MEMÓRIAS
DOS
SERVIÇOS GEOLÓGICOS DE PORTUGAL

THE CARBONIFEROUS OF PORTUGAL

Edited by M. J. LEMOS DE SOUSA and J. T. OLIVEIRA

NÚMERO 29 LISBOA 1983

DIRECÇÃO-GERAL DE GEOLOGIA E MINAS

Director : *Delfim de Carvalho*

Editors : *M. J. Lemos de Sousa* and *J. T. Oliveira*

Address : R. Academia das Ciências, 19-2.º — 1200 Lisboa — Portugal

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CONTENTS

FORWARD

The last two decades have seen an outstanding advance in our knowledge and understanding of the Carboniferous geology in Portugal. The purpose of this issue of Memórias dos Serviços Geológicos de Portugal *is to provide, at the occasion of the X International Congress of Carboniferous Stratigraphy and Geology, a basic document bringing together and reviewing the most significant scientific results of that period.*

The Carboniferous terrains are of particular interest in Portugal. The huge polymetallic massive sulphide orebodies of the Iberian Pyrite Belt and the coal deposits of the Douro region form part of Carboniferous geological units. Geology has played a decisive role in the discovery of many of those mineral deposits and will be an increasingly important factor in future mineral exploration.

I believe the articles in this volume are a fair sample of the many efforts devoted to various geological studies on our Carboniferous formations. As a matter of fact the most relevant achievements are presented here. I do also believe this is a major contribution for a better understanding of the Carboniferous Geology, the Hercynian orogeny and our Resource Base.

Lisboa, May 13, 1983

DELFIM DE CARVALHO
Director

PREFACE

As representatives of Portugal in the Organizing Committee of the X International Congress of Carboniferous Stratigraphy and Geology (Madrid, 1983) we accepted the responsability to edit an up-to-date synthesis of the main aspects of the Carboniferous of Portugal.

As a matter of fact, the last synthesis on the Carboniferous of Portugal, concerning, mainly, the Stratigraphy and the Palaeontology of terrestrial formations, was written by Carlos Teixeira in 1944.

For the next few years, however, the subject has gained considerable interest and several research teams have been engaged in the investigations of various aspects of the Portuguese Carboniferous. These investigations have been carried out by Portuguese state organizations (Serviços Geológicos de Portugal, Serviço de Fomento Mineiro, Faculdade de Ciências de Lisboa, and Faculdade de Ciências do Porto), foreign universities, and private companies.

More recently, and particularly during the planning of this volume, a considerable research effort, which beneficted from international co-operation, was made to revise the principal aspects of the Portuguese Carboniferous, namely those related to Stratigraphy, Palaeontology, Structural Geology, and Economic Geology.

The wide range of subjects shown in this book is illustrative of the breadth of the 'Carboniferous' research in Portugal, today. The papers concerning the Marine Carboniferous were co-ordinated by J. T. Oliveira, and those related with Terrestrial Carboniferous by M. J. Lemos de Sousa.

Obviously, the present volume, which in many aspects represents a new type of publication of the Serviços Geológicos de Portugal, is not exhaustive. We do believe, however, that the various papers presented here will be of interest as guide-lines to further investigations.

We wish to express our gratitude to the authors who accepted to contribute.

We are most indebted to the Director of the Serviços Geológicos de Portugal for permission to publish this volume in the Memórias dos Serviços Geológicos de Portugal, *and for allowing us to peruse archives and collections.*

Thanks are also due to the Directors of the Geology Departments of the Faculdade de Ciências e Tecnologia de Coimbra, Faculdade de Ciências de Lisboa, and Faculdade de Ciências do Porto, for the access to collections and all kinds of facilities.

Some of the investigations have been, partly or wholly, supported by NATO Research Grant n.º 85.80/D1/D2. We wish to express our thanks to the NATO Research Council for this financial assistance.

The international co-operation has involved many colleagues from different European countries. We gratefully acknowledge this participation, and the Universities and Geological Surveys concerned.

We wish to thank the staff of the Art Department of the Serviços Geológicos de Portugal who prepared the originals of most of the figures included in this volume.

Although we had decided not to mention individuals in this preface, an exception ought to be made with respect to Miss Maria Manuela Tavares who typed the manuscripts, organized the referenca lists, and the author index. Her enthusiasm and efficiency in building up this book are acknowledged.

M. J. Lemos de Sousa
J. T. Oliveira
Editors

LIST OF CONTRIBUTORS

F. J. A. S. Barriga [1]
D. Carvalho [2]
L. C. G. Domingos [3]
R. M. C. Eagar [4]
J. L. S. Freire [5]
F. Gonçalves [1, 2]
J. Munhá [1]
J. T. Oliveira [2]
E. Pereira [2]
A. Ribeiro [1, 2]
J. Brandão Silva [1]
F. Gomes da Silva [3]
M. J. Lemos de Sousa [6]
R. H. Wagner [7]

[1] Departamento de Geologia, Faculdade de Ciências, Universidade de Lisboa, Rua da Escola Politécnica 58, 1294 Lisboa Codex, Portugal.

[2] Serviços Geológicos de Portugal, Rua da Academia das Ciências 19-2.º, 1200 Lisboa, Portugal.

[3] Museu e Laboratório Mineralógico e Geológico, Faculdade de Ciências e Tecnologia, Universidade de Coimbra, 3049 Coimbra Codex, Portugal.

[4] The Manchester Museum, The University, Manchester M13 9PL, England, U.K.

[5] Serviço de Fomento Mineiro, Rua da Amieira, 4465 São Mamede de Infesta, Portugal.

[6] Mineralogia e Geologia, Faculdade de Ciências, Universidade do Porto, Praça Gomes Teixeira, 4000 Porto, Portugal.

[7] Department of Geology, The University, Beaumont Building, Brookhill, Sheffield S3 7HF, England, U.K., and E.N. Carbonífera del Sur, Peñarroya-Pueblonuevo (Córdoba), España.

MARINE CARBONIFEROUS

THE MARINE CARBONIFEROUS OF SOUTH PORTUGAL: A STRATIGRAPHIC AND SEDIMENTOLOGICAL APPROACH [1]

J. T. Oliveira [2]

Key words: South Portuguese Zone; Marine Carboniferous, Portugal; Pyrite Belt; Volcano-sedimentary Complex; Baixo Alentejo Flysch Group; Turbidites; Black shales; Dolomitic limestones; Goniatites; Conodonts; Famennian; Tournaisian; Viséan; Namurian; lower Westphalian.

Palavras-chave: Zona Sul Portuguesa; Carbonífero marinho, Portugal; Faixa Piritosa; Complexo Vulcano-sedimentar; Grupo Flysch do Baixo Alentejo; Turbiditos; Xistos negros; Calcários dolomíticos; Goniatites; Conodontes; Fameniano; Tournaisiano; Viseano; Namuriano; Vestefaliano inferior.

ABSTRACT

The marine Carboniferous stratigraphy and sedimentology of Portugal is revised in the light of recent information. The marine facies outcrop mainly in the South Portuguese zone. Three major stratigraphic units are recognised: the Volcano-Sedimentary Complex of the Pyrite Belt, the Baixo Alentejo Flysch Group and the Carrapateira Group.

The Volcano-Sedimentary Complex comprises a great variety of volcanic, sedimentary and mixed rocks, whose stratigraphic relationships and depositional environments are discussed. Its age ranges from mid Famennian to the early part of the late Viséan.

The Baixo Alentejo Flysch Group is a southward prograding turbiditic succession, divided in three lithostratigraphic units: the Mértola Formation, of late Viséan age; the Mira Formation of latest Viséan to Namurian age and the Brejeira Formation of mid Namurian to early Westphalian age. The stratigraphic and sedimentological characteristics of the turbidites in all three units, their depositional pattern and source of detritus are analysed and commented upon.

The more condensed facies of the Carrapateira Group occur only in southwestern Portugal. They comprise shales, marls, limestones and black shales. Three units are recognised: Bordalete Formation, of mid to late Tournaisian age; the Murração Formation of Viséan age and the Quebradas Formation of Namurian age. The latter two units are particularly rich in goniatites.

The character of the Devonian substrate and comparisons between the three major lithostratigraphic units allow some comments on the basin evolution during the Carboniferous.

RESUMO

No presente trabalho a estratigrafia e sedimentologia do Carbónico marinho de Portugal são revistas à luz dos resultados obtidos no decurso de recentes investigações. As fácies marinhas do Carbónico afloram principalmente na Zona Sul Portuguesa, onde são conhecidas três unidades litoestratigráficas de ordem maior: O Complexo Vulcano-Sedimentar da Faixa Piritosa, o Grupo do Flysch do Baixo Alentejo e o Grupo da Carrapateira.

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(1) Investigation partly sponsored by NATO Research Council (Grant N.° 85.80/D1/D2).

(2) Serviços Geológicos de Portugal, Rua da Academia das Ciências, 19-2.°, 1200 Lisboa, Portugal.

O Complexo Vulcano-Sedimentar compreende grande variedade de rochas vulcânicas e sedimentares cuja idade se escalona entre o Fameniano médio e a parte inferior do Viseano superior. As relações estratigráficas e os ambientes de deposição das diversas litologias são amplamente discutidos.

O Grupo do Flysch do Baixo Alentejo é uma sequência turbidítica, progradante para sudoeste, composta por três unidades litoestratigráficas: Formação de Mértola (Viseano Superior), Formação do Mira (topo do Viseano superior-Namuriano) e Formação da Brejeira (Namuriano médio-Vestefaliano inferior). A análise das características sedimentológicas dos turbiditos permite tecer algumas considerações quanto ao seu modelo de deposição e origem do material detrítico.

O Grupo de Carrapateira aflora unicamente nos antiformes de Aljezur e Bordeira e é constituído por três unidades: Formação de Bordalete (Tournaisiano médio e superior), Formação do Murração (Viseano) e Formação de Quebradas (Namuriano). As fácies dominantes são pelitos, margas, calcários em grande parte dolomíticos e xistos negros. As unidades constituintes do Grupo Carrapateira, no seu conjunto, mostram-se bastante mais condensadas do que as unidades contemporâneas do Complexo Vulcano-Sedimentar e do Grupo do Flysch do Baixo Alentejo. Este facto e as características sedimentares do substracto Devónico fornecem indicações sobre a paleogeografia da Zona Sul Portuguesa durante o Carbónico.

1. INTRODUCTION

In Portugal the marine Carboniferous is only recognised confidently in the South Portuguese Zone. North of this palaeogeographic zone (Fig. 1) the Carboniferous is essentially terrestrial (see 'Terrestrial Carboniferous' — this volume). In the Portuguese part of the Ossa-Morena Zone two lithostratigraphic units of marine facies are regarded as possibly Carboniferous, viz. the Terena Formation, a turbiditic sandstone unit which crops out between Barrancos and Estremoz and the so called 'Xistinhos' Formation, composed of light to dark grey shales, which outcrops between Vendas Novas and Alfundão. However, the Terena Fm is probably of late Devonian age (Perdigão *et al.* 1982) and the stratigraphy of the 'Xistinhos' is still poorly understood. For these reasons, the present work is limited to the South Portuguese Zone.

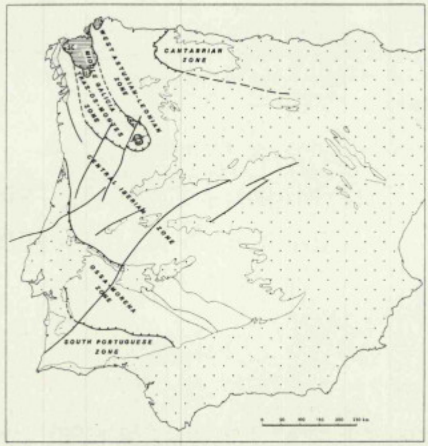


Fig. 1 — Palaeogeographic and geotectonic units in the Iberian Massif (Ribeiro *et al.* 1979)

The stratigraphic units of the South Portuguese Zone have been subdivided in two subzones (Carvalho *et al.* 1971), or in sectors (J. Oliveira *et al.* 1979). To facilitate the understanding and description of the different units, a simpler and slightly different scheme is preferred in this work (Fig. 2). The overall picture

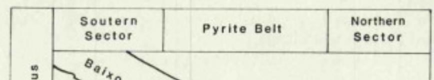


Fig. 2 — Major lithostratigraphic units of the South Portuguese Zone

of the South Portuguese Zone is shown in a geological sketch map (see appendix). This map incorporates previous work, unpublished mapping by state and private organisations and current research.

To prepare a synthesis on the marine Carboniferous stratigraphy and sedimentology has not been an easy task. The stratigraphic and sedimentological research on this topics has been only resumed recently. Many problems still remain to be solved, leading to uncertainties on stratigraphic correlations, sedimentary environments, etc. A critical approach to these difficulties will be apparent through out the text.

Although presented as a synthesis, this paper is not intended to be a simple compilation. References to recent research, kindly provided by colleagues and the author's co-workers are incorporated in the text.

2. HISTORICAL INTRODUCTION

The geology of the South Portuguese Zone, particularly the stratigraphy of the marine Carboniferous evolved in an irregular fashion.

After DELGADO's (1876, 1910) and PRUVOST's (1912/13) pioneer studies, work on the Pyrite Belt remained stagnant for about fifty years. Only in the 1960s, when state departments (Serviço de Fomento Mineiro — SFM; Serviços Geológicos de Portugal — SGP), foreign universities and private companies started a systematic mapping programme the stratigraphy began to be understood. The main contributions were those of KLEYN (1960) who mapped the volcano-sedimentary and flysch sequences of the southern termination of the Cercal Anticline; MAC GILLAVRY (1961a, b) who published the first modern approach on the Pyrite Belt stratigraphy and palaeogeography; STRAUSS (1965, 1970) who made a detailed study of the geology of the Lousal mine area; and VAN DEN BOOGAARD (1963, 1967) who recognised for the first time Famennian and Tournaisian conodonts in the Pomarão and Mértola limestones. The latter also worked out a very detailed stratigraphic column in the Pomarão Anticline. Since then, this stratigraphic column has become classic all over the Iberian Pyrite Belt. ZBYZEWSKI *et al.* (1964) published the first 1/50,000 geological map (Castro Verde Sheet); PFEFFERKORN (1968) attempted to correlate the stratigraphic units of the Mértola area with those north of the Pyrite Belt, along the Guadiana river valley; SCHERMERHORN & STATON (1969) defined the stratigraphy around the Aljustrel mine and recognised a phase of overthrusting in that area. In terms of Southern Portugal geotectonics, CARVALHO *et al.* (1971) divided the South Portuguese Zone into two palaeogeographic subzones; SCHERMERHORN (1971) wrote an interesting outline of the whole Pyrite Belt stratigraphy; CARVALHO *et al.* (1975) disputed the new interpretation (BERNARD & SOLER 1974) for the stratigraphy of the Pomarão-Puebla de Guzman Anticline; FANTINET *et al.* (1976) dated the Famennian basement of the Mértola region; CARVALHO (1976) mapped the lithostratigraphic units of all the Cercal Anticline and emphasised their similarities with those of the classic Pyrite Belt; CARVALHO *et al.* (1976a) published a synthesis of the Iberian Pyrite Belt geology; LECA (1976, 1983) carefully mapped the Alcoutim and the Castro Verde-Panoias regions; VAN DEN BOOGAARD & SCHERMERHORN (1981) discovered Famennian conodonts in the Monte Forno da Cal limestones, south of Castro Verde.

South of the Pyrite Belt the stratigraphic work began with PRUVOST's (1914) study of a collection of goniatites from Southern Portugal, housed at the Serviços Geológicos de Portugal. He determined four goniatite zones: *Prolecanites algarbiensis* PRUVOST, *Goniatites striaus* SOWERBY, *Glyphioceras beyrichianum* DE KONINK and *Gastrioceras carbonarium* BUCH. SOUSA (1919/22, 1920, 1924, 1926) following PRUVOST's study, determined upper Visean goniatites north of Azinhal village and tried to separate Dinantian from Moscovian rocks everywhere in the sedimentary basin. He also recognised the general stratigraphic succession of the Carrapateira region. After SOUSA's work, and during thirty years, only short palaeontological notes were published; FLEURY (1924) described foraminifers from Southwest Portugal (the exact localities were not shown); COSTA (1943) determined Carboniferous faunas from localities between Odemira and Grândola; FEIO (1946a, b) studied goniatites and a rich *Posidonia becheri* occurrence near Mértola; DELÉPINE (1957) studied a goniatites collection housed at the Serviços Geológicos de Portugal. The end of the 1950s mark the beginning of field mapping in the Carrapateira area, leading to the discovery of the Praia do Murração thrust (FEIO & LOMBARD 1958). Meanwhile LOMBARD (1958) described for the first time some flysch cross sections while FANTINET (1960, 1963) did the petrography and measured the flysch palaeocurrents near Grândola. Further south, FRISCHMUTH (1968) investigated the flysch sedimentology and tectonics between Alcoutim and Azinhal. Towards the end of the 1970s there was renewed interest on the Carboniferous geology and stratigraphy. PERDIGÃO (1978) attempted a general outline of the stratigraphy of the Aljezur and Bordeira areas; J. OLIVEIRA *et al.* (1979) published a first preliminary study on the entire flysch stratigraphy and sedimentology. More recently, J. OLIVEIRA (1982) commented on the main problems concerning the Devonian-Carboniferous stratigraphy of Southern Portugal and (J. OLIVEIRA *et al.* in prep.) worked out the stratigraphy of the upper Devonian and Carboniferous sediments of the Aljezur and the Bordeira Antiforms. In the Carrapateira area, RIBEIRO *et al.* (in press a) made a detailed study of the local geology. Three recent books on the general geology of Portugal discussed also the stratigraphy of the marine Carboniferous (RIBEIRO *et al.* 1979; TEIXEIRA & GONÇALVES 1980; TEIXEIRA 1981). The former of these books synthetised the subject quite correctly. Lastly (1982), the Serviços Geológicos de Portugal published a 1/200,000 geological map (sheet 7) on which part of the marine Carboniferous is represented.

3. DEVONIAN BASEMENT

The Devonian basement on which the marine Carboniferous has been laid down is dated only in the Pyrite Belt and in southwestern Portugal. In both sectors it shows similar lithologies and equivalent ages.

Throughout the northern sector the age of the lithostratigraphic units is still uncertain due to the absence of stratigraphically reliable fossils. Since the aim of this paper is the Carboniferous stratigraphy and sedimentology the problems of the Devonian are alluded to only briefly.

Northern Sector

In this sector the lithostratigraphic units presently recognised have a geological setting which seems compatible with a large anticline, dipping northwestwards (see the geological map annex). From bottom to top they are: the Pulo Fm, a highly deformed, greenschist metamorphic unit, composed of black and grey schists, quartzites and rare interbedded acid and basic volcanics; the Pulo Fm is overlain on the southern limb of the anticline by the Atalaia Fm (sandstones and phyllites) and this by the Gafo Fm (a turbiditic sandstone / shale unit). On the northern limb, the Pulo Fm grades into the Ribeira de Limas Fm (sandstones, phyllites and rare tuffites) which is overlain by the S. Iria Fm (a greywacke / shale turbiditic unit with sandstones and rare limestones intercalated). The S. Iria and Ribeira de Limas Fms together constitute the Ferreira-Ficalho Group.

The geological setting of all these units is compatible with a lateral equivalence, although possibly diachronous, between the Gafo and S. Iria Fms, on one side, and the Atalaia and Ribeira de Limas Fms, on the other (CARVALHO *et al.* 1976b). Indeed, the S. Iria and Gafo Fms show similar lithologies and yield the same kind of poorly preserved plant remains and crinoids. The Ribeira de Limas and Atalaia Fms are also similar lithologically, but are unfossiliferous, as is the Pulo Fm. The age of these units has been interpreted in different ways: the Pulo Fm was considered as Silurian (DELGADO 1904/07), Tournaisian (PFEFFERKORN 1968) and more recently late Devonian and older (SCHERMERHORN

1971; CARVALHO *et al.* 1976b); the Atalaia and Ribeira de Limas Fms were correlated with the Volcano-Sedimentary Complex of the Pyrite Belt (PFEFFERKORN 1968; CARVALHO *et al.* 1976b), that is known to be late Devonian to Viséan; the Gafo and S. Iria Fms were considered as possibly Tournaisian (CARVALHO *et al.* 1971), early to middle Viséan (SCHERMERHORN 1971; CARVALHO *et al.* 1976b) or even late Viséan (PFEFFERKORN 1968). Finally the S. Iria and Ribeira de Limas Fms were taken together as lower Devonian (CARVALHOSA 1965; TEIXEIRA & THADEU 1967) or upper Devonian (Carta Geológica de Portugal 1968, scale 1/000000).

At the moment, and as far as the author is concerned, the age of all these lithostratigraphic units is still controversal: in a recent paper (J. OLIVEIRA 1982) the S. Iria / Gafo Fms were considered as possibly late Devonian to Tournaisian, and an unconformity (or a tectonic boundary) has been suggested between these units and the underlying Ribeira de Limas / Atalaia Fms. In this case the Ribeiras / Atalaia Fms should be older than the Volcano-Sedimentary Complex of the Pyrite Belt and the Pulo Fm might even be Silurian.

Pyrite Belt

The Devonian basement of the Pyrite Belt is composed of phyllites, siltstones, quartwackes, orthoquartzites and, in some places (Pomarão, Mértola, southeast of Castro Verde), limestone lenses and nodules at the top. All these lithologies are grouped together in the so-called Phyllite Quartzitic Formation or simply PQ (SCHERMERHORN 1971). It is almost certainly up to 200 m thick.

In the Pomarão Anticline the PQ quartzites (locally designated as Eira do Garcia Fm, VAN DEN BOOGAARD 1967) are overlain by a 20 m thick pelitic unit with interbedded limestone lenses and nodules. From this unit PRUVOST (1914) determined the following fauna: *Clymenia laevigata* MÜNSTER, *Phacops granulatus* MÜNSTER, *Orthis arcuata* PHILLIPS, *Cypricardina scalaris ?* PHILLIPS and *Petraia radiata* MÜNSTER. These faunas indicated a Famennian age. In the limestones of the same unit, now called Nascedios Fm, VAN DEN BOOGAARD (1963) found conodonts of upper Famennian age (middle *velifera* and *styriacus* zones). He also identified foraminifers (*Tolypammina* sp. and *Hyperammina* sp.) and poorly preserved ostracods and fish remains.

Near Mértola, FANTINET *et al.* (1976) described a phyllite horizon with interbedded limestone and quartzitic lenses. From the limestones they recovered: *Platyclymenia richteri* WEDEKIND, *Cyrtoclymenia* sp., *Posidonia venusta* MÜNSTER, *Edmondia* cf. *philippi* HALL & WHITFIELD, *Cimitaria* HALL & WHITFIELD, and conodonts of the *velifer* and *styracus* zones. These faunas also indicate a late Famennian age.

Southeast of Castro Verde, in the Monte Forno da Cal region, VAN DEN BOOGAARD & SCHERMERHORN (1981) determined lower Famennian conodonts (lower *marginifera* zone) from limestone lenses, near the top of the PQ.

From the above assemblages it is clear that the top of the PQ is of Famennian age. Below this level no fossils have been recorded and the bottom of the unit is unknown. The stratigraphic position of many quartzitic horizons, with respect to the Volcano-Sedimentary Complex (see below) is still dubious, owing to the tectonic complexity. This question is to be discussed in conjunction with the Volcano-Sedimentary stratigraphy. In this work it is admitted that the PQ may reach a Tournaisian age.

The PQ lithologies are strongly deformed tectonically thus making it almost impossible to analyse their vertical and lateral sedimentary development. However the quartzite beds occasionally display sedimentary structures such as massive to graded beds, parallel lamination, herringbone type cross beds, planar cross bedding and synsedimentary folds. The sedimentological and environmental significance of these structures is still to be analysed. Earlier work considered the quartzites either as shallow water sandstones (MAC GILLAVRY 1961a; STRAUSS 1970; J. OLIVEIRA 1982) or as deep water sandstones (SCHERMERHORN 1971; LECA 1983).

Southwestern Portugal (Aljezur and Bordeira Antiforms)

The Devonian age of the lowermost sediments of Southwestern Portugal was discovered (1976) when J. KULLMANN, at the request of A. RIBEIRO, did a preliminary determination of a *Clymenid* sp. recovered from black shales about 100 meters below a quartzite horizon east of the village of Carrapateira. This identification has been confirmed recently (J. OLIVEIRA *et al.* in prep.). PERDIGÃO (1978), on his schematic map of the Aljezur and the Bordeira areas, considered the shales and overlying quartzites as corresponding to the Tournaisian. Recent fieldwork by RIBEIRO *et al.* (in press a) and J. OLIVEIRA (unpubl.), recorded on the Carta Geológica de Portugal, scale 1/200,000, sheet 7, 1982, showed that these shales and quartzites are widely represented. They were included in the Tercenas Formation, to which a late Devonian age has been attributed.

Recent investigations on a coarse sandstone at the top of the Tercenas Fm, near Monte Novo (5 km north of the Bordeira village) discovered brachiopods which were studied by P. RACHEBOEUF (University of Brest). They are poorly preserved but the following species could be identified: *Rugosochonetes* sp.; cf. *Syringothyris* sp. and Spiriferacea indet. These fossils do not allow an exact determination of stratigraphic age, but they suggest the early Carboniferous. This means that the top of the Tercenas Fm probably belongs to the Carboniferous. This conclusion is subject to reservation owing to the poor quality of the specimens.

The Tercenas Fm is composed of black shales at the lower part of the exposed succession (its base in unknown) grading into silty burrowed beds and quartzites. These quartzites display similar sedimentary structures to those of the Pyrite Belt. This fact led the author (J. OLIVEIRA 1982) to suggest that both were deposited in a widespread shalow-water environment.

4. CARBONIFEROUS

4.1. Volcano-Sedimentary Complex

General Framework

The Volcano-Sedimentary Complex (Pomarão Group of MAC GILLARY 1961b) comprises a variety of volcanic, sedimentary and mixed rocks which are arranged in northwest-southeast trending lineaments more or less parallel to the Ossa-Morena — South Portuguese zones boundary (this unit is also designated in other papers of this volume as 'volcanic-siliceous complex').

An outline of the stratigraphy of the Volcano-Sedimentary Complex has been given by SCHERMERHORN (1971). Since then, the work carried out by state and private institutions has solved some of the stratigraphic, petrological and tectonic problems.

As is common in a volcanic area, the Volcano-Sedimentary Complex of the Pyrite Belt (VS for short) shows a variety of lithologies and changes of facies making it almost impossible to sum up the stratigraphic succession in a single column. In order to offer the reader a general view of the VS assemblage, seven stratigraphic colums of the best studied areas are discussed here (Fig. 3).

Since the petrology, mineralisation and tectonics of the VS are described in other contributions (this volume) this discussion will focus on the stratigraphic and sedimentological items.

The Pomarão Anticline provides the more easily studied and best known stratigraphic sequence (VAN DEN BOOGAARD 1967). Here, the sequence comprises three acid volcanic episodes alternating with tuffites and terrigenous sediments. Near the base of the succession intrusive sills of diabase and keratophyres also occur (according to MUNHÁ 1983 — this volume, the kerathophyres are metandesites). The first acid volcanism (Cerqueirinha Fm) largely displays tuffs and minor agglomerates of dacitic to keratophyre composition with a thickness of about 40 m. The Touril Fm, which overlies the first volcanic episode is a mixed volcanic and sedimentary unit comprising the following members:

— Lower-Phyllite Member, composed of 50 m of black to grey shales and siltstones with chert lenses and tuffites interbedded in the lowermost part;

— Quartzite Member, a 20 m thick unit with impure sandstones, silstones and shales, and many interbedded iron and manganese carbonate nodules and lenses;

— The 10 m thick Tuffaceous Horizon of VAN DEN BOOGAARD, comprising acid tuffs and tuffites, which proved to correspond to an important acid volcanic episode in Spain (Tharsis and Calanas areas);

— Middle Phyllite Member of black and pink siliceous shales, again with iron and manganese carbonate nodules, and a thickness of about 80 m;

— Red Phyllite Member (50 m) with red to purple shales and radiolarites grading lateraly into green and pale siliceous shales, also with interbedded manganese nodules and lenses (in the following pages the term 'red shales' must be understood as including all the lithologies above described);

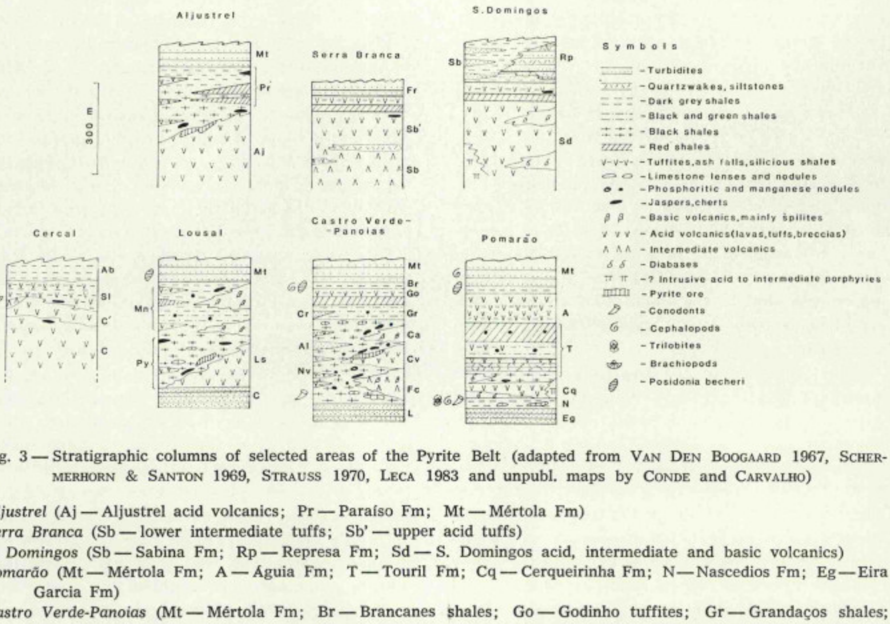


Fig. 3 — Stratigraphic columns of selected areas of the Pyrite Belt (adapted from VAN DEN BOOGAARD 1967, SCHERMERHORN & SANTON 1969, STRAUSS 1970, LECA 1983 and unpubl. maps by CONDE and CARVALHO)

Aljustrel (Aj — Aljustrel acid volcanics; Pr — Paraíso Fm; Mt — Mértola Fm)
Serra Branca (Sb — lower intermediate tuffs; Sb' — upper acid tuffs)
S. Domingos (Sb — Sabina Fm; Rp — Represa Fm; Sd — S. Domingos acid, intermediate and basic volcanics)
Pomarão (Mt — Mértola Fm; A — Águia Fm; T — Touril Fm; Cq — Cerqueirinha Fm; N — Nascedios Fm; Eg — Eira Garcia Fm)
Castro Verde-Panoias (Mt — Mértola Fm; Br — Brancanes shales; Go — Godinho tuffites; Gr — Grandaços shales; Ca — Casevel basic volcanics; Cv — Castro Verde acid volcanics; Fc — Monte Forno da Cal intermediate to basic volcanics; Nv — Neves shales; Al — Algaré Quartzites; Cr — Curral quartzites; L — Lançadoiras quartzites)
Lousal (Mt — Mértola Fm; Ls — Lousal acid and basic volcanics; Py — Pyrite Fm; Mn — Manganese Fm; C — Corona Fm)
Cercal (C — Cercal lower acid volcanics; C' — Cercal upper acid volcanics; Sl — São Luís Fm; Ab-Abertas shales)

— The third acid volcanic episode is represented by the Aguia Fm, a 100 m thick pile of reworked green and yellow acid tuffs and tuffites with some siliceous shale horizons in between. The Aguia Volcanics mark the end of volcanic activity and give way to an important turbiditic input into the now quickly subsiding basin.

Away from the Pomarão Anticline the VS succession changes (Fig. 3). First to be examined is the area where the base and top of the succession has been established, i.e. the Castro Verde-Panoias lineament and the Lousal Anticline. Along the Castro Verde-Panoias lineament (ALBOUY *et al.* 1981; LECA 1983) only one full acid volcanic episode, comprising acid tuffs and rhyodacites, is known. This episode (the Castro Verde Acid Volcanics) is underlain by the Monte Forno da Cal intermediate to basic volcanics, composed of andesites and spillites, and overlain by the Casevel basic volcanics (spilites, tuffs and minor agglomerates). Basic volcanics also occur below the first acid volcanics in the Montinho old mine area (LECA 1983), to a certain extent also in the Pomarão Anticline, and in some Spanish localities (Calanas, Sotiel, Ververde del Camino, etc.). The Casevel basic volcanics may have their equivalents in the Lousal spilites and diabases, and although not represented in the Pomarão Anticline they are also known from Spain (Tharsis, Calanas, etc).

The volcanics of the Castro Verde-Panoias lineament and the Lousal Anticline interfinger with black nodular shales (nodules of phosphatic limestone and manganese), siliceous shales, tuffites and rare limestone lenses. In the Castro Verde-Panoias lineament some sandstones (the Algaré and the Curral quartzites) also occur. These represent clastic incursions in what appear to have been a quiet depositional environment. Further north and eastwards the sandstones are even more widespread. This makes their stratigraphic position vis-à-vis the PQ quartzites open to many questions. This important point will be discussed later on.

According to LECA (1983) the red shales/Aguia Fm couple of the Pomarão Anticline has its equivalents in the Alcoutim Anticline, and in the red shales/Godinho tuffites of the Castro Verde-Panoias lineament. The same author also considers that the red shales may correspond to an environmental turning point from deep reducing to shallow oxidising environments in the basinal area and that the tuffs and tuffites may represent a distant manifestation of an important third volcanic episode in Spain (Gatos Volcanics). It has also been assumed that this volcanism should have taken place in a shallow water or even a subaerial environment, in order to explain its enormous extent all over the Iberian Pyrite Belt (LÉCOLLE 1974; LECA, 1983). Consequently, this couple has been regarded as an excellent marker horizon, particularly along the southern branch of the entire Iberian Pyrite Belt (ROUTHIER *et al.* 1980; LECA 1983).

Not withstanding the widespread occurrence of the red shales/tuffites couple, it seems questionable to assume that they represent a unique time-stratigraphic marker in such a large area. Abstraction made of tectonic complications, field geologist in Portugal are aware that, even at the larges scales, the couple cannot be used everywhere as a reliable marker horizon in geological mapping (see also CARVALHO *et al.* 1976a).

The oxidising environmental interpretation for the red shales depositional area does not explain some of the geological characteristics. If red shale deposition was only dependent upon an oxidising environment (and presumably on basinal highs), it is strange that they are not represented in the Aljezur-Carrapateira area which appears to have been a local high for a long time and where black shale sedimentation took place. The present author finds it difficult to regard these red shales as representing a normal pelagic sedimentation. Although their petrology and geochemistry is not well known, it may be admitted that they have some link with volcanic activity, assumed that at the time of volcanic activity, basinal irregularities, that is highs and depressions, did exist. On the relatively shallow oxidising areas (the highs) the pelagic sediments deposited in a volcanically contaminated environment would become red in colour; conversely pelagic sediments deposited in the reducing environment of the depressions (even though volcanically contaminated) would not become red. This possible distribution would explain the lateral gradation observed from red to black shales (as for instance near the Aljustrel volcanic center — V. OLIVEIRA SFM unpubl. Report 1972). However one problem emerges: why didn't the shales deposited in the Cercal volcanic area, which has been interpreted as a longstanding basinal high, become red? Does this mean that the volcanic activity took place here in deeper water? This question still remains unresolved and must be investigated further.

In the northern branch of the Pyrite Belt, between S. Domingos and Aljustrel, the VS stratigraphy is more poorly understood. In the author's opinion this is due to four main reasons: 1 — the Devonian basement is only well dated near the town of Mértola, an area highly affected by tectonic overthrusts; 2 — In what seem to be the main volcanic centres viz. the areas of S. Domingos, Serra Branca, Albernoa and Aljustrel, the base of the succession is unknown; 3 — The tectonic deformation along this northern branch is stronger than in the southern one; 4 — The clastic sedimentation and the lateral facies changes also appear to be more important.

The best studied stratigraphic succession is that of the Aljustrel area (Fig. 3). Here, the acid volcanics form the lowermost unit and their base has never been reached, even by boreholes. The volcanics were locally divided into four lithofacies (SCHERMERHORN & STANTON 1969): Megacryst tuffs, Green tuffs Mina tuffs and Felsites. These divisions were established on a petrological basis, that is megacrysts of potash feldspar and quartz in the Megacrysts/Green Tuffs and only albite megacrysts in the Mina Tuffs/Felsites. Overlying but also interfingering with the upper volcanic strata, there is the Paraíso Fm, composed of jaspers, black and grey shales, red shales, tuffites, and siliceous shales.

Some kilometres northeast of Aljustrel SCHERMERHORN & STANTON (1969) described the following conformable lithostratigraphic units: the Gomes Fm, composed of phyllites, quartzites, quartzwackes and silstones, the base of which is unknown; the Seixo Fm, composed of phyllites, quartzites, greywackes, siliceous slates and lenses of chert and jasper, whith a total thickness of approximately 300 m; the Vale de Água Fm, a succession of shales and a minor amount of greywackes, lenses of chert and manganese ores, with a total thickness of 400 m. The Gomes Fm lithologies were considered to be laterally equivalent to the PQ, and therefore of late Devonian age: both the Seixo and the Vale de Água Fms were regarded as lateral equivalents of the Aljustrel VS assemblage. This stratigraphic succession has been disputed by L. CONDE (SEREM [1], SMMPP [2], SMS [3] unpubl. Report 1975) to whom the Gomes Fm is the uppermost regional unit, and therefore a lateral variation of the upper Visean Culm facies (the Mértola Fm of the present paper). A recent re-examination of this question leds to the following conclusions: (1) the separation of the Gomes Quartzites from the Seixo Fm is difficult to establish, owing to the similarity of their lithologies, a situation quite common along the northern branch of the Pyrite Belt; (2) besides the previously described lithologies, the Seixo Fm in places shows red shales interbedded between the quartzwackes, a stratigraphic setting very close to that of the Represa Fm of the S. Domingos region (see below); (3) the tectonic deformation of the quartzwackes and associated lithologies (the Gomes Fm in the sense of CONDE) is much stronger than that of the upper Viséan Culm facies thus making their lateral equivalence unlikely.

The Serra Branca Antiform, north of Mértola provides a slightly different stratigraphic succession (L. CONDE-SEREM, SMMPP, SMS, unpubl. Report 1976). Below the tuffites/red shales couple (Fig. 3), this author recognised two volcanic episodes: the upper one being composed mainly of acid tuffs and the lower one comprising principally intermediate tuffs. Inter-

(1) SEREM = Société d'Etudes, de Recherches et d'Exploitations Minières, S. A.

(2) SMMPP = Sociedade Mineira e Metalúrgica de Peñarroya Portuguesa, Lda.

(3) SMS = Sociedade Mineira de Santiago.

bedded within the lower volcanic episode a thin shaly-quartzite horizon occurs. The base of this succession has not been seen.

The VS assemblage of the S. Domingos region, as it has been defined most recently (D. CARVALHO SFM unpubl. mapping 1976) also deserves attention, particularly because it is open to comment. CARVALHO described a succession comprising: a main acid volcanism (rhyolites, tuffs and agglomerates), intermediate? to acid intrusive porhyries, basic volcanics (spilites and diabases), red shales and tuffites. Overlying the volcanics two laterally equivalent units were mapped viz. the Sabina Fm, composed of shales and quartzites, and the Represa Fm, with shales and quartzwackes. These two units were considered to be conformably overlain by the Gafo Fm, a turbiditic sandstone/shale unit to which reference has been made (see Devonian basement). According to the same author, the volcanics and the Sabina/Represa Fms are of Tournasian age and the Gafo Fm of Viséan age. If this stratigraphic sequence is compared with those already discussed, the position of the Sabina quartzites and the Represa quartzwackes seems surprising. Indeed, in the above described VS stratigraphic sequences and also in most of the previously published work, such lithologies were almost always considered to belong either to the Devonian PQ or as lateral equivalents of the VS. The Represa Fm shows that there are red and green shales of the VS type interbedded in the quartzwackes. This fact and the tectonic deformation of both the quartzwackes and quartzites (two folding phases as in the VS) support the idea that these lithologies are lateral equivalents of the volcanics. However, based on presently available maps of the Pyrite Belt, there are also some reasons in favour of CARVALHO's interpretation. These lead us to the quartzite problem, which is discussed later.

The last local VS succession to be discussed here is that of the Cercal Anticline. It was only in 1976 that CARVALHO correlated this sucession with the classic Pyrite Belt stratigraphy. This author recognised the following units from bottom to top (Fig. 3): two acid volcanic episodes composed mainly of tuffs and felsites, intermediate to basic tuffs, and lenses of jasper; the S. Luís Fm, a mixed unit composed of tuffites, felsites and terrigenous shales. The base of the succession is unknown. The brachiopods determined by PAECKELMANN (*in* QUIRING 1936) were recovered from the top of the acid tuffs and from the lower part of the S. Luís Fm. The studied fossils (*Spirifer verneuilli* var. *archiaci* MURCHISON and *Productella caperata* SOWERBY) indicated a Strunian (latest Devonian) age. This means that the bulk of the volcanic pile is older than Strunian. In the southern termination of the Cercal Anticline, KLEYN (1960) reported metamorphic orthoquartzites either interbedded or orverlying the acid metavolcanics. Most of these quartzites were considered by CARVALHO (1976) as sheared acid metavolcanics.

In between the volcanic centres of the Pyrite Belt the sedimentary rocks (shales, siltstones, limestone lenses, quartzites, quartzwackes) are more common.

However their stratigraphic position relative to the PQ lithologies has proved difficult to establish.

