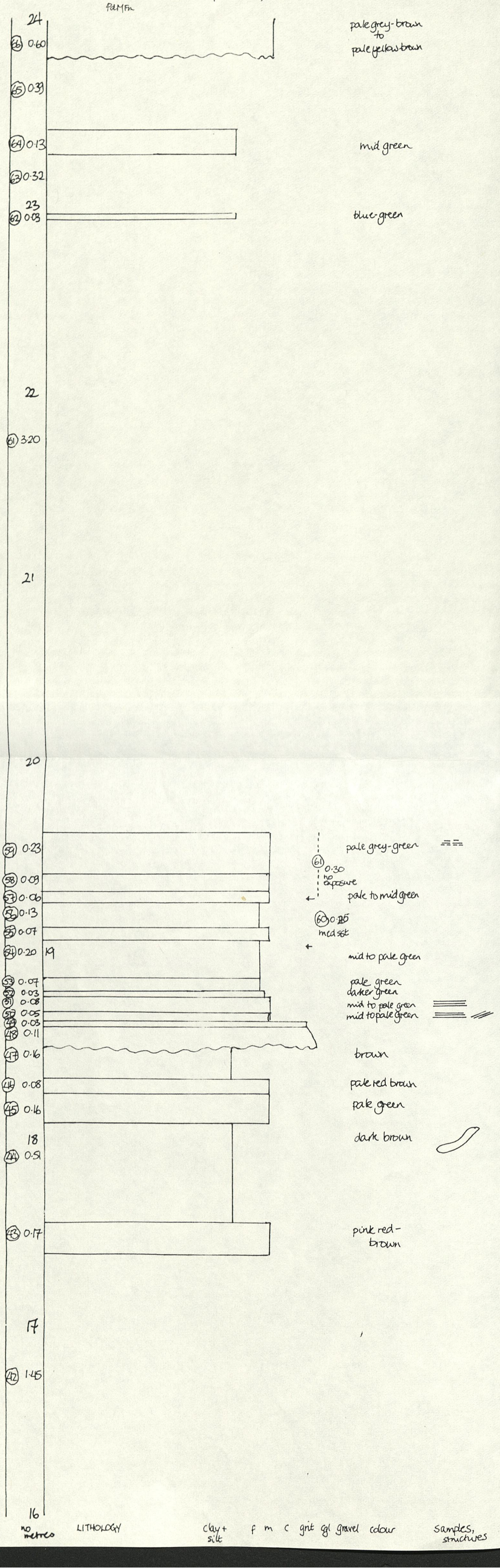
		colour	samples, structures	remarks
		brick red		patchy mottling - pale grey. a few white mica grains.
		white - pale pink		massive or weakly developed 10cm bedding
				micaceous
				½-1mm black grains common throughout.
		pale pink - purple		colour lamination in shades of grey at base, black grains 1-2mm 1mm plates of white mica. 2-10cm bedding.
		purple - pink		well cemented. ? ferruginous
		pale pink -		grains rounded
		pale purple - mauve		weak lam colour banding : purple - green.
		white to pink		more massive, slightly friable
		white - pink		qzose, micaceous. colour banding defines lamination 1-4mm bedding. mica common on bedding planes.
		white - faint grey		bedding 3-10cm
		white		qzose. 1-2% ½mm black grains weakly developed 5-10cm bedding
		white		better cemented. qzose. well sorted. Few lithoclasts
		white		similar to ① but 2-3mm green lithoclasts (subrounded)
				similar to ① weakly developed 5-10cm bedding
				massive
				weakly developed bedding plane grains subangular 1mm orange stained flecks
				massive
				moderately friable ½mm orange staining around black grains
		white	262, 263	qzose. well sorted. subangular grains muscovite. black grains 2% lithoclasts 5mm yellow micrite

Piedrahita de Muñó ② April 26th/27th, 1983.

16				
1·45	(43)	Pam Tm		
0·19	(41)		brown	No exposure. ? muds + silts, local 10cm calcareous horizons below 1m 30cm
0·11	(40)			similar to (39) but a little less resistant/nodular. rare included mica
0·20	(39)		red brown to yellow	more poorly exposed. fibrous spary calcite bedding parallel 2mm thick. mottled. subresistant. numerous 1cm blue-grey nodules, more calcareous up.
			rich brown to dark green blue	poorly exposed. mottled. Abundant ¼mm white mica in soil.
0·55	(38)			
15				
0·33	(37)		yellow brown - pink brown	bedding parallel 4cm scale colour banding. abundant 1/16 - ½ mm blocky black grains, a few more rounded pinkish brown.
0·13	(36)		pink grey - brown	½mm brown - pale green grains - weathered mafics?
0·42	(35)		white	qzose, locally pale green. numerous ½mm cubic green grains. black grains ¼mm 1%, similar red-brown
14				
0·09	(34)		rich dark brown	poorly exposed
0·03	(33)		yellow brown	resistant, hard, qzose, cemented
0·14	(32)		rich dark brown	local 1cm irregular green mottles. fissile - leaves 1-2mm. micaceous
0·09	(31)		brown-grey	resistant, similar composition to (30)
0·12	(30)		dark brown red	micaceous, poorly exposed. grains subangular. 1mm mica on bedding planes
			offwhite to pale grey-green	abundant 1mm green grains. burrows empty
0·60	(29)			
13				dark filled 8cm long burrows
0·32	(28)		mauve-green alternating	micaceous. fissile - leaves 2 to 10mm.
0·18	(27)		pink-green	very micaceous
0·48	(27)			very rich in 1-2mm dark olive green grains ?px/hbl. massive
12				
0·13	(26)			poorly exposed, micaceous, fissile. lamination folded by 11cm normal microfault
0·09	(25)		dark red white-mauve	very micaceous - abundant ½-1mm white mica, green/brown laminations. colouring reflects lamination
	(24)			
0·56	(24)			
			white-green	
0·20	(23)		green-white	friable, better cemented. Laminated basal 2cm, rest 1-2cm bedded
11				
0·16	(22)		white	includes black grains, limonite yellow grains and reflective muscovite.
0·55	(21)			fairly massive. weakly developed bedding on 20cm scale qzose. grains subangular. ½-1% subround orange grains. 2-3% subangular black grains. some muscovite.
0·22	(20)		grey to pale green	faintly mottled, hard. rare 1mm white mica included
10			grey brick red to green	mottled, nodular as (18)
0·60	(19)			scattered ¼mm white mica
0·52	(18)		dark brick red	poorly exposed. nodules 8-12cm, rounded at 32cm, light brown to grey brown. local motting blue-green
9				
0·42	(17)			2-15mm pebbles 5-10% at base, decreasing up. Lithoclasts, green 2×10mm // bedding.
0·28	(16)		purple brown - blue grey	mottled. offset 30-40cm by fault plane dipping 75° to N. (normal)
0·48	(15)			
8				

no metres LITHOLOGY | clay + silt | f | m | c | grit | cgl | gravel | colour | samples, structures | remarks

Piedrahita de Muñó (3). April 21st, 1985.



well cemented, qzose, grains subangular to angular, interlocking.
numerous ¼mm black angular grains 1-2%, subround orange-brown grains ½% similar size

no exposure. dull dark brown earth.

Poor exposure but many 2mm - 1cm leaves, a few pebbles/nodules 2-5cm mid green cemented fine sst

no exposure

rich dark brown mud/soil

no exposure.

rich dark brown soil, golden sandstone pebbles.

white mica flakes. a few golden brown ¼mm fragments.

moderately exposed. Numerous ⅛ - ½mm pale orange - brown grains. white mica flakes + ¼mm red patches.
less well exposed
2x3mm subrectangular black woody fragments. 10% golden yellow subangular ¼ - ½ mm grains
slightly resistant. grains subangular to subround, cemented. qzose but 1% black, 3-4% pale brown ¼mm grains.
rare 3x10mm black resinous lustre grains ? organic, aligned parallel to bedding.

numerous yellowish-very pale brown ½mm grains.
more poorly exposed, weathered
common ¼mm black grains, resinous lustre. lamination, colour banding + fissility all bedding parallel.
micaceous. 5-10% dull yellow weathered clasts ½ - 1½ mm. colour banding associated with lamination
similar to 43, weathered, friable, green clasts predominating, some yellow, few red
equal proportions of yellow, blue green and deep rich red rock fragments. clasts angular, equant
green mottled. bottom 1-2cm poorly exposed, fissile, remainder very poorly exposed.

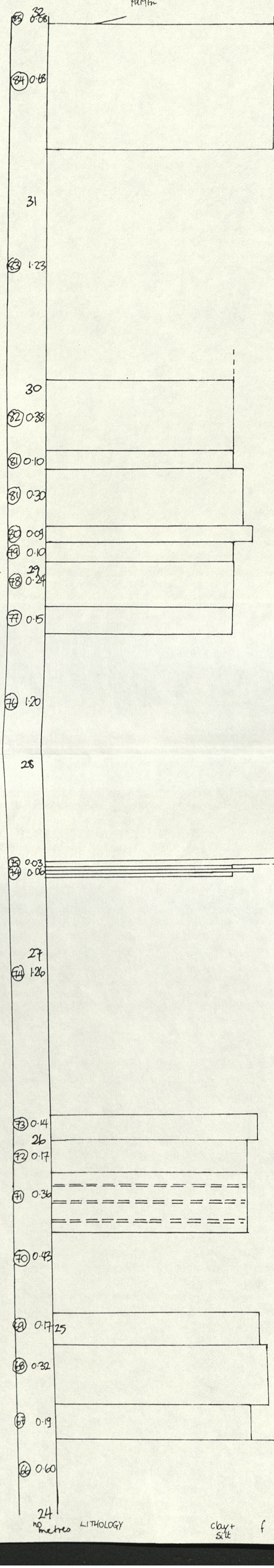
slightly less resistant. greenish blue mottles.
slightly friable. grains subangular. ¼ - ½mm black to dark brown and dark green grains 5%
mica in ½ - 2mm plates, very concentrated on bedding planes.
mud, poorly exposed. patches of rich brown fine micaceous sst at top.
a few ½mm rounded white qz.
local dull mid-green colouring, bedding parallel stripes. Top 15cm has burrows,
filled with fine brown red sst.

qzose, white mica. white sparry calcite veins ⊥ to bedding.
rare blue-grey patches. numerous cleaved transparent grains. some ½mm blocky white calcite cement.
no calcareous horizons top 15cm.

no exposure. muds + silts? local 10cm calcareous horizons below 1m 30cm

no metres | LITHOLOGY | clay + silt | f | m | c | grit | gl | gravel | colour | samples, structures | remarks

Piedrahita de Muño (4). April 27th/30; 1983.



similar. yellowish. only locally developed as separate bed.

qtose. white mica and pink-black grains present. grains generally 1mm. massive

no exposure

mid to light brown soil rich in pink to grey brown sst fragments of (84)

fissile. very poorly exposed

micaceous, ¼-½mm white mica, other grains not visible. 2-3mm bedding parallel strong fissility.

mottled pale mid green. ¼mm white mica, ¼mm cuboid very dark green grains. more fissile at top (½cm) than base (2cm)

½mm mica plates, other grains ¼mm
fissile 5mm - 1cm

slightly more resistant. ½-1mm mica. 1-2mm hard black crusts. Fissility bends around joints in sst above.

poorly exposed. micaceous. 2mm - 1cm fissile.

no exposure. mid brown soil.

abundant grey-brown sst fragments.