The stratigraphic problems of the quartzites

The stratigraphic position of many quartzites (indeed orthoquartzites, impure sandstones and siltstones) has been a disputed matter over the past few years with particular regard to those of the northern branch of the Pyrite Belt. In the Pomarão Anticline PRUVOST (1912/13) has already demonstrated their late Devonian age, more recently confirmed by VAN DEN BOOGAARD (1963, 1967). SCHERMERHORN (1971) in his outline of the Pyrite Belt stratigraphy, also assumed that most of the quartzites are Famennian and that they are the older rocks in the succession. The same stratigraphic position has also been admitted by STRAUSS (1970) for the quartzites of the Lousal area.

In the 1970's this interpretation began to be disputed. Geological mapping carried out by private companies and the SFM (CARVALHO unpubl. mapping 1976) suggests that in some areas (Albernoa, east of Alcaria Ruiva hill, Serra Branca, S. Domingos, etc.) the quartzites appear to overlie the VS assemblage, and consequently could not be of late Devonian age. In Spain also, most of the quartzites of the northern zones of the Pyrite Belt were considered as overlying the volcanics, and even to be lateral equivalents of the upper Viséan Culm facies (ROUTHIER *et al.* 1980). The most striking example in Portugal is probably that of the Albernoa area. Here, the detailed geological map (Fig. 4) shows the quartzites surrounding the VS sequence, which appear to be in the normal stratigraphic order (acid volcanics overlain by jaspers, red and black shales). As such the quartzites appear to overlie the volcanics.

Mapping in the Mértola region (still in progress) points to some interesting conclusions which may help the understanding of at least some of these 'upper quartzites'.

a — The sedimentological features of most of the regional quartzites (massive to parallel-laminated beds, small scale and herringbone cross beds, slump balls, synsedimentary folds, etc.) are very close to those of the Famennian quartzites of the Pomarão Anticline. This fact, although not proving a similar age (unfortunately no fossils have been found in the 'upper quartzites') show that both quartzites were deposited in the same kind of sedimentary environment.

b — The imbricate tectonic style locally demonstrated near Mértola (FANTINET 1971; CARVALHO SFM unpubl. map 1975) now appear to have a much wider regional significance possibly on the scale of the entire Pyrite Belt (RIBEIRO *et al.* in press b; see also RIBEIRO & SILVA 1983 — this volume). According to SILVA (*in* J. OLIVEIRA & FLORIDO in press) the quartzites which apparently overlie the Serra Branca Volcanics are allochthonous, i.e. they have been transported along an overthrust plane running at its base. Therefore, these quartzites may not be taken as being stratigraphically above the volcanics.

Allochthonous successions are also known in many other places (Aljustrel and Neves-Corvo mines, Penilhos and Sr.ª do Amparo hills, south of S. Domingos, etc.). Such tectonic complications should be kept in mind before assuming a definitive stratigraphic position for the quartzites. If the conclusions reached in the Mértola region, and particularly for the Serra Branca antiform, are to be accepted, the apparent stratigraphic position of many of the upper quartzites could be explained by such tectonic emplacements. This could be the situation in the case of the Albernoa quartzites. In the S. Domingos area tectonic overthrusts could also have played an important role.

The author is aware that this is a delicate question and that simple generalisations can lead to misleading interpretations. This means that each geological situation must be carefully investigated.

The above considerations do not preclude the existence of quartzitic horizons younger than late Devonian. Indeed, it seems reasonable to infer that the clastic sedimentation went on in VS times (early Carboniferous) and consequently there are quartzites which are contemporaneous with the volcanics (as reported by LECA in the Castro Verde area; see also geological map in appendix). However, as the volcanic activity became more important, the basin paleogeography and its depositional environments apparently changed in such a way as to reduce the clastic input. The quartzites became finer and more thinly bedded, structureless, as isolated intercalations in the shales, and, in places, mixed with volcanic material. Although not yet proved, the existence of gravity quartzite slumps cannot be excluded.

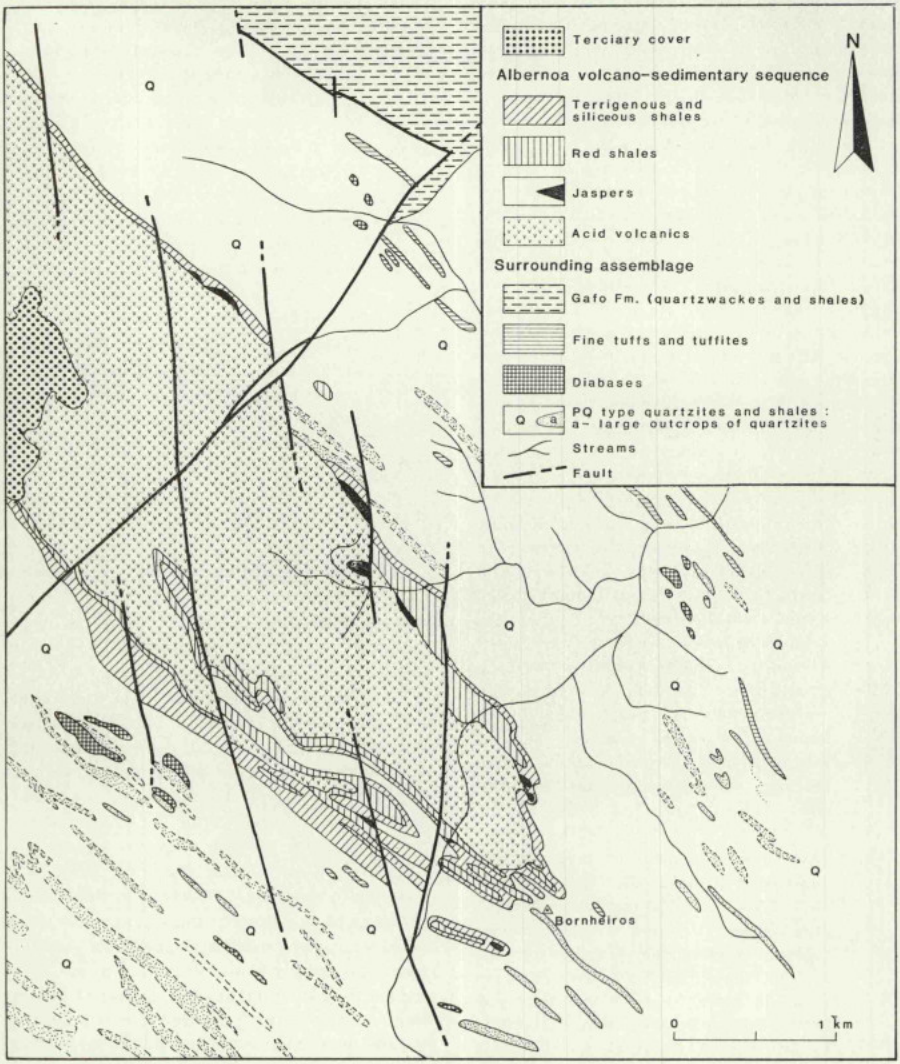


Fig. 4 — Geological map of the southern termination of the Albernoa Antiform, adaptated from unpubl. map by CONDE (SEREM, SMMPP, SMS)

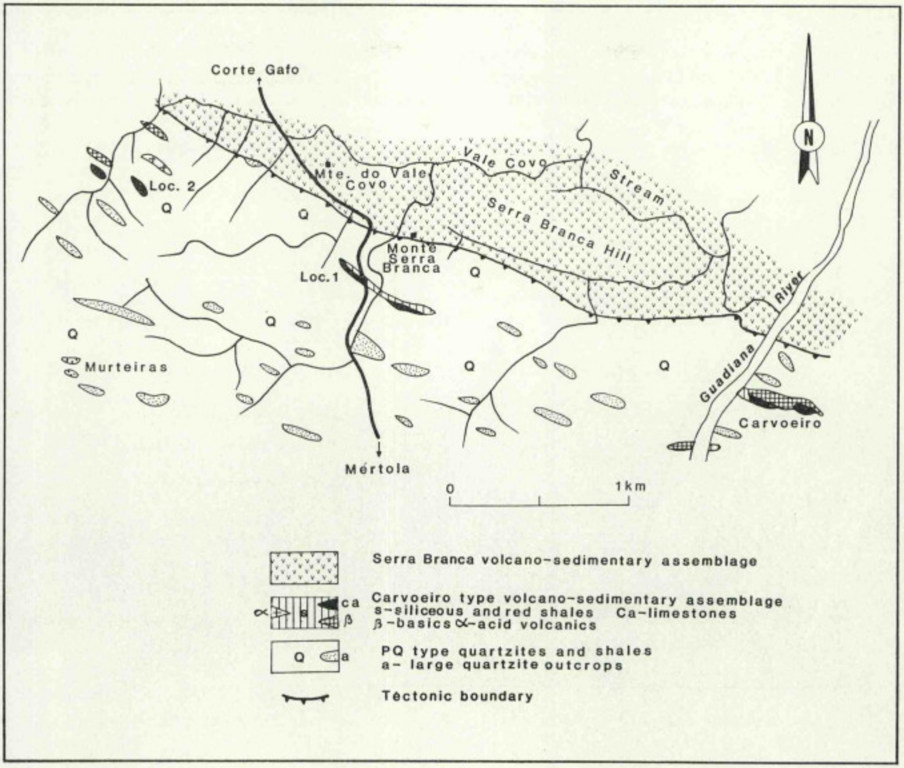


Fig. 5 — Localities with limestone bearing conodonts, south of the Serra Branca Antiform

The stratigraphic position of many quartzitic horizons is surely one of the most controversial subjects of the Pyrite Belt geology. Surprisingly, in the published work it has been neglected, and only (CARVALHO 1979; J. OLIVEIRA 1982) has it been referred to briefly. The aim of the above discussion is to call the reader's attention to this question, which is important both for its stratigraphic and palaegeographic implications as well as in base metal exploration.

Age of the Volcano-Sedimentary Complex

As seen before, the lowermost well dated rocks of the Pyrite Belt are in the upper levels of the PQ Formation, that is the Monte Forno da Cal limestones (early Famennian, VAN DEN BOOGAARD & SCHERMERHORN 1981) the Mértola limestone nodules (middle to upper Famennian, FANTINET *et al.* 1976) and the Pomarão limestone lenses (upper Famennian, VAN DEN BOOGAARD 1963). The overlying volcanics in these places have their base few metres above the dated horizons, and the boundary seems conformable. As seen before the bulk of the volcanic pile in the Cercal Anticline, is regarded as older than Strunian by CARVALHO (1976).

northern region of the Pyrite Belt, including its lower boundary, is still dubious, and the Strunian age admited for the upper levels of the Cercal volcanics, based only on two brachiopod species, needs better palaeontological

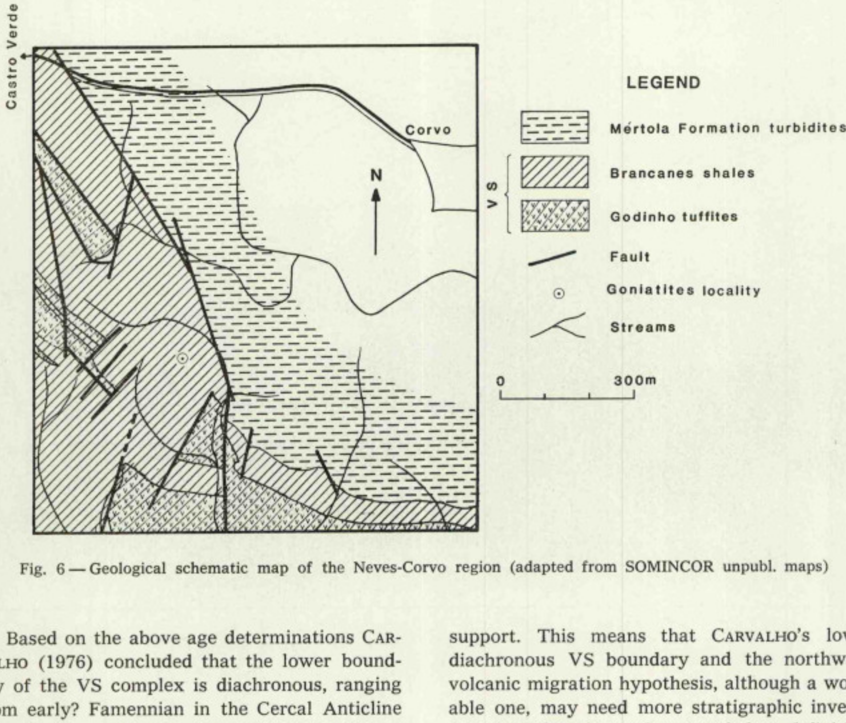


Fig. 6 — Geological schematic map of the Neves-Corvo region (adapted from SOMINCOR unpubl. maps)

Based on the above age determinations CARVALHO (1976) concluded that the lower boundary of the VS complex is diachronous, ranging from early? Famennian in the Cercal Anticline to early Tournaisian in the northern regions of the Pyrite Belt. This diachronism has been considered by this author as being the result of a volcanic migration towards the north, probably linked to an active northward dipping subduction zone.

Conodont dating from Monte Forno da Cal, Mértola and Pomarão limestones indeed point to a certain difference in age for the top of the PQ. However, outside the Pomarão Anticline, the full VS stratigraphic succession of the

support. This means that CARVALHO's lower diachronous VS boundary and the northward volcanic migration hypothesis, although a workable one, may need more stratigraphic investigations before it can be regarded as definitive.

In Spain the volcanics were considered as unconformably overlying the PQ quartzites (ROUTHIER *et al.* 1980). In Portugal this boundary has always been taken as conformable, and this is also the view expressed here.

From the VS itself age determinations are still scarce. Besides the Carvoeiro limestones (VAN DEN BOOGAARD 1963) which yielded conodonts of late Tournaisian age (*anchoralis* zone), only two more limestone lenses, very poor in conodonts, were recently dated (Fig. 5). The following faunas (kindly determined by W. EDER, Gottingen University) were recorded: from loc. 1 *Gnathodus semiglaber* BISCHOFF and *Gnathodus pseudosemiglaber* THOMPSON & FELLOWS, indicating an early Viséan age; from loc. 2 *Gnathodus semiglaber* BISCHOFF and *Gnathodus cuneiformis* MEHL & THOMPSON, indicating the upper Tournaisian (*anchoralis* zone). The limestones of loc. 1 are closely associated with silicious and red shales, those of loc. 2 appear as isolated outcrops, but in their northern area there are acid tuffs and basic rocks. As already pointed out by MAC GILLAVRY (1961b), the Carvoeiro limestones are also associated with siliceous shales, tuffites and basic rocks, appearing in the core of a small syncline overlying PQ type quartzites and quartzwackes. The three dated limestone lenses seems to be linked with the same lithostratigraphic horizon. It is worthwhile to note that limestone-bearing conodonts of early Viséan age were also identified from the Sotiel mine, Spain (VAN DEN BOOGAARD & SCHERMERHORN 1975).

The boundary between the VS assemblage and the overlying flysch turbidites of the Mértola Formation (see below) is gradual in many places. In the Pomarão Anticline, the Águia Vol-

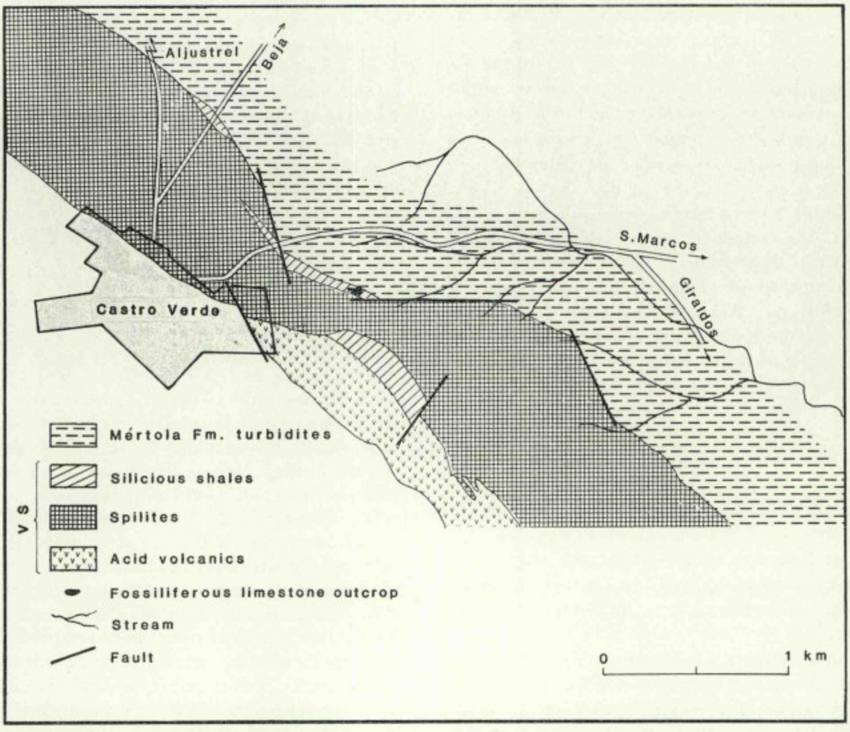


Fig. 7 — Localization of the limestone bearing conodonts, near Castro Verde (geology adapted from LECA 1983)

canics are conformably overlain by a thin bedded turbiditic unit (the Phyllite Member of VAN DEN BOOGAARD 1967), 20 m thick, and yielding *Posidonia becheri* BRONN. At the southeastern termination of the Castro Verde-Panoias lineament the VS grades upwards into the flysch facies through the Brancanes Shales, 20 m thick. These Brancanes Shales contain abundant squashed remains of *Posidonia becheri* BRONN and goniatites. One goniatite species recorded from the top of these shales (Fig. 6), as identified by C. H. T. WAGNER-GENTIS, proved to be *Goniatites hudsoni* BISAT, indicating the upper B_2 zone of the Viséan.

Two kilometres east of Castro Verde (Fig. 7) a grey to pink limestone lense (kindly shown by V. OLIVEIRA), which occur between the Castro Verde spilites and the Mértola Fm turbidites, yielded the following conodont species: *Streptognathus?* sp., *Gnathodus bilineatus* ROUNDY and *Gnathodus claviger*. This fauna indicates the upper Viséan *bilineatus* zone (determinations by W. EDER).

In the S. Francisco da Serra Anticline, southwest of Grândola, *Posidonia becheri* also occurs in the VS/flysch facies transition.

In the Cercal Anticline the upper levels of the S. Luís Shales did't supply any fossils. However in the overlying Abertas Shales (which represent the basal part of the flysch sequence) *Cravenoceras* sp. has been recorded (J. OLIVEIRA *et al.* 1979).

In the northern branch of the Pyrite Belt the VS/Mértola Fm boundary is frequently a tectonic one, but *Posidonia becheri* is found interbedded in the flysch very close to the boundary.

The above information indicates that the upper boundary of the VS is near the base of the Upper Viséan, probably close to the upper B_2 goniatite zone or *G. bilineatus* conodont zone.

In conclusion, the age of the VS of the Pyrite Belt, based on presently available data, ranges from middle Famennian to early late Viséan. A similar stratigraphic range has also been suggested by CARVALHO (1976).

4.2. Baixo Alentejo Flysch Group

The Baixo Alentejo (BA) Flysch Group is a major stratigraphic unit which extends across more than half of the South Portuguese Zone depositional area. A general description of the BA Flysch Group stratigraphy and sedimentology has been given by J. OLIVEIRA *et al.* (1979). In this paper only a synthesis of the most significant features is presented, with emphasis on recent stratigraphic and sedimentological advances.

The BA Flysch Group is composed of three lithostratigraphic units; Mértola, Mira and Brejeira Formations. All three formations trend northwest-southeastwards, more or less parallel to the Pyrite Belt and the Ossa Morena/South Portuguese zones boundary. The age of the three formations is late Viséan, latest Viséan-Namurian and middle Namurian to early Westphalian, respectively; they young southwestwards. This fact and their wholly turbiditic sandstone/shale character clearly record the progressive development of the sedimentation area.

4.2.1. Mértola Formation

Sedimentology

In terms of WALKER & MUTTI's (1973) terminology, the main facies recognised are: turbidite facies type C (mostly greywackes and shales) and facies G (mostly shales). However, other clastic resedimented facies are also represented, namely facies type A and B (massive sandstones, pebbly sandstones, microconglomerates), facies type F (conglomerates, breccias, pebbly mudstones, olistolits) and facies type E (thin-bedded structureless sandstones).

Most of the type C sandstones show the typical sedimentary features of the classic turbidites (WALKER 1978), that is Bouma divisions, sole marks, bedding regularity, etc. The individual sandstone beds have prevailing A→B Bouma divisions (although other divisions also occur), an average thickness around 30 cm (indeed it ranges from several centimetres to more than

1 metre), variable sorting and grain sizes. The thickest beds appear frequently amalgamated, with dispersed mud clasts inside.

Sandstone types A and B are usually associated with the amalgamated beds, and in places their thickness can be over 3 m.

The pelitic facies type G comprises shales, siltstones, thin limestone lenses and nodules and thin-bedded greywackes. On a whole, these lithologies define 'interturbiditic' horizons with a thickness of 30 m or more. In association with this facies rare goniatites and more frequent *Posidonia becheri* occur. Recent mapping in the Mértola region, still in progress, proved that many of these 'interturbiditic' horizons have a lateral continuity of several kilometres to tens kilometres and when mapped on a large scale they prove useful for an understanding of the flysch geology. Such an 'interturbiditic' horizon on the southern edge of the Mértola Fm, with rich goniatite faunas, could be followed in more than one hundred kilometres across the sedimentary basin and has been used as a transitional marker bed between the Mértola and Mira Formations (J. OLIVEIRA *et al.* 1979; J. OLIVEIRA & WAGNER-GENTIS in press).

The resedimented facies type F appears intercalated in the classic turbidites, either as isolated sedimentary incursions or associated with fining and thickening upward turbidite cycles. Because the sedimentological and tectonic significance of these facies has been open to question, they will be discussed below.

The facies type E appears randomly associated with all the other types of facies, particularly with thin-bedded turbidites.

Following RICCI-LUCCHI's (1975) terminology the Mértola Fm turbidites and associated facies appear vertically organised either in fining or thickening upward cycles (simple or multiple), in symmetric cycles, or without any visible vertical organisation. The turbidic cycles define in places fining an thickening upward megasequences. These sedimentological characteristics can be observed throughout the Mértola Fm depositional area.

The petrology of the Mértola Fm turbidites and associated facies has been described extensively by MAC GILLAVRY (1961b), VAN DEN BOOGAARD (1967) and SCHERMERHORN (1971). Most of the sandstones are albite-lithic greywackes containing abundant clasts of albite, quartz, rock fragments, mica flakes, and lesser potash feldspar grains which are set in a sericite-clorite-calcite matrix. The main detrital accessories are epidote, sphene, zircon, apatite, leucoxene and tourmaline.

Important to note is the composition and the percentage of the greywacke rock fragments. These comprise volcanics (felsic volcanics, acid tuffs and rarer intermediate tuffs, spilites and diabases), cherts (grey and black), jaspers, siliceous shales, quartzites, quartzwackes, siltstones, phyllites and aggregates of quartz. Many of these rock fragments match the PQ and VS lithologies, a fact already noted by MAC GILLAVRY. According to VAN DEN BOOGAARD (1967) at least 60 % of the rock fragments were derived from his Pomarão Group (PQ + VS); 20 % lack characteristic features and only 20 % have lithologies not recognised in the Pyrite Belt area.

The clasts of the conglomerates and breccias have practically the same composition as those in the greywackes, but are usually more rounded and more variable in size. However angular pebbles, cobbles and metric blocks are also common particularly as the northern branch of the Pyrite Belt is approached.

This summary of characteristics of the rock fragments shows that they were subjected to a previous erosion and transport mechanism. The question remains whence they came from, i.e. either from a rising part of the Pyrite Belt area MAC GILLAVRY (1961a) or from the Beja Massif (SCHERMERHORN 1971). Before discussing this important question, let us first analyse the Mértola Fm/VS boundary.

The basal part of the Mértola Fm is in many places (Alcoutim, Pomarão, Graça dos Padrões, etc.) composed of shales and thin-bedded turbidites grading downwards into VS shales and tuffites. This means that the basal contact is conformable. In some other places (Lousal, east of Castro Verde) the basal part of the Mértola Fm is composed of thick greywackes which are in direct contact with basic volcanics, or even with pyrite ore, as for instance in the authochthonous VS unit of the Neves-Corvo mine area. This latter type of boundary must be seen as a disconformity probably related with submarine erosion caused by the turbidity currents. In the northern branch of the Pyrite Belt the Mértola Fm/VS boundary is frequently tectonised and little stratigraphic information can be derived from it.

Summarising, it seems that both local conformities and disconformities occur. However, no clear unconformity has been proved until now.

Returning to the source area of the rocks forming detritus, two different views have been defended:

According to MAC GILLAVRY (1961a) and VAN DEN BOOGAARD (1967) part of the original sedimentary basin was subjected to rapid subsidence, whereas another part was rising. The rising part was subjected to erosion in order to provide the detritus for the subsiding area. As most of the rock forming fragments have the same composition as the PQ and VS lithologies, they may have come from the Pyrite Belt itself. In this case a stratigraphic gap or unconformity might exist in the rising area, a fact which has not yet been proven.

A different interpretation has been presented by SCHERMERHORN (1971). According to this author the Mértola Fm/VS boundary is everywhere conformable and, as the Mértola Fm turbidites are deep water sediments, no part of the Pyrite Belt would have been uplifted and subjected to erosion above sea level. Consequently the detrital material of the Mértola Fm greywackes and conglomerates could not be derived from the Pyrite Belt area.

Palaeocurrent measurements in the Mértola Fm turbidites (FANTINET 1963; VAN DEN BOOGAARD 1967; FRISCHMUTH 1968), although showing a complex pattern of current directions, suggest that the flow came predominantly from the northern quadrants. It is worthwhile to note that the flysch-like sediments of the S. Iria and Terena Fms, near the Beja Massif, and the Gafo Fm turbidites show petrological characteristics close to those of the Mértola Fm.

Putting all the above features together SCHERMERHORN assumed that the source area of the Mértola, the S. Iria and Terena turbidites must has been the Beja Massif, an uplifted geanticline which was probably active from the upper Tournaisian onwards. In order to explain the amount of rock fragments of volcanic and related rocks in all the flysch units the same author concluded that a volcano-sedimentary complex similar in size if not larger than that of the Pyrite Belt, might have existed in the Beja Geanticline. This geanticline would have been surrounded by pebble beaches and shelves where the rock fragments obtained their present high roundness and sphericity indices.

Having presented these different interpretations, it seems appropriate to make some comments on this matter particularly since neither SCHERMERHORN nor MAC GILLAVRY answered some important questions:

1—MAC GILLAVRY's assumption that the Iberian Meseta, and consequently the Beja Massif, did not play an important role as a detrital source area cannot be accepted. As aforementioned the Terena and S. Iria flysch formations in the near vicinity of the Beja Massif, and the southward more distant Gafo flysch-like unit have close petrologic and sedimentological affinities with the Mértola Fm turbidites, and were probably laid down by the same type of sedimentary mechanism. Indeed, I do believe that some of the detritus of the Mértola Fm turbidites were transported by turbidity currents flowing from source areas north of the Pyrite Belt and probably from the Beja Massif. However, I agree with GILLAVRY's statement that a large amount of the Mértola Fm detritus was derived from the Pyrite Belt itself.

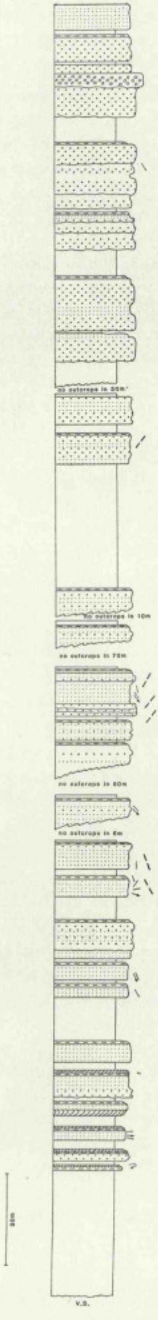
2—SCHERMERHORN's interpretation deserves also attention. Recent work on the Pyrite Belt volcanics (LÉCOLLE 1974; ROUTHIER *et al.* 1980; LECA 1983; MUNHÁ 1983-this volume) indicates that at least part of the volcanic activity took place in shallow water or even subaerial environments. As far as the turbidite deposition is concerned it is not universally accepted that they are always linked to deep water environments. But more important for this discussion is SCHERMERHORN's idea that the Mértola Fm/VS boundary is conformable and that no erosion existed below sea level. Disconformities are not uncommon in the depositional area but these may be related to submarine erosion caused by turbidity currents, a fact which is now currently accepted in the literature. Without rejecting the Beja Geanticline feeding role and the possible existence of a Tournaisian volcanic complex in that geanticline (but not with the postulated dimensions), it appears reasonable to infer that the Pyrite Belt area also acted as an important source of detritus for the Mértola Fm turbidites. Otherwise, it is difficult to explain the high percentage of Pyrite Belt type rock fragments in the Mértola Fm sandstones, conglomerates and breccias. Yet it is hard to explain so many poorly rounded and unrounded grains, pebbles, cobbles and blocks in these rocks (the Biguina conglomerate north of Aljustrel even contains manganese cobbles, which could not resist a long transport).

Furthermore, why are there no conglomerates and breccias in the S. Iria and the Gafo Formations when, on the contrary, they are so well developed along the southern edge of the north branch of the Pyrite Belt? It is suggested here that the Pyrite Belt topography was very irregular in VS times (if not already in PQ times) and parts of the VS assemblage, specially near the volcanic centres, were probably above wave base. As a result the rocks could be eroded and winnowed resulting in the development of small shelf areas. In the topographic depressions, a steady pelagic and ash fall sedimentation may have been occasionally disturbed by turbidity currents flowing in from the north. At the beginning of late Viséan times the entire area remained unstable with the approach of the southward prograding subsidence. Slumps and slides down slope led to the development of debris fow and strong turbidity currents, capable of producing submarine erosion prior to deposition. Immediately after subsidence the area became affected by southward directed compressional forces and synsedimentary faults may have developed into thrust planes prior to the principal folding. This means that sedimentation and tectonics are seen as a continuous process (J. OLIVEIRA 1982; RIBEIRO *et al.* in press b, see also RIBEIRO & SILVA 1983 — this volume).

Another important question is the sedimentological and tectonic significance of the conglomerates and breccias. In a previous paper (J. OLIVEIRA *et al.* 1979) these facies were considered as having been deposited in braided channels of the mid fan zone of a submarine fan. However, since some of the conglomerates occur near the VS/Mértola Fm boundary, which is developed as a thrust plane in many places, it has also been suggested (J. OLIVEIRA *et al.* 1979; RIBEIRO *et al.* in press b) that they could have been deposited in front of synsedimentary thrusts. Recent observations allow further comments on these ideas.

The geological map in the appendix shows that there are many conglomerate lenses, parallel to the strike, intercalated in the Mértola Fm turbidites. In light of the two hypothesis presented it is important to consider the conglomerates near the VS / Mértola Fm boundary. On the southwestern limb of the Pomarão Anticline, a detailed sedimentary log of the Formoa Conglomerate (Fig. 8) shows that it is normaly interbedded in the enclosing turbidites and that it lies more than 500 m above the conformable VS/Mértola Fm boundary (faults recorded in the log are transverse to the strike and no thrust fault has been detected). Consequently, this conglomerate may not be considered as being linked with a synsedimentary thrust plane.

Further northwest, in the Namorados area, west of Mértola, detailed mapping (Fig. 9) shows that the conglomerate horizons are cut out by the Pero da Vinha overthrust which brought the VS lithologies over the Mértola Fm turbidites. The thrust plane is presently folded and affected by one cleavage, and was therefore emplaced prior to the last main regional folding phase (see also CARVALHO *et. al.* 1976a).



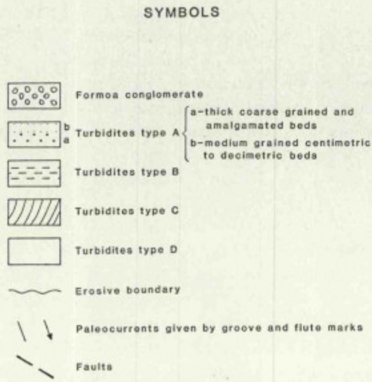


Fig. 8 — Sedimentary log of the lower part of the Mértola Fm turbidites and the Formoa Conglomerate, west of Pomarão Anticline. Represented in terms of percentage of A, B, C and D type turbides, as defined by CRIMES *et al.* (1974)

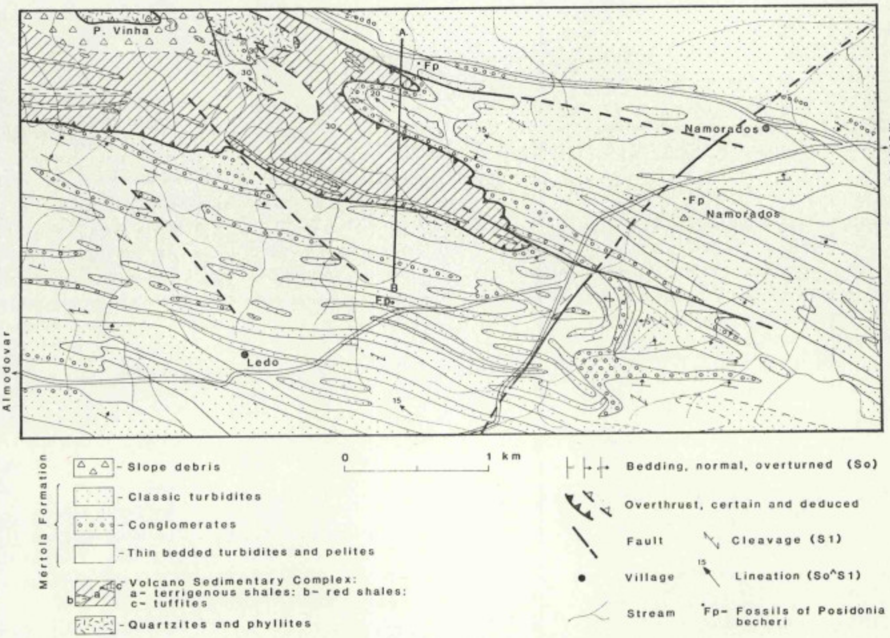


Fig. 9 — Detailed geological map of the Namorados region, west of Mértola.

Since the thrust plane cut across different conglomerate horizons, it is assumed that the conglomerates were already deposited when the allochthonous VS sheet arrived (Fig. 10).

In the São Marcos da Atabueira region the conglomerates and breccias are overlain by over 100 m of classic turbidites and the local conglomerate/turbidite sequence is facing northeast, i. e. towards the VS/Mértola Fm thrust boundary. Again, it is not easy to see the depositional links between the conglomerates and the thrust plane. A similar stratigraphic setting is observed in the Biguina Conglomerate, north of Aljustrel.

In conclusion, the close relationship between conglomerates and syn-sedimentary thrust planes is not obvious, if presently available information is considered. Petrological and sedimentological characters of the conglomerates, particularly of those close to the VS/Mértola Fm boundary, suggest that they may be related with slumping, slides and debris flows caused by tectonic instability on a sedimentary slope. To what extent this instability was already linked to compressional forces is not yet clear.

At first glance, and as mentioned before, the Mértola Fm turbidites and conglomerates were seen as mid-fan deposits of a submarine fan. Recent unpublished work by the present author indicates that this model is probably too simple to explain the entire depositional process of the Mértola Fm turbidites. Yet I do belive that most of the conglomerates were laid down in channels and that the positive and negative megasequences are related to sedimentary lobes and their progradation. However, no main feeder channel or canyon has yet been proven and with the present state of knowledge classic divisions of a submarine fan are not clearly recognised.

Stratigraphic range and fossils

In many places the basal part of the Mértola Fm consists of black to dark grey pelites grading upwards into thin-bedded turbidites. Local lithological variations and their meaning have been discussed already. *Posidonia becheri* occurs commonly.

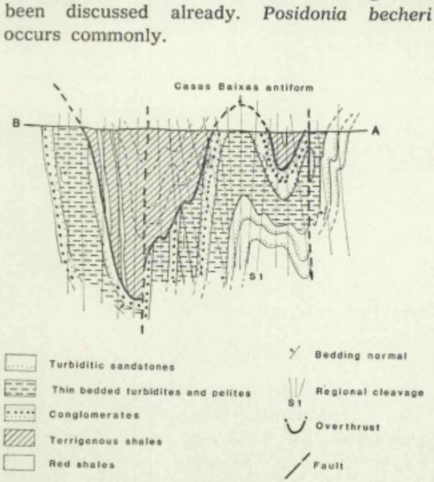


Fig. 10 — Interpreted cross section through the Pero da Vinha overthrust, Namorados region

The upper part of the Mértola Fm is composed of classic turbidites and interbedded pelites. These pelites yielded *Goniatites falcatus G. granosus, Dombarites parafalcatoides, Lyrogoniatites,* etc. (see the Mértola/Mira Fm boundary below) indicating the highest Viséan. Overlying the upper Mértola Fm turbidites there is an interturbiditic marker horizon which is taken as representing the base of the Mira Fm.

The thickness of the Mértola Fm is not known due to the incomplete biostratigraphic control and tectonisation, but it is certainly in excess of 1000 m.

A review of published work shows that the Mértola Fm provided mainly *Posidonia becheri* and much rarer goniatites. The main collecting areas are: south of Grândola (COSTA 1943), Cachopo-Barrada-C. Serranos (DELÉPINE 1957; J. OLIVEIRA *et al.* 1979); Mértola (FEIO 1946a, b; BOUCKAERT *in* MAC GILLAVRY 1961b), Azinhal-Odeleite-Vaqueiros (SOUSA 1924; DELÉPINE 1957; FRISCHMUTH 1968). All these fossils were collected in single localities and the stratigraphic control has been strictly limited. The following faunas were described:

Goniatites: *G. crenistria* PHILLIPS var. *globoides* SCHMIDT, *G. crenistria* PHILLIPS var. *globostriata* SCHMIDT, *G. striatus* SOWERBY, *G. subcircularis* MILLER, *Lusitanoceras algarviensis* SOUSA, *G. myrtilensis* FEIO, *G. falcatus* ROEMER, *G. sphaericus* MARTIN, *G. maximus* BISAT, *G. granosus* PORTLOCK, *Hibernoceras carrauense* MOORE & HODSON, *Neoglyphioceras spirale* PHILLIPS.

Bivalves: *Posidonai becheri* BRONN, *Posidoniella* aff. *vetusta* SOWERBY, *Posidoniella* aff. *laevis* BRONN, *Carnevella membranacea* M'COY.

Trilobites: *Phillipsia* aff. *westphalica* HAHN.

Others: poorly preserved brachiopods, orthocerids and crinoids.

Plants: drifted remains of *Archaeocalamites* and *Mesocalamites*.

These fossils proved the existence of upper Viséan P_1 and P_2 goniatite zones or in terms of the German goniatite zones, Go_α, Go_β, and Go_γ.

4.2.2. **Mira Formation**

General characteristics of the Mira Fm were described by J. OLIVEIRA *et al.* (1979). Since then, little work has been done on this unit. In this paper only a brief synthesis of the main features of the unit will be presented, with reference to recent work carried out on its transition to the Mértola Fm.

Sedimentology

The Mira Fm is also composed of turbiditic sandstones and pelites but their internal features are less variable than those of the Mértola Fm. Following WALKER & MUTTI's (1973) terminology, the most widespread facies are classic turbidites of type C and pelagic facies of type G. Facies types A and B and thin organized conglomerate lenses are very rare and the facies type F (breccias, desorganized conglomerates, pebbly mudstones, etc.) is absent.

The classic turbidites outcrop south of the Mira/Mértola Fms boundary along a sedimentary belt 10 to 15 km wide. Along this belt the turbidites show the characteristic Bouma divisions but with a predominance of $B \rightarrow C$ divisions. Turbidite cycles are less frequent but in some points it is possible to recognize positive and negative sequences to which the thin conglomerate lenses and thick A, B type sandstones are related. The pelagic facies type G and associated thin-bedded turbidites are most common, particularly along a belt running from Cercal to Saboia, São Marcos da Serra, São Barnabé and the Caldeirão hill. These facies are composed mainly of millimetric to centimetric turbiditic beds and black to dark grey pelites. In places this facies interfingers with positive and negative sedimentary sequences which are thought to represent prograding sedimentary lobes. For instance such a negative sequence can be seen, 7 km south of Santana da Serra on the main road to São Marcos da Serra. The boundary between the Mira and Mértola Fm is transitional (see below). In the Cercal Anticline the base of the Mira Fm is composed by black and dark grey shales (Abertas shales) with a local conglomerate (Vale Longo) at its lower levels.

Near the Mira/Brejeira boundary the thin-bedded turbidites and shales grade into thicker centimetric to decimetric sandstones and the transition to the Brejeira Fm is gradual.

Palaeocurrent measurements are scarce and from what is known it seems that the flow directions were predominantly northwestwards.

The sandstone petrology is also poorly known. The rocks still have a greywacke composition but with fewer feldspars and lithic fragments and a higher percentage of quartz. This petrological composition is probably due to reworking before final deposition.

These sedimentological characteristics suggest that the Mira Fm turbidites were deposited far from the source area. The positive and negative sequences may represent lobes of the external fan and the pelagic facies point to deposition on the basin plain.

Stratigraphic range and fossils

At the base of the Mira Fm turbidites there is an interturbiditic horizon of variable thickness but never in excess of 100 m. This horizon comprises mainly dark grey pelites and thin-bedded turbidites. As mentioned before these lithologies were taken as a transitional marker horizon between the Mértola and Mira Fms, and can be followed from S. Tiago do Cacém to west of Azinhal, via Dogueno and Taipas (Fig. 11).

Recent work carried out between Dogueno and Cabaços (J. OLIVEIRA & WAGNER-GENTIS in press) proved that the lower beds of this marker horizon still contain upper Viséan goniatites and that the remaining beds are unfossiliferous. From the collecting localities (Fig. 11) the following goniatites species were determined: *Lusitanites subcircularis* MILLER, from loc. 1; *Dombarites parafalcatoides* RUZHENCEV & BOGOSLOVSKAYA from locs 4, 5, 6, 7, and 10; *Lyrogoniatites* sp. from loc. 5. This fauna indicate the upper P_2 Zone in the highest Viséan.

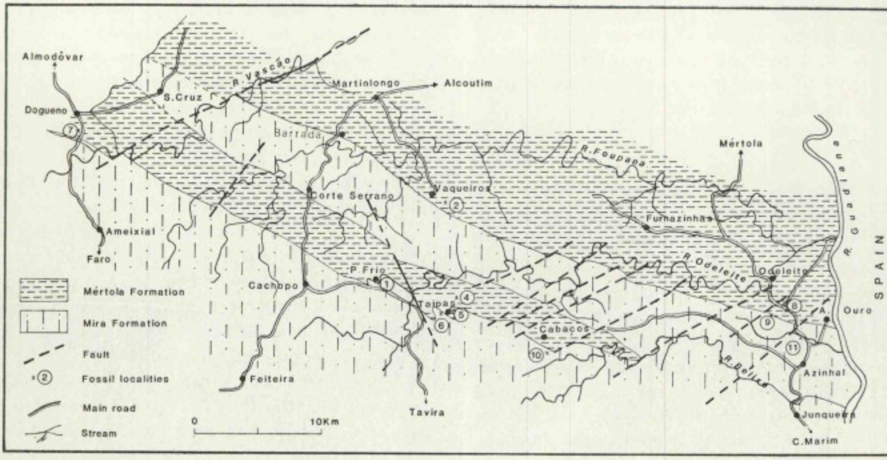


Fig. 11 — Upper Viséan goniatite localities, between Dogueno and Almada de Ouro (adapted from J. OLIVEIRA & WAGNER-GENTIS in press)

3 km north of this marker horizon, pelitic levels interbedded in classic turbidites, and outcropping between Barrada and Almada de Ouro, also yielded the same gonitatite species (locs 2, 8, 9). In addition the following goniatites species were determined: *Platygoniatites* sp. *'early form' Lyrogoniatites georgensis* MILLER & FURNISH, *Lyrogoniatites mutabilis* RUZHENCEV & BOGOSLOVSKAYA, *Lyrogoniatites* aff. *eisenbergensis* RUPRECHT, *Pachylyroceras cloudi* MILLER & YOUNQUIST, all from loc. 2; *Neoglyphioceras* cf. *gradatum* RUZHENCEV & BOGOSLOVKAYA from loc. 9. These fossils also indicate the upper Viséan P_2 Zone.

From these determinations it is concluded that the base of the transitional marker horizon and the northern Barrada-Almada de Ouro pelitic intercalations are of the same age. This repetition is seen as a result of the southward directed changes in turbiditic facies and the tectonic folding style (for details see op. cit.).

In the Cercal Anticline the base of the Mira Fm (the Abertas Shales of CARVALHO 1976), contain weathered siliceous nodules which yielded rare and poorly preserved goniatites of the Namurian E Zone (J. OLIVEIRA *et al.* 1979).

Stratigraphic research on the Mira Fm turbidites is still in its first stages. Until now the unit seems to be poor in fossils. Collecting northeast of Odemira, east and near the Mira/Brejeira formational boundary and southeast of Saboia proved the existence of the following goniatites (J. OLIVEIRA *et al* 1979): *Cravenoceratoides* cf. *edalensis* BISAT, *Cravenoceras* sp., *Eumorphoceras bisulcatum* GIRTY, *Homoceras* cf. *beyrichianum* KONINCK, *Reticuloceras compressum* BISAT & HUDSON, *Homoceratoides varicatus* SCHMIDT, *Bashkirites* sp., *Reticuloceras superbilingue* BISAT. Fossils other than goniatites are mainly indeterminable plant remains.

South of the Mértola/Mira formational boundary, i. e. towards the lower Mira Fm beds, only rare fossils have been found. Only recently (J. OLIVEIRA & WAGNER-GENTIS in press) *Girtyoceras modestum finale* RUZHENCEV & BOGOSLOVSKAYA, *Metadimorphoceras* sp.

and *Girtyoceras* sp. nov. were found near Ponte de Tabua (loc. 11, Fig. 11), 2 km north of the Azinhal village. This fauna indicate the lower Namurian E_1 Zone. In spite of the present paucity of fossils it is believed that further stratigraphic research will provide more Namurian fossils.

From the above faunal assemblage it is concluded that the age of the Mira Fm ranges from latest Viséan to late Namurian.

4.2.3. **Brejeira Formation**

The Brejeira Fm is the youngest unit of the Baixo Alentejo Flysch Group. After its stratigraphic definition (J. OLIVEIRA *et al.* 1979), very limited work has been carried out on this unit, as is the case with the Mira Fm.

Sedimentology

As with the other units of the BA Flysch Group, the Brejeira Fm is composed of turbidites and interbedded pelites. However, on the whole, the unit show marked differences with respect to the Mértola and the Mira Fms.

In a rough and tentative way the Brejeira Fm is divided into two sedimentary belts: a northern one, 5 to 10 km wide, runs immediately south of the Brejeira/Mira Fms boundary; the southern belt is represented by the remaining outcrops to the south. The boundary between these belts is ill-defined.

The sedimentary facies of the northern belt shows grey to whitish turbiditic sandstones with prevailing A→B Bouma divisions, massive sandstones and interbedded dark grey to bluish grey pelites. The individual sandstone beds have variable centimetric to metric thicknesses. Sole marks, particularly load casts, and less common flute and groove casts, are relatively frequent. Drifted plant remains are very common, both in the sandstones and in the interbedded pelites.

The interturbiditic horizons are made up of pelites and thin-bedded siltstones, usually 10 to 20 m thick.

The vertical organisation of the sandstones shows simple fining and thickening upward cycles but a random vertical organisation is also very common. The sandstone/shale ratio is > 1.

From a petrological point of view the sandstones are medium to coarse-grained impure quartzites, with a quartz content of over 80 %, few albite feldspars, some mica flakes and rare lithic fragments of sedimentary origin, which are set in a fine siliceous sericitic matrix. The massive sandstone are more mature and with an higher proportion of quartz grains. The siltstones are thin laminate rocks with fine quartz grains and clay material.

The southern sedimentary belt is much less known. The dominant sandstone facies are classic turbidites with predominant B→C Bouma divisions. Pelites and thin-bedded turbidites are better developed than in the northern belt and consequently the sand/shale ratio is < 1. Sole marks, consisting of groove and flute casts, are common as well. The average thickness of the individual sandstone beds is less than that of the impure quartzites, but in some places there are thicker sandstone beds associated with sedimentary cycles. These can be seen for instance in the Praia de Arrifana region. Here, in the cliff on the northern side of the beach, metric thick slumps, are interbedded in the sandstones (kindly shown by A. RIBIEIRO).

The sandstones have a greywacke composition, with quartz, albite, feldspars, mica flakes and rare mud clasts. Rock fragments of volcanic origin have not been seen. Near the Aljezur and Carrapateira antiforms the greywackes contain a high percentage of dark clay material.

Palaeocurrent measurements (about two hundred measured) in the sandstones of the northern quartzitic belt show a complex pattern of flow directions predominantly to the southeast and east and lesser to the southwest and north. Although still very scarce, these flow directions and the distinct sedimentary character of the sandstones suggest that the detritus is probably not derived from the north but from a source area somewhere in the present Atlantic ocean This possibility is reinforced by palaeocurrent measurements made between Arrifana and Praia do Castelejo, which show flow directions predominantly towards the northern quadrant (A. RIBEIRO pers. comm.). If so, the basin may have been supplied from both the northern and southern quadrants.

Stratigraphic range and fossils

The lithological transition between the Brejeira and the Mira Fm is gradual and the base of the Brejeira Fm is placed at the first impure quartzite beds. The boundary is well defined in the field but its biostratigraphic control needs further elucidation. South of the village of Nave Redonda the lowermost Brejeira beds yielded goniatites of the Namurian H Zone (*Homoceras beyrichianum*) these being the oldest fossils found in the unit. However, south of São Marcos da Serra, *Reticuloceras* sp. of the Namurian R_1 Zone has been found in the interbedded pelites of the quartzitic suite, in a position very close to the Mira/Brejeira Fms boundary. This may mean that the boundary is diachronous.

In the Aljezur region, the Brejeira greywackes rest on different lithostratigraphic units, either black shales of the Quebradas Fm (see the Carrapateira Group below), impure limestones and shales of the Murração Fm or even shales of the Bordalete Fm. Some of these boundaries may be related to local disconformities, such as near the Alcaria farm, 3 km northeast of Aljezur, where there is a conglomerate layer at the base of the flysch greywackes. In spite of this most of the abnormal boundaries are probably linked to local tectonic thrusts. In any case, *Reticuloceras superbilingue* BISAT of the Namurian R_2 Zone has been found in some outrops of the Quebradas shales, a few metres below the base of the Brejeira greywackes.

Further south, in the Bordeira Antiform area, the base of the Brejeira Fm is frequently tectonized, either by Hercynian thrusting or by late to post Hercynian normal faults. Nevertheless, at the southern termination of the antiform, the higher Quebradas Fm shales yielded *Gastrioceras* sp. probably of the G_1 (Yeddonian) upper Namurian.

The top of the unit is not known. *Gastrioceras listeri* MARTIN has been found in loose nodules 8 km northeast of Aljezur, where they were identified by DELÉPINE (1957). These are still the youngest fossil found in the sedimentary basin.

As in the other units of the Baixo Alentejo Flysch Group, the thickness of the Brejeira Fm is unknown. A thickness the order of 2,000 m is assumed.

From the Brejeira Fm the following goniatites have been mentioned (DELÉPINE 1957; J. OLIVEIRA *et al.* 1979): *Homoceras subglobosum* DOLLÉ, *H. beyrichianum* DE KONINCK, *H. henkei* SCHMIDT, *Reticuloceras reticulatum* PHILLIPS, *R.* cf. *circumplicatile* FOORD, *Homoceratoides divaricatus* HIND, *Ht. praereticulatum* BISAT, *Gastrioceras cancellatum* BISAT, *G. cumbriense* BISAT, *G. crenulatum* BISAT, *G. subcrenatum* SCHLOTHEIM, *G. weristerense* DEMANET, *G. listeri* MARTIN and *Anthracoceras* sp. The tentative identification of *Cravenoceras* sp. and *Eumorphoceras* sp. of J. OLIVEIRA *et al.* (1979) has not been confirmed by later research. These faunas indicate the Namurian H, R_1, R_2 and G_1 Zones and the lower Westphalian G_2 Zone.

Lamellibranchs attributed to *Pterinopecten* sp. are common and *Selenimyalina variabilis* has also been recorded.

Drifted plant remains, including logs of up to 50 cm lenght, are frequently met with. They are invariably in poor preservation. However, from a locality northeast of Portimão (the Vale de Corvos geographic marker), TEIXEIRA & PAIS (1976) obtained *Calamites (Crucicalamites)* cf. *cruciatus* STERNBERG, *Lepidodendron* cf. *obovatum* STERNBERG, *Sigillaria (Eusigillaria)* cf. *tessellata* BRONGNIART and *Syringodendron* sp. These flora was considered to be indicative of the Westphalian.

The fossils quoted show that the age of the Brejeira Fm ranges from middle Namurian (H Zone) to early Westphalian (G_2 Zone).

4.3. **Carrapateira Group**

The sedimentary facies included in the Carrapateira Group have deserved attention due to their distinct lithologies and faunal content, within the South Portuguese Zone. Recent research in the Bordeira and Aljezur Antiforms (J. Oliveira *et al.* in prep.; Ribeiro *et al.* in press a) clarified many aspects of the stratigraphic succession, particularly with regard to macrofauna. However, owing to the geological and palaeogeographical importance of these facies, further investigations are necessary, particularly on the microfaunas and the depositional environments.

The data presented here represent a preliminary synthesis of the work carried out by the above mentioned authors, with brief references to previous investigations.

The Carrapateira Group is composed of three lithostratigraphic units, the Bordalete, Murração and Quebradas Formations, the ages of which range from middle Tournaisian to late Namurian. The entire thickness is c. 250-300 m; this means that the facies are more 'condensed' than those of the almost coeval VS and Baixo Alentejo Flysch Group. The palaeogeographic significance of this change is briefly referred to later.

4.3.1. **Bordalete Formation**

This is a succession of dark grey shales with interbedded centimetric layers of calcareous siltstones and thin laminate black shales and siltstones towards the top. Calcisiltitic nodules and lenses, rich in pyrite, can be found through the entire lithological succession. In the Murração beach section the upper part of the unit shows sets of black shales and interbedded siltstones developing large scale cross bedding and small syn-sedimentary folds. The thin siltstone layers display also ill-defined graded bedding and soft sedimentary deformations (load structures). Thin pyrite layers and nodules also appear interbedded in the black shales. Meandering and simple vertical burrows are present locally.

A detailed log of this unit could not be prepared due to the absence of continuous exposure and good marker beds. Consequently, the approximate thickness of 100-200 m for this formation remains doubtful.

In previous work (Pruvost 1914; Sousa 1919/22; Delépine 1957) reference has been made to *Prolecanites algarviensis* Pruvost, *Pericyclus princeps* Koninck, *Pericyclus kochi* Holzapfel. Recent investigation also showed the presence of *Imitoceras rotatorium* (De Koninck), *Zadelsdorfia* cf. *bransoni*, *Pericyclus* cf. *blairi* Miller & Gurley and *Achegonus (Macrobole) drewerensis latipalpebrata* Osmolska. This fauna indicates the middle Tournaisian. These fossils were recovered from different localities with limited stratigraphic control. From regional stratigraphic correlations it appears likely that the upper part of the unit reaches an upper Tournaisian level. The lithological features of the Bordalete Fm point to a quiet, sometimes instable, reducing marine environment.

4.3.2. **Murração Formation**

The best exposed section of this unit crops out in the Murração beach section, southwest of Carrapateira village. The stratotypes of its two component members have been defined on this beach (Fig. 12).

Pedra das Safias Member

The stratotype is exposed in the Pedra das Safias salient, south of Murração beach. It comprises dark grey and black shales with interbedded, partly dolomitic, limestone beds and nodules; these are overlain by marly carbonate dolomitic limestones. The thickness is c. 27 m. The base of the unit is placed at the lowermost carbonate bed, close to a *Metacoceras* horizon, and the top at the first dark grey shale bed with *Posidonia becheri*. Its macrofauna is listed in Fig. 12. One of the most characteristic features of this unit is its richness in crinoids and, to lesser extent corals. A loose dark shale block containing trilobites belonging to *Tawstokia nasifrons* R. & E. RICHTER (lowermost Viséan) has been found near the lower boundary of the unit. This block is regarded as belonging to the lower levels of this member.

Elsewhere in the Carrapateira area, and presumably from the lower part of the Pedra das Safias Mb, *Cyathophilum mitratum* SCHLOTH. and *Caninia cornucopia* MICHELIN have been mentioned (SOUSA 1919/22).

The faunal content of the Pedra das Safias Mb indicate the lower and middle Viséan.

Vale Figueira Member

The stratotype of this member has been defined in the northern cliff of the Quebradas beach. The unit is composed of bioturbated dolomitic limestones, dark grey shales and coquinas, passing up into dolomitic limestones intercalated between dark grey and black shales. The top of the unit is tentatively placed at 5 m above three limestone marker beds which are alternating with dark and black shales yielding pyritized *Sudeticeras* sp.

The lithological succession, the faunal assemblage an the stratigraphic age are shown in Fig. 12.

From the Carrapateira region *Goniatites crenistria* PHILLIPS, *G. striatus* SOWERBY, *G. granosus* PORTLOCK, *G. subcircularis* MILLER, *Beyrichoceras* cf. *micronotum* PHILLIPS and *B.* cf. *obtusum* PHILLIPS have been identified (SOUSA 1919/22; DELÉPINE, 1957). From the upper Viséan limestone beds of this area poorly preserved foraminifers (*Saccammina* cf. *carteri*, *Endothyra* sp.?, *Lagena* sp.?, *Bigenerina* sp., *Textularia?* and *Fusulina* s. l.) have been referred by FLEURY (1924).

In the northern part of the Bordeira Antiform and in the Aljezur Antiform the Murração Fm has not been divided in component members. The unit is more shaly here and a division into members proved impracticable. The lithological composition of the Murração Fm in this area suggests a northeastward deepening of the sedimentary basin.

From the condensed character and lithological features of the Murração Fm it is inferred that the unit has been deposited in a open, at times reducing and saline, marine environment. The possibility of its deposition on a shelf close to a neighbouring platform is suggested here.

4.3.3. **Quebradas Formation**

The Quebradas Fm stratotype is in the Quebradas beach section (Fig. 12). It comprises black shales with interbedded carbonate layers, lenses or nodules, partly dolomitic, and two main horizons of this laminated shales and siltstones. Phosphoritic and manganese nodular layers intercalated in black shales, form a conspicuous feature of this unit. The stratotype thickness is c. 80 m.

Inland, in both the Bordeira and Aljezur Antiforms, the unit maintains the fundamental stratotype characteristics, although usually with a lesser thickness. However, in places thin ferro-manganesiferous lenses and nodules can be found. As with the Murração Fm, this unit also appears more shaly in the Aljezur Antiform, suggesting again deepening to the northeast. The black shales and the carbonate lenses and nodules show white or red-brown weathering, respectively.

In the stratotype section goniatites, although frequent, are poorly preserved, thus allowing only partial stratigraphic control of the unit stratotype (see Fig. 12). However in the Murração beach area, at some distance from the type section, and in the general area of the two antiforms, cravenoceratids, *Eumorphoceras bisulcatum* GIRTY, *Homoceras beyrichianum* DE KONINCK, *Reticuloceras* cf. *reticulatum* PHILLIPS, *Reticuloceras* cf. *superbilingue* BISAT etc., indicate the presence of all the Namurian E,H,R,G_1 Zones in Southwest Portugal.

The lithological and biostratigraphic caracteristics of the Quebradas Fm suggest that the unit has been deposited in a reducing marine environment. The laminated shale horizons may already represent the first entry of weak turbi-

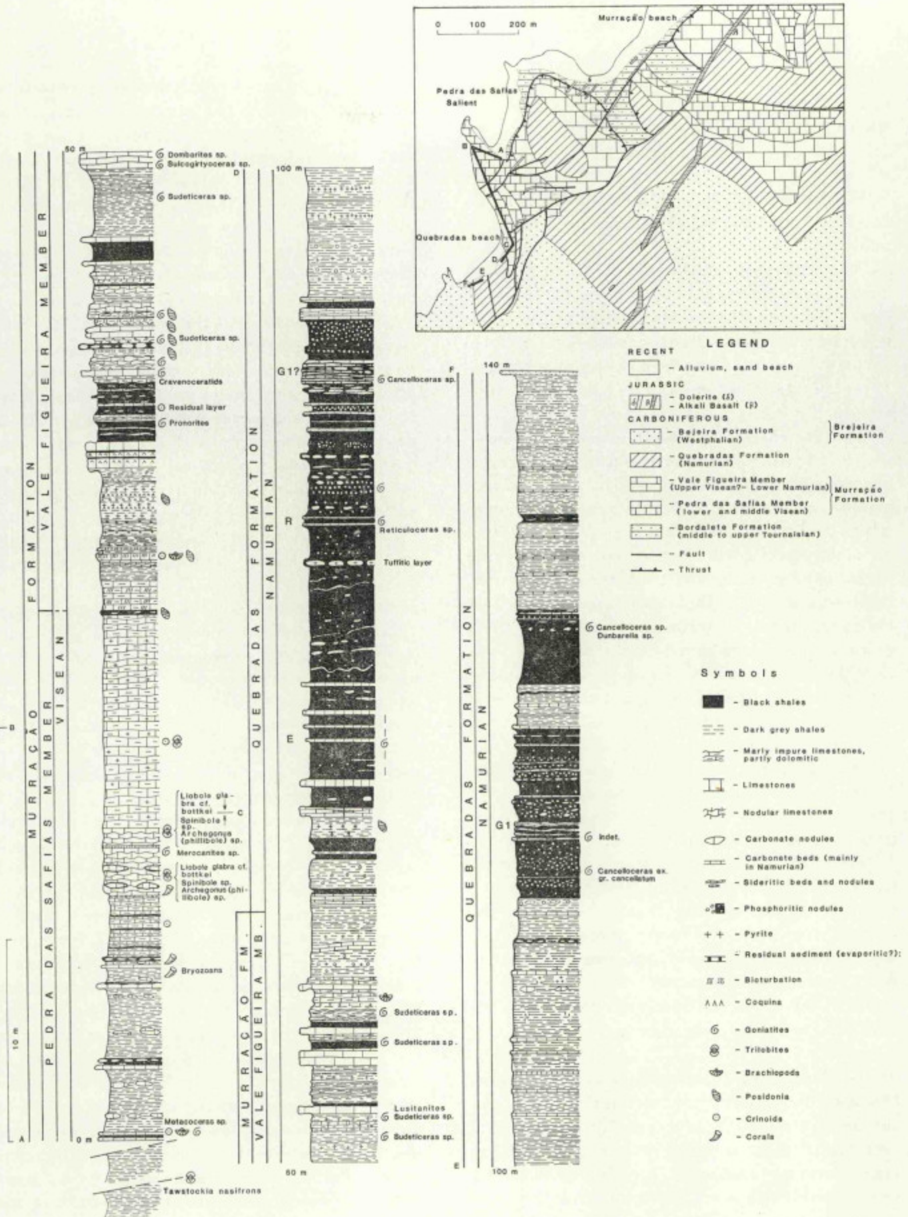


Fig. 12 — Stratigraphic log of the Murração and Quebradas Fms, Atlantic seabord south of the village of Carrapateira (adapted from J. OLIVEIRA *et al.* in prep.)

dity currents, which become stronger later on and which are responsible for the bulk of the Brejeira Fm turbidites.

5. CARBONIFEROUS PALAEOGRAPHY OF THE SOUTH PORTUGUESE ZONE

The palaeogeography of the South Portuguese Zone, during Carboniferous times is, in many respects, still a matter of conjecture. This is so because some basic aspects of the local geology have not yet been fully investigated. Three outstanding points are discussed here:

a — **The character of the Devonian basement**— in the Pyrite Belt and Southwest Portugal generally the Devonian basement is composed mainly of quartzites, quartzwackes, siltstones, shales and minor limestone lenses near the top. This suggests a continuous cover of detrital rocks over a large part of the sedimentary basin during late Devonian times. However, the depositional environment in which these detrital rocks were laid lown is an open question. Do they represent a deep or a shallow water environment? The sandstones show everywhere the same basic types of sedimentary features, the significance of which is not clearly understood due the absence of detailed studies. Consequently, the source area of the detritus can hardly be visualized in the present state of our knowledge.

From local field observations in the Pomarão, Mértola and Carrapateira areas, and taking into account the regional geology, the detrital rock features seem more compatible with deposition in shallow rather than in deep water environments. But evidence is required in order to reach a definitive conclusion.

b — **Nature of the boundary with the Ossa-Morena Zone**—at present the boundary between the South Portuguese and Ossa-Morena zones is tectonic in nature. North of this boundary the Ossa-Morena lithologies are highly sheared basic metavolcanics, peridotites and diorites on which small remmants of greywackes, quartzites and rarer limestones, are found (BATISTA *et al.* 1976). In Portugal the basic rocks were considered by ANDRADE (1977) as representing an ophiolites suite. In Spain (BARD 1977) the basic metavolcanics (Acebuches amphibolites) were taken as metatholeiites of abyssal affinity, but not true ophiolites. These rocks were recently correlated with the basic metavolcanics of Beja area (ANDRADE & V. OLIVEIRA 1982). All the authors admit a Silurian to early Devonian age for these basic rocks. South of the boundary quartzites, siltstones, greywackes, and shales of the S. Iria Fm are found, with a possible late Devonian to Tournaisian age. These rocks show low grade metamorphism. It is worthwhile to note that close to the boundary the quartzites, also sheared, are most common.

The Ossa-Morena (OMZ)-South Portuguese (SPZ) zones relationship has been interpreted in many ways: as a geanticlinal (OMZ)-geosynclinal (SPZ) couple (SCHERMERHORN 1971); as a plate tectonic boundary with associated northward subduction, between a continental plate (OMZ) and an oceanic plate over which the sediments of the SPZ are forming an accretionary prism (BARD 1971; BARD *et al.* 1973; CARVALHO 1971; RIBEIRO & SILVA 1983 — this volume); as a hinterland-foredeep-foreland realm (J. OLIVEIRA *et al.* 1979), etc.

If the Ossa Morena-South Portuguese zones relationship is that of a continental versus oceanic plate (or back arc basin as *in* RIBEIRO 1983-this volume), its present boundary should represent a suture line separating two distinct geological domains. From field data the S. Iria Fm lithologies seem to show a sedimentary polarity from flysch-like sediments in the south to quartzites and siltstones near the boundary. The remnants of quartzites and greywackes in the Ossa-Morena Zone are of the same type of those of S. Iria Fm. This fact led the author (1982) to assume that before the deposition of the S. Iria Fm sediments, the Ossa Morena and the South Portuguese zones were in proximity.

A different interpretation is given by BARD (1977) to whom the basic metavolcanics of the

Ossa-Morena Zone, near the tectonic boundary, are possibly contemporaneous with a rift (or micro-ocean) separating the two palaeogeographic zones.

c — **The significance of the condensed facies of southwestern Portugal** — The sedimentary facies of the Carrapateira Group show that they have been deposited in reducing, sometimes saline, marine environments. To what extent these environments were linked to a local submarine high (J. OLIVEIRA *et al.* 1979) or to a neighbouring platform to the south (J. OLIVEIRA 1982) is not clear. Their deposition on a shelf area in front of a marine platform (of uncertain position) is only tentatively suggested here.

The points discussed above (some others could also be emphasised, such as the geotectonic significance of the Pyrite Belt volcanics), show clearly the difficulties which remain to be solved in order to gain a better understanding of the local palaeogeography and geodynamics.

In more general context, the South Portuguese Basin has been seen as representing the western continuation of the North European sedimentary basins, via southern England, northern France and north Germany. This continuation can be obtained, either by closing the Bay of Biscay, (COGNÉ 1977; JULIVERT *et al.* 1980, etc.) or by a clockwise rotation of 90° or more of the Iberian Peninsula (BLESS *et al.* 1977; PAPROTH 1982). This last hypothesis, which is based mainly on biostratigraphic comparisons, would imply the complete separation of Iberian Peninsula from France in pre-Permian times.

In this case, the role of the Ibero-Armorican arc, as it has been interpreted by many workers (MATTE & RIBEIRO 1975; RIES & SHACKLETON 1971; etc.) would be open to question.

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HERCYNIAN MAGMATISM IN THE IBERIAN PYRITE BELT

J. Munhá [1]

Key words: Iberian Pyrite Belt; Hercynian magmatism; Tectonic setting.

Palavras-chave: Faixa Piritosa Ibérica; Magmatismo hercínico; Ambiente tectónico.

ABSTRACT

Widespread Hercynian magmatic activity in the Iberian Pyrite Belt corresponds to the igneous stage in the development of the South Portuguese Zone geosynclinal, from late Devonian to lower Carboniferous times.

Petrographic and geochemical data indicates that the dominant volcanic rock types, basalts, andesites (minor) and the more voluminous rhyolites are not linked by fractional crystallization.

Early basaltic lavas are tholeiitic with chemical characteristics transitional to arc tholeiites (high LILE/HFSE ratios), similar to some basalts erupted during the initial stages of back-arc spreading; towards the top of the volcanic sequence the mafic rocks display enrichment in incompatible elements (Ba, Nb, P, Zr, LREE), such that the upper lavas are typical 'within plate' alkaline basalts. Petrogenetic moddelling of minor and trace element chemistry of the various basalts demonstrates that they cannot be related solely by fractional crystallization. Even with partial melting mechanisms at least two different mantle sources are required to explain the compositional differences between the major basaltic types. The results suggest that the mantle source feeding the Pyrite Belt basalts was heterogeneous with respect to both minor and trace elements, and that their mantle sources were enriched in lithophile elements.

Andesitic rocks have trace element abundances that are consistent with derivation by partial melting ($< 10\,\%$) of an hydrated LREE enriched upper mantle source, (compositionally equivalent to that estimated for contemporaneous tholeiitic basaltic lavas), followed by varying degrees of fractional crystallization. Geochemical changes across strike of the Pyrite Belt are difficult to reconcile with the activity of a northward dipping subduction zone underlying the South Portuguese Zone during the Hercynian orogeny.

New trace element data for the felsic volcanics supports inferences from Sr-isotope results and all suggest a derivation by crustal anatexis.

It is purposed that the particular rock types which occur in the Iberian Pyrite Belt, including andesites, transitional arc tholeiites and within-plate alkaline basalts, reflect the transient geochemical nature of the mantle under a former active continental margin combined with complex melting relationships attending the initial stages of an attempt for ensialic back-arc spreading.

RESUMO

A actividade magmática Hercínica na Faixa Piritosa representa o estádio ígneo da evolução da Zona Sul Portuguesa, ocorrendo do Devónico superior ao Carbónico inferior.

Os dados petrográficos e geoquímicos indicam que os litotipos vulcânicos mais frequentes, basaltos, andesi-

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[1] Departamento de Geologia, Faculdade de Ciências, Universidade de Lisboa, Rua da Escola Politécnica 58, 1294 Lisboa Codex, Portugal.

tos e riólitos (dominantes), não são relacionáveis por processos de cristalização fraccionada.

As lavas basálticas mais antigas são toleíticas apresentando características químicas transicionais para os toleítos de arco insular (altos valores da razão LILE/HFSE), e são semelhantes a certos basaltos produzidos durante os estádios iniciais de expansão nas bacias marginais; para o topo da sequência vulcânica, as rochas máficas mostram enriquecimento significativo em elementos incompatíveis (Ba, Nb, P, Zr, Tr-leves), de tal modo que as lavas máficas superiores são basaltos alcalinos 'intra-placa' típicos. Os modelos petrogenéticos da geoquímica de elementos menores e em traços, dos vários tipos de basaltos, demonstra que estes não podem ser simplesmente relacionados por processos de cristalização fraccionada. Mesmo envolvendo mecanismos de fusão parcial, são necessárias, pelo menos, duas fontes mantélicas distintas, para explicar as diferenças composicionais entre os diferentes tipos de basaltos. Os resultados sugerem que as fontes mantélicas de onde derivaram os basaltos da Faixa Piritosa eram heterogéneas do ponto de vista químico e mineralógico.

A geoquímica dos elementos em traço das rochas andesíticas é consistente com a sua formação por fusão parcial ($< 10\,\%$) de uma fonte mantélica hidratada e enriquecida em terras raras leves (com composição semelhante à estimada para os basaltos toleíticos), seguida de graus variáveis de cristalização fraccionada. As variações geoquímicas regionais são difíceis de conciliar com a hipótese da existência de uma zona de subducção, com polaridade para norte, sob a Zona Sul Portuguesa, durante a orogenia Hercínica.

A geoquímica de elementos em traço e isotópica, das rochas félsicas, sugere anatexia crustal.

Sugere-se que os vários tipos de rochas vulcânicas que ocorrem na Faixa Piritosa Ibérica, reflectem evolução geoquímica do manto superior sob uma (antiga) margem continental activa, bem como processos complexos de fusão parcial, durante os estádios iniciais de expansão numa bacia marginal ensiálica.

1. INTRODUCTION

Because of their close relations with ore deposits, the Iberian Pyrite Belt volcanic rocks have been subjected to many studies (for references see STRAUSS 1970); nevertheless, some problems still remain open to discussion, particularly in what concerns their nature, genesis and tectonic setting (see for instance, CARVALHO 1971; SCHERMERHORN 1975; ROUTHIER *et al.* 1977). Recent work (MUNHÁ 1979) indicates that the volcanism in the Iberian Pyrite Belt is essentially representative of a bimodal association of tholeiitic to alkalic basalts and rhyolites and has suggested an ensialic rifting model as a probable tectonic setting for the Hercynian magmatism in the area. This paper outlines the geologic characteristics of the Pyrite Belt volcanism, documents further petrographic and geochemical evidence for this (MUNHÁ 1979) hypothesis and, by means of petrogenetic modelling, provides constraints on how the (crust/mantle) source regions, the degree of partial melting and the fractional crystallization history may have changed during the period of Hercynian igneous activity in the Iberian Pyrite Belt.

2. GEOLOGICAL SETTING

Extremely large pyritic massive sulphide ore bodies are located in the Iberian Pyrite Belt. The latter extends from near Sevilla in southwest Spain to the Atlantic coast in south Portugal and constitutes the intermediate sub-zone (CARVALHO *et al.* 1971) of the South Portuguese Zone (LOTZE 1945), an external lower Carboniferous eugeosyncline flanking the central Hercynian block of Iberia.

The Iberian Pyrite Belt contains three major lithostratigraphic units (SCHERMERHORN & STATON 1969; SCHERMERHORN 1971; CARVALHO *et al.* 1971; J. OLIVEIRA *et al.* 1979; an outline geological map of the South Portuguese Zone may be found in the introductory chapter of this volume), from base to top, the Phyllite-Quartzite Fm (PQ), the Volcanic-Sedimentary Complex (VS), and the Mértola Fm (Culm facies). The VS Complex comprises sediments, volcanic rocks and the stratiform sulphidic and manganese deposits (STRAUSS *et al.* 1977; BARRIGA & CARVALHO 1983 — this volume). The Mértola Fm is a flysch sequence of slates and greywackes. The exposed strata are of upper Devonian and lower Carboniferous age (J. OLIVEIRA 1983 — this volume). All these rocks were deformed an regionally metamorphosed during the Hercynian orogeny. The tectonic framework is characterized by an imbricated structure facing southwest due to high-angle reverse faults post-dating the main cleavage, of pre-Westphalian D age (RIBEIRO & SILVA 1983 — this volume).

The Volcanic-Sedimentary Complex: A summary of its geologic characteristics

The Volcanic-Sedimentary Complex (VS) is a heterogenous rock group with rapidly varying thickness from place to place and quickly changing volcano-sedimentary facies in both lateral an vertical directions. The thickness of VS ranges from only a few tens of meters up to about 800 meters; its rock sequence varies from almost entirely volcanic, as at Rio Tinto (RAMBAUD 1969; GARCIA 1974) and other important eruptive centers, to non-volcanic, as for instance northeast of Aljustrel (SCHERMERHORN & STATON's (1969) Corte Vicente Anes Group).

The VS is the only lithostratigraphic unit with abundant acid and basic volcanic rocks, and it is the exclusive host for all the stratabound sulphide and manganese mineralizations.

The volcanic rocks occur in lenses and levels up to several hundred meters thick, contained in the sedimentary framework of VS. The latter is formed mostly by slates, with varying admixtures of terrigenous detritus and of biogenic, chemical and volcanic components.

The volcanics are felsic, mainly meta-rhyolites (quartz keratophyres), and mafic, meta-basalts and meta-dolerites (spilites, albite-diabases), with occasionally some intermediate rocks (see SCHERMERHORN 1975). The felsic rocks largely outnumber the mafic ones and comprise about 70 % of the total outcropping volcanic areas. Although there is some overlap of felsic onto mafic rocks, or vice versa, the mafic and felsic eruptive centers are well separated and no lithological transitions have been described so far (SCHERMERHORN 1970, 1975; SOLER 1973; ROUTHIER *et al.* 1977).

Felsic volcanics are mainly pyroclastics of variable grain sizes and are grouped around eruptive centers along fissure-type volcanic lineaments (STRAUSS & MADEL 1974), which are outlined by volcanic breccias, agglomerates, isolated lava flows and, occasionally, volcanic chimneys (MADEL & LOPERA 1976; CARVALHO 1976, 1979). Laterally these volcanics form well-bedded tuffs or tuffaceous shales and/or interfinger rapidly with sediments, most frequently shales and radiolarian cherts, indicating a submarine deposition. At a few places felsic lavas exhibiting columnar jointing and ignimbritic flows have been mapped (ROUTHIER *et al.* 1977; LOPERA 1978; LECA pers. comm. 1976), that might indicate transient subaerial or extremely shallow water submarine volcanic activity.

In spite of its usual heterogeneity there are large areas in the Pyrite Belt where it is possible to establish a generalized lithostratigraphic sequence for VS felsic volcanism. One may refer as a typical example to the zone located between Pomarão (Portugal) and Sotiel (Spain); there, three major felsic volcanic episodes have been recognized which show an impressive regularity all along the strike of the large Puebla de Guzman anticlinal structure (for detailed geological descriptions see: VAN DEN BOOGAARD 1967; STRAUSS & MADEL 1974; CARVALHO *et al.* 1976b; MADEL & LOPERA 1976; ROUTHIER *et al.* 1977). For other areas, however, this type of sequence is not strictly applicable and one or even two of the above mentioned felsic volcanic episodes may be missing.

Mafic volcanics appear in many parts of the Pyrite Belt, however, their greatest extent and thickness is always observed at larger distances from the centers of the felsic volcanism and hence most frequently associated with slates and siliceous sediments.

Mafic lava flows and tuffs show their largest development in the west of the Pyrite Belt, occupying a zone which stretches from Castro Verde to north of Lousal, in south Portugal. Here they from the upper part of VS, locally overlain by the Mértola Fm slates (SCHERMERHORN 1970). In Spain mafic flows occur at various levels but more commonly near or at the base of VS (see FEBREL 1967; RAMBAUD 1969; STRAUSS & MADEL 1974). The occurrence of radiolarian chert filling the interstices in pillow lavas (WILLIAMS 1966; SCHERMERHORN 1970) indicates that the volcanism took place in a submarine environment.

Mafic intrusives form sills and less often irregular masses. According to SCHERMERHORN (1970) doleritic sills represent high level intrusions injected, in part into semi-consolidated sediments at no great depths bellow the sea floor. Although they tend to be more widespread than 'spilites' it is usually observed that the greatest volume of intrusions was emplaced in areas where mafic flows were also abundant. Very large intrusions have been occasionally mapped (see AYE 1974), but individual sheets are rarely over 25 meters thick. At the contacts the sills are generally concordant with the host strata, show chilled margins and produce contact metamorphism which ranges from silght induration of the wall rocks to cordierite-andalusite hornfelses and spotted slates around larger bodies.

The age of the VS Complex has been defined on the basis of fossil evidence (QUIRING 1936; COSTA 1943; VAN DEN BOOGAARD 1963; PFEFFERKORN 1968; VAN DEN BOOGAARD & SCHERMERHORN 1975, 1981) and with reference to the well dated underlying PQ Fm and overlying Culm facies, as ranging from the uppermost Devonian to lower Viséan. A somewhat wider time span for volcanic activity was recently recognized by CARVALHO (1976), who also advocated a northward migration of the volcanic front, with volcanism starting in the upper Devonian at Cercal-Odemira region and lasting until upper Viséan times in the northernmost part of the Pyrite Belt. However, this migratory pattern may not have been so simple since metavolcanic rocks are known from both Pulo (ZBYSZEWSKI *et al.* 1964) and Xistos do Freixial (equivalent to Ribeira de Limas formation; CARVALHO *et al.* 1976a) (V. OLIVEIRA *et al.* 1977) formations, that suggests that volcanism, although not so widespread as in Cercal, may also have started earlier in the north (see also J. OLIVEIRA 1973 — this volume).

Deposition of the VS Complex started with widespread felsic volcanism producing thick piles of mainly submarine tuffs in many places. Between and away from the felsic eruptive centers, the prevalence of clayey sediments, with intercalated mafic volcanics and biogenic/chemical admixtures, suggests pelagic deposition. It seems thus, that the paleogeographic environment for the Pyrite Belt during this period may have been characterized by submarine volcanic highs (locally emergent) alternating with pelagic troughs where sedimentation went on essentially undisturbed except by tuffitic, biogenic and chemical interbeds as a result of nearby volcanism. The incoming of Culm-type greywackes, towards the top of VS, marks the onset of tectonic instability and strong differential subsidence which culminates during the deposition of the succeeding flysch sequence (see also J. OLIVEIRA 1983 — this volume).

3. PETROGRAPHY AND GEOCHEMISTRY

3.1. INTRODUCTION

Difficulties arise when one studies the igneous petrogenesis of older volcanic rocks, since they commonly loose more or less their primary igneous features during periods of metamorphism and deformation. This, of course, is one of the main reasons why the petrogenesis of ancient eugeosynclinal rocks remain poorly understood.

The Iberian Pyrite Belt volcanics generally exhibit clear igneous textures. However, with the exception of clinopyroxenes in mafic and intermediate rocks, the primary mineralogy is rarely preserved. The volcanic rocks have metamorphic assemblages characteristic of the prehnite-pumpellyite/lower greenschist facies and available geochemical data indicates that they have experienced significant redistribution of several major and trace elements during the alteration processes (MUNHÁ 1976, 1979, 1892a; MUNHÁ & KERRICH 1979/80, 1980; MUNHÁ *et al.* 1980); as pointed out by several authors (c. f. R. SMITH 1968; VALLANCE 1974a, b), this has major implications for any discussion of their magmatic affinity. For this reason, samples used in this study have been selected from among those showing the least degree of secondary alteration, and the geochemical data has been restricted to those elements (e. g. Al, P, Sc, Ti, V, Cr, Ni, Y, Zr, Nb, RE and FeO^t/MgO ratios) which are generally thought to be relatively immobile during weathering and metamorphism (R. SMITH 1968; CANN 1970; PEARCE & CANN 1971; HERRMAN *et al.* 1974; FLOYD & WINCHESTER 1975; R. SMITH & S. SMITH 1976; COISH 1977; CONDIE *et al.* 1977; G. HANSON 1980).

The systematic covariance of the supposed 'immobile' elements and their congruent relationships with the chemical characteristics of relict high temperature igneous phases (MUNHÁ 1983) suggest that they indeed represent original igneous features. The discussion in this paper is based on this premise.