1cm

yellow-green and pink brown 1-2mm clasts set in friable med micaceous sst. matrix. alternation fissile silt/shale and micaceous fine sandstone.

poor exposure. dull brown earth.

slightly more resistant, but still poorly exposed. ½-1mm mica flakes. other grains ¼mm.

cf interbeds of (71), less resistant, micaceous

2-3cm beds more fissile, darker brown, mica ¼-½mm plates

no exposure. dull mid to dark brown soil rich in finely micaceous silt/f. sand

micaceous. possibly grading up into silt as exposure deteriorates.

massive. white mica. ½mm black grains quite common.

poorly exposed, locally mid green to blue mottled, abundant ½mm flakes white mica.

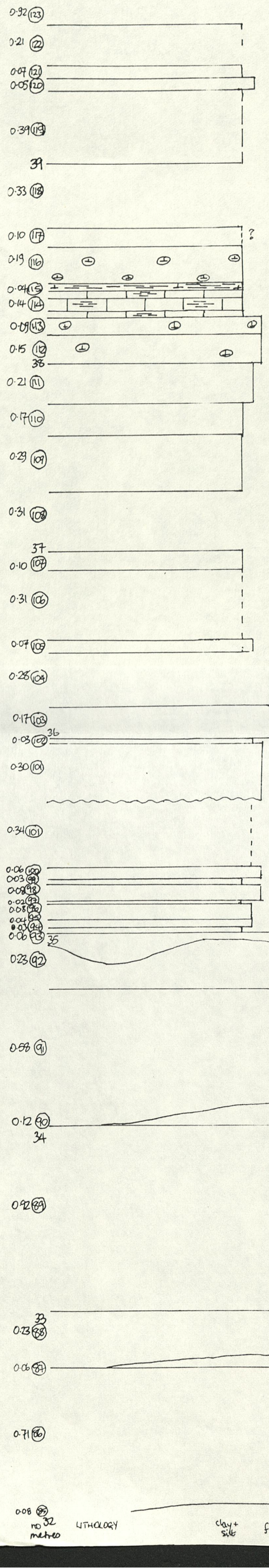
well cemented, qtose, grains subangular to angular, interlocking.
numerous ¼mm black angular grains 1-2%, subround orange-brown grains ½% similar size

remarks

Piedrahita de Muño (3) April 2, 1985

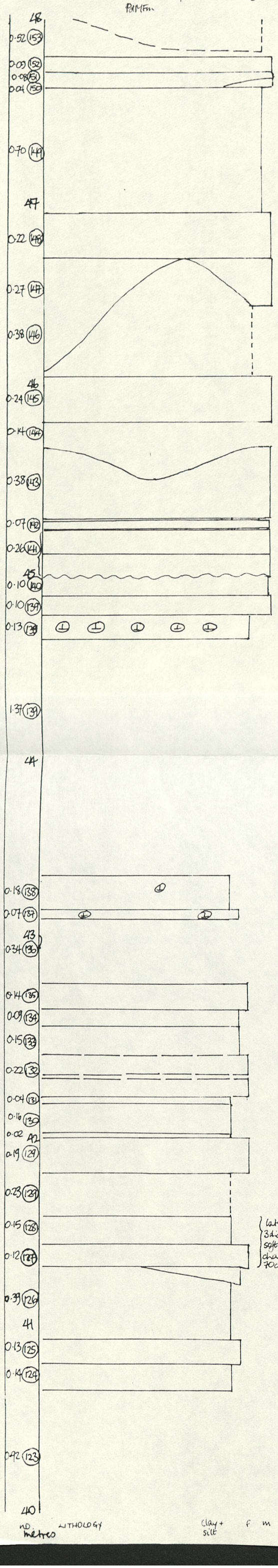
PdMTm

40



	green	no exposure. rich brown soil only.
	brown to khaki green	2-10mm fissile shales at 10-13 cm, 18-21 cm. otherwise rich brown soil only.
	pale green to pale brown.	1-2mm fissile. 1-2mm oval dull grey patches ? ostracods. V. rich in white mica on bedding planes (½-1mm).
	?	poorly exposed alternations of sdst, green + rich brown shales. 1-2mm fissile. clay rich. white mica.
		poor exposure. grey green (locally brown mottled) soil.
	v. dk grey green?	Poor exposure. greenish grey muddy soil.
	light to mid green	rare ¼mm pale orange qz grains. occasional 2-5mm nodules, opaque, grey, calcareous.
	grey-green	similar to (114), but poorly exposed, very marly. hard impure limestone. scratches steel (just).
	pale pink to green	similar to (112). nodules ½-1cm, very rare.
	pale pink to pale grey brown	micaceous. rare 1cm oval nodules
	mid to purple brown	mid green mottling. weak 3cm to moderate 5mm fissility. ¼mm white mica. other grains barely visible. locally more massive
	mid to light brown	as (107)
	mid grey	green colour patches as (107)
		no exposure. dark brown to purple soils.
	mid to light brown	slightly more resistant. a few faint green patches, 1-2mm. discontinuous.
	mid to dark brown	poorly exposed shales. ¼mm white mica.
	mid to dull green	1mm white mica
		poor exposure. mid to dark brown soil, abundant white mica.
	purple pink to brown	green patches 1-2cm. ¼mm white mica.
	mid brown	less resistant. abundant mica. two shaley partings separated by 2½-3cm v. fine sandstone. micaceous
		poor exposure. ½-¼mm white micaceous sst exposed locally. fissility 1mm at top, 1cm at base
	mid to dk brown	1-2mm white mica abundant on bedding planes. occasional patchy mid green colouring.
	mid brown to mid green	1mm white mica plates
	mid to dk brown	1mm white mica, not resistant.
	dark brown	very weathered. clay rich muddy shale
	mid to dark br.	more resistant, very micaceous - abundant ½mm white mica on bedding planes
	dark brown	½-1mm white mica. ¼mm qz
	Br ?	½mm white mica channelled base 14cm relief. Top surface irregular, very micaceous, possibly weakly rippled.
	rich pink / grey	well cemented, similar to (91), colour banding parallel to bedding
	rich pink	a little white mica in ½-1mm plates. many pink-red brown grains.
	-dip 5 to 10° NW	approx 6cm bedding parallel colour banding - pink to pale yellow brown.
	rich pink	cemented. qzose, some white mica, only locally present as separate bed.
	pale pink	white mica. mid grey-brown weathering.
	pale pink	similar to (84).
	pinkish white	only locally developed as separate bed. Similar to (84), (85), (86).
	off white	similar. white mica. rare 1mm dark grains, a few pinkish brown grains.
		similar, yellowish, only locally developed as separate bed.

Piedrahita de Muñó (6), April 30th, May 1st, May 3rd; 1983



rich golden brown
mid brown
gold-brown

golden brown

white-yellow

white

beige brown

pale yellow-brown

yellow

pale yellow-brown to dark grey
pale yellow-brown to dark grey

pale brown-yellow
green-grey

mid green

dark brown

mid green to dark rusty brown
dark rusty brown
dark brown

dark brown
dark brown

dark rusty brown

rich brown

lateral equivalent 3 distinct 6-8cm soft sst beds channel 70cm to 0.
dark purple-brown

rich brown

mid green to rich brown
rich brown

no exposure. mid brown soil, v rich in golden sst.

grose. slightly friable. white mica ½mm plates on bedding planes. occasional dark orange staining grains subangular.
rich ochre orange grains, oval 1mm "muddy".
lens. qz pebbles, white to grey, 5-10mm 5-10%.

grose, cemented, grains subangular. white mica ½-¼mm.
2-6mm cracked white qz, subrounded + voids.
a few ¼mm round to subround black grains. Irregular (weathered) top surface.

more massive, white mica. a few ¼-½mm white mica.

heavily channelled massive med sst., several 5mm qz, numerous 1mm fine orange lined voids.
a few more angular voids, black lined 1mm. ? ostracods?

poor exposure, blueish beige muddy soil, numerous yellow-blue brown sst fragments.
rock ? fine beige brown sst, ¼mm white mica, qz ½mm. friable

finely mottled. ½-1mm white mica present, qz, very rare pale green grey grains.

poor exposure. 1cm sst fragments. muddy blue-beige soil.

white mica 1mm plates, abundant on bedding planes

variable thickness 5-7cm. 1cm soft shaley partings above + below.
similar to (140), a little white mica in ½mm plates. a few v. rare pale green ½mm grains. a few dark orange grains, 1% black grains.

grose, grains ½-1mm, subangular to angular. a few ≤¼mm black grains some yellow-orange stained qz.
burrow/nodule 1cm
hard sst, local, 10×3mm nodules/clasts of hard grey lst

no exposure.

light brown, rich brown, yellow blue, beige, marly soils.

rare patchy mottling (mid brown). One included 8cm clast/nodule cemented hard green siltstone. angular.

locally nodular, hard. some mid green patches — including ½mm green to black crystals.
poor exposure, rich dark brown soil, abundant dark brown and some pale green silt/shale fragments

mottled, micaceous
as (133), ½mm white mica.
resistant, local pale green mottling. ¼mm white mica.

poor exposure rich brown soils. lateral equivalent 2×10cm hard sst beds, thin shale interbeds

as (130), patchy 1-2mm green colouring.
2-4mm poor fissility

1-2cm soft shale parting locally
patchy green mottling. ½mm plates white mica. possibly laterally discontinuous.

poorly exposed. patchy 1-2cm green mottling. 1-2mm fissile.

local mid green mottling. 2-3mm fissile. numerous 3-4mm diameter, 3-4cm long vertical burrows. extend into top 3-4cm of bed below.
mildly channellised, hard. ¼mm white mica present.