3.2. PETROGRAPHIC FEATURES OF THE METAVOLCANIC ROCKS

The metavolcanic rocks in the Iberian Pyrite Belt comprise a wide variety of rock types ranging from mafic (locally ultramafic) to felsic in composition.

STRECKEISEN (1967) has summarized the diverse opinions concerning the nomenclature of igneous rocks and proposed a very logical classification based on several mineralogical and chemical paramters. Unfortunatelly this or any other classification scheme cannot be strictly applied to altered igneous rocks since knowledge of their original chemical and/or mineralogical composition is required, an information which is often difficult if not impossible to obtain for many metamorphosed rocks. In view of this constraint the Iberian Pyrite Belt metavolcanic rocks have been simply subdivided into mafic, intermediate and felsic according to chemical criteria and inferred igneous mineralogical composition.

The mafic rock group comprises essentially rocks of basaltic composition and corresponds to the 'spilites' and 'albite-diabases' of previous authors (SCHERMERHORN 1970, 1975; SOLER 1973). The intermediate group ($SiO_2 = 57 - 64$ wt % anhydrous) corresponds to andesites, and the felsic division ($SiO_2 > 64$ wt %) to most definitions of dacite and rhyolite (see also EWART 1979). According to the nomenclature adopted by previous workers most of the intermediate rock samples studied here could be designated as keratophyre (KLEYN 1960;

VAN DEN BOOGAARD 1967; SCHERMERHORN 1970, 1973) and most of the felsic rocks as quartz-keratophyre and quartz-kalikeratophyre (SCHERMERHORN 1970, 1973, 1975; SOLER 1973).

a) *MAFIC ROCKS*

Mafic metavolcanic rock samples selected for this study include both extrusive and intrusive rocks which occur at various levels of the VS stratigraphic sequence.

a.1) *Lower Mafic Lavas*

The lower mafic lavas (LML) are representative of mafic extrusive magmatic activity at the onset of VS times and were collected mainly from an area between ENE Calanas and Rio Tinto in southwest Spain (see also, FEBREL 1967; RAMBAUD 1969; GARCIA & MARTIN 1976).

Massive flows

The great majority of LML are fine-grained to aphanitic massive flows containing abundant primary igneous minerals or their pseudomorphs. Probably, more than 70 % by volume of these flows were originally composed of plagioclase, clinopyroxene and Fe-Ti oxides, the rest being amygdules and interstitial (glassy) material. In contrast with the upper mafic lavas, the amygdules are of small size (typically less than 2 mm) and constitute in general less than 3 % of the flow volume. The observed textures are predominantly intergranular to subophitic or, more rarely, slightly porphyritic; phenocrysts are of (albitised) plagioclase and, not so often of clinopyroxene (typically a pale coloured augite variety ranging in composition from $Fs_{10}En_{50}Wo_{40}$ to $Fs_{28}En_{34}Wo_{38}$; see Fig. 1) which also occurs as anhedral grains interstitial to plagioclase microlites of the groundmass. Fe-Ti oxides comprise about 5-10 % of most massive flows and occur as late stage micrograins of small lamellar (ilmenite) crystals largely replaced by sphene.

Pillow lavas

Pillow-lava structures were observed at various places near the base of the VS Complex; samples selected for this study were collected from an outcrop near the Zalamea la Real village, about 8 km west of Rio Tinto.

The pillows are tightly packed, 'closed packed pillow lava' as described by CARLISLE (1963), with the matrix (now consisting largely of secondary minerals) comprising only a very small percentage of the total volume. Pillow-lavas have a finer grain size than most massive flows and they also seem to have been more reactive to metamorphic recrystallization since no relict igneous minerals were observed. Pillow interiors typically show a blasto-intergranular (locally variolitic) texture, with albitised plagioclase microlites in a completely recrystallized siliceous matrix. Divergent actinolite needles may represent pseudomorphs after intergranular clinopyroxene or tachylitic material. Irregular chloritic patches probably corresponds to former glass. Towards the pillow margin the average plagioclase size decreases, the matrix/plagioclase microlites volume ratio increases and the microlites adquire a plumose shape; all these features suggesting an increase in the cooling rate. Pillows have a thin chilled rim, which is distinct in colour and texture; although in its pristine state the pillow rim was most probably a glassy material, it is now entirely replaced by pumpellyite, epidote, actinolite and other metamorphic minerals.

Mafic pyroclastics

Mafic tuffaceous rocks, probably formed by fragmentation of basalt rapidly quenched by sea water, occur associated with the pillow lavas. The tuffs are well foliated and commonly enriched in carbonate and iron oxide minerals. They seem to have been essentially composed of glassy fragments, now mostly flattened owing to deformation, which in most cases include euhedral plagioclase phenocrysts or glomerophenocrysts. The mineralogy is now completely metamorphic and nothing of its original constitution can be determined. Similar pyroclastic rocks have been described by RAMBAUD (1969) from the Rio Tinto area and by MIDDLETON (1960) from the Hercynian of SW England.

a.2) *Upper Mafic Lavas*

The upper mafic lavas (UML) occur near the top of the VS sequence and were collected from the Castro Verde — Ourique region in south Portugal (see also SCHERMERHORN 1970).

UML are very fine-grained to aphanitic and highly vesicular. The amygdules, often filled by calcite, chlorite and hematite, are spherical to almond shaped (when not deformed) and can reach up to several cms in diameter. Due to their high porosity the UML commonly show a very high degree of alteration (see MUNHÁ 1979; MUNHÁ & KERRICH 1979/80, 1980), and consequently relict igneous minerals are rare. Strongly deformed lavas have been transformed to greenschists consisting of an aggregate of chlorite, calcite, hematite, white mica and albite, with sphene/leucoxene (± epidote, actinolite, magnetite, sulphides and occasionally K-feldspar). Less deformed samples are made up of albitised plagioclase microlites, often in a random arrangement, occasionally accompanied by phenocrysts, floating in a groundmass of chlorite and finely granular sphene/leucoxene, iron oxides, calcite and epidote. In the few exceptional samples in which relict igneous minerals may still be observed the texture is best described as pilotaxitic with laths of plagioclase and grains of pyroxene set in an altered finely crystalline mesostasis. Invariably the pyroxene is a purplish to purplish-brown titaniferous salite ($Fs_{12}En_{38}Wo_{50}$ to $Fs_{16}En_{34}Wo_{50}$; see Fig. 1) sometimes, partially replaced by kaersutite and biotite. A few chloritic pseudomorphs after olivine (?) and abundant sphene/leucoxene micrograins (after Fe-Ti oxides) were also observed. The presence of titaniferous salite in the UML contrasts with the clinopyroxene varieties observed in the LML (see Fig. 1) and indicates that the UML have affinities with alkali basalts (G. MACDONALD & KATSURA 1964).

a.3) *Doleritic Sills*

Doleritic rocks have a dark green to grey-green colour (when fresh) and, except for very thin sills or the chilled margins of thick ones, are coarser than the lava flows.

The composition and texture of the meta-dolerites are not uniform throughout the whole of a sill. In the outer part (up to a few meters) the rock consists of

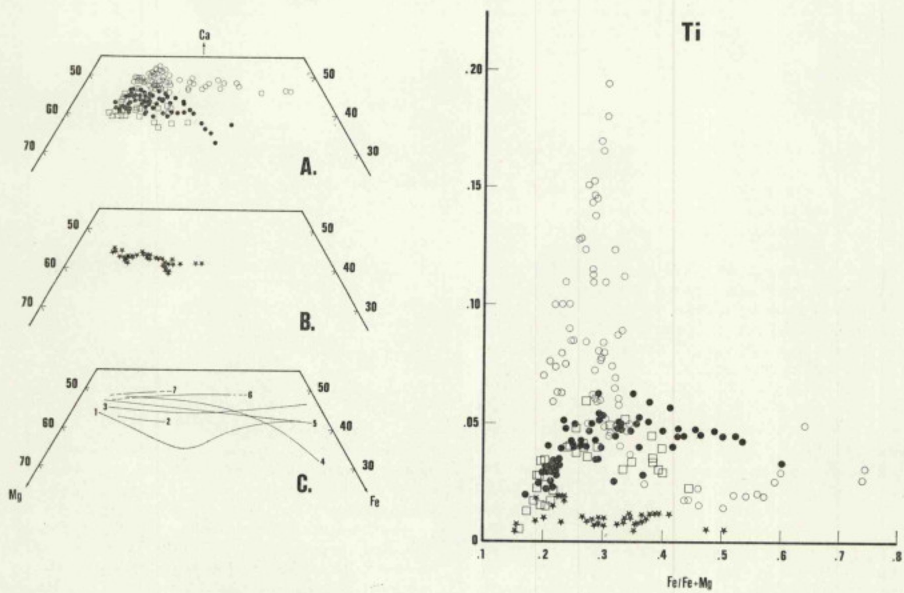

Fig. 1 — On the left: Ca-Mg-Fe compositional variations of clinopyroxenes occurring as relict phases in the Iberian Pyrite Belt metavolcanic rocks (after MUNHÁ 1983).

A. — Lower mafic lavas (squares); type-A dolerites (closed circles); type-B dolerites + upper mafic lavas (open circles).

B. — Intermediate (andesitic) rocks (stars).

C. — Comparative data from: (1) Skaergaard, (2) Kap Edward Holm, (3) Shiant Island, (4) Shonkin Sag, (5) Japanese alkaline rocks, (6) Canary Islands, (7) Monte Somma (VESUVIO). For references see MUNHÁ (1983).

On the right: Ti variations with Fe/(Fe + Mg) ratio in relict clinopyroxenes, (symbols are the same as above).

small albitised plagioclase crystals set in chlorite-carbonate $\pm$ white mica-iron oxides matrix which probably corresponds to former glass and/or ferromagnesian minerals. In their inner parts the doleritic sills consist largely of medium-grained plagioclase and clinopyroxene crystals (or their replacement products). Relicts of calcic plagioclase occur sporadically, and many samples contain additional small amounts of (igneous) ilmenite, Ti-magnetite, amphiboles, biotite, K-feldspar and quartz. The texture is sometimes ophitic, but most commonly subophitic to intergranular.

Ultramafic cumulates (picritic dolerites?) may also occur within some doleritic sills (see STRAUSS & MADEL 1974). Petrographic study (MUNHÁ 1982b) suggests that these cumulate rocks (now intensely serpentinised) were originally essentially composed of olivine (Fo_{80}), plagioclase and minor spinel (Cr-pleonaste to Al-chromite), with intercumulate clinopyroxene ($Fs_{9-10}En_{44-46}\ Wo_{45-46}$) and late stage kaersutite-richterite, phlogopite, Cr-titanomagnetite and ilmenite.

Detailed petrographic examination of relict igneous phases in a large number of samples has permitted recognition among the sills of two doleritic rock types.

Type A — characterized by ilmenite-plagioclase (up to $An_{61.3}Ab_{38.3}Or_{0.4}$)-(pale coloured) augite ($Fs_{10}En_{47}\ Wo_{43}$ to $Fs_{38}En_{24}Wo_{38}$; see Fig. 1). Sporadic occurrence of quartz + K-feldspar as late stage granophyric intergrowths suggests tholeiitic affinities for this rock group (G. BROWN 1967).

Type B — characterized by Ti-magnetite (rare)-ilmenite-plagioclase (up to $An_{60.4}Ab_{37.4}Or_{2.2}$)-titaniferous salite to sodian ferrosalite ($Fs_{11}En_{41}Wo_{48}$ to $Fs_{48}En_{7}Wo_{45}$; see Fig. 1)-kaersutite-biotite. Type B dolerites tend to intrude at high levels within the VS sequence, being sometimes associated with upper lava flows; their relict mineralogy also indicates affinities with the alkali basaltic rocks (G. MACDONALD & KATSURA 1964; WILLKINSON 1974; G. BROWN 1967).

b) *INTERMEDIATE ROCKS*

Although subordinate on a regional scale, intermediate rocks also occur and are locally important, especially in the northernmost part of the Iberian Pyrite Belt.

The studied samples were collected from Monte do Forno da Cal (Almodovar), from Pomarão (VAN DEN BOOGAARD's (1967) keratophyres) and from various places along CARVALHO's (1976) volcanic lineament D, which runs parallel to the regional strike going through S. Domingos mine (see Fig. 3 *in* CARVALHO 1976). Samples from Monte do Forno da Cal are extrusive rocks occurring near the base of the VS Complex (SCHERMERHORN pers. comm. 1976) while those from Pomarão include both sills, intrusive into the Cerqueirinha and Nascedios Formations, and extrusive rocks within the Touril Formation (see VAN DEN BOOGAARD 1967). Other samples are mainly extrusive rocks probably occurring at stratigraphic levels similar to those of the Pomarão region.

The intermediate rocks have a distinctly porphyritic texture. Phenocrysts of albitised plagioclase (up to 25 % modal) and clinopyroxene (up to 10 % modal; $Fs_{9}En_{47}\ Wo_{44}$ to $Fe_{30}En_{29}Wo_{41}$, see also Fig. 1), often grown together into clusters, occur in a trachytic textured groundmass chiefly consisting of plagioclase microlites and interstitial chlorite, iron oxides, quartz, carbonate and sphene/leucoxene. Hornblende partially replaces some clinopyroxene phenocrysts. Minor biotite (largely replaced by chlorite) as well as traces of apatite were also observed in some samples.

c) *FELSIC ROCKS*

The petrography and field relations of the felsic metavolcanics in the Iberian Pyrite Belt have been described in several recent publications (c. f. SOLER 1973; SCHERMERHORN 1975, 1976; ROUTHIER *et al.* 1977). In view of these earlier detailed studies no attempt has been made to perform a systematic petrographic investigation of the felsic rocks. Therefore, only a brief summarized description will be presented here.

The felsic rocks range in composition from dacite to rhyolite, and vary in texture from felsitic and felsophyric to coarse-grained tuffs and agglomerates; highly siliceous rhyolitic tuffs are dominant.

Although texturally similar to some intermediate rocks, typical dacites may be distinguished by the common occurrence of quartz phenocrysts and by their lower content of ferromagnesian minerals. Besides quartz, albitised plagioclase phenocrysts occur almost invariably whereas clinopyroxene microphenocrysts are rare. The groundmass is usually composed of very small plagioclase microlites; however, in some specimens it may grade into an almost irresoluble devitrified felsite.

In the rhyolitic rocks, albite (originally oligoclase-andesine) and quartz are the most common phenocryst phases; in addition K-feldspar may also appear, sometimes attaining large sizes as in the Megacryst Tuff at Aljustrel (SCHERMERHORN 1976). Biotite phenocrysts (largely replaced by chlorite) and rare chloritic pseudomorphs after garnet were also reported by some authors (SCHERMERHORN 1976; ROUTHIER *et al.* 1977; F. BARRIGA pers. comm. 1983). Most frequent accessory minerals include apatite, zircon, allanite and tourmaline.

The great majority of the samples collected for this study correspond in many respects to the fine grained granular tuffs as described by SCHERMERHORN (1975). Typical specimens are composed of small grains arranged in a greywacke type texture with the components closely packed in a very fine volcanic matrix; the fragmental components are generally small (devitrified) glassy particles and variable amounts of quartz and feldspar crystals. The rocks may be aphyric but most commonly are albite- and/or quartz-phyric.

3.3. GEOCHEMISTRY

3.3.1. ANALYTICAL METHODS

Major elements together with Rb, Sr, Ba, Sc, Y, Zr, Nb, Cr, Ni and V were determined mainly by X-ray fluorescence spectrometry (NORRISH & HUTTON 1969). Li, Cu and Zn were analysed by conventional AAS. REE were determined by a thin film X-ray fluorescence technique as described by FREYER (1977). Analyses of P and REE are updated and revised relative to those presented by MUNHÁ & KERRICH (1980).

Because cf the large volume of data available the results for each lithotype are presented in terms of average (and range) compositions which are listed in Tables 1, 3 and 4. Although this approach obscures the peculiarities of particular rocks, it provides a useful framework for more detailed studies.

3.3.2. GEOCHEMISTRY OF THE MAFIC ROCKS

3.3.2a) GEOCHEMICAL FEATURES OF THE IBERIAN PYRITE BELT BASALTIC ROCKS

Major Elements

The great majority of the basaltic rocks from the Iberian Pyrite Belt are hyperstene-olivine normative, but a few type B dolerites have nepheline in their norms, suggesting alkaline affinities. However, normative compositions are highly susceptible to the effects of secondary alteration and, in consequence, no further discussion of these results will be presented here.

The analysed samples display a considerable range of $Fe_2O^t_3$ and TiO_2 values (see Table 1), and plots of FeO^t/MgO as a function of TiO_2 and FeO^t (Fig. 2A) show that there is a general tendency for a positive correlation. Such relationships are similar to those observed in many non-orogenic basalts (c. f. SHIDO *et al.* 1971; BARBIERI *et al.* 1975; MIYASHIRO & SHIDO 1975) and indicate that the fractionation trends are not calc-alkaline (KUNO 1959; MIYASHIRO 1974).

Ti, P, Zr, Y and Nb Variations

Ti, P, Zr, Y and Nb are all elements with high field strength and, as a result, they are not usually transported in aqueous fluids (see ALDERTON *et al.* 1980) and tend to remain unaffected in rocks which have suffered metasomatic alteration. This property, together with their characteristic variations in fresch lavas, means that these elements can be used to study metamorphosed basalts whose mineralogy and chemistry has otherwise been greatly affected (PEARCE & CANN 1971, 1973; FLOYD & WINCHESTER 1975, 1978; WINCHESTER & FLOYD 1976, 1977).

As shown in Fig. 2B the analysed samples display a considerable range of P_2O_5/Zr and Y/Nb values. B-dolerites and UML are enriched in P_2O_5 and Nb ($P_2O_5/Zr \sim 30$, $Y/Nb \sim 0.6$; see also table 1) whereas, in contrast, Pulo Formation greenschists, LML and A-dolerites tend to be slightly depleted in P_2O_5 and Nb ($P_2O_5/Zr \sim 14$, $Y/Nb \sim 8\text{-}11$) relative to upper mantle P_2O_5/Zr and Y/Nb values of about 18 and 8, respectively (RINGWOOD 1975; NESBITT & SUN 1976; SUN & NESBITT 1977). Both P_2O_5/Zr and Y/Nb (Fig. 2B) indicate that the mafic metavolcanic rocks from the Iberian Pyrite Belt have chemical features ranging continously from those characteristic of alkali basalts to those typical of the tholeiitic series. Pulo Fm greenschists/LML/A-dolerites and UML/B-dolerites have clear affinities with tholeiitic and alkali basalts respectively, in accordance with petrographic and mineral chemistry (see MUNHÁ 1983) relations previously discussed.

Incompatible elements, such as Ti, P, Zr, Y and Nb, maintain constant ratios in those liquids formed by large degrees of partial melting and hence reflect, to a certain extent, the chemical characteristics in their upper mantle sources (TREUIL & JORON 1975; SUN & NESBITT 1977; JORON *et al.* 1978; SUN 1982). Figs 3 and 4 examine the relative enrichment/depletion order among these elements. Pulo greenschists, LML and A-dolerites have TiO_2/P_2O_5 ratios approaching mantle values (10-12; RINGWOOD 1975; SUN 1982), but UML and B-dolerites, as well as all rocks of

Table 1 — Major and trace element concentrations in mafic metavolcanic rocks from the Iberian Pyrite Belt

	PULO FMT. GREENSCHISTS			LOWER MAFIC LAVAS			UPPER MAFIC LAVAS			A — DOLORITES			B — DOLERITES		
	Average	Range	N	Average	Range	N	Average	Range	N	Average	Range	N	Average	Range	N
SiO_2 (a) wt. %	50.03	47.01 — 53.21	6	53.15	47.23 — 56.31	10	47.72	46.35 — 50.54	7	49.72	46.24 — 55.20	35	47.65	45.25 — 49.65	7
TiO_2	1.51	1.31 — 1.77	6	1.95	1.51 — 2.29	10	2.66	2.05 — 3.45	7	1.81	1.00 — 2.94	35	2.74	1.65 — 3.21	7
Al_2O_3	15.16	14.37 — 15.69	6	15.56	14.01 — 17.53	10	17.16	14.36 — 20.47	7	16.55	14.30 — 19.66	35	16.50	14.70 — 18.01	7
Fe_2O_3 (b)	10.95	9.89 — 12.51	6	11.28	8.94 — 15.48	10	11.80	8.82 — 13.57	7	11.35	7.30 — 14.50	35	12.74	10.97 — 14.74	7
MgO	7.40	6.96 — 7.96	6	5.71	4.54 — 7.61	10	6.65	4.70 — 10.85	7	8.01	4.66 — 11.08	35	7.25	3.83 — 11.24	7
MnO	0.19	0.16 — 0.22	3	0.24	0.16 — 0.62	10	0.23	0.17 — 0.36	7	0.20	0.13 — 0.29	31	0.24	0.16 — 0.36	7
CaO	12.18	8.24 — 14.16	6	8.42	4.33 — 11.30	10	8.48	4.92 — 12.21	7	8.99	3.99 — 13.21	35	8.00	6.22 — 9.01	7
Na_2O	2.53	1.86 — 4.15	6	3.33	1.76 — 5.91	10	3.10	1.78 — 4.58	7	2.81	1.19 — 5.82	35	3.31	1.53 — 5.16	7
K_2O	0.05	nd — 0.12	6	0.14	nd — 0.44	10	1.70	nd — 3.54	7	0.41	nd — 1.66	35	0.99	0.22 — 2.54	7
P_2O_5	0.11	0.09 — 0.15	6	0.22	0.13 — 0.30	10	0.50	0.23 — 1.03	7	0.16	0.03 — 0.37	35	0.58	0.23 — 1.21	7
Trace Elements															
Cr ppm	297	215 — 373	6	138	15 — 285	9	192	92 — 291	7	237	60 — 574	34	126	7 — 293	7
Ni	76	63 — 87	6	41	10 — 85	9	133	85 — 211	7	78	11 — 157	35	76	9 — 179	7
V	295	255 — 344	5	288	220 — 340	9	262	198 — 293	6	263	127 — 403	26	245	163 — 313	7
Cu	na			30	20 — 42	3	45		2	76	35 — 172	16	na		
Zn	na			52	44 — 58	3	60		2	85	59 — 160	16	na		
Li	na			31	30 — 33	3	53		2	57	48 — 67	5	na		
Rb	3	nd — 15	6	6	1 — 14	10	36	12 — 88	6	13	1 — 58	35	25	5 — 61	7
Sr	139	77 — 188	6	201	66 — 359	10	423	241 — 766	7	217	62 — 721	35	636	419 — 879	7
Ba	4	nd — 12	6	73	10 — 216	10	515	191 — 1181	6	159	11 — 555	35	487	123 — 830	7
Sc	39	34 — 42	3	36	30 — 44	9	26	24 — 28	5	36	20 — 53	33	22	8 — 30	7
Y	33	24 — 43	6	39	32 — 50	10	27	21 — 38	7	33	13 — 57	35	29	20 — 46	7
Zr	79	60 — 103	6	147	105 — 201	10	161	104 — 261	7	119	45 — 231	35	193	85 — 337	7
Nb	3	2 — 6	6	5	3 — 7	10	55	25 — 92	7	6	1 — 18	34	57	16 — 115	7
La	na			15.1	9.7 — 20.3	4	37.3	28.3 — 45.5	3	15.0	8.99 — 27.8	11	36.8	13.2 — 79.2	7
Ce	na			31.5	20.7 — 43.7	4	68.0	50.1 — 84.6	3	31.1	18.2 — 56.5	11	68.9	25.4 — 161	7
Nd	na			21.7	15.1 — 29.5	4	33.8	24.8 — 40.9	3	20.0	11.8 — 35.8	11	36.6	14.7 — 92.6	7
Sm	na			5.61	4.21 — 7.53	4	6.55	5.23 — 7.42	3	5.13	3.27 — 8.99	11	7.40	3.47 — 17.5	7
Eu	na			1.77	1.38 — 2.25	4	1.92	1.61 — 2.14	3	1.50	0.99 — 2.27	11	2.14	1.13 — 4.95	7
Gd	na			6.25	4.89 — 8.44	4	5.68	4.69 — 6.19	3	5.46	3.52 — 9.31	10	6.65	3.37 — 15.1	7
Yb	na			3.13	2.59 — 4.23	4	1.84	1.41 — 2.09	3	2.75	1.68 — 4.76	11	1.90	1.35 — 2.43	7
Elements Ratios															
FeO^t/MgO	1.32	1.17 — 1.56	6	1.80	1.27 — 2.21	10	1.73	1.09 — 2.51	7	1.32	0.77 — 2.65	35	1.79	0.95 — 2.61	7
Al_2O_3/TiO_2	10.24	8.12 — 11.98	6	8.18	6.51 — 10.97	10	6.63	4.88 — 8.22	7	10.04	5.01 — 19.66	35	6.32	4.67 — 10.41	7
P_2O_5/Zr	14.1	12.2 — 15.7	6	14.3	12.4 — 15.2	10	29.5	18.9 — 39.5	7	13.5	6.6 — 20.5	35	29.5	26.3 — 35.9	7
P_2O_5/Ce				71.9	66.9 — 82.1	4	76.7	70.8 — 87.5	3	56.2	36.2 — 70.1	11	85.7	73.6 — 98.3	7
Zr/Nb	27.4	17.2 — 33.0	6	31.6	24.0 — 40.0	10	3.1	2.8 — 4.2	7	28.3	4.9 — 62.5	34	3.8	2.8 — 5.3	7
Zr/Y	2.45	2.28 — 2.64	6	3.83	3.00 — 4.17	10	5.88	4.36 — 7.72	7	3.59	1.80 — 4.63	35	6.63	4.05 — 8.86	7
Y/Nb	11.23	6.83 — 13.00	6	8.40	6.31 — 11.67	10	0.55	0.35 — 0.84	7	8.13	1.06 — 20.00	34	0.66	0.35 — 1.31	7
Ti/Zr	116	98 — 131	6	82	56 — 97	10	105	74 — 146	7	96	58 — 151	35	92	55 — 116	7
Ti/V	31	30 — 35	5	42	36 — 45	9	63	56 — 70	6	43	33 — 50	26	70	49 — 114	7
La/Nb				3.15	2.64 — 4.06	4	0.63	0.55 — 0.68	3	2.38	1.16 — 3.97	11	0.67	0.51 — 0.83	7
(La/Ce) cn				1.24	1.20 — 1.30	4	1.43	1.39 — 1.48	3	1.24	1.10 — 1.31	11	1.40	1.27 — 1.51	7
(La/Sm) cn				1.63	1.41 — 1.75	4	3.45	3.30 — 3.74	3	1.78	1.41 — 2.05	11	3.05	2.32 — 4.06	7
(La/Yb) cn				3.18	2.42 — 3.77	4	13.4	12.0 — 14.8	3	3.63	2.91 — 5.50	11	12.6	6.46 — 21.5	7

(a) — All major elements calculated on a volatile-free basis; (b) — All Fe calculated as Fe_2O_3.

alkaline affinity in general (NOCKOLDS 1954; SUN & G. HANSON 1975b; KAY & GAST 1973), have TiO_2/P_2O_5 less than 10, suggesting that P is preferentially enriched in these magmas. Zr, like P, also seems to be enriched relative to Ti, whereas Y is relatively depleted; this leads to an enrichment of Zr over Y, as shown on the Zr/Y vs Zr diagram (Fig. 3). In this diagram the Pulo Formation samples are characterized by Zr/Y values similar to those commonly reported for 'normal' type MORB (Zr/Y = 2.5; SUN *et al.*, 1979), in contrast with both LML/A-dolerites (Zr/Y = 3-4) and UML/B-dolerites (Zr/Y = 6-7) which are enriched in Zr relative to the chondritic ratio (Zr/Y = 2.5; SUN & NESBITT 1977; WOOD *et al.* 1979c). Fig. 4 (see also Table 1) shows that Zr/Nb range widely, from an average value of about 30 (similar to many island arc, some back-arc basin basalts and N-MORB) in Pulo greenschists, LML and A-dolerites, down to about 3 (characteristic of most within-plate basalts) in UML

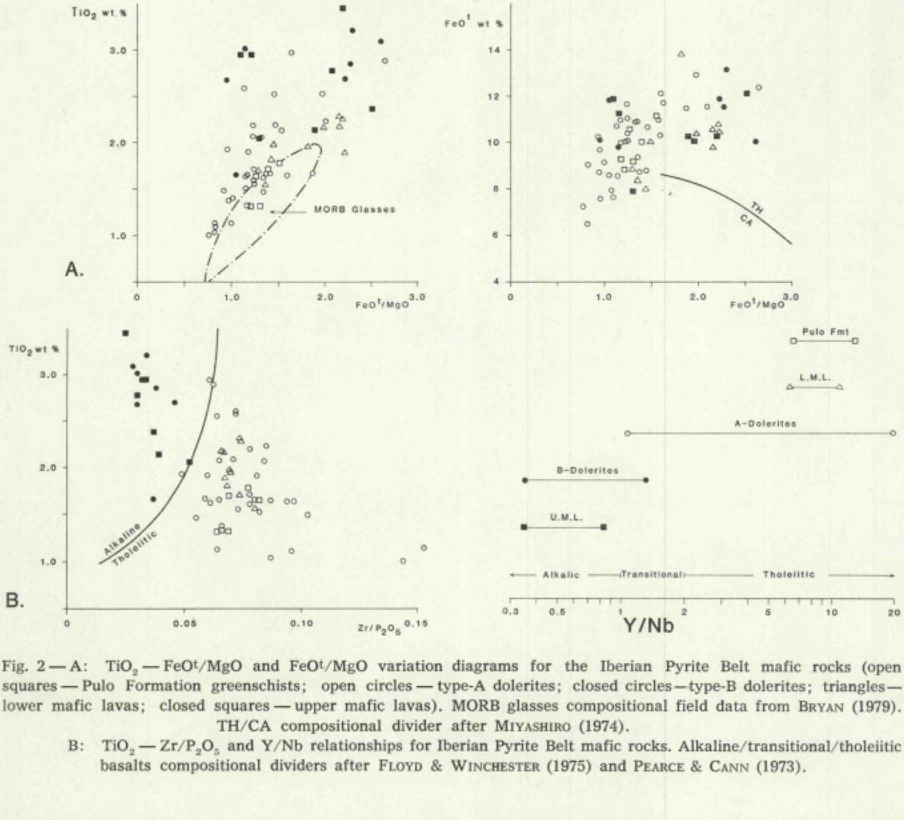

Fig. 2 — A: TiO_2 — FeO^t/MgO and FeO^t/MgO variation diagrams for the Iberian Pyrite Belt mafic rocks (open squares — Pulo Formation greenschists; open circles — type-A dolerites; closed circles — type-B dolerites; triangles — lower mafic lavas; closed squares — upper mafic lavas). MORB glasses compositional field data from BRYAN (1979). TH/CA compositional divider after MIYASHIRO (1974).

B: TiO_2 — Zr/P_2O_5 and Y/Nb relationships for Iberian Pyrite Belt mafic rocks. Alkaline/transitional/tholeiitic basalts compositional dividers after FLOYD & WINCHESTER (1975) and PEARCE & CANN (1973).

and B-dolerites. It is evident that for UML and B-dolerites Nb is much more enriched than Zr, thus the sequence of relative enrichement is, $Nb > P > Zr \geqslant Ti > Y$; a similar enrichment

The Ti-Zr diagram (Fig. 3) shows that the A-dolerite/LML and Pulo Formation samples group in or near the ocean-floor basalt field but that most UML/B-dolerite samples extend

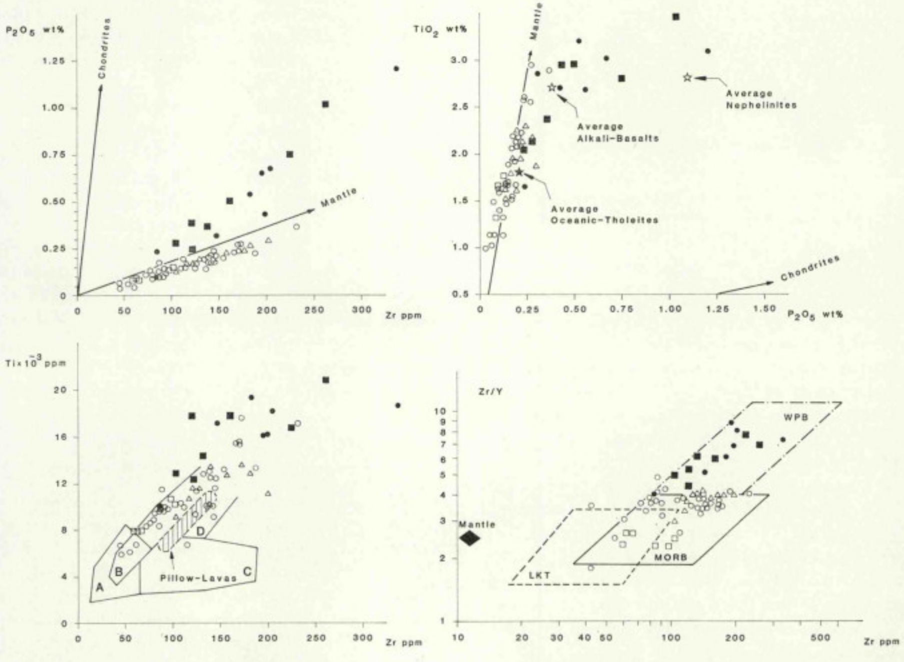

Fig. 3 — P_2O_5 — Zr, TiO_2 — P_2O_5, Ti — Zr and Zr/Y — Zr relationships for Iberian Pyrite Belt mafic rocks.

Ti — Zr diagram: tectonic subdivisions after PEARCE & CANN (1973). A, B: low-K tholeiites; B, C: calc-alkalic basalts; B, D: ocean floor basalts. 'Pillow lavas' refer to highly metasomatised mafic pillow lavas (not included in table 1 LML average composition) occurring at the base of VS Complex near Zalamea (Spain); average Ti/Zr = 74 (8 samples).

Zr/Y — Zr diagram: compositional fields after PEARCE & NORRY (1979).

pattern is true for many alkaline rocks (e. g. GERASIMOVSKY 1974) and reflects the order of decreasing $D^{s/l}$ values for common residual mantle mineralogies.

beyond its limits at high Ti and Zr values. On the Zr/Y - Zr discriminant diagram (Fig. 3) all UML/B-dolerites do fall in the within-plate basalt field (see also MUNHÁ 1979; FLOYD 1982:

note that Monte do Forno da Cal samples reported by FLOYD are andesites).

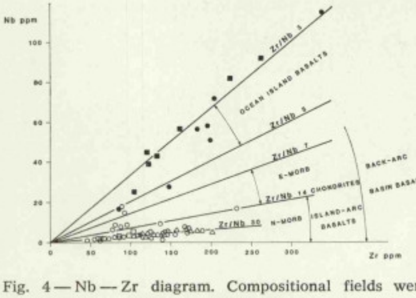

Fig. 4 — Nb — Zr diagram. Compositional fields were compiled from various sources (mainly, WOOD *et al.* 1979b; 1981).

Rare Earth Elements (REE)

REE Masuda-Coryell plots (Fig. 5) of the Iberian Pyrite Belt basaltic rock averages indicate that in all mafic lithotypes the light REE (LREE) are enriched with respect to heavy REE (HREE) compared to chondrites. Although a considerable variation in the degree of REE fractionation is apparent for the whole set, within each group the chondrite-normalized REE patterns are essentially parallel (except for a few B-dolerites) and there is a general positive correlation between REE abundances and abundances of other trace and minor elements.

UML and B-dolerites show the highest degree of LREE/HREE fractionation (see Fig. 5 and Table 1) and their steep REE distribution patterns closely resemble those of alkali basalts and basanitoides reported by KAY & GAST (1973) and SUN & G. HANSON (1975b). Since Nb, P and Y tend to follow the LREE and HREE respectively and Zr is similar to Sm in geochemical behaviour (BESWICK & CARMICHAEL 1978; SUN *et al.* 1979; SUN 1982), the high Nb/Zr, P_2O_5/Zr and Zr/Y values are consistent with their LREE/HREE ratios, in particular the high La/Sm and La/Yb values.

LML and A-dolerites display (Fig. 5) REE distribution patterns similar to many LILE-enriched [1] tholeiites from both continental and oceanic environments (FREY *et al.* 1968; SCHILLING 1971; BARBIERI *et al.* 1975; SHIBATA *et al.* 1979). The most significant feature of LML and A-dolerites REE data is the pronounced LILE (LREE, Ba) enrichment relative to Nb (Fig. 5; see also Table 1), which distinguishes this basaltic suite from the other Pyrite Belt mafic rocks as well as from most MORB [2] and within-plate basalts (WOOD *et al.* 1981). Trace elements which, like Nb (e. g. Zr, Hf, Ta), have high charge/ionic radius are characteristically depleted in island arc basalts relative to within-plate and many MORB (PERFIT *et al.* 1980; HICKEY & FREY 1982), and their concentrations have been successfully used to discriminate basalts from differing tectonic settings (PEARCE & NORRY 1979; WOOD *et al.* 1979a). It should be noted however, that many back-arc basin and continental rift basalts also have sources with island arc affinities regarding LILE/HFSE [3] relationships (c. f. GILL 1976a, b; SAUNDERS & TARNEY 1979; SAUNDERS *et al.* 1979; WEAVER *et al.* 1979; MORRISON *et al.* 1979; THOMPSON *et al.* 1980; COISH *et al.* 1982).

Ti-V Variations

Considerable variations are also evident for Ti/V ratios (Fig. 6; Table 1). As shown in Fig. 6, individual suites of basaltic rocks from world wide sources plot in well defined, but different, fields of Ti/V, which has been interpreted as a consequence of different physical conditions of melting and/or large scale inhomogeneity of Ti and V in the mantle (LANGMUIR *et al.* 1977; WASS 1980; SHERVAIS 1982) possibly, correlative with tectonic setting. Pulo Formation greenschists, LML and A-dolerites plot in the field for mid-ocean ridge basalts and continental tholeiites ($Ti/V = 20\text{-}50$) whereas, UML and B-dolerite samples show $Ti/V > 50$ which are characteristic of alkali basalts. As on the Zr/Y - Zr diagram (Fig. 3), Pulo Formation greenschist samples constitute again a some-

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(1) LILE = Large Ion Lithophile Elements

(2) MORB = Middle Ocean Ridge Basalts

(3) HFSE = High Field Strength Elements

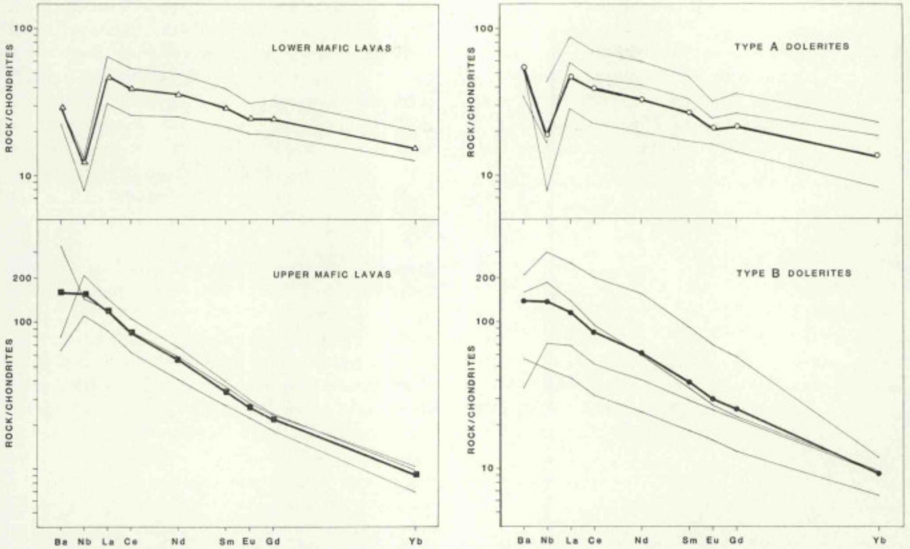


Fig. 5 — Average (and representative) chondrite-normalized Ba, Nb and REE patterns of the Iberian Pyrite Belt mafic rocks. Chondrite normalizing values after SUN & G. HANSON (1976) and SUN & NESBITT (1977).

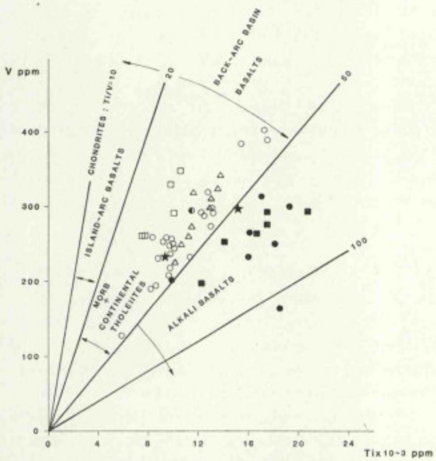


Fig. 6 — Plot of V-Ti for Iberian Pyrite Belt basaltic rocks. Compositional fields after SHERVAIS (1982); (stars: comparative data from AYE (1974)).

what individualized group showing the lowest Ti/V values (average Ti/V = 31); LML and A-dolerites display intermediate Ti/V ratios (~ 43) similar to those reported by JOHNSON *et al.* (1978) for the Bismarck sea marginal basin.