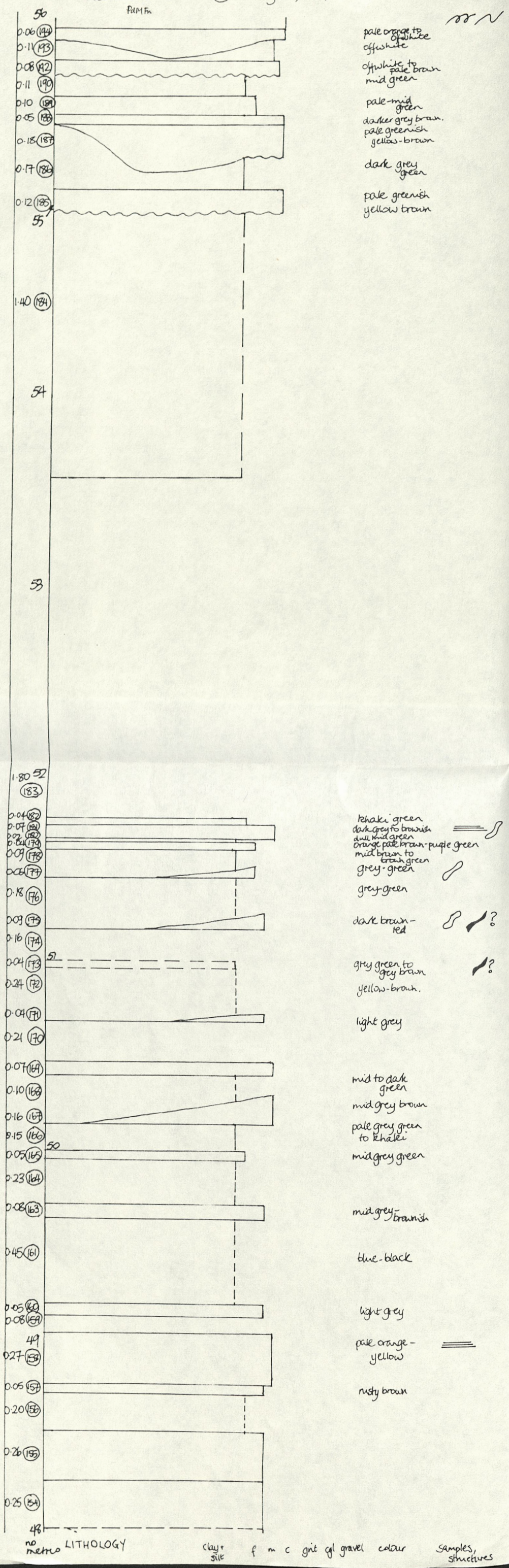
occasional green mottling. 10cm lens light to dark brown fine sst/coarse siltstone. ¼-½mm mica.

mottled. ½mm white mica.
¼mm white mica visible. 2-3mm fissile.

no exposure, rich brown soil only.

no. metres | LITHOLOGY | clay + silt | f | m | c | grit | gg | gravel colour | structures, sample | remarks.

Piedrahita de Muña (7). May 3rd, 5th, 1983.



orange stained grains 10-15%. Black frags 1-2mm, rarely 5-6mm, irregular. ¼mm white mica. Fairly well sorted Qz subangular to ang slightly friable. qz grains subangular ¼-½mm. white mica ¼-½mm. organic flakes 2-15mm on bedding planes.

cemented. goose, but orange clayey material (weathering of fst or mafics?) 5%. a few black blocky. ¼-½mm grains fissile, organic black grains especially abundant up to 1cm, nearly always parallel to bedding orange staining 5%

similar to (186). white mica ½-1mm plates. a few subrounded black ?mafics ¼mm, black organic flak orange staining 30%.

similar to (187). white mica. local very strong orange-brown rusty weathering around dark nucleus.

similar to (185), but top especially a little richer in organic material including one 30×2mm flake.

a few ½mm angular pale-flesh qz at base which is less organic rich

very rich in black flakes 2-10mm, greasy lustre. Larger flakes fissile ? black shale. (30-40%) others may be woody.

goose. friable. abundant 1mm white mica flakes.
numerous black elongate or subrectangular grains/flakes 2-8mm. greasy/resinous lustre - organic or woody? up to 1cm long, parallel to bedding.
orange brown 5-10mm staining around ? no organic larger black grains.
patchy 3-4cm larger scale diffuse orange staining.

no exposure.

greyish brown to greenish grey brown soil.

clays underneath. ~40cm bedding parallel stripes
alternating rich red brown / grey green clays.

no exposure.

rich brown to reddish brown soil.

black angular organic flakes 2-5mm, a little white mica.
slightly less well cemented, white mica in 1mm plates heavily concentrated on bedding planes.
shalely parting. white mica burrows 7-8mm diameter.
cemented very few 8mm white mica, more abundant ½mm blocky grains extend at least 4cm up from bed base
1-2mm fissile parallel to bedding. ¼mm white mica present.
lens. a little white mica in ¼-½ mm plates. Burrow 7mm diameter, one 1cm angular rip up clast. grey green
no exposure – rich mid brown to grey soil, fragments of burrowed sandstone. grey-green silt underneath

local lens, 31cm lateral extent. white mica ½-1mm plates, heavy concentration on bedding surfaces.
no exposure, pale brown to greyish soils. burrows poorly preserved 5mm, tubular subhorizontal to 45°.

local - 21cm lateral exposure.

no exposure, mid brown to grey soils. occasional yellow-brown silts exposed beneath.

local, only 15cm lateral extent. hard, cemented. a few ¼mm black grains.
mid brown to khaki green soil, loose fragments of dull grey-green fine white-mica sst..

fairly well cemented, similar to (167)
poor exposure. hard. white mica + some qz visible with a hand lens.

hard, cemented. qz pale grey to white (transparent), black blocky, pink to purple subround, ¼-½mm white mica plate
silts. 2mm fissile. poorish exposure. white mica ¼mm. a few dark green/black grains ¼-⅛mm

white mica ¼-½mm, rare ¼mm black flakes, v few ¼mm rounded pink grains.
poor exposure. mid to khaki green soil. ?mid grey-green to green brown shales underneath. white mica.

local? but at least 25cm lateral extent, hard, cemented. black grains ⅛mm, a few sglt orange ½m
rare transparent pink grains ¼mm. a little mi
no exposure. grey-brown, locally blue-black soil, golden sst. fragments. Soil soft.
?blue-black clay underneath.

bottom 1cm ?ferruginous, mid to darker brown, rusty, hard.
no exposure. mid grey-brown muddy soil.
well sorted, partially cemented. grey-white clear qz, ½mm white mica flakes. many sglt orange weathered grains. rare ¼mm black grains.
4-5cm thin dark horizons, poorly developed.

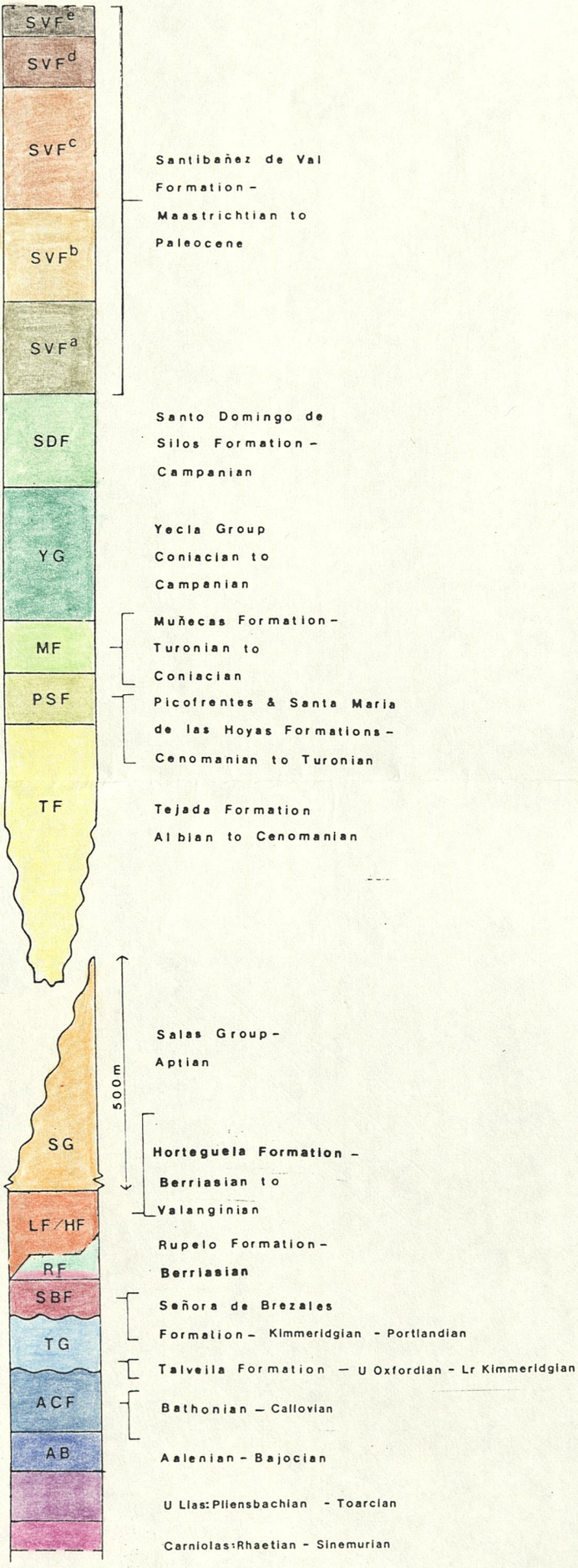
hard, crusty, cemented. qz, white mica.
poor exposure. exposed nearby as friable sst/siltstone (orange, ochre) with 1cm black sglt coaly, bituminous horizons.

similar to (154), black grains 10-20% throughout matrix. Often clustered.

Varying thickness, ½mm white mica present, many ¼mm black blocky grains. ⊥ to bedding joints 2-3cm

remarks

Stratigraphic Column



Piedrahita de Muño (8), May 5th. 1983.

PdMFm

64

63

219 4·20

no exposure,
rich brown to reddish, clay-rich soil
poor exposure 5m above road.

62

140 218

61

gap – no exposure.
brown soil, rarely greyish

60

217 — dark grey brown to pinkish purple — hard, cemented.
a little $\frac{1}{8}$-$\frac{1}{4}$ mm white mica, other grains just visible.
mottling, purple brown to mid green

0·12 216 — brownish grey — hard, cemented. rare $\frac{1}{4}$-$\frac{1}{2}$ mm white mica plates. occasional dark to mid green $\frac{1}{2}$ cm patches.

0·12 215 — greenish grey-brown — poor exposure. 2-3mm fissile.

0·11 214 — grey-beige — $\frac{1}{2}$mm white mica.

0·14 213 — green — weakly developed 2-3cm bedding parallel fissility. white mica $\frac{1}{4}$mm plates.

0·19 212 — pale greenish beige — very soft, 1cm bedding // fissility.

59

0·09 211 — mid green — soft, friable. one dull green 1cm stain/clast?

0·12 210 — as 208 but less friable.

0·27 209 — pale grey to offwhite — gpose. orange grains. white mica – a few $\frac{1}{2}$mm plates. $\frac{1}{4}$mm black grains. friable

0·25 208 — mid grey to patchy offwhite — more massive. qz subrounded. brown-orange grains $\frac{1}{2}$-1mm 10-15%, a little white mica, black grains, 1cm voids at top, grey-green fine grain

0·28 207 — white — friable. white mica. orange fragments $\frac{1}{2}$-1mm locally 40%, generally 5-10%. Black blocky grains rarely up to 1mm

0·04 206 — offwhite — orange grains as 204, voids elongate, oval 2cm long.

0·07 205 — white — gpose, qz $\frac{1}{2}$-$\frac{3}{4}$ mm, orange + black grains $\frac{3}{4}$-1mm. solution holes oval, elongate, parallel to bedding up to 1cm in length.

58

0·08 204 — very dark grey weathering

0·20 203 — similar to 201 but more friable. orange-green to black grains 20%. qz subangular, just less than 1mm. burrow 5mm diameter, 5cm long.

0·18 202 — mid grey to patchy green v. weathered — similar to 201

offwhite to pale grey — better cemented. orange/black grains up to 1mm 10%.

0·53 201 — locally medium-fine sand, slightly friable.

57

pale grey — or

pink — black grains scattered ayg.

200 — powdery pink — orange/black grains 30%

199 — offwhite to beige — alternations of 2cm fine cgl separated by 4-5cm coarse sand (cgls: 2-10mm irregular grey-granfissile clasts)

198 — pinkish red — (sands: 1-6mm greenish orange clasts, $\frac{1}{4}$mm black grains)

0·22 198

0·12 197 — pale brick to orange yellow — friable. grains $\frac{1}{2}$mm, subangular to angular. orange stained dull grains $\frac{1}{2}$-1mm, 3%, white mica, black grains

R λ=12cm a=3cm passym. flow ? to SW — offwhite to pale green/pink. — grains angular to subangular. moderately cemented/slightly friable. $\frac{1}{4}$-$\frac{1}{8}$mm, clustered. burrow 2cm, 45°

? R (base) — scattered $\frac{1}{4}$mm black grains, blocky, angular, orange weathered grains $\frac{1}{4}$-$\frac{1}{2}$mm also.

0·30 196

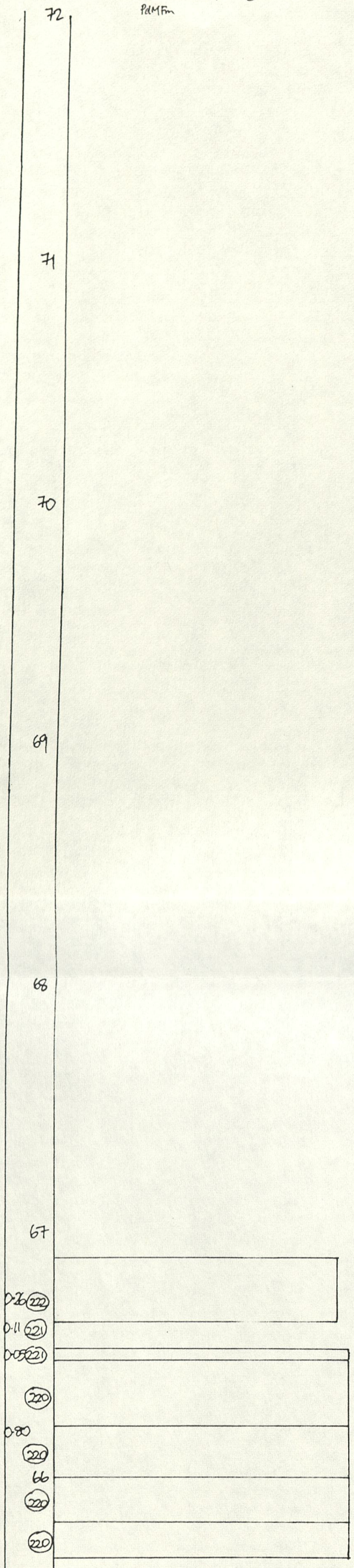
0·45 195 — offwhite to pale brick brown (weathered) — 2-3cm sets dip component to SW. — fairly massive. sugary white medium qz, sometimes cemented, other times friable. scattered $\frac{1}{4}$mm black blocky grains. fairly numerous orange $\frac{1}{4}$mm soft clayey orange grains.

56

no metres | LITHOLOGY | clay+silt | f | m | c | grit | cgl | gravel | colour | samples, structures | remarks

Piedrahita de Muño (9), May 5th, 1983.

PdM Fm



dark brown-grey (purple brown fresh)

rear brown to dull greenish grey

grey green

pale brown

pale brown

qtose, cemented. grains subangular. black grains.
rare 2-3mm voids - parallel traces low angle or parallel to bedding - qz fallen out of bed?

no exposure

5cm as (220)
cemented.

qtose. grains subangular and including scattered ½mm - 1mm black grains.
rare pinkish grains, some pale brown-green clear qz.
rare dull offwhite - weathered felspar.
a few dull orange to dark brown grains.
white mica present (not abundant) 1-2mm plates.

Weak 3-4mm lamination associated with 1cm colour banding (grey : green)
bedding parallel weakly developed fissility in 3rd bed (richer in black grains?).

no exposure,

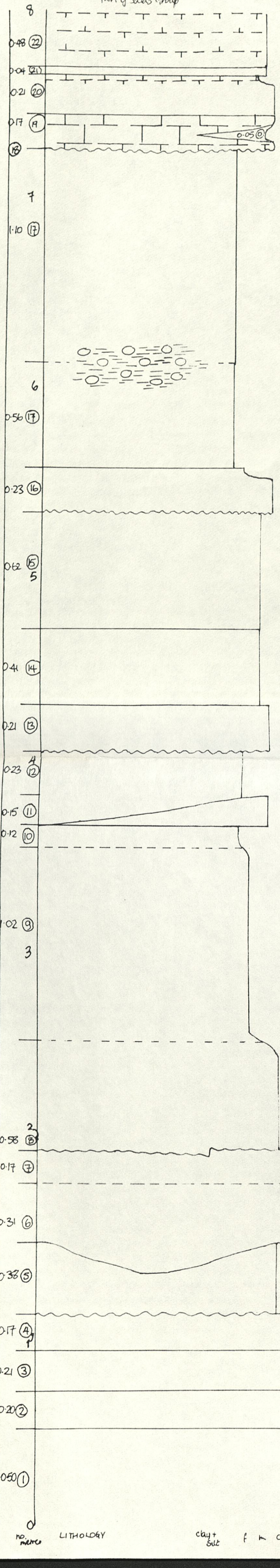
rich brown to reddish clay-rich soil.

poor exposure 5m above road.

no metres	LITHOLOGY	clay + silt	f	m	c	grit	gravel	cgl	colour	samples, structures	remarks

Cubillo del César (1). September 16th, 1983.

Part of Salas Group



white

dark brown - ochre
light grey
dull brown

pale green to yellow

rich dark purple brown

rich dark purple brown

rich lime green
very pale green to white
? base

purple

purple to grey green, white, occasionally orange

pale golden yellow-white

purple/green

white to very pale yellow brown
purple

lime green to rich purple

pale green

pale golden brown

large scale poorly defined cross-bedding.

pink-orange

light rusty brown

pink

pink.

pink

oncoids - white, occasionally fragmented in interstices,
- size 4 to 4mm. concentric. some detrital nuclei
others spar nuclei.

cemented, hard, calcareous. floating grains, occasional 1cm subrounded qz algal rims. reworked oncoids

local friable - originally organic rich - voids (or shale flakes)
top 4cm hard, cemented, fine, included 1cm oval clasts - ½ to 1cm oncoids fragmented on weathered surface.
rip-up clast - 7cm, irregular, dark sst. occasional 1mm rounded qz.
qzose - many round voids 1mm → originally carbonaceous material?

hard, cemented, calcareous. included qz pebbles round, irregular, cracked 1-4cm
5cm lens, reworked oncoids, ovoid or elongate? incomplete 1-2cm weakly oriented parallel to bedding

(18) gradational top of (17) - white or pale green calcareous ⇒ hard.
floating 1-4mm rounded qz, rare 1cm round qz pebbles.

local patchy grey-white mottling, white/green veining
diffuse 1cm scale cracking.

matrix - fine purple shale with 1mm x 5mm white branching rootlets.

(17) - palaeosoil profile?

calcareous, cemented, hard silty shale, cracked chalky marly mottling 19cm
occasionally lime green streaky mottling. white calcareous coating to shale blocks (1-2mm thick)
cracked pattern - 1 to 2cm scale, flakey/brecciated.
very rare 1cm qz pebbles (round)

? possibly brecciated top. top 3cm richer in mafics.
included qz clasts 1-3cm.
local concentration of black/green mafics.

pale lime green mottling.
fissile 1cm scale.
included qz clasts round, ½cm, plus ½-1mm rounded to subangular white qz clasts in matrix.

included qz clasts, med sand, grains, biotite 1%, angular to subangular.
qz clasts 1cm, rounded to subrounded, previously cracked, abrasion pitted, 5% locally more.
poorly exposed.

cemented, hard, qzose, well sorted subangular grains.
qz clasts included ½-3cm, cracked, subrounded, poorly sorted, various colours 5-10%.
locally fenuginous veins aligned NE-SW on bedding surface, ½-1% biotite (locally 10-15%).
occasionally purple mottled. blocky fracture.

mottled.
included qz clasts (occasionally irregular), 1-2cm. ? cracked followed by rounding. qz clasts <10%.
weathering purple to orange.

42cm lens. cemented, hard. pebbles - qz, 1 to 2·5cm, supported in sand matrix ? 10%, subrounded (angular when cracked)
included cracked qz clasts - round to subrounded, 1-3cm - common.
rounded, ½-2cm, randomly scattered.

poorly exposed
friable, well sorted. subrounded grains.
weathers → clay-rich

mottled - weathes purple-green orange to yellow.

very rare included qz clasts.

massive, friable, moderately cemented, subangular to subrounded grains, qzose, rare ½mm biotite.