3.3.2b) PETROGENETIC DISCUSSION

Characteristic geochemical features of the Iberian Pyrite Belt basaltic rocks can constraint to some extent the petrogenetic processes responsible for their formation, as well as the nature of the parental mantle. These features may be summarized as follows (see also Table 1):

1) Consistent enrichment in LREE, and crossing REE patterns.
2) Large variations in the incompatible and compatible element abundances.
3) Large variations in some incompatible element ratios.

b.1) *Fractional Crystallization*

The considerable variation in the FeO^t/MgO ratio and Ni contents for A-dolerites and LML (see Table 1) as well as the systematic and congruent decrease of Ni concentrations with increasing La contents (Fig. 7) suggest an

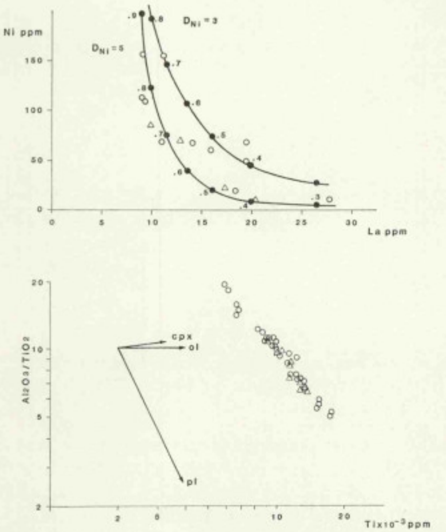


Fig. 7 — Ni — La and Al_2O_3/TiO_2 — Ti relationships for lower mafic lavas and A-dolerites. Solid lines on the Ni — La diagram correspond to fractional crystallization models assuming Co(Ni)=300 ppm, Co(La)=8 ppm and Ni bulk solid/liquid partition coefficients of 3 and 5 (closed circles: fraction of residual liquid). Vectors on lower diagram were modelled for $\simeq$ 70 % fractionation.

early fractionation of olivine. With increasing FeO^t/MgO, TiO_2 increases (Fig. 2) while Al_2O_3/TiO_2 decreases (Fig. 7); this indicates plagioclase fractionation, a feature supported by negative Eu anomalies for the most fractionated samples (Fig. 5). Similarly, some of the variations in Cr and Sc contents (Table 1) can be explained by clinopyroxene and Cr-spinel crystallization. It is thus concluded that shallow level Rayleigh fractionation involving olivine, clinopyroxene and plagioclase, might have been responsible for a significant part of the geochemical differentiation observed for A-dolerites and LML.

However, the variation diagrams (see Figs 1, 3, 4, 5, 6) indicate that B-dolerites and UML deviate considerably from the trends defined by A-dolerite and LML samples. It is not likely that simple closed system shallow level fractionation of olivine, plagioclase and clinopyroxene, could generate the A-dolerites, LML, B-dolerites and UML samples from a common parent magma.

Open system crystal fractionation models, involving continual replenishment of a magma chamber and only partial extraction of fractionated liquid, were tested using the equation for enrichment of a trace element in the steady state liquid (O'HARA 1977). While this process can produce significant changes on the relative fractionation among trace elements of the residual magmas, to generate meaningful heterogeneities, in terms of the studied samples, the increase in absolute trace element concentrations would largely exceed the observed variations among the Pyrite Belt basaltic rocks (see also PANKHURST 1977). Clearly, this process is incapable of producing all the observed chemical variation.

Alternative mechanisms for relating these basaltic rocks are fractional crystallization at high pressures and partial melting processes of upper mantle peridotites. These models are examined bellow.

b.2) *Fractional Crystallization at High-Pressures*

Eclogite fractionation, i.e. high-pressure fractionation of garnet and clinopyroxene (YODER & TILLEY 1962; GREEN & RINGWOOD 1967; O'HARA 1968, 1973) is an effective process to change the light/heavy REE ratio in a residual liquid without greatly changing the major element composition. In order to produce the REE content differences between LML and UML estimated parental magma compositions (see Table 2, compositions (1) and (2)) about 60-70 % eclogite fractionation is required, and garnet must comprise approximately 40-50 % of the cumulate assemblage to keep the Yb bulk solid/liquid partition coeficient at 1.6-1.8 (see FREY *et al.* 1978, P. C. set 3). The approximately 60-70 % fractionation of garnet and clinopyroxene, required by the REE calculations, would, however, almost completely deplete the residual liquid in Sc (and also Cr) (see also, GAST 1968; FREY *et al.* 1974; BLANCHARD *et al.* 1976); consequently, it is concluded that fractional crystallization at high-pressures could generate REE distribution pattern variations similar to those observed for the analysed samples, but compatible trace element concentrations are not consistent with this model.

b.3) *Partial Melting of Upper Mantle Peridotite*

Upper mantle processes used to explain the genesis of basaltic magmas include batch melting (PRESNALL 1969), fractional (and incremental) melting (BOWEN 1928; PRESNALL 1969), continuous melting (LANGMUIR *et al.* 1977; WOOD 1979), zone refining (P. HARRIS 1974) and the generation of source heterogeneities (O'NIONS & PANKHURST 1974; BOUGAULT *et al.* 1979). I will now present these models and evaluate their utility for explaining the Iberian Pyrite Belt data.

i) *Batch Partial Melting*

It is possible to generate complex relationships between the degree of LREE enrichment and HREE abundances by batch partial melting of a peridotite containing a phase(s) that preferentially enriches HREE, for example garnet. Fig. 8A models non-modal batch melting of a LREE-enriched garnet lherzolite source; it shows that an approximately 5 % equilibrium partial melt of this source has La/Yb close to that of 'primitive' B-dolerite/UML (see table 2), whereas 15-20 % equilibrium melts have La/Yb similar to those of most A-dolerites and LML. Thus, these calculations

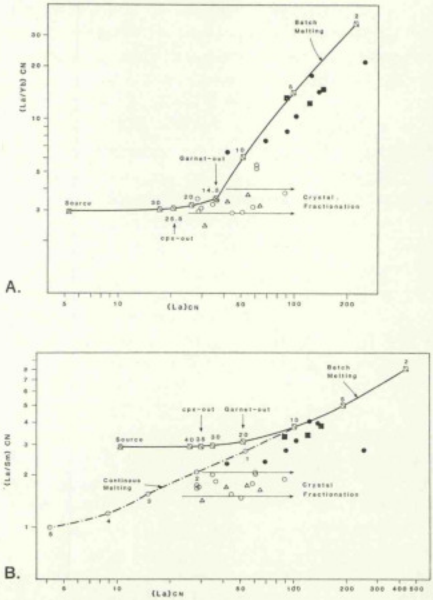


Fig. 8 — A: (La/Yb)cn — (La)cn relationships in liquids modelled by batch melting compared to the Pyrite Belt basaltic rocks. Initial peridotite mantle source mineralogy = 54.5 % olivine + 24 % orthopyroxene + 15 % clinopyroxene + 6.5 % garnet. Up to 14.5 % melting, the melt is 5 % olivine + 5 % orthopyroxene + 45 % clinopyroxene + 45 % garnet; from 14.5 % to 25.5 % melting, the melt is 5 % olivine + 5 % orthopyroxene + 90 % clinopyroxene; after 25.5 % melting the melt is 50 % olivine + 50 % orthopyroxene.

B: Continuous melting model of a LREE enriched garnet lherzolite source that yields Pyrite Belt upper mafic lavas parental REE abundances by $\simeq$ 10 % batch melting. REE patterns similar to those of most A-dolerites and lower mafic lavas can be generated by the second 3 % melting increment if 3 % basaltic liquid remains in the residue. Residual source after 10 % batch melting: 55 % olivine + 25 % orthopyroxene + 15 % clinopyroxene + 5 % garnet. During continuous melting, the melt is 5 % olivine + 10 % orthopyroxene + 40 % clinopyroxene + 45 % garnet.

indicate that it is possible to derive, from a common mantle source, tholeiitic and alkali basaltic magmas with REE distribution patterns similar to those of the Pyrite Belt basalts.

Batch partial melting cannot account for all the data, however, because of the very large incompatible trace element ratio variations among the studied samples (see Table 1; see also, ERLANK & KABLE 1976; LANGMUIR *et al.* 1977; SUN & NESBITT 1978; PEARCE & NORRY 1979; WOOD *et al.* 1979b, c), unless accessory minerals (e.g. apatite, ilmenite, rutile) are included in the residual mantle assemblage. If there was apatite in the residue P_2O_5 concentrations of the melts would be essentially independent of the extent of melting, because P is a stoichiometric element in apatite (G. HANSON & LANGMUIR 1978), but (for example) Ce concentrations would be inversely proportional to the extent of melting and percent of apatite present (SUN & G. HANSON 1975a, b). Table 1 indicates that P_2O_5 and Ce are roughly covariant for the analysed samples, indicating that apatite is not a mineral phase in the residue at the time of magma segregation. Similarly, increasing Nb contents and decreasing Zr/Nb ratios from LML/A-dolerites to UML/B-dolerites also provide arguments against the presence of residual rutile and/or ilmenite (McCALLUM & CHARETTE 1978). Thus, accessory minerals are probably not important and highly incompatible element (e.g. La, Ce, P, Nb) ratios are not likely to have been significantly modified, during batch partial melting, relative to those of their mantle source.

ii) *Fractional, Incremental and Continuous Melting*

Fractional and/or incremental melting provides an effective way of changing the incompatible trace element ratios in basaltic melts derived from an homogeneous sources (SHAW 1970). Fig. 8B models continuous melting (i.e. an incremental melting process in which a portion of the generated liquid always remain in equilibrium with the residue; see LANGMUIR *et al.* 1977) of a LREE enriched garnet lherzolite source that yelds the Pyrite Belt UML parental magma REE distribution pattern by 10 % batch partial melting; liquids with REE abundances similar to most A-dolerite/LML can be generated by the second 3 % melting increment of the residue of the initial source if 3 % basaltic liquid remains in the residue. In spite of the rather good fit for the REE data, the model is not consistent with available field relations, because the most incompatible element enriched basaltic rocks are concentrated towards the top of the volcanic sequence. If repeated melting from one source occurred in the Pyrite Belt, fractional, incremental or continuous melting seem unlikely.

iii) *Zone Refining*

Zone refining may play some role in the variation of incompatible trace element abundances, but this process cannot account for the crossing REE patterns as shown by the Pyrite Belt basaltic rocks (see Fig. 5) because zone refining requires increasing LREE enrichment with increasing HREE abundances (G. HANSON 1977).

iv) *Source Heterogeneities*

Trace element and isotopic studies of basalts (GAST 1968; KESSON 1973; SCHILLING 1973, 1975; SUN & G. HANSON 1975a); O'NIONS *et al.* 1977; TATSUMOTO 1978; WOOD *et al.* 1981) as well as the inspection of ultramafic nodules and rocks of presumed mantle origin (VARNE & GRAHAM 1971; FREY & GREEN 1974; FREY & PRINZ 1978; BOETTCHER & O'NEIL 1980; IRVING 1980; WASS & RODGERS 1980) indicate that chemical heterogeneities, on various scales, are very common in both subcontinental and suboceanic mantle.

From the above discussion of the various mantle processes and their effects on the trace element distribution in basalts, it seems difficult to produce the observed variations if all the basaltic rocks were derived from a homogeneous mantle source. Therefore, it is suggested that the variations in some trace element ratios were

Table 2 — Source composition models for Iberian Pyrite Belt basaltic rocks

	% Melt	(ppm) Ba	Nb	P	Zr	Ti	Y	Sc	V	Ni	La_{cn}	Sm_{cn}	Gd_{cn}	Yb_{cn}
Lower mafic lavas (1)		64	2.4	594	95	8585	28	28	201	363	24.9	17.7	15.2	10.2
Source for LML	15	10	0.4	90	15	1566	4.7	21	75	2334	3.8	2.9	2.5	1.8
Upper mafic lavas (2)		478	56	2277	164	14823	22	21	228	350	110.8	30.2	19.2	7.6
Source for UML	9	47	5.5	205	18	2183	4.7	25	96	2130	10.4	3.7	2.7	2.3
Primordial mantle (3)		7.56	0.62	90.4	11	1527	4.87	14-20	83-110	2000	2.0-2.3	2.0	2.0	1.6-2.5
Chondrites (4)		3.51	0.39	1000	5.6	620	2.2	5	49	9500	ppm 0.315	ppm 0.192	ppm 0.259	ppm 0.208

Olivine and plagioclase/basaltic liquid partition coefficient data for Nb, Zr, Ti and Y from PEARCE & NORRY (1979), for Ba, Sc, and V from FREY *et al.* (1978), for REE from SCHNETZLER & PHILPOTTS (1970) and DRAKE & WEILL (1975), D(P) and D(Ni) assumed as 0 and 10, respectively (see FREY *et al.* 1978). In upper mantle melting models partition coefficient data for the REE, Ba, P, Zr, Y, Sc and V from FREY *et al.* (1978; P. C. set 3), for Ti from SUN *et al.* (1979) (olivine, orthopyroxene) and PEARCE & NORRY (1979) (garnet), D(Nb) for clinopyroxene from McCALLUM & CHARETTE (1978) otherwise equal to 0.

(1) Lower mafic lava sample E-16A + 20 % (ol. + pl.). Source calculated assuming that $\overset{P}{Co} = 90$ ppm (RINGWOOD 1975; SUN *et al.* 1979; residual upper mantle mineralogy after 15 % melting = 64 % ol. + 28 % opx. + 8 % cpx.

(2) Average of 5 relatively unfractionated (Ni ⩾ 100 ppm) upper mafic lava and type-B dolerite samples + 6-13 % olivine. Source calculated according to the techniques of SUN & G. HANSON (1975b) and MINSTER & ALLEGRE (1978), assuming that (CoHREE) cn = 2.5 (KAY & GAST 1973; SUN & G. HANSON 1975b); residual upper mantle mineralogy after 9 % melting = 55 % ol. + 25 % opx. + 15 % cpx. + 5 % gt.

(3) Sources of data: RINGWOOD (1975), SUN & NESBITT (1977), WOOD *et al.* (1979b).

(4) Sources of data: SUN & G. HANSON (1976), LANGMUIR *et al.* (1977), SUN & NESBITT (1977), KAY & HUBBARD (1978), WOOD *et al.* (1979c).

(at least in part) inherited from the mantle source. This means that within the Iberian Pyrite Belt, basaltic liquids derived from different sources were once available to be either injected as sills and/or erupted as lavas.

The following discussion is a preliminary attempt to characterize and contrast the trace element geochemistry of the main basaltic rock group's mantle source and to investigate the mechanisms responsible for their chemical variability,

3.3.2c) SIGNIFICANCE OF MANTLE SOURCE HETEROGENEITIES

The trace element characteristics of the mantle source for the Iberian Pyrite Belt basalts can be estimated from the composition of the samples with most primitive incompatible element levels (see Table 2). Consideration of Ti, Zr, P and HREE concentrations in 'primitive' LML and UML samples in comparison with estimated values for the primordial mantle (SUN & NESBITT 1977; WOOD *et al.* 1979b) indicates that these rocks may represent 15 % and 9 % mantle melting, respectively. Assuming these degrees of melting, suitable residual mantle mineralogies and appropriate distribution coefficents the trace element chemistry of the mantle source for these basalts can be calculated (Table 2). Both of the modelled mantle sources are LREE enriched, but the LML modelled source is enriched in LIL elements over the other incompatible elements showing a strong relative depletion in Nb.

That most incompatible element enriched basalts have long term time integrated LREE and Rb/Sr depleted (or chondritic) sources (O'NIONS *et al.* 1977, 1978; DEPAOLO & WASSERBURG 1976; RICHARD *et al.* 1976) obviously requires secondary (relatively recent) enrichment events in their mantle sources (LLOYD & BAILEY 1975); such events can be reconciled with a veined upper mantle source region (WOOD *et al.* 1979b), in which a refractory host mantle is permeated by veins of an incompatible element enriched melt (FREY & GREEN 1974; IRVING 1980). While the UML parental magma source composition could be successfully modelled by about 5 % veining a N-type MORB source with an undersaturated basic melt of its own composition, it is evident that this process is not an effective one to generate the La/Nb fractionation as observed for LML and most type A dolerites (see Tables 1, 2 and Fig. 9).

La and Nb are not readly fractionated by partial melting processes — even by incipient melting. Nevertheless, all subduction zone magmas and some back-arc basalts have very low relative Nb contents and La-Nb fractionation has clearly occurred during their genesis (SAUNDERS *et al.* 1980; WOOD *et al.* 1981; see also Fig. 9). Current models of magma generation at subduction zones (c. f. RINGWOOD 1974; BEST 1975; DEPAOLO & JOHNSON 1979; KAY 1980) require chemical alteration of the overlying mantle wedge by addition of small proportions of melt and/or fluid phase derived by incipient melting and dehydration of the subducting lithospheric slab. The apparent decoupling of HFS and LIL group incompatible elements, which is perhaps the most distinctive feature of orogenic basalts (see Fig. 9), may reflect higher solubilities of LILE with consequent entry into fluids — thus, enriching the overying mantle —, while HFSE are retained in the residual assemblages (see ALDERTON *et al.* 1980).

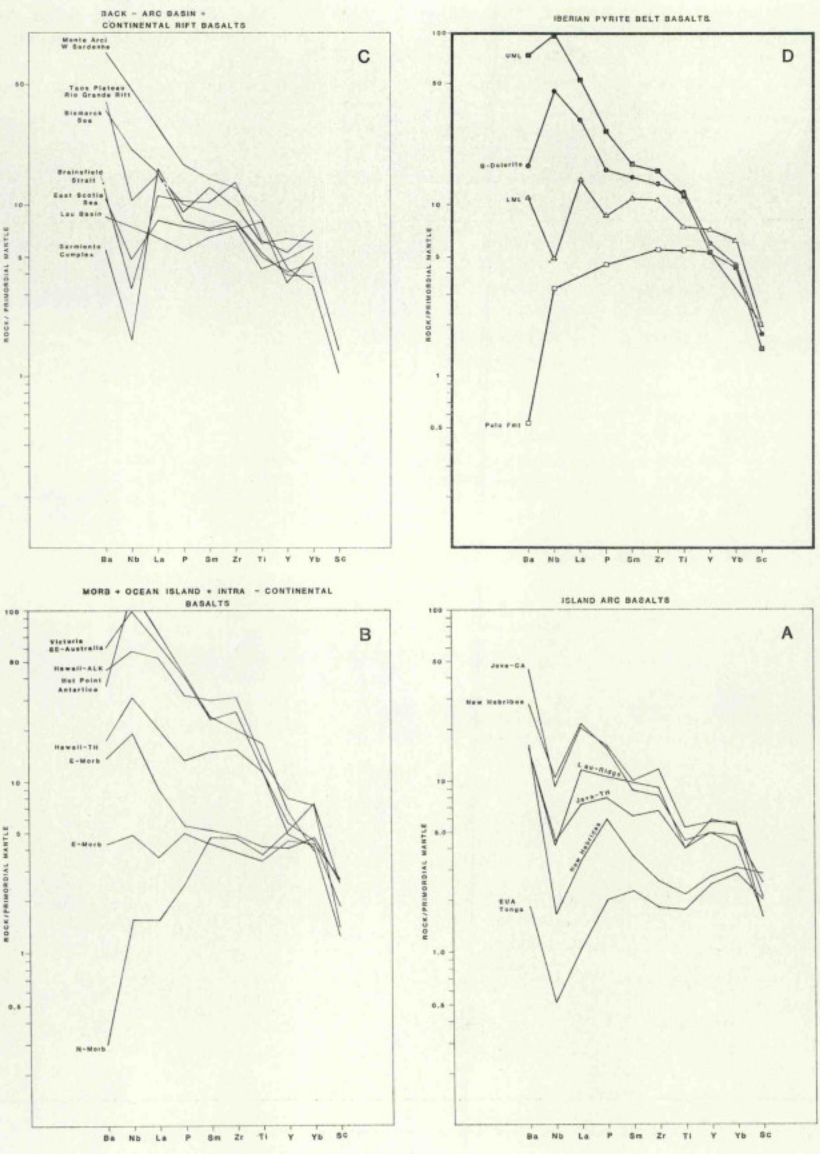
Fig. 9 — Trace element abundances in basalts normalised to primordial mantle abundances.

A: Island Arc basalts. Data sources are: EWART & BRYAN (1972), GILL (1976a), WHITFORD *et al.* (1979), DUPUY *et al.* (1982).

B: MORB, Ocean Island and Intra-Continental basalts. Data sources are: FREY *et al.* (1978), SUN *et al.* (1979), KYLE (1981).

C: Back-Arc Basin and Continental Rift basalts: data sources are: GILL (1976a), JOHNSON *et al.* (1978), SAUNDERS & TARNEY (1979), SAUNDERS *et al.* (1979), BVSP (1981), DOSTAL *et al.* (1982).

D: Iberian Pyrite Belt basaltic rocks.


BACK - ARC BASIN + CONTINENTAL RIFT BASALTS
C
Monte Arci W Sardenha
Taos Plateau Rio Grande Rift
Bismarck Sea
Brainsfield Strait
East Scotia Sea
Lau Basin
Sarmiento Complex
ROCK/ PRIMORDIAL MANTLE
Ba Nb La P Sm Zr Ti Y Yb Sc
IBERIAN PYRITE BELT BASALTS
D
UML
B-Dolerite
LML
Pulo Fmt
ROCK/PRIMORDIAL MANTLE
Ba Nb La P Sm Zr Ti Y Yb Sc
MORB + OCEAN ISLAND + INTRA - CONTINENTAL BASALTS
B
Victoria SE-Australia
Hawaii-ALK
Hut Point Antartica
Hawaii-TH
E-Morb
E-Morb
N-Morb
ROCK/PRIMORDIAL MANTLE
Ba Nb La P Sm Zr Ti Y Yb Sc
ISLAND ARC BASALTS
A
Java-CA
New Hebrides
Lau-Ridge
Java-TH
New Hebrides
EUA Tonga
ROCK/PRIMORDIAL MANTLE
Ba Nb La P Sm Zr Ti Y Yb Sc

Although the LML/A-dolerite fractionation trends are distinctily tholeiitic (see Figs 2, 3 and 9) there must be some reason for their assuming 'calc-alkaline' (orogenic) trace element (La/Nb) characteristics. The tectonic situation of back-arc spreading however is one where there is also a potential for involvement of fluids from the subducting slab. These may

Table 3 — Average major and trace element concentrations in andesitic rocks from the Iberian Pyrite Belt

	GROUP 1: NORTHERN OUTCROPS			GROUP 2: SOUTHERN OUTCROPS		
	Average	Range	N	Average	Range	N
SiO_2 (a) wt. %	61.45	56.61 — 64.02	22	60.47	57.03 — 64.28	15
TiO_2O_3	0.94	0.66 — 1.28	22	0.74	0.55 — 0.85	15
Al_2O_3	16.64	12.99 — 19.67	22	18.65	16.11 — 21.07	15
Fe_2O_3 (b)	7.08	4.89 — 8.29	22	5.70	4.63 — 6.96	15
MgO	3.67	1.38 — 6.94	22	4.18	2.40 — 6.52	15
MnO	0.10	0.04 — 0.21	22	0.07	0.04 — 0.12	14
CaO	4.95	3.39 — 7.13	22	3.86	0.39 — 6.92	15
Na_2O	3.48	2.29 — 5.66	22	5.42	3.27 — 8.06	15
K_2O	1.54	nd — 2.70	22	0.77	0.01 — 1.66	15
P_2O_5	0.15	0.06 — 0.51	22	0.14	0.07 — 0.20	15
	Trace Elements					
Cr ppm	79	11 — 195	22	56	37 — 108	15
Ni	8	nd — 39	22	24	nd — 81	15
V	174	109 — 223	22	106	87 — 134	15
Cu	49	25 — 85	9	22	nd — 65	11
Zn	77	68 — 84	9	58	20 — 113	11
Li	na			52	32 — 73	6
Rb	37	nd — 65	22	14	nd — 43	15
Sr	213	74 — 486	22	432	236 — 742	15
Ba	223	78 — 370	21	113	6 — 242	15
Y	29	22 — 44	22	16	13 — 19	15
Zr	141	116 — 206	22	107	95 — 125	15
Nb	9	3 — 18	22	7	4 — 14	15
La	20.5	16.3 — 24.8	3	14.4	12.7 — 16.1	2
Ce	40.5	33.6 — 47.2	3	31.4	26.8 — 35.9	2
Nd	21.6	17.8 — 25.0	3	16.5	14.4 — 18.6	2
Sm	4.85	4.17 — 5.51	3	3.43	2.84 — 4.02	2
Eu	1.27	1.12 — 1.39	3	0.82	0.71 — 0.93	2
Gd	5.03	4.38 — 6.16	3	2.87	2.65 — 3.09	2
Yb	3.53	3.41 — 3.68	3	1.16	1.12 — 1.20	2
	Elements Ratios					
FeOt/MgO	1.95	0.97 — 4.69	22	1.29	0.81 — 2.22	15
Zr/Y	4.94	3.91 — 5.82	22	6.93	5.05 — 9.61	15
(La/Sm)cn	2.55	2.38 — 2.74	3	2.59	2.44 — 2.73	2
(La/Yb)cn	3.84	2.82 — 4.68	3	8.24	6.99 — 9.49	2

(a) — All major elements calculated on a volatile-free basis; (b) — All Fe calculated as Fe_2O_3

affect the magma generation processes; occurrence of back-arc basalts with island arc affinities reflects their mantle source characteristics, and is most probably a geochemical reflection of the subduction process (see SAUNDERS & TARNEY 1979; SAUNDERS *et al.* 1980). Their geochemical analogies with both LML and A-dolerites from the Iberian Pyrite Belt seem evident (see Fig. 9).

3.3.3. GEOCHEMISTRY OF THE INTERMEDIATE ROCKS

Major element geochemistry of the Iberian Pyrite Belt intermediate rocks have been previously described by ROUTHIER *et al.* (1977).

Intermediate rock samples studied here were selected to be representative of the geochemical variations across the Pyrite Belt; group 1 samples come from CARVALHO's (1976) volcanic lineament D and correspond to the northern outcrops whereas group 2 samples were collected from near Almodovar and Pomarão villages corresponding to the southern outcrops (see Table 3) — the distance between northern and southern sampled outcrops ranges from 15 to 30 km across strike.

The purposes of this section are:

a) to outline the geochemical characteristics of the Pyrite Belt andesitic rocks.

b) to discuss their petrogenetic implications.

c) to test as to what extent the now available geochemical data agrees/disagrees with current plate tectonic models for the South Portuguese Zone.

3.3.3a) GEOCHEMICAL DATA

Major Elements

The content of SiO_2 (on an anhydrous basis) in the volcanic rocks described here varies within a relatively narrow range (see Table 3), with the majority of the samples having 57-64 wt. % SiO_2; according to most systems of classification they could be referred to as andesites.

The andesitic rocks from the Iberian Pyrite Belt show many similarities with comparable volcanics from orogenic belts. They are characterized by relatively high Al_2O_3 contents, but low $Fe_2O^t_3$ and TiO_2 concentrations (see Table 3). The analysed samples do not show significant Fe enrichement and fall on the SiO_2 vs FeO^t/MgO and FeO^t vs FeO^t/MgO diagrams (Fig. 10) into the field of the calc-alkaline rock series (MIYASHIRO 1974; see also MUNHÁ 1983). In detail, however, there are differences between the two rock groups and, as it is shown in Fig. 10, group 1 samples are enriched in TiO_2 and FeO^t relative to those of group 1.

As it should be expected, K_2O and Na_2O contents show a considerable variation. Nevertheless, the majority of andesites have K_2O concentrations bellow 2.5 wt. %, suggesting that they belong to the normal calc-alkaline series rather than to the high-K calc-alkaline series (c. f. JAKES & WHITE 1972).

Trace Elements

The contents of Nb, Zr, Y and Sr (considering only the less altered samples regarding Sr contents) in the analysed samples are similar to those of comparable rocks from other calc-alkaline series (c. f. JAKES & WHITE 1972; WHITFORD *et al.* 1979; see also Table 3). Rocks of group 1 tend to be higher in Zr and Y but lower in Sr than those of group 2 (Fig. 11). Y shows a positive correlation with Zr for group 1 samples; Sr concentrations decrease with increasing Y contents.

Ni and Cr concentrations decrease with increasing FeO^t/MgO values (Fig. 11); however, at low FeO^t/MgO values, rocks of group 1 have higher Cr and lower Ni contents than those of group 2. Samples analysed for V indicate a concentration range of 85-225 ppm (see Table 3), very similar to that reported for other calc-alkaline andesites (TAYLOR & WHITE 1966; JAKES & WHITE 1972); group 1 andesites are enriched in V relative to those of group 2 (Fig. 10).

REE abundances in the Iberian Pyrite Belt andesitic rock averages are plotted normalized to chondrites in Fig. 12A. The average pattern of group 1 samples show a moderate enrichment in LREE and only small fractionation of HREE, features which are typical of calc-alkaline rocks of island arcs (TAYLOR *et al.* 1969a; YAJIMA *et al.* 1972; WHITFORD *et al.* 1979). Andesites of group 2 are also enriched in LREE but have distinctly fractionated HREE; their REE distribution patterns are similar to those of some continental margin andesitic rocks (THORPE *et al.* 1976).

3.3.3b) PETROGENESIS OF THE ANDESITIC ROCKS

Some possible processes which may account for the genesis and the observed chemical variations of the andesitic rocks in the Iberian

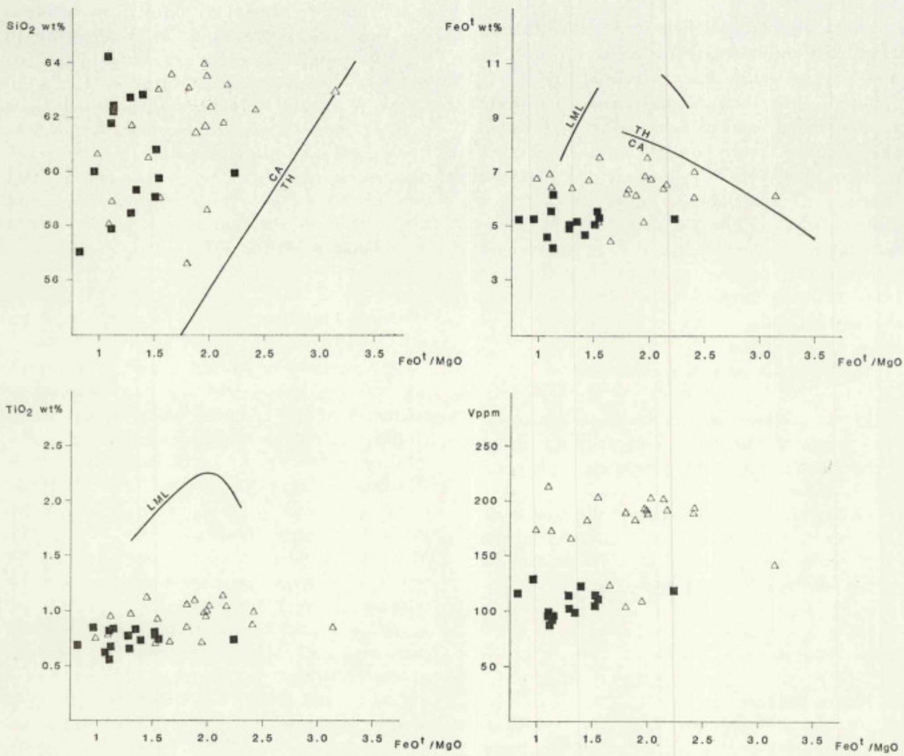


Fig. 10 — SiO_2, FeO^t, TiO_2 and V — FeO^t/MgO variation diagrams for the Iberian Pyrite Belt andesitic rocks (group 1: triangles; group 2: squares). Tholeiitic and calc-alkaline rock series compositional fields after MIYASHIRO (1974). LML — trends for lower mafic lavas.

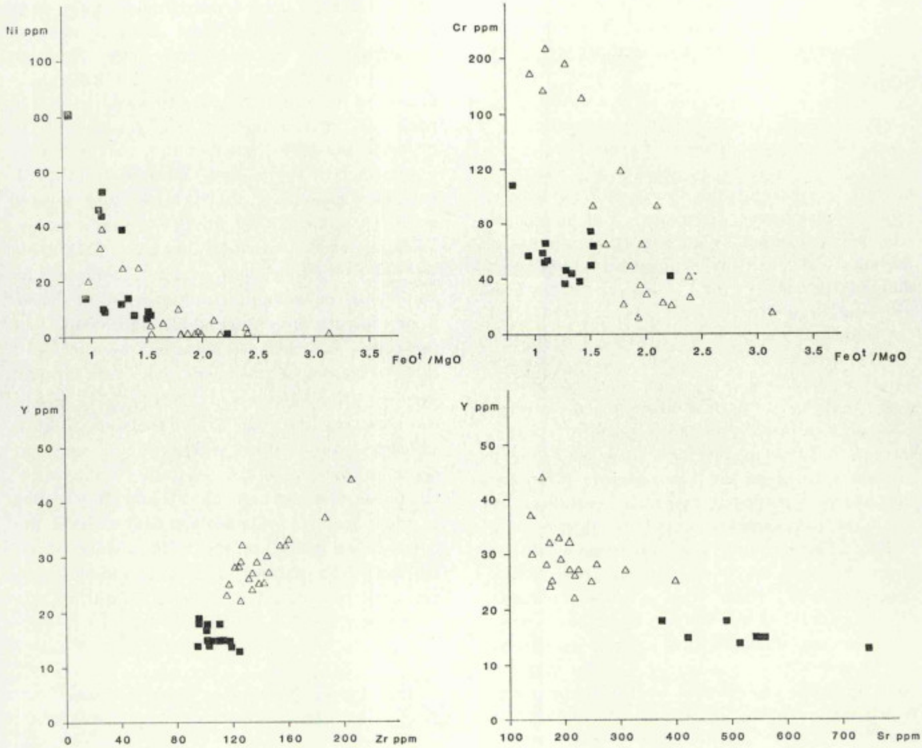


Fig. 11 — Ni, Cr — FeO^t/MgO, Y-Zr and Y-Sr variation diagrams for the Pyrite Belt andesitic rocks (Y-Sr diagram: least altered samples — non-carbonated, CaO $>$ 4 % on an anhydrous basis).

Pyrite Belt are now evaluated in the light of the geochemical data. The models examined were recently reviewed by BOETCHER (1973), RINGWOOD (1974) and GILL (1981).

i) *Anatexis of Sialic Crust*

All theories involving development of andesitic magmas from sialic material (e. g. PICHLER & ZEIL 1972; FERNANDEZ *et al.* 1973; TAYLOR & HALLBERG 1973) are difficult to reconcile with the occurrence of andesites in island arcs without pre-existing continental crust and with the spatial chemical variations in lavas across the arcs and their relationships with subduction zones (HATHERTON & DICKINSON 1969). Alternatively, mixing of basalts with crustal derived acid magmas has been suggested as a means of generating andesitic magmas (e. g. EICHELBERGER 1974, 1975); however, there is not field evidence for magma mixing in the Pyrite Belt and available geochemical data is not appropriate for evaluating this model. Therefore, this hypothesis will not be discussed further.

ii) *Crystal Fractionation of Basaltic Magmas at Shallow Levels*

Low pressure crystal fractionation involving olivine, plagioclase, pyroxene and magnetite has been purposed as an important process in deriving andesite from a more primitive basalt magma (OSBORNE 1969). Such an origin was also suggested by ROUTHIER *et al.* (1977) for the andesitic rocks in the Iberian Pyrite Belt.

Clinopyroxene and whole rock chemical variation diagrams (Figs 1 and 10) show, however, that basalt and andesite trends diverge considerably; also, high Ni, Cr and V concentrations observed in some andesites (see Table 3), precludes an origin by low pressure fractionation of basalt involving extensive crystallization of olivine, pyroxene and magnetite (TAYLOR *et al.* 1969b). Furthermore, the geochemical differences between the two andesitic rock groups described here render such a process unlikely for the genesis of the suite as a whole. I therefore conclude that andesites are not derived by low pressure fractionation of basaltic magmas such as those represented by the mafic rocks which occur in the Iberian Pyrite Belt. It is possible, however, that some of the within group chemical variation might reflect modification by low-P fractionation involving plagioclase, clinopyroxene and, probably, olivine and magnetite.

iii) *'Eclogite' or Amphibole Controlled Fractionation*

As summarized by NOBLE *et al.* (1975), 'recent studies have cast serious doubt on the assumption that andesite and related rocks of the calc-alkaline suite are largely unmodified products of partial melting of subducted ocean floor basalt'. Typically, models incorporating amphibole and/or garnet + clinopyroxene as residual phases, involve partial melting (30-40 % melting) of subducted ocean floor tholeiite in amphibolite or eclogite mineralogy. At such high degrees of melting the relative REE abundances of the liquid approach those of the source rock and, in consequence, it is not possible to generate the LREE fractionation that is typical of the Iberian Pyrite Belt andesites from LREE depleted ocean floor tholeiite (see also THORPE *et al.* 1976; DOSTAL *et al.* 1977). Likewise, the available trace element $D^{s/1}$ data (see GILL 1978) are not compatible with andesite generation by amphibole or eclogite fractionation of ocean floor tholeiite. It is thus highly improbable that the parent magmas of andesites, there included those occurring in the Pyrite Belt, were directly derived from subducted ocean floor basalt. Nevertheless, the high Sr contents of group 2 andesites would reflect segregation at sufficient depth so that plagioclase was not a stable residual phase and the high La/Yb and Zr/Y ratios as well as low Y contents would suggest the presence of garnet in the residue. In contrast the flat HREE pattern of group 1 samples indicates that garnet did not play a significant role in the petrogenesis of these rocks.

iv) *Partial Melting of Hydrous Upper Mantle Peridotite*

The strong objections to the amphibolite/ecologite melting models indicate that the ultimate origin of calc-alkaline lavas must be within the mantle. Recent experimental evidence suggests that melting of hydrous mantle could play an important role in the genesis of andesitic magmas (KUSHIRO 1974; NICHOLS 1974; MYSEN & BOETCHER 1975a, b; TATSUMI 1982). Magmatic liquids equilibrated with upper mantle peridotite should be expected, however, to have higher contents of Ni and Cr than those observed in the andesitic volcanics from the Iberian Pyrite Belt (see TATSUMI & ISHIZAKA 1982); thus, it is probable that partial melting was followed by some degree of fractional crystallization.

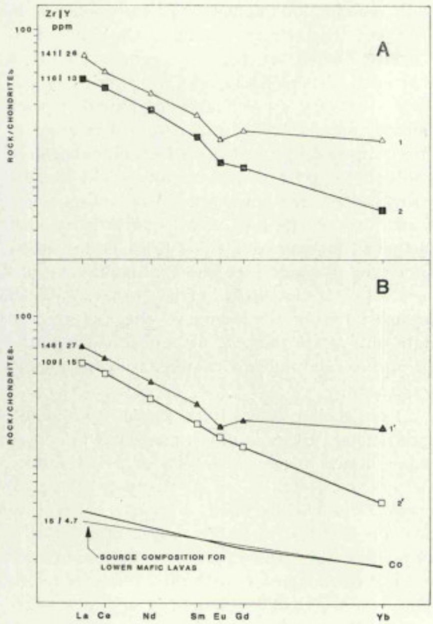


Fig. 12 — A: Average chondrite-normalized REE patterns of andesitic rocks from the Iberian Pyrite Belt. 1) Group 1 samples; 2) Group 2 samples.

B: Petrogenetic models for the Iberian Pyrite Belt andesitic rocks. Co — Initial upper mantle source composition.

1': 8 % batch partial melting (residue: 60 % olivine + 20 % orthopyroxene + 10 % clinopiroxene + 10 % amphibole) followed by 35 % fractional crystallization (30 % olivine + 55 % plagioclase + 25 % clinopyroxene + 10 % amphibole).

2': 8 % batch partial melting (residue: 53 % olivine + 20 % orthopyroxene + 10 % clinopyroxene + 10 % amphibole + 7 % garnet) followed by 10 % fractional crystallization (40 % olivine + 20 % plagioclase + 40 % clinopyroxene). Solid/liquid partition coefficient data from LOPEZ *et al.* (1977), FREY *et al.* (1978) and PEARCE & NORRY (1979).

Derivation of andesites with the trace element abundances described earlier from mantle peridotites with a (x2-3) chondritic trace element pattern raises a problem analogous to that of derivation from ocean floor tholeiite basaltic source, i.e., very low degrees of melting (<3 %) would be required to produce the observed abundances of LILE, while several refractory phases would have to be involved to explain the Nb, P, Ti and Zr contents of the andesitic rocks. Therefore, rather than to purpose a very low degree of melting, I prefer to consider the evidence obtained from the basaltic rocks (see Table 2) and, accordingly, suggest that the andesites may be also products of partial melting of a LILE enriched upper mantle peridotite. For example, if allowance is made for magmas to segregate at different depths (i.e., above and bellow garnet stability field) then (see Fig. 12B) about 8 % melting of an hydrated LREE enriched mantle source, compositionally equivalent to that estimated for the LML, will yeld liquids with trace element distribution patterns similar to those of the andesitic volcanics. As already discussed in relation to the petrogenesis of the Pyrite Belt basaltic rocks, a plausible mechanism for generating an enriched source in orogenic regions is contamination of overlying peridotite by a fluid phase derived from the downgoing subducted plate.

From experimental studies on the melting relations of hydrous peridotite MYSEN & BOETCHER (1975a, b) concluded that it would be possible to produce almost any liquid, from andesite to olivine nephelinite, by partial melting of mantle peridotite given the appropriate starting material composition, temperature, pressure, fH_2, fO_2, fH_2O and fCO_2.

Andesites, LML and type-A dolerites in the Pyrite Belt are closely related in space and time; they also exhibit similar, 'anomalous', low concentrations of some incompatible elements (e. g. Nb) and available geochemical data suggest that they could be derived from a (compositionally) similar mantle source. Thus, although apparently related from the point of view of their genesis, fractional crystallization is hardly the mechanism that relates them. An alternative hypothesis, within the framework of MYSEN & BOETCHER (1975a, b) (see also TATSUMI 1982) model, would be to derive the andesitic magmas as the initial H_2O-rich peridotite melt fraction while lowering of aH_2O, for exemple by increasing the degree of melting, would result in basaltic liquids. Such andesite and basalt will exhibit close similarity in the geochemical parameters listed above.

3.3.3c) GEOCHEMICAL CONSTRAINTS REGARDING PLATE TECTONIC MODELS FOR THE SOUTH PORTUGUESE ZONE: DISCUSSION

Most plate tectonic models for the Iberian Variscan chain (BARD 1971; CARVALHO 1971, 1976; BARD *et al.* 1973; VEGAS & MUÑOZ, 1976) interpreted, with minor variants, the tectonic and volcanic features of the South Portuguese Zone as characteristic of destructive plate margins. Specifically, VEGAS & MUÑOZ (1976) equated the Iberian Pyrite Belt volcanism with the initial evolutionary stages of an island arc whereas, CARVALHO (1971, 1976) used an 'Andean type' model and interpreted the volcanic space-time relations in terms of a northward dipping subduction zone underlying an Hercynian continental margin lying somewhere south of the Pyrite Belt.

Modern volcanic rocks of the calc-alkaline series occur characteristically, but not exclusively (see R. HANSON & AL-SHAIEB 1980), in island arcs and active continental margins that overlie subduction zones (MIYASHIRO 1974). The most common volcanic rock type in many of these orogenic belts is andesite.

A zonal arrangement in the petrography and geochemistry of island arc volcanic rocks has been recognized for many years, particularly in the circum Pacific arcs (c. f. KUNO 1959; SUGIMURA 1968). Comprehensive recent surveys of the geochemical variations across island arcs and active continental margins (c. f. DOSTAL *et al.* 1977; WHITFORD *et al.* 1979; GILL 1981) emphasized, among other features, the increase in Sr contents, Zr/Y and La/Yb ratios and decrease in Y, HREE and maximum Fe contents of lavas with increasing depth to the Benioff zone.

If the spatial chemical variations in lavas across modern arcs and their relationships with subduction zones are extrapolated to interpret the tectonic setting in the South Portuguese Zone then, from the geochemical data discussed above (3.3.3a; see also Table 3), it is clear that the (presumable) subduction zone polarity should have been reversed relative to that proposed by most plate tectonic models. In addition, the geochemical features of the basaltic rocks which are contemporaneous with andesites in the Pyrite Belt, clearly differ from what is typical for calc-alkaline and tholeiitic basalts occurring in island arcs and continental margins (see also MUNHÁ 1979; FLOYD 1982). Thus, it seems difficult to reconcile the petrological data with plate tectonic models proposing simple direct relationships between subduction and volcanism.

The geochemistry of the Iberian Pyrite Belt intermediate rocks is not consistent with single stage models suggested for the origin of andesitic volcanics such as simple melting of typical ocean ridge basalts or of peridotite believed to be representative of the (average) upper mantle (x2 chondrites). The volcanic rocks are enriched in LILE in comparison with the calculated melts. Thus, the high LILE contents in andesitic rocks of the Iberian Pyrite Belt may suggest their derivation from an enriched source. Model calculations and experimental data indicate that a suitable source for andesites could be the hydrous equivalent to upper mantle peridotite source of contemporaneous LML and A-dolerite basaltic rocks. Interestingly enough, these basalts also share some geochemical features with orogenic lavas (e. g. high La/Nb values), although still showing fundamental differences relative to calc-alkaline and low-K tholeiite series of active continental margins and island arcs. Rock suites displaying these complex chemical characteristics seem to be typical of the initial stages of back-arc spreading (see GILL 1976a, b; WEAVER *et al.* 1979) and, presumably, reflect the progressive return of sub-arc (or sub-continental) mantle to physical and chemical conditions unaffected by subduction. Within such an environment, fluids derived from the dehydrating subducted slab may penetrate locally the source regions of the back-arc

Table 4 — Major and trace element concentrations in felsic metavolcanic rocks from the Iberian Pyrite Belt

	DACITES			RHYOLITES			High-SiO_2 RHYOLITES			CERCAL RHYOLITES		
	Average	Range	N	Average	Range	N	Average	Range	N	Average	Range	N
SiO_2 (a) wt. %	67.27	66.25 — 68.20	6	74.64	69.24 — 77.47	17	79.92	77.76 — 81.94	12	78.54	75.46 — 83.41	16
TiO_2	0.67	0.38 — 0.98	6	0.28	0.12 — 0.60	17	0.08	0.03 — 0.37	12	0.08	nd — 0.13	16
Al_2O_3	16.09	14.87 — 17.26	6	13.11	12.09 — 14.64	17	11.47	9.98 — 13.86	12	10.66	8.63 — 12.87	16
Fe_2O_3 (b)	4.47	1.25 — 6.33	6	2.67	0.93 — 4.68	17	1.05	0.29 — 1.89	12	1.68	0.06 — 3.90	16
MgO	1.81	0.41 — 3.85	6	0.86	nd — 2.69	17	0.34	nd. — 1.27	12	0.11	nd — 1.09	16
MnO	0.20	0.01 — 0.54	6	0.05	0.01 — 0.25	17	0.09	0.01 — 0.95	12	0.02	nd — 0.16	16
CaO	1.42	0.25 — 2.51	6	1.09	nd — 5.06	17	0.32	nd — 1.92	12	0.01	nd — 0.11	16
Na_2O	4.04	1.34 — 5.55	6	3.59	0.40 — 5.83	17	5.63	0.64 — 8.21	12	0.39	0.09 — 1.74	16
K_2O	4.03	1.19 — 10.79	6	3.66	nd — 8.21	17	1.09	nd — 4.28	12	8.59	6.72 — 10.94	16
P_2O_5	0.12	0.07 — 0.18	6	0.05	nd — 0.14	17	$<$ 0.01	nd — 0.05	12	$<$ 0.01	nd — $<$ 0.01	16
Trace Elements												
Cr ppm	17	5 — 41	6	18	nd — 64	17	6	nd — 18	12	2	nd — 10	16
Ni	nd		6	3	nd — 25	17	nd		12	6	nd — 16	16
Rb	77	37 — 168	6	82	nd — 172	17	32	nd — 131	12	219	166 — 304	16
Sr	125	23 — 253	6	99	21 — 473	17	177	16 — 554	12	15	nd — 35	16
Ba	344	237 — 575	6	350	22 — 684	17	253	19 — 952	12	393	87 — 605	16
Y	52	31 — 65	6	38	12 — 61	17	41	25 — 63	12	92	68 — 113	16
Zr	242	171 — 324	6	145	93 — 214	17	75	49 — 146	12	360	177 — 456	16
Nb	14	7 — 17	6	15	5 — 39	17	22	15 — 31	12	50	38 — 75	16
La	49.4	37.2 — 59.2	3	42.3	32.0 — 48.2	3	17.6	12.3 — 22.8	3	28.1	1.21 — 52.9	7
Ce	98.6	74.7 — 118	3	82.1	63.2 — 98.4	3	36.4	23.4 — 48.9	3	60.9	3.24 — 121	7
Nd	50.5	37.9 — 60.8	3	36.7	25.4 — 45.5	3	18.7	11.4 — 26.3	3	30.2	1.36 — 65.6	7
Sm	10.7	7.74 — 12.9	3	6.93	5.02 — 8.76	3	5.01	3.60 — 6.40	3	8.01	0.94 — 17.0	7
Eu	1.89	1.06 — 2.48	3	1.35	0.97 — 1.62	3	0.35	0.26 — 0.48	3	0.40	0.01 — 1.05	7
Gd	9.87	7.06 — 12.4	3	6.04	5.09 — 7.32	3	5.08	3.80 — 6.32	3	9.86	1.86 — 19.3	7
Yb	6.03	4.29 — 7.40	3	3.21	2.62 — 4.09	3	4.15	3.64 — 4.90	3	11.9	9.4 — 14.1	7

(a) — All major elements calculated on a volatile-free basis; (b) — All Fe calculated as Fe_2O_3

basalts producing the necessary conditions for andesitic magma generation.

3.3.4. GEOCHEMISTRY OF THE FELSIC ROCKS

In terms of relative volume of erupted magma, the felsic rocks are greatly dominant in the Iberian Pyrite Belt. Their major element geochemistry has been described by several previous studies (RAMBAUD 1969; STRAUSS 1970; SOLER 1973; SCHERMERHORN 1976; ROUTHIER *et al.* 1977) and Sr isotope data was provided by HAMET & DELCEY (1972) and PRIEM *et al.* (1978).

Opinions diverge in what concerns the origin of felsic volcanics in the Iberian Pyrite Belt. Thus, SCHERMERHORN (1970, 1975, 1976), SOLER (1973) and HAMET & DELCEY (1972) recognizing the essentially bimodal character of the Iberian Pyrite Belt volcanism and the great dominance of the felsic over the mafic/intermediate lithotypes, suggested a derivation by anatexis of continental crust, whereas ROUTHIER *et al.* (1977) favoured, in contrast, an origin by fractionation of a more primitive basaltic magma. It is expected that the new geochemical data provided here (see Table 4) will help to elucidate some of the petrological problems involved in the formation of the felsic volcanics in this province.

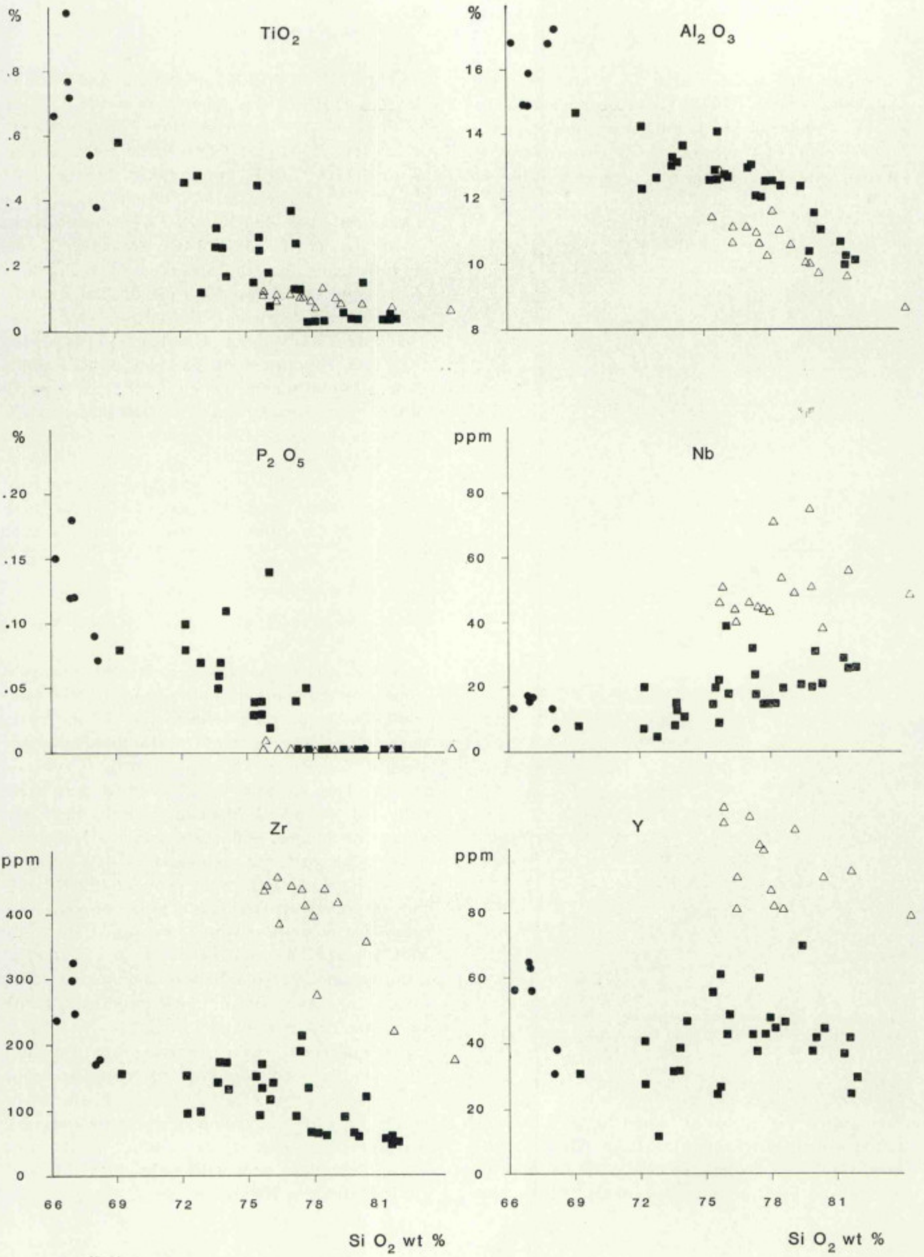
All the rock specimens studied here have suffered some degree of secondary alteration (hydrothermal/regional metamorphism) involving considerable changes on the alkali and alkali-earth element concentrations (see MUNHÁ *et al.* 1980), parameters that are fundamental when dealing with felsic rock igneous geochemistry. These features preclude a detailed petrogenetic discussion about the Pyrite Belt felsic volcanics. Therefore, we will only examine the most obvious petrological consequences of the 'immobile' element data.

As pointed out by previous authors (c. f. SCHERMERHORN 1975) volcanic rocks of dacitic composition seem to be subordinate in the Pyrite Belt, siliceous rhyolites being the dominant felsic lithotype. The rhyolitic rocks rarely contain less than 70 wt. % SiO_2 and many samples contain more than 77 wt. % SiO_2 (see Fig. 13). High SiO_2 contents also seem to be characteristic of rhyolites of other bimodal basalt-rhyolite suites which are petrologically distinct from rhyolites of calc-alkaline predominantly andesitic suites (CHRISTIANSEN & LIPMAN 1972), the latter containing usually less than 72 wt. % SiO_2. Thus, it seems possible that many of the rhyolites are not directly related to dacites (or to more primitive rocks).

Relative to most rhyolites, dacitic rocks are enriched in Al_2O_3, TiO_2 and P_2O_5 (see Fig. 13 and Table 4), and, except for the evolved rhyolites from the Cercal-Odemira region, they also range up to higher Zr and Y concentrations; Nb contents are, however, lower than those of most high-SiO_2 rhyolites (see Fig. 13). The average REE abundances are plotted normalized to chondrites in Fig. 14A which shows that the dacites REE relative distribution patterns are virtually identical to those of group 1 andesites, although displaying higher total REE abundances and larger Eu negative anomalies; higher total REE correlates well with Zr and Y concentrations and suggest that some dacites could be derived by fractional crystallization of an andesitic magma such as that represented by group 1 andesites.

In the rhyolites, Al_2O_3, TiO_2 and P_2O_5 all decrease with increasing SiO_2 contents (Fig. 13; see also Table 4). Zr, Y and Nb generally increase with the increase of SiO_2; however, the most siliceous samples are depleted in Zr and Y. The abundances of Zr (except for the Cercal-Odemira rhyolites, to be discussed later) are similar to those of the rhyolites of New Zealand (EWART *et al.* 1968b) with comparable SiO_2 contents. On the other hand, the Pyrite Belt

Fig. 13 — TiO_2, Al_2O_3, P_2O_5, Nb, Zr, Y — SiO_2 relationships for the Iberian Pyrite Belt felsic rocks. Closed circles: dacites; closed squares: rhyolites; triangles: Cercal rhyolites.



%
TiO2
.8
.6
.4
.2
0
%
Al2 O3
16
14
12
10
8
%
P2 O5
.20
.15
.10
.05
0
ppm
Nb
80
60
40
20
0
Zr
ppm
400
300
200
100
0
Y
ppm
80
60
40
20
0
66
69
72
75
78
81
Si O2 wt %

rhyolites tend to be higher in Y and Nb than equivalent rocks from New Zealand (EWART *et al.* (1968b) being similar, or only slightly lower than the average granites (TAYLOR & WHITE 1966); high-SiO_2 rhyolites (see Table 4)

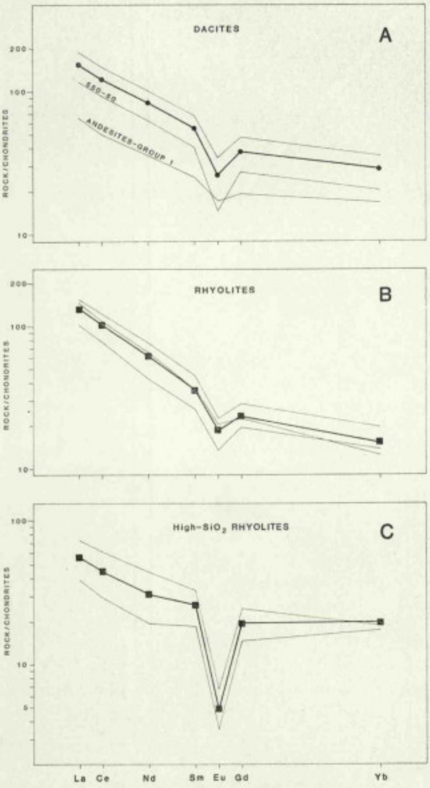


Fig. 14 — Average (and representative) chondrite-normalized REE patterns of Iberian Pyrite Belt felsic rocks.

have Zr, Y and Nb contents which compare favourably with those of some highly differentiated subalkaline rhyolites from Glass Mountain, California (NOBLE *et al.* 1972). Chrondrite normalized REE abundances of the Pyrite Belt rhyolites show LREE enrichment and variable degrees of HREE fractionation (Fig. 14B), similar to those of rhyolites from New Zealand (EWART *et al.* 1968b). High-SiO_2 rhyolitic rocks display a moderate LREE enrichment, though with lower REE concentrations, larger negative Eu anomalies and higher HREE concentrations (lower La/Yb) (see Fig. 14C). High HREE contents indicate that the observed Zr depletion is not related to the fractionation of zircon, which is distinctly enriched in HREE (NAGASAWA 1970; see also MAHOOD & HILDRETH 1983). The high SiO_2 rhyolitic rock patterns of REE abundances fall within the range of REE patterns for other high-silica rhyolites (HILDRETH 1979; BACON & DUFFIELD 1981), some evolved granites (EMMERMAN *et al.* 1975), and aplites (MITTLEFEHLDT & MILLER 1983) and (if reflecting their original composition) suggest that similar processes have contributed to the fromation of such magmas.

Rhyolites from the Cercal-Odemira Region

Felsic rocks from the Cercal-Odemira region constitute the southernmost occurrences of Hercynian volcanic rocks within the South Portuguese Zone. The geology of the area has been studied by KLEYN (1960) and by CARVALHO (1976). The volcanic rocks studied here were collected at an abandoned quarry, near the village of S. Luís, and are representative of the lower felsic tuff unit, as described by CARVALHO (1976). Thin sections often show quartz-K feldspar intergrowths and aggregates of very fine grained iron oxides, which are similar to those that replace original granular aggregates of groundmass ferro-magnesian silicates in silicic lavas that have cooled under oxidizing conditions (see NOBLE *et al.* 1978).

Compared to other Pyrite Belt rhyolites with similar SiO_2 contents, the rhyolitic rocks from the Cercal-Odemira region seem to be highly evolved and some of their geochemical features (see Table 4) are similar to those of high-SiO_2 mildly peralkaline rhyolites (MAHOOD 1981; MAHOOD & HILDRETH 1983). The rocks are somewhat low in Al_2O_3 (Fig. 13, compare NOBLE 1968) and show significant enrichment in Zr, Y and Nb over calc-alkaline rhyolites/granites (TAYLOR & WHITE 1966) although considerably less so than many comendites, pantellerites (EWART *et al.* 1982a; NOBLE & PARKER 1975; L. SMITH *et al.* 1977) and peralkaline granites (BOWDEN & TURNER 1974). REE relative distribution patterns (Fig. 15) are similar to those of many peralkaline silic rocks being characterized by strong HREE enrichment and marked Eu depletion (see BOWDEN & WHITLEY 1974; NOBLE *et al.* 1979). High contents of Zr, Y and HREE have been interpreted as a consequence of the excess of alkalies in the magma, which stabilizes the highly charged cations within the melt (DIETRICH 1968; WATSON 1979)

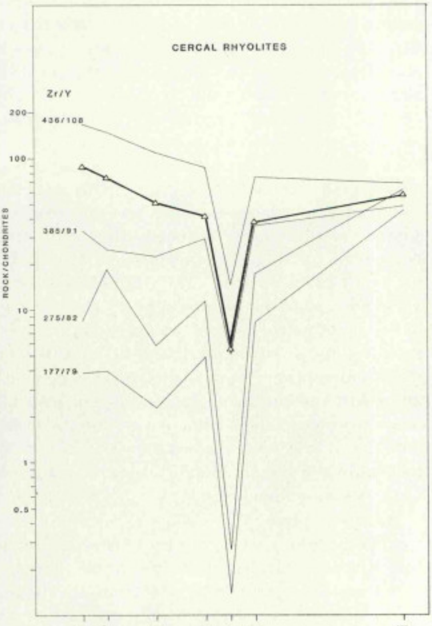


Fig. 15 — Average (and representative) chondrite-normalized REE patterns and Zr, Y contents of Cercal rhyolites.

while very extensive feldspar fractionation, under low fO_2, result in extreme Eu depletion, as it is typical of many peralkaline oversaturated magmas. The most strikingly Eu depleted specimen has Eu/Eu* of about 0.02 (compare NOBLE *et al.* 1979); this, and other similar samples, also have lower Zr contents and display appreciable LREE depletion (Fig. 15) being characterized by complex REE patterns, with a general positive slope on chondrite-normalized plots, similar (though with lower absolute abundances) to those observed for some peralkaline granites (see BOWDEN & WHITLEY 1974; N. HARRIS & MARRINER 1980).

In view of the extensive hydrothermal alteration exhibited by these rocks (see MUNHÁ *et al.* 1980) one might consider that the observed trace element variations could result from secondary processes; however, ALDERTON *et al.* (1980) have shown that the development of secondary K-feldspar, which is a characteristic feature of the Cercal rhyolites alteration, does not change appreciably the general shape of the REE distribution patterns (and Zr abundances) (REE data — Fig. 14A — for K-enriched sample 550-50 apparently supports their observations) and, in consequence, it is possible that the peculiar trace element characteristics described above reflect primary igneous features rather than the secondary processes.

PETROGENESIS

The two most frequently invoked models for the origin of calc-alkaline magmas of dacitic to rhyolitic composition are fractional crystallization of basaltic or andesitic magma and partial melting of crustal rocks (c. f. EWART & STIPP 1968; EWART *et al.* 1973).

a) *Fractional Crystallization*

According to ROUTHIER *et al.* (1977) crystal fractionation of a parental basaltic magma could be a viable mechanism to produce the whole spectrum of volcanic rocks — from basalts to rhyolites — occurring within the Pyrite Belt. Geochemical data discussed in previous sections of this study shows, however, that a simple fractional crystallization model cannot account for the observed mineralogical and chemical heterogeneities both within and between basic and intermediate rock groups. One of the main arguments invoked by ROUTHIER *et al.* (1977) to derive the rhyolites by crystal fractionation is based on the considerable scatter displayed by these rocks relative to the minimum melt composition within the system Qz — Ab — Or (TUTTLE & BOWEN 1958). Obviously, this argument does not consider the extreme Na_2O/K_2O changes suffered by the felsic volcanics during hydrothermal metamorphism (AYE 1974; MUNHÁ *et al.* 1980). Moreover, many geological data suggest that rhyolitic melts do not form under the minimum melt conditions found by TUTTLE & BOWEN (1958), (J. BROWN & FYFE 1970).

There is, however, geochemical evidence that some dacitic rocks may be produced by differentiation of an andesitic magma. On the other hand, the large volume of rhyolites places a constraint on the possible mechanisms of their origin. The estimated volumetric relationships between basaltic-andesitic rocks and rhyolites in the Pyrite Belt, where rhyolites are about 3-4 times more abundant on the surface, argue against the generation of rhyolites by fractional crystallization of basaltic or andesitic magmas (see also EWART *et al.* 1968b; COULON *et al.* 1978). Furthermore, the relative differences in the trace element abundances between dacites and rhyolites are not readily explained by the fractional crystallization process.

b) *Partial Melting of the Crust*

Numerous experimental studies (e. g. J. BROWN & FYFE 1970; WINKLER 1976; WYLLIE *et al.* 1976; WYLLIE 1977) have shown that common lithological constituents of the continental crust are potential source rocks for rhyolitic magmas. Initial $^{87}Sr/^{86}Sr$ as high as 0.7135 (PRIEM *et al.* 1978; see also HAMET & DELCEY 1972) suggests that an origin by crustal anatexis may be also applicable to the Iberian Pyrite Belt rhyolites. Locally enhanced geothermal gradients within the continental crust, probably related to ascending basic magmas, could easly produce the necessary heat supply for crustal melting. Such mechanism might also explain the close association of felsic and mafic volcanics in the Pyrite Belt.

In order to evaluate the compatibility of this process with the observed abundances of trace elements in rhyolites, a partial melting model for REE has been calculated (Fig. 16). The average REE composition of the continental crust (TAYLOR 1964) was taken as the parental material and mineral assemblages of the parents tested correspond to garnet-bearing rocks of granitic to tonalitic composition. It is apparent (Fig. 16) that partial melting of any of these source rocks can produce the REE patterns of rhyolites but relatively high degrees of fusion must be envisioned to explain the absolute REE abundances of the resulting melts. Low concentrations of P and Zr in the Pyrite Belt rhyolites indicate that the corresponding original melts must have been saturated with respect to apatite and zircon (see WATSON 1979) and the occurrence of these minerals in the residual source would have reduced the estimated degrees of melting. The calculations also allow for small amounts of residual garnet, in accordance with its rare occurrence as phenocrysts in the rhyolitic rocks. Furthermore, the origin of rhyolites by anatexis of granites, tonalites or their metamorphic equivalents is also compatible with geophysical data that suggests a felsic to intermediate composition for the crust underlying the volcano-sedimentary cover of the South Portuguese Zone (see MUELLER *et al.* 1973; SCHERMERHORN 1975).

Differentiation of silicic magmas has commonly been ascribed to processes of crystal fractionation and, specifically, to the gravitational settling of phenocrysts from the upper to the lower parts of the magma chambers (see COX *et al.* 1979). HILDRETH (1979), however, has shown that the vertical chemical variations within the magma that erupted to produce the

Bishop tuff cannot be explained by the crystallization and settling of the phenocrysts now present in the unit and, in consequence, he has suggested that most aspects of compositional variations within very silicic magmas were produced by differentiation in an essentially liquid state as a result of diffusion driven by gradients in temperature and gravitational potential. Several chemical differences among closely associated rhyolite samples from the Pyrite Belt are very similar to those described by HILDRETH (1979) and may reflect similar processes; however, the available data do not allow to determine if differentiation was produced by crystal fractionation, diffusion or by a combination of the two mechanisms (see also MAHOOD & HILDRETH 1983; MITTLEFEHLDT & MILLER 1983).

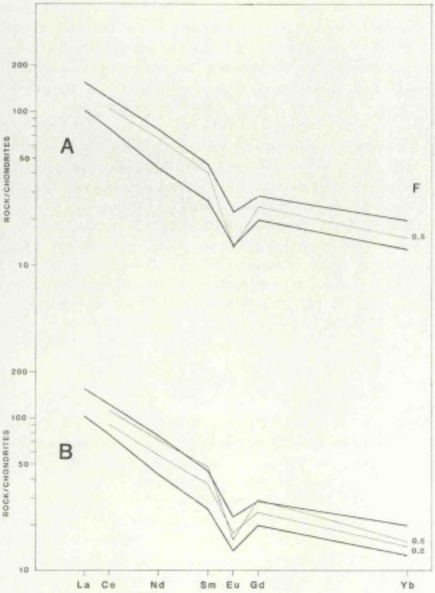


Fig. 16 — Chondrite-normalized REE contents of Iberian Pyrite Belt rhyolite rocks compared to those of melts produced by anatexis of (A) granitic (residual mineralogy: 30 % quartz + 40 % plagioclase + 20 % K-feldspar + 8 % biotite + 2 % garnet) and (B) tonalitic rocks (residual mineralogy: 20 % quartz + 60 % plagioclase + 14 % biotite + 5 % amphibole + 1 % garnet). REE contents of the parental source are those of the average continental curst (TAYLOR 1964). Partition coefficients of ARTH (1976).

The rhyolitic rocks from the Cercal-Odemira region appear to be unrelated genetically to the remaining, largely predominant, felsic volcanics in the Iberian Pyrite Belt. It is not possible to model satisfactorily the trace element variations among these rocks by fractional crystallization and significant differences in Zr, Y, Nb and REE at a given SiO_2 content strongly suggest that the Cercal rhyolites had a different petrogenetic history. These rhyolites are clearly evolved rocks and their trace element composition has many of the characteristics of highly differentiated silicic magmas erupted from crustal reservoirs. The processes acting to produce these extreme compositions have, however, obscured any record of parental magma and source rocks.

4. MAGMATIC EVOLUTION OF THE IBERIAN PYRITE BELT: SOME CONCLUDING REMARKS

Hercynian volcanism in the Iberian Pyrite Belt is essentially representative of a bimodal association of tholeiitic to alkalic basalts and siliceous rhyolites, with only subordinate andesitic/dacitic lithotypes. Although closely associated, the felsic and mafic volcanics originated and evolved separately; no lithological transitions occur, the volcanic centers are distinct, and available petrological data indicates that basalts, andesites and rhyolites are not linked by fractional crystallization. The source for the mafic magmas must be sought in the upper mantle whereas the felsic volcanism derives from magma chambers developed by melting in the crust, possibly by heat supplied by rising mafic magmas.

Geochemical studies show that the basaltic lavas occurring at the base of the volcanic sequence are tholeiitic with geochemical characteristics transitional to arc tholeiites, similar to some basalts erupted during the initial stages of back-arc spreading. However, toward the top of the VS Complex basalts/dolerites display a progressive enrichment in incompatible elements (P, Zr, Nb, LREE) and the upper mafic lavas are typical 'within-plate' alkaline basalts characteristic of continental rift zones and some ocean islands. The continuum of chemical compositions displayed by the pyrite Belt basaltic rocks cannot be accounted for in terms of fractional crystallization. It is considered that the basalts originated initially as a series of partial melts of heterogeneous mantle peridotite over a range of depths and with varying degrees of melting, followed by varying degrees of independent fractionation within each primitive magma. Petrological modelling indicates a complex chemical evolution for the Pyrite Belt basalts source regions, probably reflecting progressive introduction of non-depleted material into the upper mantle as a result of differentiation processes involving the movement of incompatible trace element rich melts or fluid phases.

The type of volcanic activity which characterizes the Iberian Pyrite Belt is not what we usually find as subduction related volcanism in orogenic areas. Basalt-rhyolite associations, similar to that described above, are commonly found in areas of extensional tectonics produced by rifting or back-arc spreading, within and/or near continental plate margins (NOBLE 1972; R. MACDONALD 1975; RANKIN 1975). By analogy with recent examples (e. g. CHRISTIANSEN & LIPMAN 1972; BEST & BRIMHALL 1974; GILL 1976a; L. SMITH *et al.* 1977; SAUNDERS & TARNEY 1979; CAMERON *et al.* 1980), it is suggested that the particular volcanic rock types which occur in the upper Paleozoic geosynclinal sequence of the Iberian Pyrite Belt, including andesites, transitional arc tholeiites and 'within plate' alkaline basalts, may reflect the transient geochemical nature of the mantle under a former active continental margin combined with complex melting relationships attending the initial stages of an attempt for ensialic back-arc spreading. Available geological data (see RIBEIRO & SILVA 1983 — this volume) is compatible with such a tectonic setting for the Iberian Pyrite Belt. Similar volcano-tectonic associations also seem to be characteristic of other upper Paleozoic areas within the Rheno-Hercynian Zone of the mid-western European Hercynian chain (see HERRMAN *et al.* 1974; BEBIEN 1976; BEBIEN *et al.* 1974, 1980; FLOYD 1976, 1982).

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STRUCTURE OF THE SOUTH PORTUGUESE ZONE

A. Ribeiro [1, 2] and J. Brandão Silva [1]

Key words: Folding; Geodynamics; Non-coaxial deformation; Shortening; South Portuguese Zone; Thin-skinned style; Thrust belt.

Palavras-chave: Contracção; Deformação não-coaxial; Dobramento; Faixa de carreamentos; Geodinâmica; Tectónica pelicular; Zona Sul Portuguesa.

ABSTRACT

The South Portuguese Zone is a typical thin-skinned thrust belt in the SW branch of the Iberian Variscan Orogen. Folds and thrust display a SW vergence in upper Devonian and Carboniferous sediments and volcanics. The age of deformation and flysch deposition is younger towards the SW and both facts reflect a tectonic and palaeogeographic polarity. Geodynamic considerations favour a model of opening of a back-arc basin since the middle Devonian, soon followed by closure and collision between the basement of the Ossa-Morena Zone and the SW Foreland, leaving on the surface the South-Portuguese accretionary prism.

RESUMO

A Zona Sul-Portuguesa constitui uma típica faixa de carreamentos pelicular no ramo SW do Orógeno Hercínio Ibérico. As dobras e carreamentos mostram vergência para SW nos sedimentos e vulcanitos do Devónico superior e Carbónico. A idade de deformação e da deposição do Flysch é mais recente para SW e ambos os factos reflectem uma polaridade tectónica e paleogeográfica. Considerações geodinâmicas favorecem um modelo de abertura de uma bacia pós-arco desde o Devónico médio, cedo seguida por fecho e colisão entre os socos da Zona de Ossa-Morena e do ante-país a SW, preservando à superfície o prisma acrecionário da Zona Sul Portuguesa.

1. INTRODUCTION

The South Portuguese Zone is a typical, slightly arcuate, thrust belt (Ribeiro *et al.* in press). The dominant structures change from a N-S direction on the Atlantic seaboard to an E-W direction near the Spanish border. The vergence is towards the SW: thrusts and folds interfere but the transport is systematically in that direction. Deformation increases towards the NE, as the contact with the Ossa-Morena Zone is approached. The structural polarity is reflected in the younging of flysch sediments towards the SW and an increasing metamorphic grade northeastwards.

2. DESCRIPTION OF GEOTRAVERSE

A synthesis of the structure of the portuguese segment of this zone can be given if one considers a NE-SW geotraverse from Beja to cape São Vicente (Fig. 1).

(1) Departamento de Geologia, Faculdade de Ciências, Universidade de Lisboa, Rua da Escola Politécnica 58, 1294 Lisboa Codex, Portugal.

(2) Serviços Geológicos de Portugal, Rua da Academia das Ciências 19-2.°, 1200 Lisboa, Portugal.

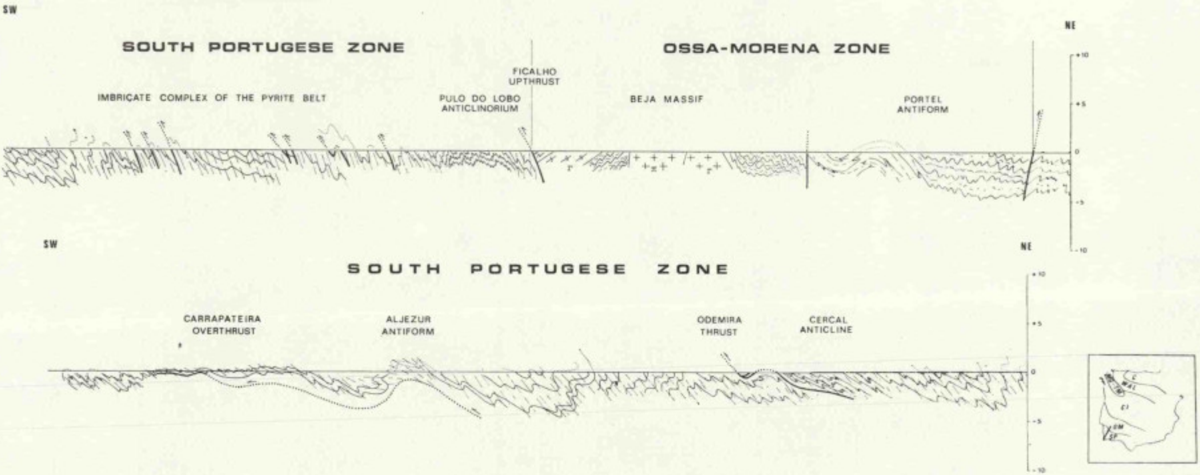

Fig. 1 — A geotraverse through the South Portuguese Zone (from RIBEIRO *et al.* 1979)

2.1. **Pulo do Lobo Subzone**

The NE contact of the South Portuguese Zone is a major thrust affecting the western and southern borders of the igneous and metamorphic terrains that form the Beja-Aracena Massif. In its present state the thrust is a reverse fault dipping more than 60° to the NE, corresponding to a ductile shear zone, with mylonitisation in a band 1 km wide that later evolved into a more brittle style marked by the presence of pseudotachylites and breccias.

The Pulo do Lobo Subzone is an anticlinorium with the SW limb sheared and thinned. The NE limb is formed by the Santa Iria Formation, a possible time equivalent of the Culm (Mértola Fm) of the next subzone, which shows a dominant slaty cleavage dipping steeply to the NE. The nature of its contact with older formations is still disputed. The underlying formations (Ribeira de Limas and Pulo do Lobo) show a much more intense deformation with slaty cleavage (S_1) transposed by a penetrative recumbent crenulation cleavage (S_2) and refolding by a more open crenulation with steep dips to the NE (S_3). It is not certain whether this contact is due to a rapid downward deformation and a regional metamorphism gradient or if the Santa Iria Formation unconformably overlies an already cleaved ($S_1 + S_2$) block, with S_3 passing upwards to form the primary cleavage in its cover.

2.2. **Pyrite Belt** (CARVALHO *et al.* 1976; SILVA 1980, 1981)

The tectonic complexity of the Pyrite Belt is mainly due to the presence of thrusts showing variable relationships with the regional folding. One can recognise (1) synsedimentary thrusts, for instance the Pero da Vinha-Biguina thrust preceded by a very continuous olistrostome deposited on top of the local turbidite Culm succession (Mértola Fm); (2) thrusts folded by the regional cleavage but showing folding earlier than cleavage in the allochthon, for example, the frontal part of the Mértola thrust, (3) thrusts folded by regional cleavage and showing a cleavage earlier than thrusting in the allochthon, such as the Galé and Vale de Évora thrusts.

The variable relationship between folding and thrusting can be explained by the accepted «thin skinned» model: the «piggy-back» thrust sequence in time is showed by the space arrangement when one moves from lower to upper structural levels.

The earliest stage is shown by emplacement of an internally deformed allochthon over a pratically undeformed autochthon. Depending upon the structural level we have, from top to bottom: no internal deformation near the surface, flexural folds without cleavage and flattened flexural folds or similar folds with cogenetic cleavage (S_{1a}) steeply dipping to the NE and flattening as one approaches the thrust plane, showing that the latter acts as a ductile shear zone (SILVA 1980, 1981). With the emplacement of the next thrust below the first, the previous thrust plane is itself folded with cogenetic production of the regional cleavage (S_{1b}) in the autochthon, itself concomitantly affected by the first penetrative deformation. Depending on the strutural level we have, from top to bottom: in the uppermost allochthon S_{1b} is the only visible cleavage, coplanar with the cleavage in the autochthon; in the intermediate level the flexural slip folds are flattened by the S_{1b} cleavage; at the deepest levels the regional cleavage is S_1 (a + b) if S_{1a} and S_{1b} are subparallel or S_{1b} crenulates S_{1a} (Fig. 2). This model is similar to that proposed by SCHERMERHORN & STATON (1969) for the Aljustrel overthrust. The only difference is the style of the pre-cleavage folds in the allochthon. The postulated structure by those authors is that of a folded klippe in which an antiform within the allochthon appears to be incompatible with the synformal setting of the allochthon upon the autochthon. It is suggested here that in Aljustrel the anticlines are in fact synforms affecting the reverse limb of a fold which has been cut by the thrust plane (Fig. 3).

The displacement along thrust planes can reach 20 km (allowing for later refolding); the

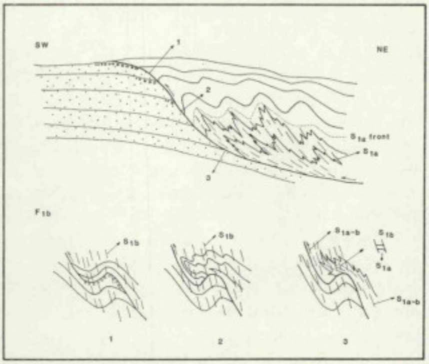

Fig. 2 — Thrust and folding relationships in the Pyrite Belt

last stage is also accompanied by high angle reverse faults, subparallel to S_{1b} which gives rise to the thrusting of the allochthon by the autochthon. These high-angle reverse faults could be the superficial expression of thrust planes at lower, non-exposed levels.

In the Pyrite Belt the folds are transected by regional cleavage S_{1b} (BORRADAILLE 1978; RIBEIRO *et al.* 1979). In fact the slaty cleavage is axial planar to the folds in some domains but

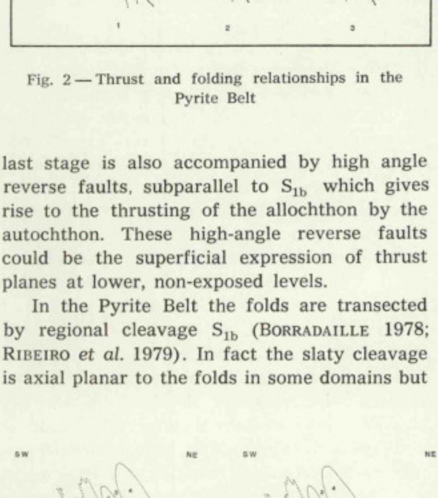

Fig. 3 — Alternative tectonic interpretations for the Aljustrel overthrust (on the left, after SCHERMERHORN & STANTON 1969; on the right, this study)

elsewhere it transects the NW-SE axial planes of the folds, in a clockwise sense. Thus, there is a left lateral shear component added to the flattening, perpendicular to the axial plane of the folds, which is also expressed by the «en échelon» arrangement of macroscopic folds in the Pyrite Belt. It can also be shown that transection increases with depth, which implies syntectonic sedimentation. This is well expressed by the flysch character of the Culm sediments (Mértola Fm). All these facts can be explained by a model invoking an almost continuous process from sedimentation to hard-rock deformation through slumping and soft-rock synsedimentary deformation. A crenulation cleavage, S_2 typically post-metamorphic, develops after S_1, mostly in the domains where that cleavage is less steep.