Bed! 70° DIP to NNE.

clast supported, moderately sorted, qz pebbly. max clast size 3cm. abundant subrounded high sphericity qz, ½-2cm.
colour as before. matrix: poor to moderately cemented, med-coarse pink-orange sand. grains subang. to subround.
rare muscovite and ? biotite

qz pebbly conglomerate. qz pebbles rounded to subrounded, heavily cracked, clast supported.
matrix: cemented med sand.
qz clasts: grey green, dark grey, occasionally orange. also: dark green shale/phyllite.

qzose, grains subangular. 1% ¼-½mm mafic grains, friable, poorly cemented, locally fenuginous.

20-25cm: weakly concentrated pebbles - qz, 1-2cm rounded, subspherical to discoid.

low angle cross bedding basal 12cm. lowest 3cm includes rare 1-2cm qz pebbles from unit below.

qz pebbles, subangular, cracked, 2-4cm.
qzose matrix, coarse sand, including abundant ½cm rounded qz pebbles, matrix supported (just)
moderately cemented, rare 1mm mafic grains ? subangular to angular biotite.

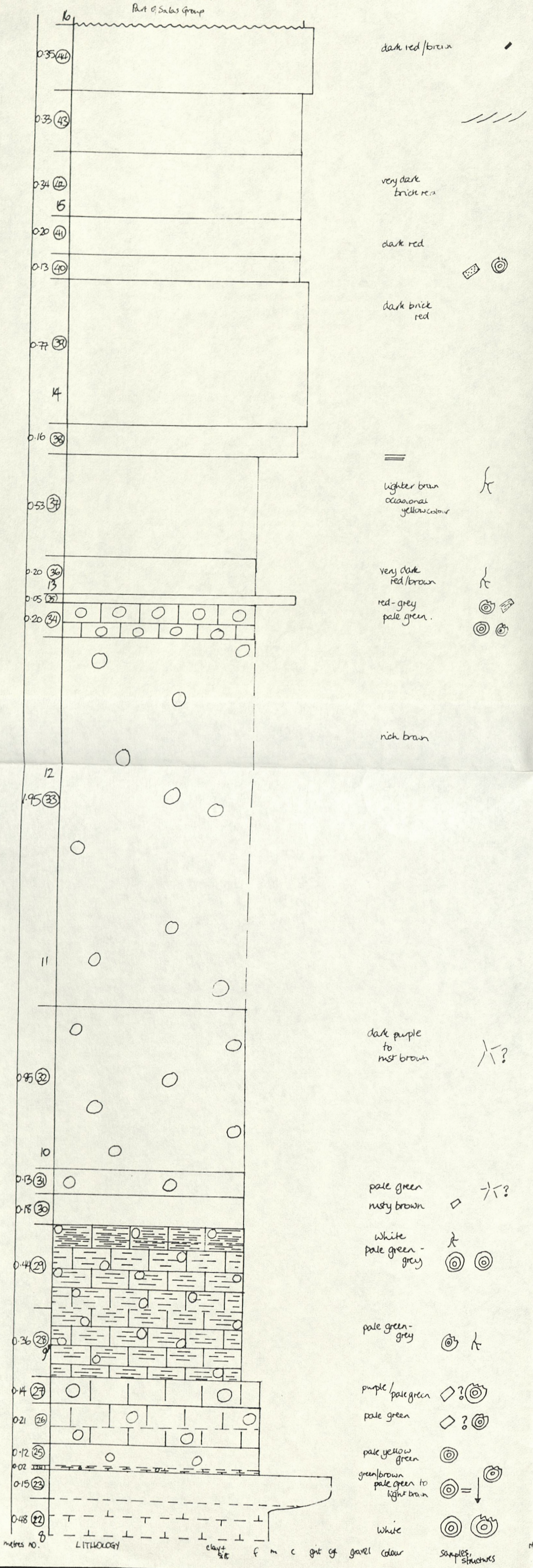
as (1), max clast size 10cm.

as (1) but rarely matrix supported, qz clasts 3-5cm, one with mica and cleavage (Cambrian metasediment?)

clast supported.
qz clasts cracked, ~5cm, rounded to subrounded, subspherical
: grey green, purple-red, occasionally orange.
no imbrication, poor grading, weak bedding only.
matrix: medium-coarse qzose sand, (2-5mm) qz shards abundant, some white mica.
other clasts: 1mm laminated/colour banded qz. angular
rare white med. friable sst..

Cubillo del César (2). September 16th, 1983

Part of Salas Group



porous, voids + void strings.
rare qz pebbles, black organic matter, some limonite staining and infils.
some larger, angular carbonate clasts (nodules) in strings also.
string bedding separation 3-4cm

alternating 2cm beds: (a) porous sst, pebble void strings, pebbles ~ 1cm (qz, rare white carbonate, rounded)
(b) well cemented qtose ssts, no voids, well sorted subangular grains.

Similar to (40). abundant voids, limonite filling.

Similar to (40)
abundant 1mm rounded voids
qz grains, moderately sorted, 1mm, orange/red, glazed.

poorly cemented, qtose, grains subrounded. including one 1cm rip up clast of fine sandstone (dark red)
abundant round ½-1cm voids, qz pebbles and ? oncoids. qz pebbles subparallel to bedding up to 3cm.

poorly cemented, very well sorted.
dark ? magic grains ~5-10%, subround to subangular.
pebble string at 42cm, discontinuous, qz pebbles ½, 1½, 3cm, separated by 5 or 6 cm.

pebble string at 13cm, recognised by 1-2cm round voids ? qz pebbles or dissolved oncoids.

grains subround qz orange to pale brown translucent,
poorly cemented, well-sorted.
weak parallel lamination lower 3cm.
Similar to (36), poorly exposed.
qz grains more abundant, 1mm subrounded translucent orange-red, glazed.
? paleosoil.

organic rich (black particles). abundant subcylindrical voids – rootlets?
diffuse brecciated aspect. ? palaeosoil.

qtose, well cemented. reworked oncoids, concentric structures. fine red silt rip-up clasts angular.
oncoids concentrated on bedding planes, 4cm apart, homogeneous med sand between.
oncoids generally entire, dark grey micritic, 1-2cm
fragmented oncoids fill interstices.

locally green mottled, brecciated – angular fragments up to 1cm.

occasional 1-3cm irregular pale green limestone nodules – hard, locally purple mottled.

exposure poor, abundant powdery calcite in cracks.

limestone/marl nodules irregular → 3cm, abundant calcareous coating to shale blocks – cracks.

pale green mottling.
½mm rounded voids, occasionally limonitic yellow. occasional 1mm rectangular white irregular calcareous clasts.
top 13cm less well consolidated, marly, fewer oncoids, almost white. 1mm round voids – rootlets
more loosely cemented, rich in oncoids – concentric dull grey micrite 1-3cm, often ovoid, long axes horizontal, clustered in pockets.
darker in oncoid rich areas.

cemented limestone/marl.
2cm dull grey micrite oncoids (fragments).
limestone set in light purple-mauve shale. shale includes 1mm dull grey calcareous oncoids.
rootlet brecciated.

similar to (26), more purple mottling, more clay rich, locally nodular aspect; nodules poorly defined 1-2cm

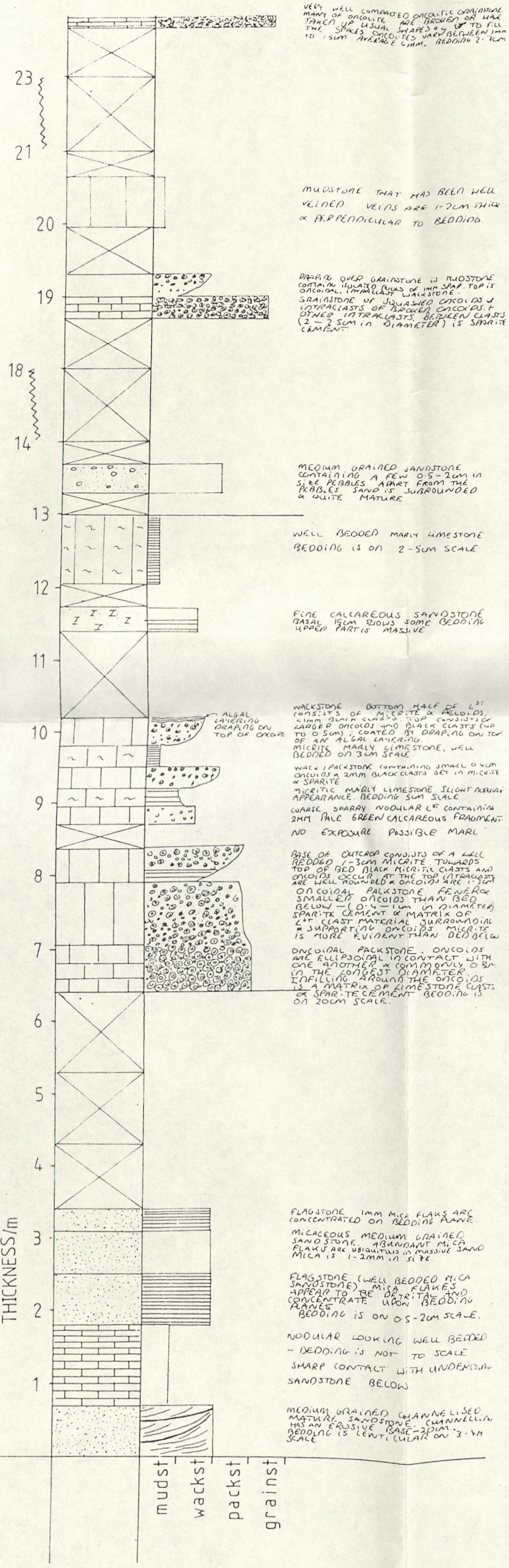
hard, calcareous, locally purple mottled. greenish gold colours around 2mm circular ? weathered oncoids / limestone.
at top one 1.5cm round golden brown reworked clast ? oncoid.

locally purple mottling. local golden brown weathering ? around oncoids.

purple mottling. calcareous matrix. included ½mm rounded qz, cemented. 3mm scale fissility. rare 1cm ? reworked oncoids
arenite siliceous matrix.

oncoids 3-26mm, increasing size to top of bed, interior chalky white.

MAMBRILLAS DE LARA (NORTH)



VERY WELL COMPACTED OOLITIC GRAINSTONE. MANY OF OOLITE ARE BROKEN OR HAVE TAKEN UP USUAL SHAPES TO FILL THE SPACES. OOLITES VARY BETWEEN 1MM TO 1.5CM AVERAGE 6MM. BEDDING 2-7CM

MUDSTONE THAT HAS BEEN WELL VEINED. VEINS ARE 1-2CM THICK & PERPENDICULAR TO BEDDING.

PACKST OVER GRAINSTONE IS MUDSTONE CONTAIN ISOLATED RUDS OF 1MM SPAR. TOP IS OOLOIDAL, LAMINATED WACKSTONE. GRAINSTONE OF SQUASHED OOLOIDS & INTRACLASTS OF BROKEN OOLOIDS. + OTHER INTRACLASTS. BETWEEN CLASTS (2 – 3 SCM IN DIAMETER) IS SPARITE CEMENT.

MEDIUM GRAINED SANDSTONE CONTAINING A FEW 0.5-2CM IN SIZE PEBBLES APART FROM THE PEBBLES SAND IS SUBROUNDED & QUITE MATURE.

WELL BEDDED MARLY LIMESTONE BEDDING IS ON 2-5CM SCALE

FINE CALCAREOUS SANDSTONE BASAL 15CM SHOWS SOME BEDDING UPPER PART IS MASSIVE

WACKSTONE. BOTTOM HALF OF LST CONSISTS OF MICRITE & PELOIDS. <1MM BLACK CLASTS. TOP CONSISTS OF LARGER OOLOIDS AND BLACK CLASTS (UP TO 0.5CM); COATED BY DRAPING ON TOP OF AN ALGAL LAMINATING. MICRITE MARLY LIMESTONE, WELL BEDDED ON 3CM SCALE.

WACK / PACKSTONE CONTAINING SMALL 0.4CM OOLOIDS & 2MM BLACK CLASTS SET IN MICRITE & SPARITE.

MICRITIC MARLY LIMESTONE. SLIGHT NODULAR APPEARANCE. BEDDING 5CM SCALE.

COARSE SPARRY NODULAR LST CONTAINING 2MM PALE GREEN CALCAREOUS FRAGMENTS.

NO EXPOSURE. POSSIBLE MARL

BASE OF OUTCROP CONSISTS OF A WELL BEDDED 1-3CM MICRITE. TOWARDS TOP OF BED BLACK MICRITIC CLASTS AND OOLOIDS OCCUR AT THE TOP INTRACLASTS ARE WELL ROUNDED & OOLOIDS ARE 1-5CM

OOLOIDAL PACKSTONE. GENERALLY SMALLER OOLOIDS THAN BED BELOW – (0.4 – 1CM IN DIAMETER) SPARITE CEMENT & MATRIX OF LST CLAST MATERIAL SURROUNDING & SUPPORTING OOLOIDS. MICRITE IS MORE EVIDENT THAN BED BELOW

OOLOIDAL PACKSTONE. OOLOIDS ARE ELLIPSOIDAL IN CONTACT WITH ONE ANOTHER & COMMONLY 0.8CM IN THE LONGEST DIAMETER. INFILLING AROUND THE OOLOIDS IS A MATRIX OF LIMESTONE CLASTS & SPARITE CEMENT. BEDDING IS ON 20CM SCALE.

FLAGSTONE. 1MM MICA FLAKES ARE CONCENTRATED ON BEDDING PLANE.

MICACEOUS MEDIUM GRAINED SANDSTONE. ABUNDANT MICA FLAKES ARE UBIQUITOUS IN MASSIVE SAND MICA IS 1-2MM IN SIZE.

FLAGSTONE (WELL BEDDED MICA SANDSTONE) MICA FLAKES APPEAR TO BE DETRITAL AND CONCENTRATE UPON BEDDING PLANES. BEDDING IS ON 0.5-2CM SCALE.

NODULAR LOOKING WELL BEDDED – BEDDING IS NOT TO SCALE

SHARP CONTACT WITH UNDERLYING SANDSTONE BELOW

MEDIUM GRAINED CHANNELISED MATURE SANDSTONE. CHANNELLING HAS AN EROSIVE BASE-20CM. BEDDING IS LENTICULAR ON 3-4M SCALE

Hortigüela Formation

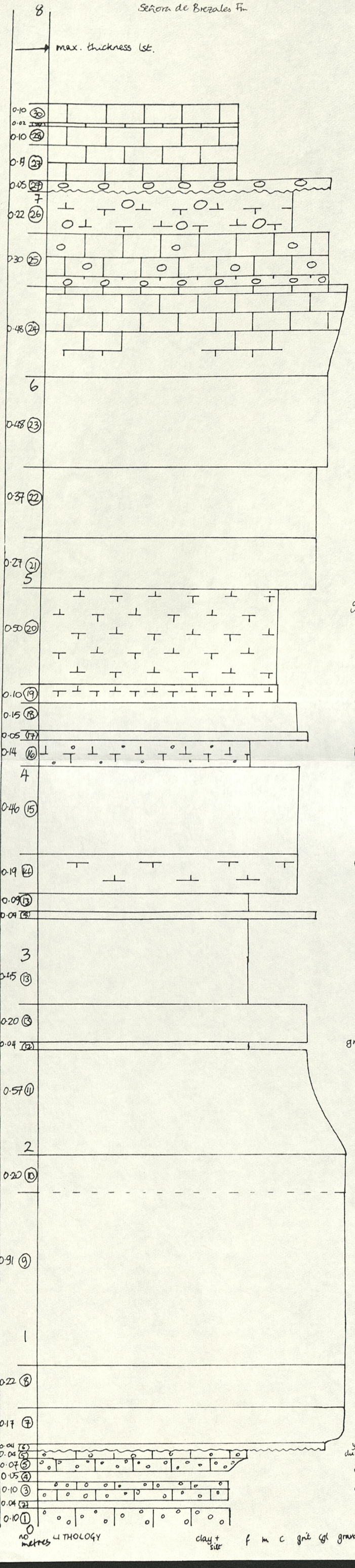
(Zurramujeres Member)

(San Martin Member)

20m

Quintanilla de las Viñas ①, April 26th, 1984.