2.3. **Cercal — Castro Marim Subzone**

The structural pattern is different on either side of the Messejana Fault, suggesting that this major fault acted as a left-lateral wrench fault during early phases of the Variscan Orogeny.

To the W of the Messejana Fault the Cercal Anticlinorium is cut, on its western limb, by the Odemira thrust with a WSW displacement of at least 4 km (D. CARVALHO unpubl. results). As one approaches the thrust plane the regional slaty cleavage, S_1, becomes flatter and more penetrative, implying that the thrust plane is a ductile shear zone coeval with the first Variscan cleavage. Where S_1 is subhorizontal a later folding phase occasionally generated conjugate crenulation cleavages.

To the E of the Messejana Fault only the general pattern of S_1 can be seen; a major thrust plane cannot be positively identified, but the structure does become imbricate (by splaying the major thrust plane to the surface). The E side of the fault represents a higher structural level and the absence of marker horizons in the Flysch (Mira Fm) does not allow a more detailed tectonic analysis.

To the SW the regional slaty cleavage (S_1) becomes steeper and less intense.

2.4. Aljezur-Carrapateira Sector

In the extreme SW of the South Portuguese Zone the Carrapateira overthrust forms one of the main regional tectonic features. This thrust has a minimum displacement of 10 km but probably over 20 km (for details see RIBEIRO 1983 — this volume).

The thrust is more ductile at its deeper levels (to the NE) and more brittle to the SW. The S_1 slaty cleavage becomes more intense and flatter towards the thrust plane, showing that it has the character of a brittle-ductile shear zone. In the more brittle final stage of emplacement, the first phase folds are sharply transected by the thrust plane, with the generation of breccias in the more competent stratigraphic levels. The thrust dies out frontally giving way to recumbent folds facing to the SW i.e. the Flysch (Brejeira Fm) has typical chevron profiles. To the NE the thrust plane must be at shallow levels within the Aljezur Antiform because S_1 becomes flatter and is refolded by a later homoaxial but steeper crenulation cleavage. Away from the thrust plane the S_1 cleavage becomes less penetrative and eventually is reduced to the hinge zones of F_1 folds.

Locally a F_3 phase is developed with conjugate kinks and chevron folds, resulting from a WNW-ESE directed compression (Bordeira Antiform). The same phase is visible in other local areas throughout the South Portuguese Zone.

3. GEOTECTONIC INTERPRETATION

The South Portuguese Zone is a typical thrust belt developed in sediments younger than middle Devonian. The cores of major anticlines never expose beds older than upper Devonian, suggesting the presence of a major décollement at the base of the imbricate complex with a virtually undeformed basement in conjunction with its pre-upper Devonian cover.

This hypothesis is supported by the interpretation of data obtained in deep seismic profiles through the SW of the Iberian Peninsula (MUELLER *et al.* 1973; PRODEHL *et al.* 1975).

These (Fig. 4) show a gradual increase on V_p down to a depth of 8-9 km (imbricate complex), followed by a low velocity channel (décollement zone) after which V_p values ~ 6.5 km/s prevail down to 15 km (lower Palaeozoic cover of autochthonous character) and, again, an increase to V_p values around 6.8 until 35 km, which is the depth of the Moho (Precambrian basement with crust of intermediate type). The Moho is deeper to the SW (35 km) than to the NE (30 km) suggesting the presence of a thinned continental margin in that direction.

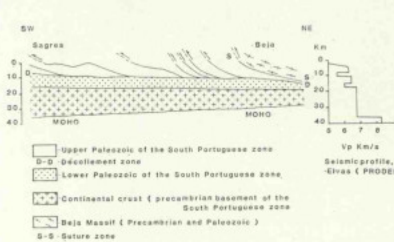

Fig. 4 — Deep structure of the South Portuguese Zone

The interpretation of tectonic and seismic data favour a «thin-skinned» model for this segment of the orogenic belt.

From a knowledge of the depth of the décollement plane a crustal shortening of around 45 % can be estimated using balanced cross-sections (HOSSACK 1979).

Following the analysis of tectonic profiles the regional tectonics will now be considered (Fig. 5). The dominant feature is an arcuate belt with stretching in *a* on the inner part, then a pure flattening narrow core and slight stretching in *b* on the outer arc. This suggests crustal buckling by a tangential longitudinal strain deformation mechanism (RIBEIRO *et al.* 1979). The sinistral sense of shear indicated by the transected folds is a general feature on the SW branch of the Ibero-Armorican arc and can be linked to the genesis of the arc.

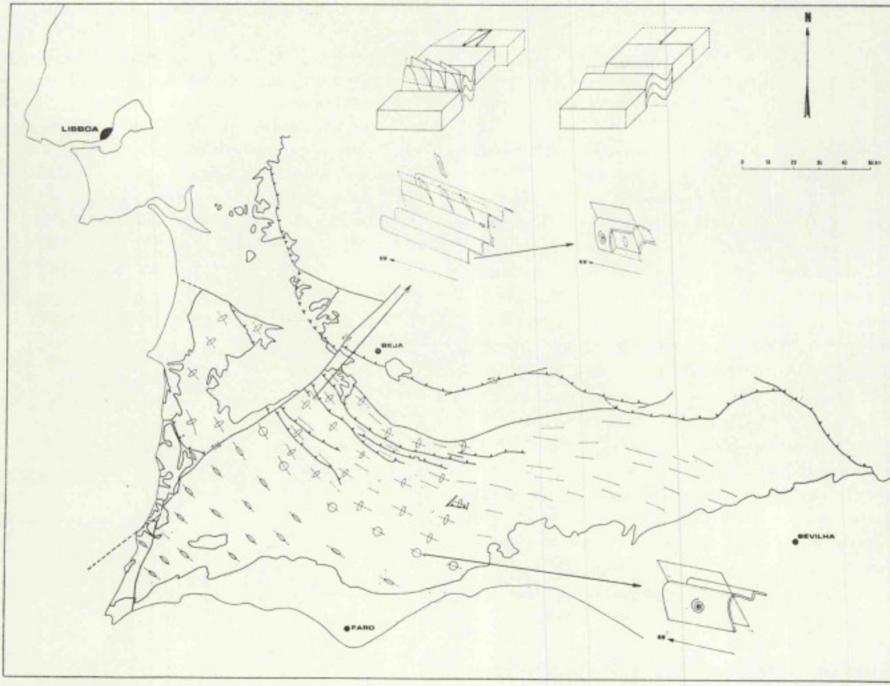

Fig. 5 — Relationship between fold axis, axial planar cleavage and stretching during the first Hercynian phase in the SW Iberian arc (from Ribeiro *et al.* 1979)

As in other thrust belts the driving mechanism of the orogeny can be lateral spreading or push from the rear zones. We favour the second model taking into account a geodynamic model for the Iberian Variscan Fold Belt as a whole (Ribeiro *et al.* in press). Palaeomagnetic (Bonhommet *et al.* 1982) and tectonic data suggest the presence of a major suture between the Ossa-Morena and Centro-Iberian zones, as a consequence of continental collision during the early Carboniferous. This collision was preceded by subduction to the SW (present geographic coordinates) and antitethic orogeny in the middle Devonian. Data from igneous petrological studies (Munhá 1981) suggest that the South Portuguese Zone is a back-arc basin related to this subduction event towards the SW. This back-arc basin has progressively closed immediately after the Viséan by means of subduction now directed to the NE (Bard 1971; Bard *et al.* 1973; Carvalho 1971). The sediments of the South Portuguese Zone formed an accretionary prism, detached from the thinned continental and back-arc oceanic crust, where they were deposited, and escaping subduction of that crust under the Beja-Aracena Massif. The process continued until final collision of the more continental-type crust of the Ossa-

Morena zone on the NE side and the thinned continental margin of the former back-arc basin to the SW, the latter now being situated below the décollement zone at the base of the imbricate thrust belt. In this model the driving mechanism would be shearing induced by subduction, and the folding would be a subordinate effect due to the response of layered materials to the compression induced by shearing.

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STRUCTURE OF THE CARRAPATEIRA NAPPE IN THE BORDEIRA AREA, SW PORTUGAL

A. Ribeiro [1]

Key words: Allochthon; Autochthon; Brittle-ductile shear; Carrapateira Nappe; Carrapateira thrust plane; Flat; Folding; Mechanism of nappe emplacement; Pressure solution; Ramp; Shearing; Thin-skinned style.

Palavras-chave: Alóctone; Autóctone; Chã (= 'Flat'); Cisalhamento; Cisalhamento dúctil-frágil; Dobramento; Manto da Carrapateira; Mecanismo de instalação de mantos; Rampa; Solução por pressão; Superfície de Carreamento da Carrapateira; Tectónica pelicular.

ABSTRACT

The Carrapateira Nappe forms the major part of the Bordeira area. The basal thrust plane is well exposed in some localities and dies out towards the frontal part. The tangential style is typically thin-skinned and the internal deformation of the allochthon is clearly related to the geometry of the basal thrust plane which acted as a zone of brittle-ductile shear.

The displacement along the thrust is probably more than 10 km considering the presence of an inferred near-surface thrust in the Aljezur Antiform. The mechanism of emplacement is possibly the uplift of the internal zones and underthrusting of the external zones.

RESUMO

O manto da Carrapateira ocupa a maior parte da área de Bordeira. O carreamento basal está bem exposto em algumas localidades e exibe amortecimento frontal. O estilo tangencial é tipicamente pelicular e a deformação interna do alóctone está claramente relacionada com a geometria do carreamento basal que funcionou como cisalhamento frágil-dúctil.

O deslocamento no carreamento é provavelmente superior a 10 km se se considerar a presença de um carreamento subaflorante inferido no antiforma de Aljezur. O mecanismo de instalação é provavelmente levantamento nas zonas internas e subcarreamento das zonas externas.

1. INTRODUCTION

The aim of the present paper is to give a synthesis of the structure of the Carrapateira Nappe in the Bordeira area (SW Portugal), as an example of the 'thin-skinned' tectonic style of the South Portuguese Zone (Ribeiro & Silva 1983 — this volume). A more detailed paper on the structure of the Bordeira area will be published in the near future (Ribeiro in prep.).

The stratigraphy of the Bordeira area is described in a companion paper (Oliveira *et al.* in

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(1) Departamento de Geologia, Faculdade de Ciências de Lisboa, Rua da Escola Politécnica 58, 1294 Lisboa Codex, Portugal; and Serviços Geológicos de Portugal, Rua da Academia das Ciências 19-2.°, 1200 Lisboa, Portugal.

prep. a) and a cross section in the Praia de Murração region is given in Fig. 1.

The geology of the Variscan and post-Variscan rock units is described briefly in RIBEIRO *et al.* (in prep.).

Tangential tectonics in the Bordeira area were first described by FEIO & LOMBARD (1958), for the Murração sector, and afterwards extended to the whole Bordeira Area (FEIO & RIBEIRO 1971). However, the real significance of the Carrapateira Nappe was only established during the course of geological field work (1976-8) for the 48-D (Bordeira) sheet of the 1/50,000 scale Geological Map of Portugal (RIBEIRO in prep.).

The Variscan Tectonics of the Bordeira area show the effects of three phases of deformation.

The first phase, F_1, is responsible for the main structural pattern visible in the area. The folds have NW-SE oriented axes and their axial planes dip at varying angles to the NE. These folds are accompanied by a cogenetic axial plane or fanning slaty cleavage of variable intensity; in the less deformed areas the cleavage is restricted to the hinge zones and dips steeply to the NE. As one approaches the Carrapateira thrust plane the cleavage becomes subhorizontal, more intense and penetrative. This fact shows that the thrust is a shear zone coeval with F_1; it dies out where the frontal part of the thrust passes into a recumbent fold facing SW.

The second fold phase, F_2, is responsible for the refolding of earlier structures, with NW-SE axes and subvertical axial planes. Shortening is moderate and the S_2 crenulation cleavage is restricted to very small areas.

The third fold phase, F_3, generated the NNE-SSW Bordeira Antiform, with a box-fold profile and limbs cut by normal faults. Shortening is more intense in the core and limbs of this antiform, and the steeply dipping S_3 cleavage with NNE-SSW direction is restricted to those areas and is virtually absent outside the antiform.

The stress field that generated the F_3 structures was probably also responsible for the conjugate WNW-ESE, sinistral, and ENE-WSW, dextral wrench-faults.

If one refers to the regional context (RIBEIRO *et al.* 1979) this F_3 phase is probably the same as that which affected the upper Stephanian C beds of Buçaco, near Coimbra.

Post-Variscan structures in the Bordeira area include the following elements: W of the N-S Carrapateira Monocline the downthrown Keuper to Kimmeridgian cover of the coastal Carrapateira outlier is preserved. A swarm of alkaline basal dykes crop out parallel to the monocline and the dip of the individual dykes is subvertical onshore and decreases eastwards as one approaches the monocline.

This suggests that the Monocline was generated in a tensional regime, related to the opening of the Atlantic during the Kimmeridgian, since volcanic agglomerates of that age occur on top of the Carrapateira cover section. The Monocline is also delineated by an alignment of alkaline basalt stocks, and a tholeiitic dolerite dyke also crops out nearby. The latter is probably coeval with the upper Jurassic Messejana dolerite (SCHERMERHORN *et al.* 1978).

The generation of the Monocline was probably accompanied by the reactivation of NW/SE trending graben structures which plunge to the W in a «touche de piano» style.

A period of N-S reverse faulting at Cerro da Vigia, caused superposition of the cover on the E side by the basement on the W side. This is probably due to the detumescence phase of evolution of the rift-related monocline.

The Ribeira de Sinceira graben (FEIO 1949, 1951) is a segment of the S. Teotónio-Aljezur-Seiceira-Ingrina graben system; this fracture zone is a late Variscan NNE-SSW wrench system (conjugate or splay of the Messejana wrench fault?) reactivated in Plio-Pleistocene times. Geomorphological (FEIO 1949, 1951) and other geological evidence of neotectonic activity can also be found in the Carrapateira Monocline and possibly also in other graben structures subparallel to the Ribeira da Sinceira Fault, such as the Ponta dos Carneiros graben.

2. THE CARRAPATEIRA NAPPE

The Carrapateira thrust plane is exposed at the following places, from NE to SW:

a) Coastal cliff between Barranco do Aipo and Pedra Ruiva.
The thrust plane approaches the surface by means of a ramp under a spectacular anticline the short limb of which is cut by the thrust. In this region the thrust (very well exposed) corresponds to a shear zone 0.5 m wide with sheath folds and stretching lineation showing a southwestward direction of movement. Where it transects the more competent horizons, (quartzites of the Monte Novo member of the Tercenas Fm) the thrust becomes more sharply defined with subordinate shears penetrating the lower block.

b) Tercenas Window
A dome caused by the interference of F_2 and F_3 phases shows the autochthonous Monte Novo and Barranco Velho members of the Tercenas Fm, over which the Bordalete Formation has been thrusted. The thrust plane transects the folds in the authochthon and in some places is accompanied either by extensive brecciation of the quartzites of the Monte Novo member, by tectonic inclusion of those quartzites in the slates of the Bordalete Fm, or by retransposition of S_1 by a very penetrative and intense crenulation cleavage coplanar with the thrust plane.

c) Coastal ciff at Monte do Eengenheiro — Praia de Murração — Pedra das Safias.
The thrust plane is sharply defined and accompanied by extensive fault breccias in the competent beds of the Vale Figueiras and Pedra das Safias members of the Murração Fm, and fault gouge in the incompetent beds of the Quebradas and Bordalete formations. The folds are clearly cut by the thrust plane, but the displacement diminishes gradually to the SW in such a way that the thrust dies out and passes into an almost recumbent fold at Praia das Quebradas (Fig. 1). As the thrust plane is approached, the S_1 cleavage becomes subhorizontal and more penetrative and the X_1 stretching lineation more pronounced, with an average orientation of NNE/SSW.
The stretching is shown by a linear fabric, especially in slates, locally obliterated by a homoaxial crenulation lineation, (constriction by channel flow), by syntaxial fibre growth around pyrite nodules and by stretched crinoids. In some places the S_1 cleavage is deformed by sigmoidal shear planes showing movement in a SSW-ward direction.

The correlation between the exposures of the Carrapateira Thrust plane show that the style of shear movement is more ductile to the NE and more brittle to the SW. This fits well, in a palinspastic reconstruction, with the thrust being deeper in the rear NE side, and nearer the surface in the frontal SW side.

The Autochthon is only exposed in very small areas below the Carrapateira thrust plane. The deformation increases upwards, as shown by the decrease in the angle between bedding and S_1 cleavage planes, as the thrust plane is approached. The Allochthon (Carrapateira Nappe) shows a variety of tectonic styles, depending on structural level and its relationship to the thrust plane. In the uppermost structural level, mainly represented by the Flysch sequence of the Brejeira Formation, the folds have steeply NE dipping axial planes, the profiles are frequently of chevron type and show the typical pattern of geometric accomodation in the thickest beds: the hinges bulge and the limbs are sheared (Ramsay 1974).

The S_1 slaty cleavage is restricted to the pelites and fold hinge zones. The greywackes show pressure-solution cleavage, and less frequent tension gashes. The former is more intense along conjugate «en échelon» tension gashes corresponding to potential or expressed brittle shears. The pressure solution cleavage forms

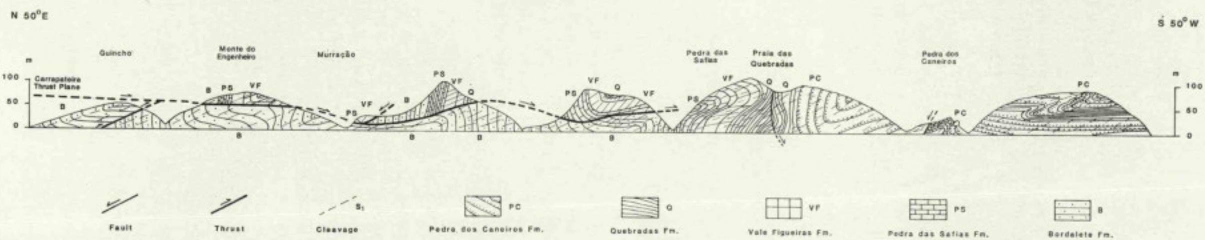

Fig. 1 — Cross section between Monte do Engenheiro and Praia das Quebradas

convergent fans around the flattened flexural-slip folds, and the gashes are displaced by slip on bedding planes; the solution seams are occasionally observed to dissolve the «en échelon» gashes. It can be concluded that the sucession of deformation mechanisms in the greywacke beds is as follow:

$\longleftarrow$ Shearing $\longrightarrow$
$\leftarrow$ Pressure solution $\rightarrow$
$\longleftarrow$ Folding $\longrightarrow$

These deformation mechanisms are coeval with flattening in the pelite beds.

The existence of a non coaxial component of deformation is shown by the fact that the ENE-WSW sinistral brittle shears are more frequent than the N-S dextral brittle shears. This means that the sinistral family of shears is syntethic and the dextral antitethic. This can be explained (RIBEIRO & SILVA 1983 — this volume) by adding a simistral component of shear to the shortening direction perpendicular to the axial traces of F_1 folds. The study of the syntectonic crystal growth in gashes (DURNEY & RAMSAY 1973) confirms this interpretation.

In the frontal part of the Carrapateira Nappe the axial planes of F_1 folds become recumbent. Spectacular folds, frequently of chevron type, with flat axial planes, can be seen between Praia das Quebradas and Praia do Castelejo, W of Vila do Bispo. Axial planes steepen again from this locality to Ponta do Telheiro, where the flysch (Brejeira Fm) is unconformably overlain by the Keuper red beds («Grés de Silves» Fm).

Moving down in the rock succession, and approaching the thrust plane, the S_1 slaty cleavage becomes subhorizontal and more intense and the folds tighten; the stretching lineation, subperpendicular to the fold axes, becomes prominent. The more competent layers are folded by tangential longitudinal strain, with neutral points developed by contact strain. During progressive deformation the neutral points migrate and generate primary crenulation cleavage in the outer arc that grades to slaty cleavage in the inner arc when traced along the hinges of the folds.

The relationship between folding and thrusting is clearly illustrated in the Carrapateira Nappe. A cross section (RIBEIRO *et al.* in prep.) shows the alternation of segments with only minor folding but intensive slicing due to second order thrusting and segments with frequent folds but virtually no thrusts, particularly along the frontal part of the nappe. A straightforward explanation can be applied (Mc CLAY & PRICE 1981): the fold-dominated segments develop above ramps and the tabular segments correspond to the flats of the thrust plane. This agrees with the fact that the roof thrust climbs up the stratigraphic succession from the NE (where it lies below the Monte Novo member) — to the extreme SW, around Praia de Murração (where it occurs at the base of the Bordalete Fm).

The geometry of the thrust plane does not favour gravity sliding as a possible emplacement mechanism for the nappe. This problem will be discussed below.

3. MECHANISM OF EMPLACEMENT OF THE CARRAPATEIRA NAPPE

The emplacement of the Carrapateira Nappe probably occurred shortly after or even during the sedimentation of the Brejeira Fm turbidites. These show ample evidence of softrock synsedimentary deformation (synsedimentary folds) sometimes with associated dewatering, producing axial planar cleavage, and slumps on various scales, e.g. the metre scale slump bed of Arrifana beach.

On the other hand the deformation structures suggest that they are due to hydraulic facturing in a sequence that had developed high pore-fluid pressures during sedimentation (BEACH 1977). This occurred prior to folding and continued to evolve during folding. Near the thrust plane one can find slip planes parallel to the bedding, which are folded with axial planar S_1 cleavage. Slickenslides on these slip planes show an orientation almost identical to that inferred for the thrust sheet. This suggests slip parallel to the bedding on undisturbed beds at the front of the advancing nappe. Some palaeocurrent measurements point to a deviation of direction in the front of the nappe, suggesting emplacement during sedimentation of the Flysch.

The Carrapateira Nappe has regional significance, as the displacement is probably in the order of 10 km, or more. In fact mapping of the Aljezur/Bordeira area (OLIVEIRA *et al.* in prep. b) which is situated about 15 km NNE of Bordeira, shows that a major thrust plane is possibly near the surface in the F_2 Aljezur Antiform; the S_1 cleavage becomes more intense and flat-lying within the core of that structural element. The obvious correlation is with the Carrapateira thrust, taking into account the presence of an open synform structure between the Bordeira and Aljezur areas; this interpretation was favoured in the geotraverse through the South Portuguese Zone (OLIVEIRA *et al.* in prep. b).

The tectonic style of the Carrapateira Nappe is coherent with a thin-skinned model for the genesis of tectonic structures in the South Portuguese Zone. However, the driving mechanism for those structures and particularly for the Carrapateira Nappe remains debatable, as is the case in many other 'thin skinned' thrust belts (i.e. push from the rear or lateral gravitational spreading?).

A peculiar feature of the Carrapateira Nappe as compared with many other similar structures is the fact that it dies out frontally. But this feature is ambiguous from the point of view of the driving mechanism, because it can be explained by underthrusting, that is, moving the authochthon underneath a stable allochthon, or by gravity gliding in a subsiding foredeep trough.

A nearly source for gravity slides can be excluded because there is no evidence of tectonic denudation in the rear of the Carrapateira Nappe. But the possibility of uplift of the Beja Massif (Ossa-Morena Zone) inducing the gravitational slides and a successive southwestwards development of thrust faults remains valid with abnormal fluid pressure favouring both thrust initiation and further movement. The problem is then the driving mechanism for the Beja Massif uplift.

Subduction of the South Portuguese Zone under the Ossa-Morena Zone (RIBEIRO & SILVA 1983 — this volume) is the probable mechanism that can produce uplift of internal zones and underthrusting of external zones. Simultaneously the absence of extensional faults also implies that the role of gravity was not dominant for the emplacement of the thrust sheets.

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CARBONIFEROUS VOLCANOGENIC SULPHIDE MINERALIZATIONS IN SOUTH PORTUGAL (IBERIAN PYRITE BELT)

F. J. A. S. Barriga [1] and D. Carvalho [2]

Key words: Volcanogenic polymetallic sulphide deposits; Iberian Pyrite Belt; Ore genesis; Hydrothermal alteration; Stockwerk rock; Jaspers; Carboniferous

Palavras-chave: Sulfuretos polimetálicos vulcanogénicos; Faixa Piritosa Ibérica; Metalogenia; Alteração hidrotermal; Rocha de 'stockwerk'; Jaspes; Carbonífero

ABSTRACT

More than 400 million metric tonnes of stratiform, polymetallic massive sulphide deposits formed in the Portuguese half of the Iberian Pyrite Belt in early Carboniferous times, especially at Aljustrel, Neves-Corvo, S. Domingos and Lousal. Ore deposits were formed at the waning stages of important episodes of felsic, explosive, submarine volcanism which took place in discrete centres, generating many different lithostratigraphic columns. Metalliferous aqueous solutions raised through footwall rocks, producing prominent hydrothermal alteration and from which metal sulphides precipitated at or near the sea floor. The sulphide deposits are therefore syngenetic, volcanogenic, exhalative and submarine. Some are autochthonous, rooted in their stockwerk feeder zones, where as others were redeposited after movement down volcano flanks.

Clues to ore genesis are present in the deposits themselves, in the hydrothermal alteration surrounding autochthonous deposits, in the pervasive alteration of the volcanic rocks and also in chemical metalliferous and siliceous sediments that occur on and laterally with respect to the sulphide accumulations.

Despite some difficulties, available data suggest that the deposits were formed as a consequence of hydrothermal metamorphism of the volcanic rocks. Heat dissipation from the volcanic piles and possibly from underlying magma chambers may have driven sea water convection through the highly permeable volcanic rocks, extracting metals from them and producing both the present spilitic and quartz-keratophyric compositions of the volcanic rocks and a sea water derived mineralized fluid. Focussed return flow of this fluid and appropriate conditions at the site of discharge may have led to large scale precipitation of most metals, generating the massive sulphide deposits and the closely associated siliceous and metalliferous sediments, according to patterns of thermal, redox and pH variations at discharge sites. However, a magmatic fluid contribution to the ore forming fluid cannot be ruled out.

RESUMO

No decurso do Carbónico Inferior depositaram-se na parte portuguesa da Faixa Piritosa Ibérica mais de 400 milhões de toneladas de sulfuretos maciços polimetálicos estratiformes, principalmente em Aljustrel, Neves-Corvo, S. Domingos e Lousal. Os depósitos são contem-

[1] Departamento de Geologia, Faculdade de Ciências, Universidade de Lisboa, Rua da Escola Politécnica 58, 1294 Lisboa Codex, Portugal.

[2] Serviços Geológicos de Portugal, Rua da Academia das Ciências 19-2.º, 1200 Lisboa, Portugal.

porâneos das fases finais de importantes episódios de vulcanismo félsico, explosivo e submarino, que tiveram lugar em centros vulcânicos individualizados, originando sequências litoestratigráficas diversas. Soluções aquosas metalíferas ascenderam através das rochas a muro dos jazigos, produzindo nestas marcada alteração hidrotermal e transportando os metais até perto da interface rocha — água do mar, onde estes precipitaram sob a forma de sulfuretos. Os jazigos são portanto singenéticos, vulcanogénicos, exalativos e submarinos. Alguns são autóctones, enraizados em 'stockwerks' que representam as zonas de ascenso de fluido mineralizante, enquanto outros foram redepositados depois de escorregamento pelas encostas de vulcões.

Indícios da génese dos jazigos encontram-se nos depósitos em si, na alteração hidrotermal que acompanha os jazigos autóctones, na metassomatização que afecta as rochas vulcânicas e também nos sedimentos metalíferos siliciosos que acompanham as massas de sulfuretos.

Os elementos actualmente existentes sugerem, apesar de algumas dificuldades, que os jazigos se formaram como consequência de metamorfismo hidrotermal das rochas vulcânicas. A libertação da energia nelas contida e de possíveis câmaras magmáticas subjacentes, teria provocado a circulação convectiva de água do mar nas rochas permeáveis, produzindo-se trocas térmicas e mássicas responsáveis pelas actuais composições espilíticas e quartzo-queratofíricas e também pelo aparecimento de fluido mineralizado derivado da água do mar enriquecida de metais extraídos das rochas.

Refluxo ascensional concentrado deste fluido e condições apropriadas nas zonas de descarga podem ter motivado a precipitação de grande parte dos metais, originando os depósitos de sulfuretos maciços e os sedimentos metalíferos e siliciosos que lhes estão associados, de acordo com gradientes térmicos, redox e do pH nas zonas de descarga. Contudo, não se pode excluir a possibilidade de uma contribuição de fluido magmático na origem do fluido mineralizante.

1. INTRODUCTION

The Iberian Pyrite Belt (Fig. 1) is an outstanding metallogenic province where large concentrations of base metal sulphides were deposited within an extensive Volcanic-Sedimentary Complex, mainly of Viséan age, described elswhere in this volume. The main part of this province extends over a distance of 230 km by about 30 km width and contains several deposits of massive sulphides, some of them known for more than 30 centuries. Tartessians, Phoenicians and Romans exploited gossans for gold and silver, and copper from the supergene enrichement zone.

More than 1,000 million tonnes of massive sulphides is the estimated figure for the original resources of known deposits and only 20 % of this amount has been mined out. Mineral resources and reserves established so far exceed 700 million tonnes, which represent the largest stock of base metals in Western Europe. The main mines are, from west to east: in Portugal, Lousal, Aljustrel and Neves-Corvo; in Spain, Tharsis, Sotiel, La Zarza, Rio Tinto and Aznalcóllar.

The ore-bodies occur at or near the top of the better developed felsic volcanic sequences, or on nearby laterally equivalent sediments. The modern approach to the study and exploration for these important mineral resources is to consider them *rocks* that form and are conserved under specific geologic conditions. Successful exploration has been intimately related with a proper understanding of the Geology of the Iberian Pyrite Belt, especially for the last 20 years. It is therefore adequate to dedicate a Chapter of this Volume to the geology, stratigraphy and genesis of the massive sulphide deposits.

2. GEOLOGY OF THE SULPHIDE DEPOSITS

The so-called 'pyrite bodies' are stratiform polymetallic sulphide concentrations, with pyrite predominant, accompanied by chalcopyrite, galena, sphalerite and several other minerals. This mineralization is a product of felsic submarine volcanism that took place during the eugeosynclinal stage of the upper Devonian-lower Carboniferous evolution of the South Portuguese Zone. This volcanism overlies a sequence of phyllites and quartzites of Famennian age and is overlain by flysch deposits ranging from the Viséan up to the Namurian-lower Westphalian (Fig. 1). During the Hercynian orogeny the whole sequence was folded and slightly metamorphosed.

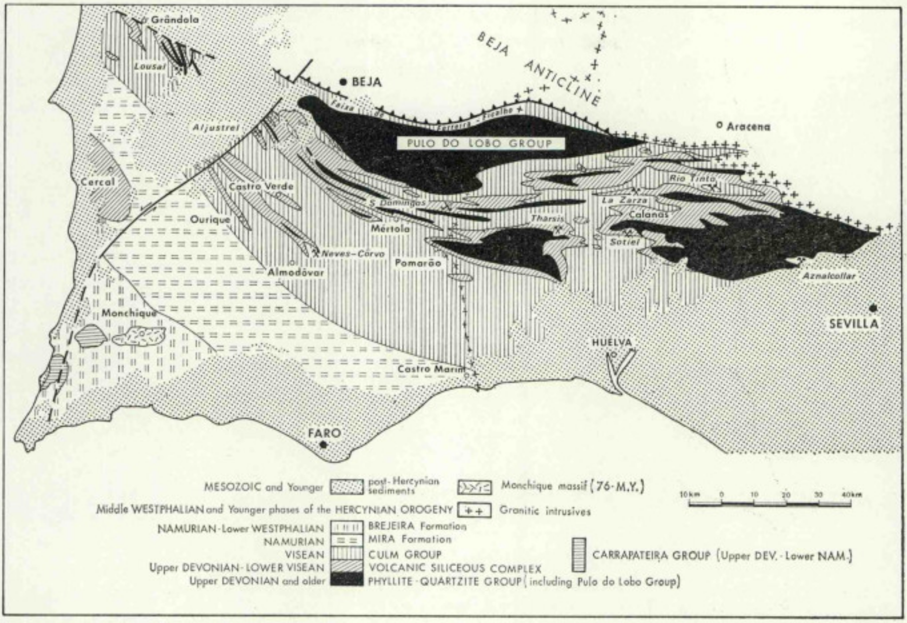

Fig. 1 — General geology of the Iberian Pyrite Belt (adapted from CARVALHO *et al.* 1976; OLIVEIRA *et al.* 1979)

The felsic volcanism is a major feature of the South Portuguese Zone. This volcanism was explosive and generated a wide variety of tuffs, a few lavas, breccias and agglomerates. It seems that the main volcanic centres were located along submarine strike faults, at areas of intersection with NE-SE lineaments (CARVALHO 1974). The main outcropping volcanism is known along five lineaments, which proved to be productive in sulphide mineralization. Chronostratigraphic and lithostratigraphic data suggest a migration of the volcanic front from South to North, ranging from the Upper Devonian to Viséan (CARVALHO 1976).

Sulphide deposition took place at or near the top of felsic piles, at the end of volcanic cycles, during the waning stage of the explosive phase. In each productive volcanic centre the sulphide mineralization occurs in a particular stratigraphic level and is usually associated with a specific tuff facies. The timing of sulphide deposition varies from centre to centre, ranging from upper Devonian (Salgadinho) to lower Viséan (Sotiel Coronada). Volcanic sequences and ore lithostratigraphic columns are different from mine to mine, so their correlation is a misleading exercise.

Most orebodies occur in sub-vertical limbs of folds, exceptionally at syncline hinges, or as lenticular beds in flat lying structures (Neves-Corvo).

Jaspers and cherts are common chemical sediments that occur in close association to the sulphide mineralization. Very often certain jaspers contain manganese oxides and hydrated oxides as products of the alteration of rhodo-

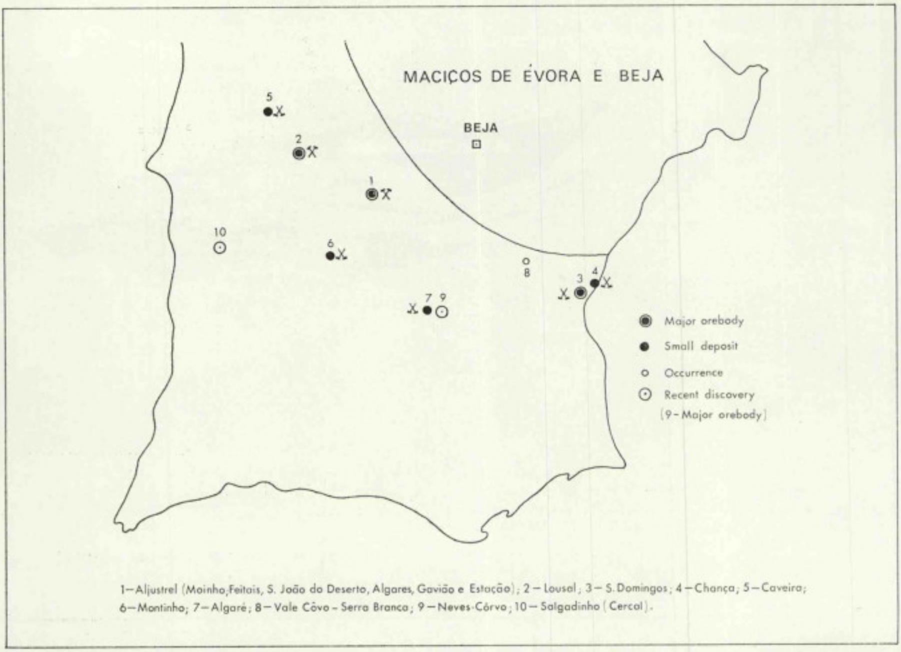

Fig. 2 — Polymetallic sulphide deposits in southern Portugal — Iberian Pyrite Belt (from CARVALHO 1979)

chrosite and rhodonite, sometimes forming small concentrations that were exploited as manganese ores.

Up to now there are 10 known areas of sulphide mineralization in the Portuguese part of the Iberian Pyrite Belt (Fig. 2): Aljustrel, Lousal, Caveira, S. Domingos, Chança, Serra Branca-Vale Covo, Montinho, Algaré, Neves-Corvo and Salgadinho (Cercal). Aljustrel, with 250 MT distributed in 6 deposits — Algares, Moinho, S. João do Deserto, Feitais, Estação and Gavião — is nowadays the largest concentration of massive sulphides in the Pyrite Belt and certainly one of the largest in the world. Neves-Corvo is a new orebody already famous for its abnormaly high Cu grades. The start of production is scheduled for 1986. Lousal is nearly exhausted, while S. Domingos (mined out) was a Cu rich, very important producer in the past.

The orebodies are stratiform lenses or sheets of sulphides of different sizes, up to 4.5 km long by 1.5 km wide and up to 80 m thick, possibly the original dimensions of Rio Tinto (Spain), which corresponds to an original tonnage of more than 500 MT (WILLIAMS *et al.* 1975). A comprehensive analysis regarding size and geometry on 60 sulphide orebodies of the Huelva province totalling 750 MT (ALVAREZ 1974), led to the conclusion that four categories could be considered:

1 — Very large deposits: $>$ 20 MT (850 m long $\times$ 80 m thick $\times$ 350 m wide) $=$ $=$ 475 MT

2 — Large deposits: 5 – 20 MT (480 m × × 40 m × 260 m) = 170 MT
3 — Medium deposits: 1 – 5 MT (350 m × × 25 m × 170 m) = 95 MT
4 — Small deposits: < 1 MT (200 m × × 12 m × 75 m) = 10 MT

Following this classification and including the Portuguese deposits it turns out that sulphide concentrations have the following distribution:

very large deposits	≃ 75 %
large deposits	≃ 17 %
medium deposits	≃ 8 %
small deposits	≃ 1 %

The average composition of known resources and reserves is (CARVALHO *et al.* 1976):

Sulphur	46 %	Zinc	2,9 %
Iron	40 %	Lead	1,1 %
Copper	0,7 % [1]	Gold	0,8 g/t
		Silver	30 g/t

A host of alloy metals such as Sn, Cd, Co, Hg, Bi, Se and many others is also present, in concentrations ranging from tens to hundreds of ppm. Many of these metals (including Pb and Zn) are often not recovered because of the fine grained nature of most ores. This, in turn, is responsible for the expression 'Iberian *Pyrite* Belt' and for frequence incorrect quotations in the literature of the orebodies as 'barren pyrite' and 'pyrite orebodies'.

Excluding the Neves-Corvo orebody in all the other deposits exceeding 30 MT the average Cu grade is low, 0,3-0,7 % (CARVALHO *et al.* 1976) and the average Zn-Pb is relatively high, up to 4-5 %. Commonly, high copper grades are found only in the smaller deposits and, locally, as oreshoots, in the larger ones. That is the case of Gavião (Aljustrel) 25 MT — 1,5 % Cu. However, in Neves-Corvo there are bodies with some tens of MT exceeding 7 % Cu. Up to now, this is the exception to the rule.

Three main ore types may be considered (STRAUSS *et al.* 1977):

- massive polymetallic pyritic ore with 35-51 % S or 66-96 % pyrite equivalent.
- disseminated polymetallic pyritic ore ('safrão') with less than 35 % S or 66 % pyrite equivalent, generally associated with massive ore.
- stockwerk pyritic ore with 5-25 % S or 10-50 % pyrite equivalent.

Massive polymetallic ore is invariably fine grained, generally compositionally banded or microbanded. Framboidal and colloform textures are often present (Plate 1A), but in other cases metamorphic recrystallization obliterates primary textures, with complete recrystallization. Pyrite and arsenopyrite become idiomorphic, sometimes fractured, whereas chalcopyrite, sphalerite and galena deform plastically, occupying intra and inter spaces with respect to the non plastic sulphides (pyrite, arsenopyrite), as illustrated in Plate 1B. It is interesting to note that primary and recrystallization textures often coexist in the same orebody.

Certain orebodies are characterized by the ubiquitous presence of clastic sedimentary structures, (Plate 1C) indicative of rather significant sedimentary displacement. The presence of these structures in often enormous bodies of sulphides and the normal stratigraphic relationships with interbedded volcanics and sedimentary rocks led SCHERMERHORN (1970, 1971b) to consider these sulphide ore deposits a rock type proposing the name 'pyritite' as a descriptive and non-genetic rock designation for redeposited massive sulphides. Pyritite is indeed a most common ore type in the Iberian Pyrite Belt.

Stockwerk ore generally occurs at the footwall of bodies of the other ore types, as a network of pyrite + chalcopyrite (±pyrrothite) + + quartz + chlorite hosted in generally highly altered volcanic rocks. The pyrite-chalcopyrite

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[1] Not considering Neves-Corvo. Already defined reserves at this orebody shift the Cu overall average of the IPB to nearly 1.1 %.

ratio is usually much smaller (often locally < 1) than in other ore types. Stockwerk zones correspond to the channelways of mineralizing solutions responsible for ore deposition.

Pyrite is the more abundant sulphide mineral in all types of ore. Sphalerite, chalcopyrite, galena and arsenopyrite are the remaining common sulphide minerals. Many other metallic minerals occur in minute quantities, such as tetrahedrite, bournonite, tennantite, pyrrothite, cobaltite, stanite (?), boulangerite, greenockite, meneghinite, magnetite, hematite and native bismuth (GASPAR & CONDE 1978). Cassiterite has also been reported from some deposits (AYE & PICOT 1976). Non-metallic minerals include quartz, several carbonates, sericite, chlorite and barite. Mineralogical zonation is frequently prominent, with chalcopyrite, chlorite and sericite concentrated towards the footwall (and stockwerk), sphalerite, galena and barite abundant near the hanging wall and in peripheral zones, and pyrite, arsenopyrite, quartz and carbonates scattered throughout.

The orebodies can be classified into three main types (CARVALHO 1979), as follows (Fig. 3):

— Autochthonous (in situ, rooted in stockwerk)
— Transitional
— Allochthonous (redeposited, stockwerk absent).