Señora de Brezales Fm

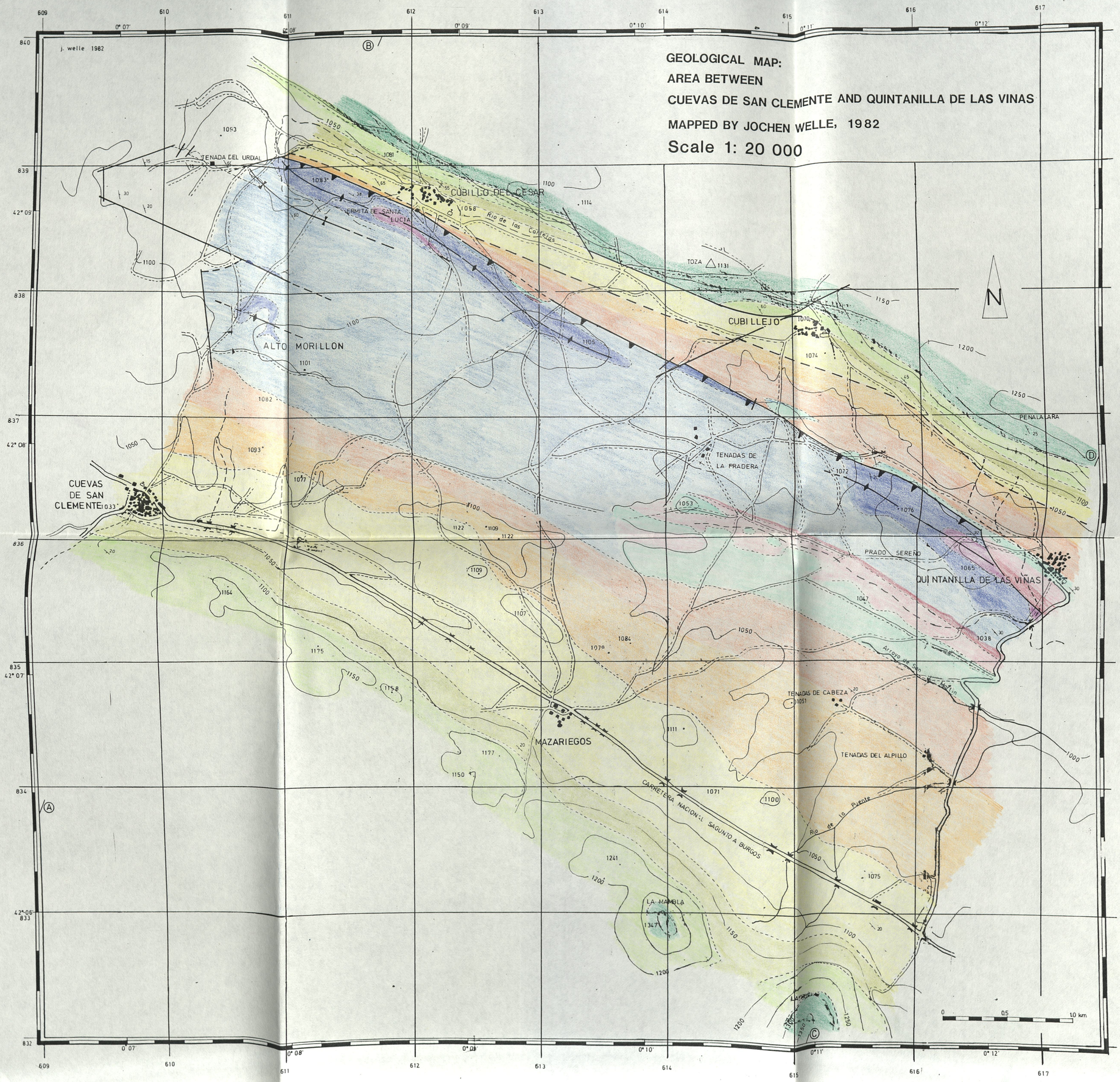


light to mid grey		similar. 1mm subhorizontal white sparry veins.
grey		similar to (27) (28)
grey	S?	more massive
grey	S *	series of complex 1-2cm thin micrite crusts, wavy, not laterally continuous.
brick red	?? M?	reworked nodules light grey beige micrite complex interdigitation of sst and carbonate, interstitial carbonate cemented sst,
brick red		matrix - purple-flesh mottled micrite, impure. pale grey micrite clasts up to 2cm, including sparite veins 1mm some orange clasts 1-5cm. base - abundant 2-3cm clasts, becoming rarer to top.
		grey limestone rubble. clasts up to 30cm set in red slightly sandy matrix.
		clasts - grey micritic limestone - strongly purple mottled.
grey		grey limestone rich conglomerate grey + orange carbonate clasts up to 10cm - generally 1-3cm. some 1-2cm red clasts. strongly clast supported except for local matrix rich patches.
grey	M?	similar to (21). clast supported, fine sandy matrix, numerous ½-1mm white carbonate dots ? Microcodium.
grey	M?	max clast size 5cm, mean 1cm. burrow? top surface. 1cm diameter, 8cm long clasts: mostly grey, some red, a few orange, subrounded equant sometimes irregular. strongly clast supported with interstitial smaller clasts → relatively little matrix. some qz up to 1½mm some ½-1mm micritic carbonate dots.
grey-purple		no exposure - lateral equivalent 2m west is "coarse sst as (19)"
grey-purple (yellow-brown matrix)		polymict - grains orange, grey, red/purple. probably carbonate cemented. clast + matrix supported. clasts 1-3cm mostly grey, some orange, some red, irregular - dissolved clast margins. also 1-2mm qtz on weathered surfaces. smaller clasts rounded, irregular, equant.
purple-yellow	M?	clasts: yellow orange Bath/Call, fine grey carbonate points/strings within matrix. mottled, 1-2mm subrounded qtz., fine light carbonate, dark purple marly silt, yellow mottling as 2cm patches. very many thin ¼mm carbonate strings - Microcodium?
red		clasts ½-3cm, mostly grey weathering limestone, some yellow weathering. rich in many 1-2mm grey carbonate clasts
red grey		clasts orange, red, black, yellow. rounded to subrounded, set in finer white/pink? carbonate rich matrix.
438	?* M?	sst again, numerous fine white dots sandy matrix, clasts 1-2cm: red + grey mostly, a few orange
		sst only
439	* M?	no exposure - lateral equivalent - matrix-rich conglomerate, numerous grey carbonate strings + dots.
grey to red	* M?	mottled, abundant grey carbonate veins - strings ½mm wide to 1mm
	M?	max clast size 41cm. cgl becoming more matrix-rich, red fine sand. clasts - more irregular margins, mottled and ?1-2mm spar coatings.
		max. grain size 26cm. some grey weathering pebbles are also finely crystalline microsparite. grey weathering finer pebbles becoming more abundant up section. now possibly alternating 20cm beds of finer conglomerate and coarse rubble clasts but organisation not clear. some clasts of yellow orange coated, orange finely crystalline (Bath-Call) fossiliferous lst. up to 50cm. many grey and red weathering pebbles are probably Bath/Call. too, fossiliferous, sparry areas and uniform fine arenite grain size.
		similar clast compositions to (7). clast size variable - areas of 10-20cm clasts heavily clast supported, interstitial pockets of 1-2cm clasts in deep red sandy matrix, not clast supported. clasts of: grey-coated limestone (pure white beige micrite thin sparry traces and diffuse orange staining ?c) much lst finely recrystallised. : brown orange lst, fine grained, recrystallised (Bath/Call) : red micrite rich in 1-2cm fossils (weathered Bath/Call). poorly sorted. clasts 2-5cm, larger clasts up to 30cm. clasts 1cm, irregular, mauve grey carbonate, rounded edges. supported by yellow red fine sand matrix. qtz grains 2-3mm subrounded. largest shell fragment 1cm. more recrystallised microspar texture to top.
yellow-red dull orange-brown		no exposure rhynchonellids 4cm, irregular, vugs (filling brachs + fragments with spar) up to 4cm irregular. Mottled purple patches, diffuse.
dull orange-brown		no exposure
yellow orange	* r, t	qtz grains ½-1mm abundant, spar veins 2cm × 4mm purple mottling, especially around shells a few ½cm pale orange lst intraclasts. rhynchonellids 2cm numerous, terebratulid fragments angular, fractured, 1cm.
yellow-orange	r, t	remarks

GEOLOGICAL MAP:
AREA BETWEEN
CUEVAS DE SAN CLEMENTE AND QUINTANILLA DE LAS VIÑAS

MAPPED BY JOCHEN WELLE, 1982

Scale 1: 20 000

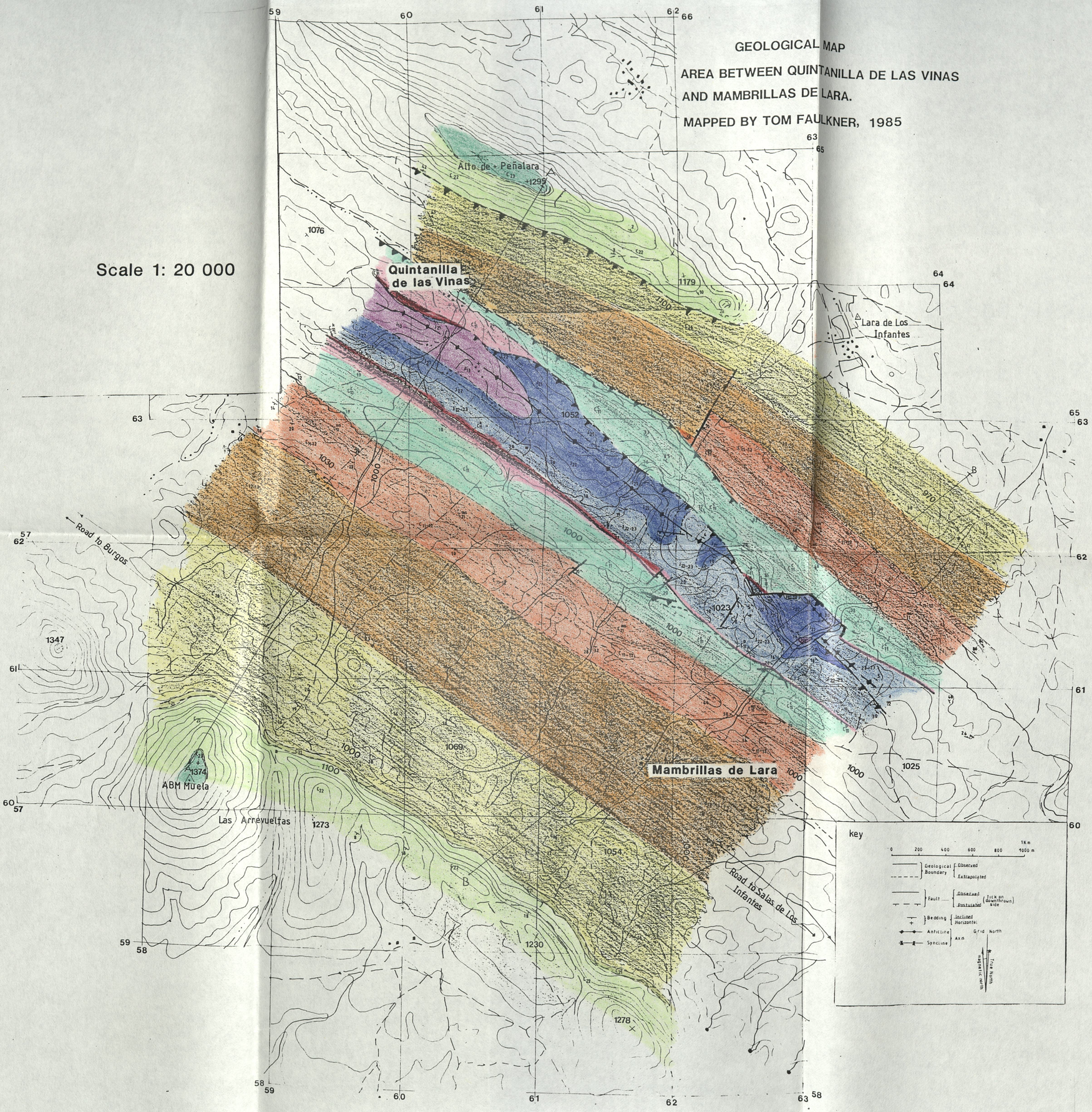


GEOLOGICAL MAP

AREA BETWEEN QUINTANILLA DE LAS VINAS AND MAMBRILLAS DE LARA.

MAPPED BY TOM FAULKNER, 1985

Scale 1: 20 000



The Geology South of Santo Domingo de Silos, Provincia de Burgos, Spain.

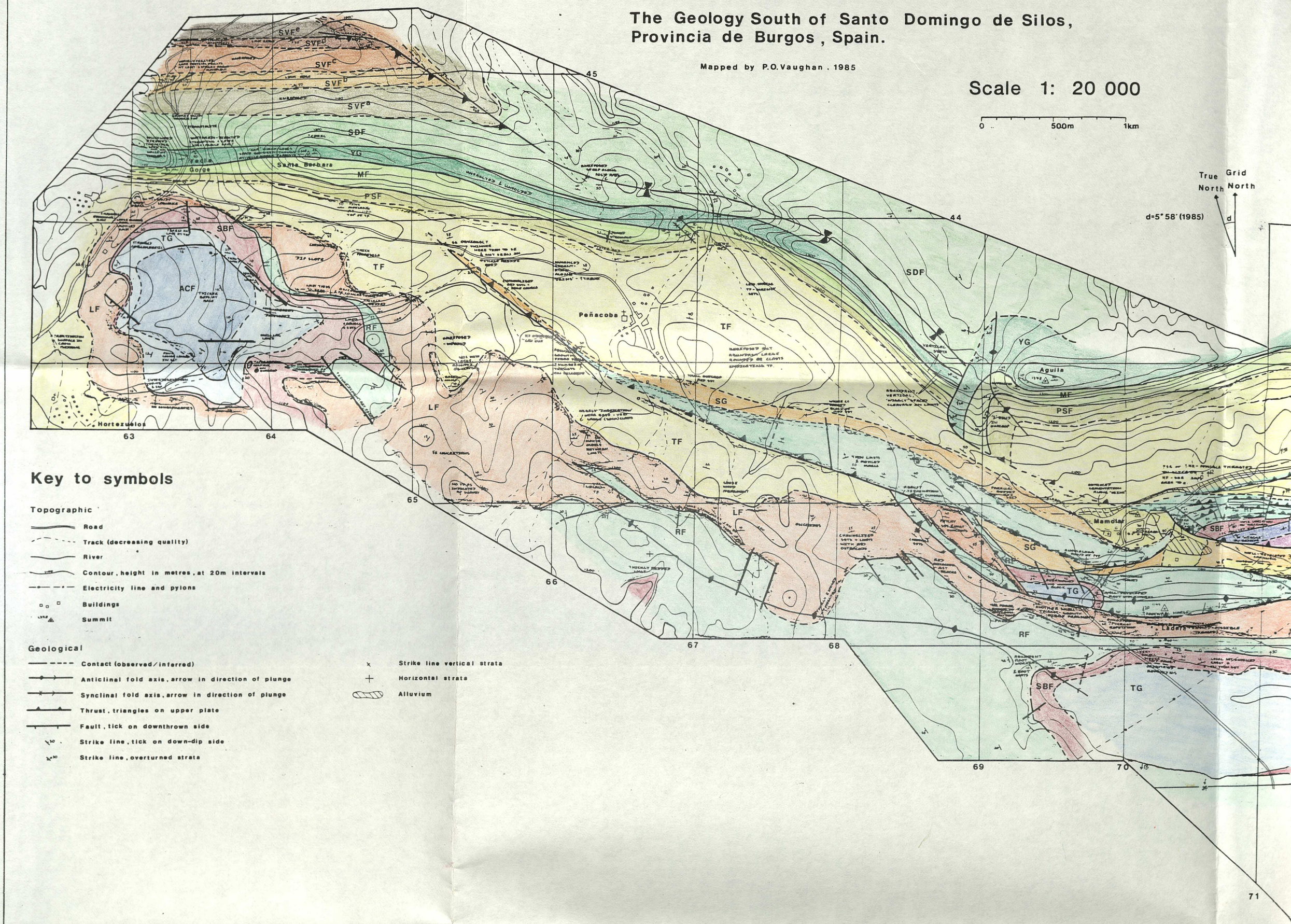
Mapped by P.O. Vaughan. 1985

Scale 1: 20 000

0 500m 1km

True North / Grid North

d=5°58'(1985)



Key to symbols

Topographic

- Road
- Track (decreasing quality)
- River
- Contour, height in metres, at 20m intervals
- Electricity line and pylons
- Buildings
- Summit

Geological

- Contact (observed/inferred)
- Anticlinal fold axis, arrow in direction of plunge
- Synclinal fold axis, arrow in direction of plunge
- Thrust, triangles on upper plate
- Fault, tick on downthrown side
- Strike line, tick on down-dip side
- Strike line, overturned strata
- Strike line vertical strata
- Horizontal strata
- Alluvium

GEOLOGICAL MAP OF THE SAN LEONARDO FAULT ZONE (WEST).

Nigel H. Platt (University of Oxford), June 1986.

The Geology South of Santo Domingo de Silos, Provincia de Burgos, Spain.

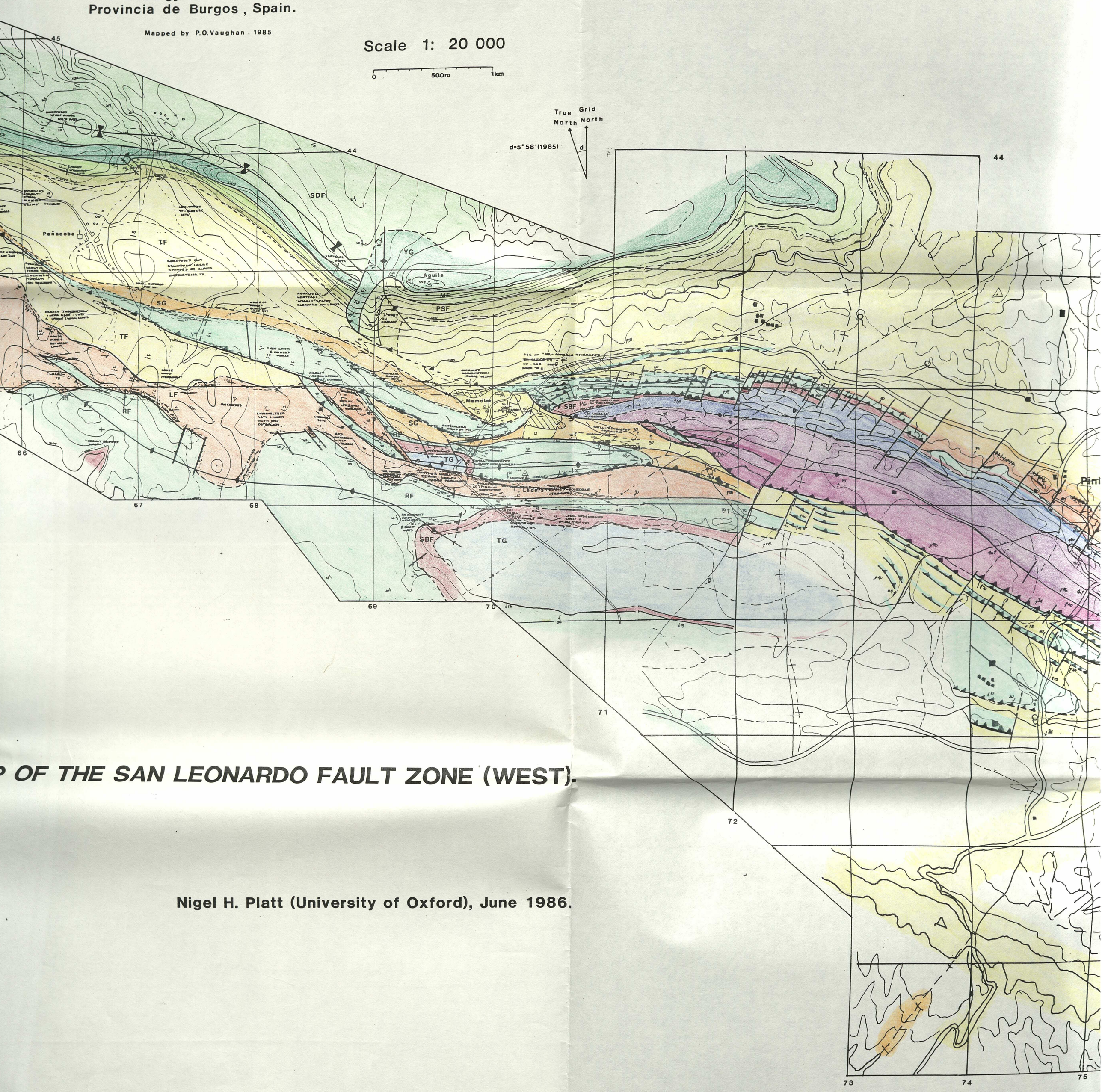
Mapped by P.O.Vaughan. 1985

Scale 1: 20 000

0 500m 1km

True North / Grid North

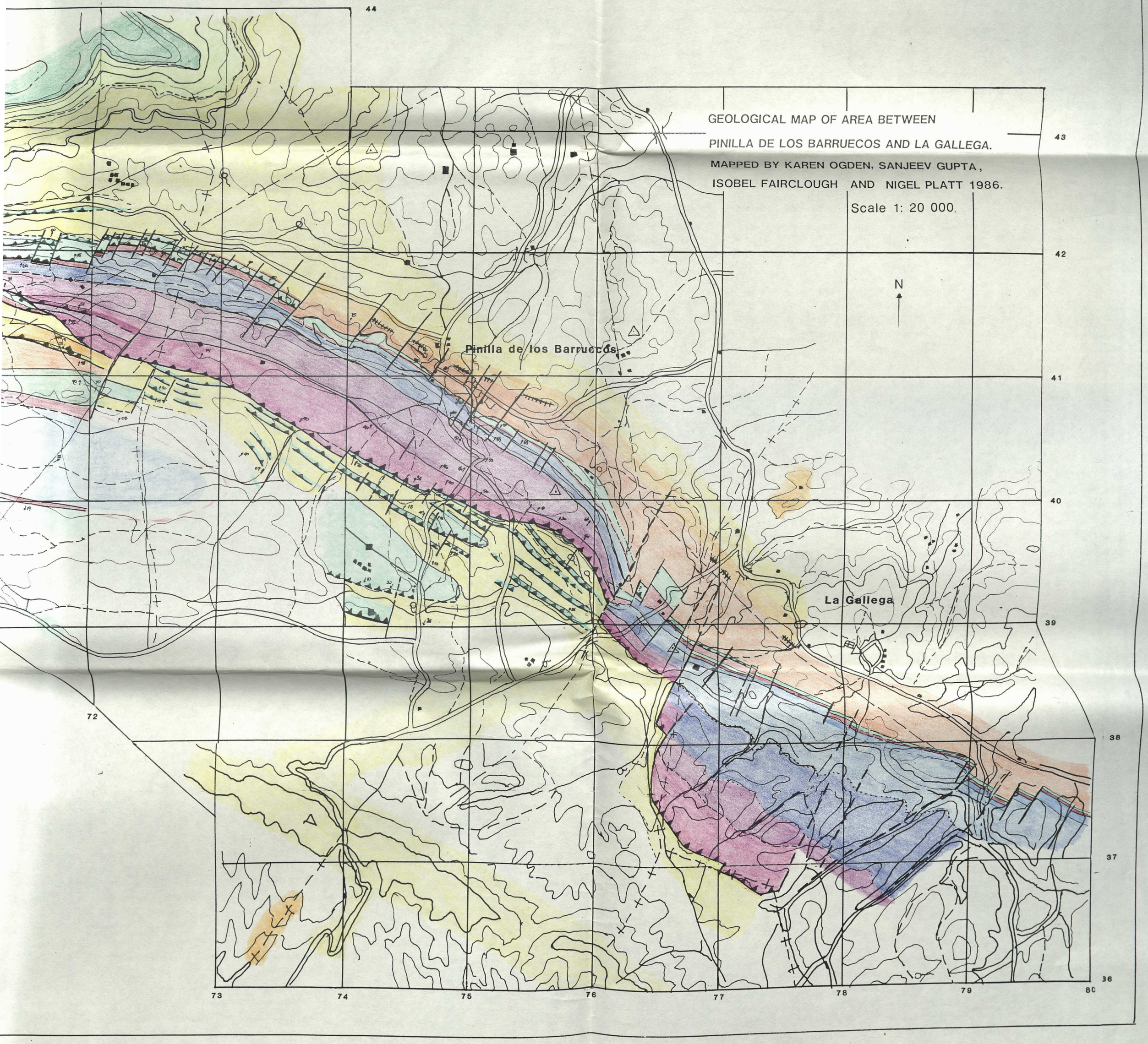
d=5°58' (1985)



OF THE SAN LEONARDO FAULT ZONE (WEST).

Nigel H. Platt (University of Oxford), June 1986.

GEOLOGICAL MAP OF AREA BETWEEN PINILLA DE LOS BARRUECOS AND LA GALLEGA.

MAPPED BY KAREN OGDEN, SANJEEV GUPTA, ISOBEL FAIRCLOUGH AND NIGEL PLATT 1986.

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