The former occurs close to the volcanic centres, overlies igneous rocks and has a footwall stockwerk (feeder zones) with strong chloritization and silicification. Quartz-sericitic alteration is widespread especially at the hangingwall. The latter corresponds to deposits where no direct connections to volcanic centres can be observed and, in extreme cases, they may be completely interstratified between sedimentary rocks of penecontemporaneous age with volcanism; these deposits do not have stockwerk; hydrothermal alterations may be completely missing and the pyritite is usually rich in sedimentary structures. The transitional type includes all the intermediate situations.

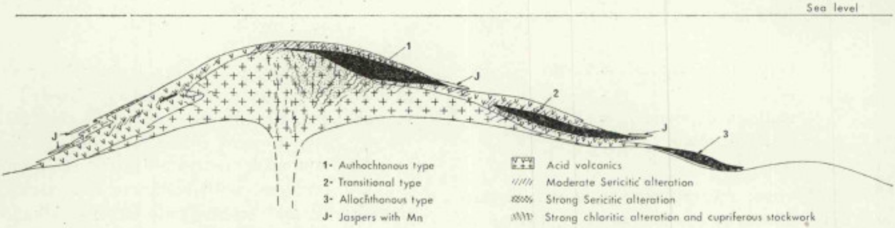

Fig. 3 — Types of massive sulphide deposits in the Iberian Belt (from CARVALHO 1979)

The Feitais, Corvo, S. Domingos and Chança (and Rio Tinto and La Zarza in Spain) are autochthonous orebodies, Moinho and Gavião (and Sotiel) are transitional type orebodies and Lousal and Montinho (and Tharsis) are allochthonous deposits. The Salgadinho disseminated deposit may correspond to a stockwerk without an overlying massive orebody, possibly because of the sedimentary redeposition of the latter elsewhere.

3. HYDROTHERMAL ACTIVITY AND ORE DEPOSITION

The Iberian Pyrite Belt massive sulphide deposits are unanimously considered to be syngenetic, in the sense that they formed at the same time as the rocks that host them, and are not later (epigenetic) additions to the stratigraphic sequences, as evidenced by the well defined horizons where they occur, the common sedimentary features they depict and also because they experienced the same metamorphic and tectonic history as their host rocks (STRAUSS 1965; RAMBAUD 1969; SCHERMERHORN & STANTON 1969; SCHERMERHORN 1970, 1971a; CARVALHO *et al.* 1971a, b, 1976; CARVALHO 1979; STRAUSS *et al.* 1977; ROUTHIER *et al.* 1980). The close spatial and temporal relationship of sulphide mineralization with the last stages of the main felsic volcanic episodes is presently well established: the sulphide orebodies are clearly volcanogenic. The fact that the volcanism was dominantly submarine indicates that ore deposition was also submarine, in agreement with the subaqueous sedimentary textures depicted by the ores themselves (STRAUSS 1965; SCHERMERHORN 1970, 1971a, b).

These sulphide deposits are also exhalative, that is, originated from metalliferous aqueous solutions rising through the footwall rocks, as autochthonous orebodies are usually chemically zoned (WEBB 1958; PINEDO 1963; CARVALHO *et al.* 1976; CARVALHO 1979; ROUTHIER *et al.* 1980; ALBOUY *et al.* 1981), underlain by stockwerk ore zones and surrounded by extensive hydrothermal alteration of the host lithologies (WEBB 1958; RAMBAUD 1969; STRAUSS & MADEL 1974; SOLER 1974; CARVALHO 1976; PLIMER & CARVALHO 1982; SALPETEUR 1976; BARRIGA *et al.* in press). Also of genetic significance is the fact that the sulphide deposits are usually surrounded or even capped by siliceous and metalliferous sediments: manganese lenses, cherts, jaspers, hematitic purple shales (SCHERMERHORN 1971b; CARVALHO *et al.* 1976; CARVALHO 1979).

The above considerations clearly indicate that ore genesis is a consequence of hydrothermal activity penecontemporaneous with volcanism, expressed in the sulphide ores themselves, the metalliferous and siliceous sediments and by the hydrothermal wall rock alteration around the ore zones. Additionally, recent data suggest that the pervasive hydrothermal metamorphism that affects most (if not all) volcanic rocks in the Iberian Pyrite Belt may also be intimately related to ore genesis (MUNHÁ & KERRICH 1980; MUNHÁ *et al.* 1980; BARRIGA 1983). We will briefly summarize the possible genetic significance of each of these aspects.

The Sulphide Ores

The Iberian Pyrite Belt sulphide deposits are composed of massive lens shaped bodies underlain (in the case of autochthonous orebodies) by a crosscutting zone of stringer and disseminated mineralization. The latter corresponds to the channelways of the mineralizing solutions (WEBB 1958; WILLIAMS *et al.* 1975; GARCÍA 1977). Oxygen isotope data (BARRIGA & KERRICH 1981; MUNHÁ & KERRICH 1981; BARRIGA *et al.* in press) indicate formation temperatures of 215 to 270°C for the stockwerk of Feitais-Estação (Aljustrel), Chança and Cerro Colorado (Rio Tinto), and 340°C for the chalcopyrite-pyrite disseminated deposit of Salgadinho. Ore zone temperatures for Feitais-Estação are around 150°C and the overlying jaspers formed at ~ 110°C (BARRIGA & KERRICH 1983). The ore zones are therefore characterized by steep thermal gradients at the time of ore formation. Ore zonation with Cu-rich basal domains, Zn-Pb intermediate zones and late stage precipitation of sulphates, followed upwards by Mn-cherts, indicates that as temperature decreased upwards the oxygen potential increased in the same direction. This conclusion is further supported by microprobe data revealing significant stratigraphically controlled compositional variation in sulphide and gangue minerals: Fe/Mn ratios in chlorite and FeS contents in sphalerites both decrease upwards (BARRIGA *et al.* in press). These thermal and chemical gradients are probably due to mixing of the mineralizing fluid with sea water (see KRAUSKOPF 1957; HELGESON 1970; LARGE 1977).

Metalliferous and siliceous sediments

Massive sulphide deposits of volcanic association throughout the world are often (if not generally) closely accompanied by metalliferous and siliceous sediments, both laterally equivalent to the sulphide ores and partly or totally capping them. They are often of considerable lateral extent, to the point of becoming important mineral exploration guides in some massive sulphide districts. The 'key tuffites', 'exhalites' and 'cherty iron-formations' of the North American literature on massive sulphide deposits (ROBERTS 1975; RIDLER 1971; HUTCHINSON 1973) and the Japanese Kuroko 'ferruginous beds' (LAMBERT & SATO 1974) are good examples of this association of sulphide mineralization and siliceous metalliferous sediments.

According to CARVALHO (1979) '... jaspers are perhaps the most typical rocks of the Iberian Pyrite Belt. Several types are known, which can occur at one or more levels of the volcanic sequences. The facies more closely related with sulphide mineralization stands out as it is generally light grey and contains scattered pyrite crystals. At Feitais (Aljustrel) these jaspers contact directly with the sulphide orebody. They grade laterally into another type of jaspers, predominantly red, with hematite and manganese oxides.

Similar 'siliceous chemical sediments', Fe- and Mn- bearing, are also present in the other volcanogenic massive sulphide-bearing districts and must reflect a sequentional evolutive trend of the mineralizing solutions whose mechanism and significance must be clarified' (translated from the Portuguese). Studies in progress (BARRIGA 1983) fully confirm these views, and provide evidence for a hydrothermal origin of the Fe-Mn cherts at Aljustrel (mainly on oxygen isotopes and on REE and transition metals abundances, see KNAUTH & LOWE 1978; BONATTI 1981 and CRERAR *et al.* 1982). Regarding the jasper facies variation with distance to sulphide mineralization, mineralogical, textural and geochemical data show that it is due to hydrothermal alteration of the jasper itself by reducing fluids expressed mainly in the vein-controlled reduction of hematite to magnetite and to pyrite and chloritization (BARRIGA 1983). The Aljustrel jaspers are therefore believed to have formed during the early stages of hydrothermal activity, in a predominantly oxidizing environment, as the primary Fe phase is hematite. Subsequent alteration near the sulphide orebodies shows that hydrothermal fluid venting continued thereafter. The alteration mineral assemblages indicate that the hydrothermal fluid was strongly reducing.

Ore zone hydrothermal alteration

During the first half of the twentieth century ore zone hydrothermal alteration was an essential aspect of both ore genesis and exploration oriented studies in the Iberian Pyrite Belt. COLLINS (1922), EDGE (1928), BATEMAN (1927), WILLIAMS (1934) and WEBB (1958) all stressed this aspect. In the late fifties (OFTEDHAL 1958) the sedimentary-exhalative theory originally advocated by KLOCKMAN (1894) gained new adherents in the Iberian Pyrite Belt literature. Since then sedimentary features received most of the atention, at the expense of hydrothermal aspects. It is certainly true that the proper recognition of the determinant factors in the conditions of sedimentation can be essential in ore exploration (SCHERMERHORN 1970). However, we maintain that hydrothermal effects are the key to the understanding of ore genesis. Ore zone hydrothermal alteration was rather neglected during about 15 years of modern studies in the Iberian Pyrite Belt with the exception of only a few, usually brief mentions (STRAUSS & MADEL 1974; SOLER 1974; SALPETEUR 1976; AYE & STRAUSS 1975). A new stage commenced with CARVALHO (1976), who showed that detailed mapping of alteration zonalities can contribute significantly to ore discovery; he has repeatedly emphasized this aspect, so carefully studied in other similar metallogenic provinces (CARVALHO 1976, 1979). Within the framework of the above discussion ore zone hydrothermal alteration must be envisaged as the result of interaction of mineralizing fluids, wallrocks and sea water, with the production of a complex set of alteration mineral assemblages according to a more or less well zoned, concentric pattern. The possibility that the more peripheral aureoles might be formed by essentially isochemical thermal metamorphism cannot be excluded, and it should also be kept in mind that subsequent regional metamorphism (see MUNHÁ in press) may have changed some of the lower grade alteration mineral assemblages. Despite these difficulties some coherent patterns are beginning to emerge. The clearest is that stockwerk ores are hosted in footwall rocks that are intensely hydrothermally altered, often to such extreme degrees that the original texture and mineralogy are lost and the rocks become aggregates of exclusively alteration minerals, usually quartz, chlorite and sulphides (chalcopyrite, pyrite, minor pyrrhotite, sphalerite), sometimes with significant sericite and/or carbonates. This lithotype is here named *stockwerk rock* (Plate 1D). Most stockwerk is a replacement of felsic volcanic rocks, usually of quartz-keratophyric composition, and gradational transitions exist from quartz-keratophyre to stockwerk rock. Intermediate areas are often sericite and carbonate rich. Original volcanic textures are progressively obliterated by corrosion of the original minerals and the concomitant precipitation of the alteration phases. This type of alteration is often present also above the orebodies, or laterally, and a gradual decrease of the degree of alteration with distance to the orebodies (along and up strike) is usually discernible. As already mentioned, hanging wall hydrothermal alteration is particularly striking at Feitais-Estação (BARRIGA 1983; see also SCHERMERHORN 1978; CARVALHO 1979). It is interpreted as meaning that hydrothermal fluid venting was taking place after deposition of the immediate hanging wall rocks (jaspers and 'siliceous schists').

The Salgadinho (Cercal) disseminated chalcopyrite deposit was found partly as a result of alteration studies (CARVALHO 1976). Additional work (PLIMER & CARVALHO 1982) provided new data on the alteration assemblages. The deposit is stratigraphically controlled, and was formed by porosity filling and replacement of the uppermost tens of metres of the Cercal felsic volcanic rocks, accompanied by extreme hydrothermal alteration of the host rhyolitic tuffs. The alteration zone is stratiform, and spreads away laterally from the area of the deposit. The principal alteration minerals are a pale green celadonitic fluoro-muscovite, ankerite and pyrite, accompanied by chlorite (7 % MgO, 1,4 % MnO) in the most intensely altered Upper Acid Tuffs. These tuffs (ore hosts) are overlain by a shale formation (S. Luís shales) depicting similar hydrothermal alteration in its lowermost 5 metres, again suggesting that hydrothermal activity persisted after the main phase of felsic volcanism, and until the beginning of deposition of the S. Luís shales.

Regional hydrothermal alteration

The Iberian Pyrite Belt volcanic rocks are essentially of spilitic and quartz-keratophyric composition (SCHERMERHORN 1975; SALPETEUR 1976; MUNHÁ 1981). Recent detailed studies leave little room for doubt that these are not primary igneous rocks, but rather the product of hydrothermal metamorphism of basaltic and rhyolitic or dacitic precursors (MUNHÁ 1979; MUNHÁ & KERRICH 1980; MUNHÁ *et al.* 1980; BARRIGA 1983; BARRIGA & KERRICH 1983). These studies also show that the fluid involved in this regional hydrothermal alteration is sea water, as evidenced by an impressive array of data that includes oxygen isotope compositions, variations in the oxidation states of iron, hydrothermal metamorphic mineral assemblages and heat and mass transfer considerations. Hydrothermal metamorphism was strongly non-isochemical, and produced a wide spectrum of compositions out of basaltic rocks (spilites) and out of acid rocks (quartz-keratophyres). Metamorphic grades change abruptly (sometimes in tens of metres) from very low grade up to lower amphibolite facies. The water content in the rock was generally high, sometimes evolving to low overall values (such as at Aljustrel, see BARRIGA & KERRICH 1983). Data also suggest that sea water was progressively modified in the course of the hydrothermal metamorphic process, from cold, oxydized, Na, Mg, and K rich into hot, reduced and Si and transition metal bearing, as evidenced by reciprocal changes in the rocks. Particularly significant are the analyses of a doleritic sill at Garrochal (MUNHÁ & KERRICH 1980) which show that its margins, more susceptible to sea water access, are preferentially hydrated, oxidized and enriched in alkali metals, and depleted in Co, Cu and Zn, as well as mineralogical and geochemical data on the Aljustrel volcanic rocks (BARRIGA 1983) showing striking Mg fixation in oxidized rocks and Fe, Zn and Cu depletion in reduced rocks, as well as complex and evolving behaviour of the alkaline metals.

This regional hydrothermal alteration is analogous to that found in ophiolites (see SPOONER & FYFE 1973; SPOONER *et al.* 1974; COLEMAN 1977; MOODY 1979; McCULLOCH *et al.* 1980), and is elegantly explained by the convective circulation of sea water through the volcanic piles, a process known to occur through sea floor rocks and which is responsible for primary oceanic metamorphism (see ELDER 1965; PALMASON 1967; LISTER 1972, 1981; WILLIAMS *et al.* 1974; ANDREWS 1978; FYFE & LONSDALE 1981; BARRIGA *et al.* 1983). It should also be mentioned that experimental data confirm the geochemical changes found (HAJASH 1975; BISCHOFF & DICKSON 1975; SEYFRIED & BISCHOFF 1977; HAJASH & ARCHER 1980; MOTTL & SEYFRIED 1980).

Ore genesis

We have seen that there is little doubt that the Iberian Pyrite Belt massive sulphide deposits precipitated at or near the coeval sea floor, either during the waning stages of local volcanic activity or shortly thereafter, from hot, reducing, slightly acidic metal-bearing solutions, either by a drop in temperature or mixing with cold sea water, or both. These are general conclusions that apply equally well to other volcanogenic massive sulphide deposits throughout the world. Genetic models for this class of deposits have varied widely through time, but currently most authors favour a hydrothermal metamorphic model whereby convective circulation of sea water through footwall rocks, driven by heat released from within the rocks themselves and from below (magma chambers, intrusions) eventually modifies sea water into a hot, reduced acidic, mineralized brine which, upon focused return flow to the ocean and given appropriate conditions at the site of discharge, may lead to the fast and concentrated precipitation of metal sulphides (see SPOONER & FYFE 1973; OHMOTO & RYE 1974; SOLOMON 1976; HEATON & SHEPPARD 1977; ANDREWS & FYFE 1976; HUTCHINSON *et al.* 1980; FRANKLIN *et al.* 1981). The model was recently confirmed by spectacular findings of hot springs issuing mineralized brines at mid ocean ridges and precipitating sulphides upon contact with sea water (BALLARD & GRASSLE 1979; FRANCHETEAU *et al.* 1979; CORLISS *et al.* 1979; RISE Project Group, 1980). Theory (BARNES & CZAMANSKE 1967; KRAUSKOPF 1967; HELGESON 1970), experiments (HAJASH 1975; BISCHOFF & DICKSON 1975; BISCHOFF & SEYFRIED 1978; MOTTL & HOLLAND 1978) and observation of natural hydrothermal solutions (Salton Sea brine, WHITE *et al.* 1963; Red Sea brine, CRAIG 1966; Galápagos Springs, CORLISS *et al.* 1979; East Pacific Rise 'black smokers', RISE Project Group, 1980) strongly suggest that metal transport is as chloride or mixed halide complexes (see also ANDREWS & FYFE 1976; HUTCHNSON *et al.* 1980; PLIMER & CARVALHO 1982), which seems to preclude significant amounts of reduced sulphur species in the mineralizing solution (BARNES & CZAMANSKE 1967, p. 355). Given that the latter is known to be reduced (from the ore zone alteration assemblages), it cannot transport SO_4^{-2} either. A separate sulphur source is therefore indicated. This hypothesis was first postulated by LOVERING (1961), and study of the Red Sea and East Pacific Rise ore systems shows that in the case of massive sulphide deposits the sulphur source is unmodified sea water, a readily available fluid at or near the sea floor. The convectively circulated modified sea water which transports the ore metals seems to loose its sulphur content in the process, as is indicated by experimental data (quoted above) and by the common occurrence of scattered pyrite in input rocks (ANDREWS 1978; FYFE & LONSDALE 1981). It is therefore upon return flow (output) of the modified, metalliferous sea water back to contact with cold, more or less pristine sea water that ore precipitation conditions are met. The necessary reduction of the sea water sulphate can be accomplished by inorganic high ($>200°C$) temperature mechanisms or by biogenic activity (indicated in some important deposits, such as Kidd Creek, Canada, where abundant graphite is found in the ore zone, see WALKER *et al.* 1975). For details on this aspect see HUTCHINSON *et al.* 1980 and FYFE & LONSDALE 1981.

There is firm evidence to support the applicability of this general model for the genesis of the Iberian Pyrite Belt massive sulphide deposits. As already mentioned, it is proved that pre-orogenic sea water hydrothermal metamorphism affected most (if not all) volcanic rocks in the area, and that abundant siliceous and metalliferous sediments are usually associated with ore. This speaks for the hydrothermal supply of Si and transition metals to the coeval sea water (see WOLERY & SLEEP 1976). Furthermore, MUNHÁ & KERRICH (1980) have shown that the sulphur isotope composition of the orebodies can easily be attributed to a sea water origin for the sulphur. Ore fluid oxygen isotopic signatures from Rio Tinto (Spain), Chança, Feitais-Estação and Salgadinho (Portugal) have been interpretd as of sea water derivation (BARRIGA & KERRICH 1981; MUNHÁ & KERRICH 1981; BARRIGA *et al.* in press). It should be mentioned that a contribution of magmatic mineralized water, although not necessary, is perfectly compatible with the data and the model.

Some objections can be raised against the sea water hydrothermal metamorphic model briefly outlined above. For example, all known sulphide deposits in the Pyrite Belt are associated with felsic volcanism, none with mafic volcanic rocks, and this cannot be explained by any feature of the rocks or ores themselves. Another fact is that sometimes several stacked cycles of felsic volcanism are present, and only one (the better developed) contains significant sulphide mineralization, despite the presence of abundant metalliferous sediments in other cycles as well. Also, sulphide mineralization seems to have taken place in geologically short time-spans as evidenced by the lack of sedimentary dilution of most orebodies. Genetic models need not explain all the facts, but should accomodate them. Further research is needed, particularly on time constraints on hydrothermal processes. At present, little more than a guess can be made to explain the felsic association of the Iberian Pyrite Belt massive sulphide deposits. It is certainly very tempting to base a classic orthomagmatic model on this fact, but care must be taken, however, as there is no evidence at all that purely igneous processes could concentrate the ~100 million tonnes of transition metals that are present, for example, in the Aljustrel orebodies, from a reasonably sized plutonic body (see for example KRAUSKOPF 1967). Even in porphyry copper systems 'no clear-cut evidence exists to relate fluids of either predominantly magmatic or meteoric origin to hypogene mineralization' (BEANE & TITLEY 1981). Another possibility is that sulphide mineralization indeed requires plutonic bodies, but essentially to provide heat from below to drive the sea water convective system beyond the initial stage of dissipation of heat from within the volcanic rocks themselves, and that the mafic magmas could come from the mantle without formation of upper crustal magma chambers. Again, further research is needed.

4. CONCLUDING STATEMENT

Our knowledge concerning the Iberian Pyrite Belt massive sulphide deposits has increased greatly during the last twenty years or so. Despite several still unsolved aspects, first rate field geology presently constitutes a solid base on which metallogenic studies are founded.

The deposition of enormous amounts of stratiform base metal sulphides is one of the most outstanding features of the Carboniferous evolution of the Hercynian Chain. These deposits formed as a consequence of specific Carboniferous geotectonic constraints and illustrate eloquently that ore deposits are the result of normal geological processes, and that their understanding requires contributions from essentially all the different fields of geology as represented in this Volume.

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PLATE

A. Colloform textures in sulphide ore (Feitais, Aljustrel). Pyritic framboidal nucleii surrounded by concentric alternating layers of silicates, pyrite (py), sphalerite (sph), chalcopyrite (cpy) and galena (gal).

B. Recrystallized massive ore (Feitais, Aljustrel). Largely euhedral pyrite with interstitial and intracrystalline chalcopyrite in a sphalerite (+ silicates).

C. 'Turbidite textured' ore (Tharsis, Spain), with clear graded bedding, clastic pyrite and black shale fragments in a fine matrix of pyrite + sphalerite + chalcopyrite + + silicates. Note penetrative cleavage and plastically disposed chalcopyrite preferentially concentrated in the coarser layers.

D. Ore grade stockwerk rock (Feitais, Aljustrel): Quartz-chlorite rock containing a chalcopyrite-rich crosscutting vein (right, dark grey) and scattered chalcopyrite and pyrite.

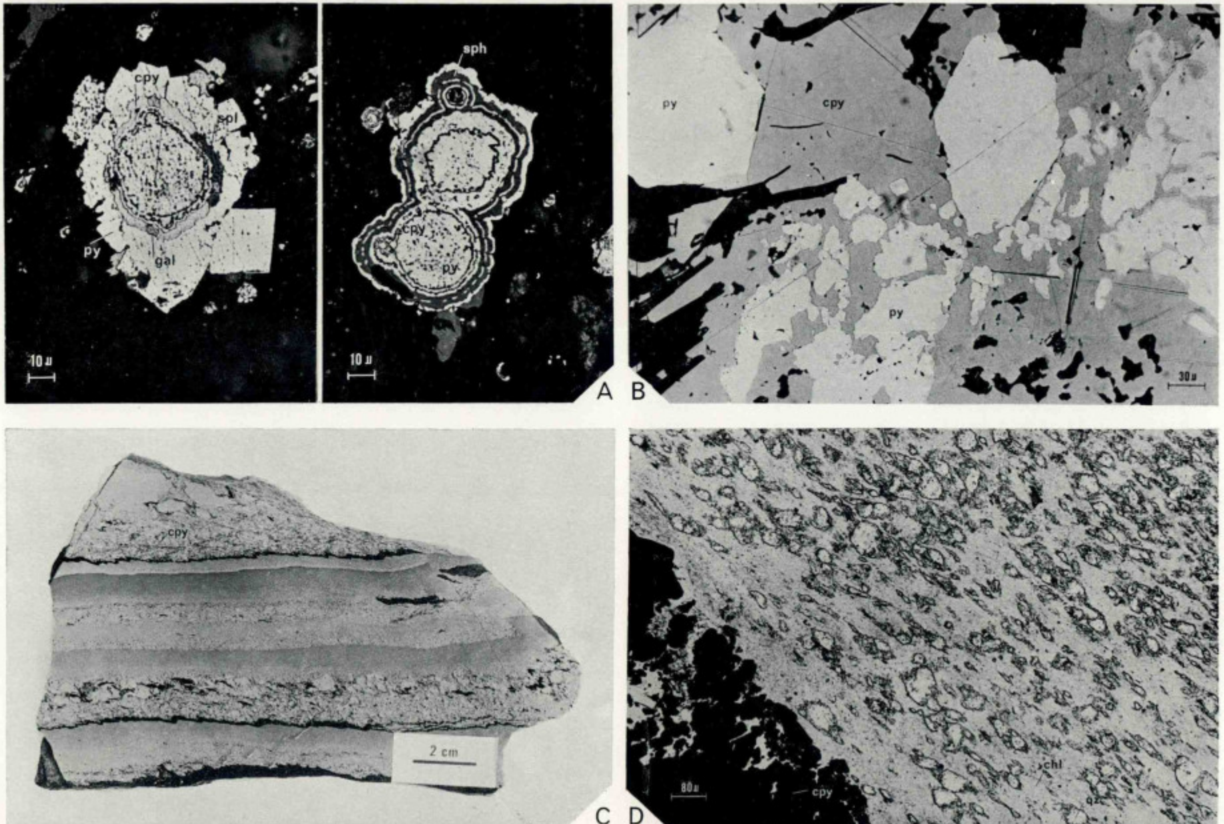
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Plate I

TERRESTRIAL CARBONIFEROUS

GENERAL DESCRIPTION OF THE TERRESTRIAL CARBONIFEROUS BASINS IN PORTUGAL AND HISTORY OF INVESTIGATIONS [1]

M. J. Lemos de Sousa [2] and R. H. Wagner [3]

Key words: Terrestrial Carboniferous, Portugal; Douro Basin; Douro Coalfield; Buçaco Basin; Santa Susana Basin; Westphalian C; Westphalian D; Stephanian C.

Palavras-chave: Carbonífero de fácies continental em Portugal; Bacia do Douro; Bacia Carbonífera do Douro; Bacia do Buçaco, Bacia de Santa Susana; Vestefaliano C; Vestefaliano D; Estefaniano C.

ABSTRACT

Terrestrial Carboniferous deposits occur in three different areas, viz.

(1) General vicinity of Oporto: a narrow NW-SE aligned strip of Westphalian C(?) and upper Westphalian D strata together with the lower Stephanian C deposits of the Douro Coalfield, altogether between Criaz and Mioma (northeast of Viseu). These strata constitute thrust slices between Lower Palaeozoic rocks, and are badly sheared as well as cleaved. The different outcrops are those of Criaz-Serra de Rates, Casais-Alvarelhos, Ervedosa and the Douro Basin. The latter contains meta-anthracites.

(2) Buçaco Basin, north of Coimbra: a N-S aligned basin alongside the Oporto-Coimbra-Badajoz-Córdoba Fault which separates the Central Iberian and Ossa-Morena zones. It shows tectonically induced alluvial fan and fluviatile deposits with a 40 m thick intercalation of floodplain deposits containing floral remains. This is of latest Stephanian C age.

(3) Santa Susana Basin in Alto Alentejo: another N-S aligned basin linked to a major fracture zone, in this case separating the Ossa-Morena and South Portuguese zones. It contains uppermost Westphalian D coal-measures.

General geological descriptions are accompanied by brief accounts of published investigations.

RESUMO

Os depósitos do Carbonífero de fácies continental ocorrem em três áreas diferentes a saber:

(1) Vizinhanças do Porto.

Entre Criaz (nordeste da Póvoa de Varzim) e Mioma (nordeste de Viseu) estende-se estreita faixa de terrenos carboníferos ao longo da qual é possível identificar afloramentos de diferentes idades: Criaz-Serra de Rates (Vestefaliano ?), Casais-Alvarelhos(Vestefaliano C ?), Ervedosa (Vestefaliano D superior) e Bacia do Douro ou afloramento Dúrico-Beirão (Estefaniano C inferior).

Trata-se de bandas fortemente laminadas entre afloramentos de rochas do Paleozóico inferior. A presença de xistosidade é, igualmente, evidente. A Bacia Carbonífera do Douro que constitui a parte NW do afloramento Dúrico-Beirão contém camadas de metantracite.

(2) Bacia do Buçaco a norte de Coimbra.

Aflora com orientação N-S ao longo da falha Porto--Coimbra-Badajoz-Córdova que separa as zonas Centro-

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(2) Mineralogia e Geologia, Faculdade de Ciências, Universidade do Porto, Praça Gomes Teixeira, 4000 Porto, Portugal.

(3) Department of Geology, The University, Beaumont Building, Brookhill, Sheffield S3 7HF, England, U. K., and E. N. Carbonífera del Sur, Peñarroya — Pueblonuevo (Córdoba), España.

-Ibérica e de Ossa-Morena. A série sedimentar inicia-se por depósitos de leque aluvial, induzidos pela tectónica, a que se seguem cerca de 40 m de depósitos de planície aluvial contendo uma flora fóssil cuja idade corresponde ao Estefaniano C mais superior. A série termina por depósitos fluviais.

(3) Bacia de Santa Susana no Alto Alentejo.

Aflora, também com orientação N-S, ao longo da falha principal que separa as zonas de Ossa-Morena e Sul Portuguesa. A idade das formações corresponde ao Vestefaliano D mais superior.

Faz-se a descrição geológica geral de cada uma das ocorrências citadas, em cada caso acompanhada da história das investigações anteriormente levadas a efeito.

1. INTRODUCTION

Terrestrial Carboniferous deposits occur in three different areas (Fig. 1) (compare Geological Map of Portugal (scale 1:500,000) — TEIXEIRA 1972). These are, from north to south, (i) the 130 km long, narrow strip of upper Westphalian and Stephanian C deposits in the general vicinity of Oporto; (ii) the Buçaco Basin north of Coimbra;and (iii) the Santa Susana Basin in the Alto Alentejo.

The general aspects of these occurrences have been dealt with by COSTA (1931) and TEIXEIRA (1944a, 1945a, 1952b, 1954, 1960, 1968). TEIXEIRA's thesis, published in 1944, is particularly important. One should also refer to the relevant chapters in textbooks and handbooks on the geology of Portugal, e.g. TEIXEIRA (1981), TEIXEIRA & GONÇALVES (1980), RIBEIRO (1979a, b). All these occurrences are linked to major fault zones. The general tectonic aspects have been discussed by COSTA (1945, 1951b, 1952b, c, 1953, 1954), FLEURY (1919/22), NORONHA *et al.* (1979, 1981) and TEIXEIRA (1942c, 1943a, b).

2. OCCURRENCES NEAR OPORTO

The most important and structurally most complex area is that found north, east and southeast of Oporto (Fig. 1). This contains cleaved rocks of two different late Westphalian ages and the lower Stephanian C strata, also cleaved, of the Douro Basin which contains the meta-anthracites worked in the Douro Coalfield. The Douro Basin (Bacia do Douro) occurs in a 90 km long strip extending from São Pedro

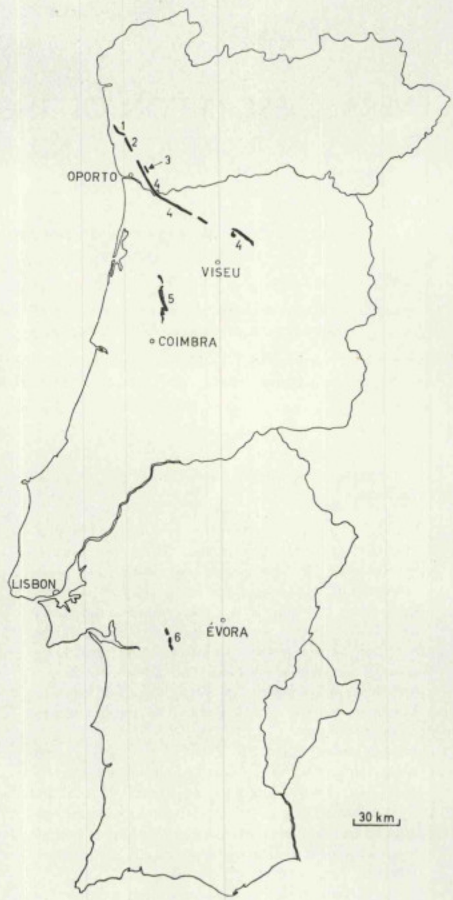

Fig. 1 — Terrestrial Carboniferous occurrences in Portugal: 1 — Criaz-Serra de Rates (Westphalian?); 2 — Casais-Alvarelhos (Westphalian C?); 3 — Ervedosa (upper Westphalian D); 4 — Douro Basin (lower Stephanian C); 5 — Buçaco Basin (uppermost Stephanian C or lowermost Autunian); 6 — Santa Susana Basin (uppermost Westphalian D).

Fins (just east of Oporto) to Mioma (northeast of Viseu). This strip is interrupted halfway by the Castro Daire Granite. The most important mines of the Douro Coalfield, in the northern part of this strip, are those of São Pedro da Cova (abandoned in 1972) and Pejão (still worked at the present time).

The first reference to the geology of the Douro Basin (in the region of São Pedro da Cova) is found in ESCHWEGE (1833). Further records were provided by SHARPE (1834, 1849; see also PINTO 1932), SCHMITZ (1852), C. RIBEIRO (1861, 1862, 1863), AZEVEDO (1858, 1866, 1867). These contained general descriptions of the coal-bearing strata. In the 1940's and 1950's several papers were published on different aspects of the basin, including general geological descriptions of the coals and coal-bearing strata and more specific studies of conglomerates and porphyrites (R. ANDRADE 1955; FONSECA 1953; FREIRE 1947; NEIVA 1943; J. PEREIRA 1945; TEIXEIRA & FONSECA 1945; TEIXEIRA *et al.* 1942).

Borehole campaigns in the Douro Coalfield were described by FREIRE (1943, 1981), MELLO & BETTENCOURT (1939) and in *Reconhecimento complementar por sondagens do jazigo carbonífero de S. Pedro da Cova* (1962).

The petrology of the meta-anthracite coals was studied by SOUSA (1983 — this volume and earlier papers).

Descriptions of the geology of the Douro Basin can also be found in several regional geological studies, e.g. COSTA (1929, 1958), MEDEIROS (1945), SCHERMERHORN (1956), SCHMITZ (1895), SLUIJK (1963) and WESTERVELD (1956). Detailed geological maps of the region were published by COSTA & TEIXEIRA (1957), DELGADO (1908), MEDEIROS *et al.* (1964), E. PEREIRA *et al.* (1980), SCHERMERHORN (1980), TEIXEIRA *et al.* (1972). The area is also covered in part by the following sheets of the Geological Map of Portugal (scale 1:50,000): 9C (Porto), 13B (Castelo de Paiva), 13D (Oliveira de Azeméis), 14C (Castro Daire) and 14D (Aguiar da Beira). A paper dealing specifically with the tectonic structure of the São Pedro da Cova area (Douro Coalfield) was published by COSTA (1951a, 1952a).

Most recently, two syntheses of earlier investigations were published by FREIRE (1981) and SOUSA (1978).

The coal-bearing strata of the Douro Basin have yielded a flora of early Stephanian C age. The succession commences with a basal conglomerate, apparently associated with a palaeotopography and in angular unconformity with Precambrian-Cambrian rocks (the so-called 'Complexo Xisto-Grauváquico') on the southwestern side of the occurrence. The lower Stephanian C strata are limited northeastwards by a faulted contact with Lower Palaeozoic sediments of the Valongo Anticline. The latter is a major structure comprising Ordovician to Devonian strata resting unconformably on Precambrian-Cambrian rocks ('Xisto-Grauváquico') which forms the core of the anticline.

In part of the Douro Basin it is clear that the lower Stephanian C deposits are folded into a synclinal structure which is not, however, linked to the Valongo Anticline. Not only does this synclinal structure correspond to a different folding phase, but its contact with the Valongo Anticline is by means of a major thrust fault (or, rather, shear zone) which accompanies the entire strip of Carboniferous deposits along its total length of 130 km. The direction of thrusting is southwestwards, which agrees with the vergence of the Valongo Anticline. The coal-bearing strata are badly sheared, as follows most clearly from the variable separation of coal seams as shown by the mine plans of São Pedro da Cova in the so-called 'bacia clássica' (i.e. the western coal-mining area, where the Stephanian C strata are in unconformable contact with the 'Xisto-Grauváquico'). The shearing is also evident from the thrust contact between slices of Silurian and Stephanian rocks in the so-called 'bacia oriental' (i.e. the eastern coal-mining area) (FONSECA 1946, 1959; SOUSA 1977).

In a small area in the northern part of the Douro Coalfield, between Tanjarro and the Montalto Sanatorium, in the general area of Ervedosa, a narrow slice of upper Westphalian D strata has been caught up in the thrust zone. This outcrop has yielded fossil plants but no coal

(TEIXEIRA 1944a, 1945a). It is structurally unconformable with the adjacent Devonian rocks but has lost its stratigraphic contacts.

Westphalian deposits, probably belonging to the Westphalian C (WAGNER & SOUSA 1982) occur further north at Casais-Alvarelhos, in what appears to be a structural continuation of the Ervedosa outcrop (TEIXEIRA 1957; TEIXEIRA *et al.* 1965; Geological Map of Portugal (scale 1:50,000), sheet 9A: Póvoa de Varzim). It may be assumed that both represent partial remnants of a Westphalian C-D succession, which was laid down unconformably on Lower Palaeozoic and Precambrian rocks and which was overlain in turn, also unconformably, by the lower Stephanian C coal-bearing strata of the Douro Basin. The tectonised nature of the present-day occurrences does not allow any speculation on the size and shape of the two successive Carboniferous basins.

Some inadequately dated Carboniferous outcrops follow the fault zone northwestwards to Criaz-Serra de Rates (DELGADO 1908; MEDEIROS undated publication; TEIXEIRA *et al.* 1965; WAGNER & SOUSA 1982; Geological Map of Portugal (scale 1:50,000), sheet 9A: Póvoa de Varzim).

3. BUÇACO BASIN

This refers to a tightly folded syncline in uppermost Stephanian C or possibly lowermost Autunian strata, which occurs north of Coimbra in an area immediately east of the Porto-Coimbra-Badajoz-Córdoba Fault, one of the principal early fracture zones in the Iberian Peninsula.

The first references to the Carboniferous of the Buçaco Basin are found in C. RIBEIRO (1850, 1853a, b), whose data were summarised in the more general work of A. SIMÕES (1853, 1895). The stratigraphic age of this basin was first discussed by LIMA (1888/92b), who compared with the Rotliegend of Germany. NEIVA (1943) and CARVALHO (1949) studied the conglomerates of the Buçaco Basin. The latter reported a weak aeolian effect on some of the conglomerate clasts. CHARNAY (1962) and COURBOULEIX (1972, 1974) incorporated the southern and northern parts of the Buçaco Basin in their regional studies of the area near Buçaco, and COURBOULEIX & ROSSET (1974) discussed the tectonic structure of the basin. A general map of the Buçaco Basin had already been published by DELGADO (1908) and a more detailed map by the same author was published posthumously by COSTA (1950).

Stratigraphic studies undertaken by WAGNER *et al.* (in prep.) and which are based on a detailed geological map produced by GOMES DA SILVA, indicate the following stratigraphic succession for the northern part of the basin studied earlier by COURBOULEIX. This succession has been established in the eastern flank of the syncline which is practically undisturbed (the western flank is sheared and does not show a complete succession). The first deposit found in the Buçaco Basin is a fanglomerate of variable thickness. The mapping has indicated that it is linked to a palaeotopography. The presence of a mobile basin margin on which active erosion took place is evident from the nature of the deposit, its red colouration and the subsequent deposition of 180-200 m of red mudflow deposits and debris flow conglomerates. These show a gradual transition to 42 m of grey floodplain deposits (alternating with the massive red beds in the lower part). This interval has yielded practically all the plant fossils recorded from the Buçaco Basin. It is followed by a little over 600 m of largely conglomeratic fluviatile deposits which are provisionally interpreted as a braided river complex. It is noted that the plant assemblages found in the grey interval show both floodplain and allochthonous associations with a relatively large number of extrabasinal elements in the latter. This reinforces the impression of topographic relief being present in the near vicinity.

The sheared western flank of the syncline shows the same stratigraphic units but the relatively thin grey interval is often missing. Precambrian phyllites underlie the upper Stephanian C (or basal Autunian) deposits unconformably on the western side which also shows contacts with the overlying Triassic red beds.

On the eastern side the Precambrian-Cambrian 'Xisto-Grauváquico' Complex is found. In other parts of the basin there is an unconformable contact with different elements of the Lower Palaeozoic succession in this area.

4. SANTA SUSANA BASIN

This basin has also been referred to in the older literature as the coalfield of Moinho da Ordem. The Carboniferous outcrops are found in three outliers, with a north-south orientation. They are positioned alongside the fault which separates the igneous massif of Beja from Devonian sediments and which is regarded as the separation between the Ossa-Morena and South Portuguese palaeogeographic and structural zones of LOTZE (1945). In fact, the Carboniferous rocks of Santa Susana (which are dated on floral remains as very late Westphalian D) rest unconformably on quartz porphyrites on the eastern (Ossa-Morena) side, whilst there is a faulted contact with Devonian sediments on the western side.

The earliest records regarding the Carboniferous of Moinho da Ordem are in LIMA (1895/98), where stratigraphic information occurs together with palaeobotanical data. Other papers dealing specifically with Santa Susana are those of CHOFFAT (1921) and J. SIMÕES (1921). The conglomerates in the lower part of the sucession of Santa Susana were studied successively by NEIVA (1943) and ASSUNÇÃO (1948), who determined the provenance of the granite boulders in the Carboniferous conglomerates.

The northern outlier, that of Jongeis, is the only one to have supported an active coal mine, which was worked for many years and finally abandoned in 1944. Apart from conglomerates there is little outcrop in the Jongeis area, but mine and borehole data show the presence of several hundred metres of partly conglomeratic coal-bearing strata with numerous plant remains and apparently of entirely terrestrial facies (C. ANDRADE 1927/30, 1955a, b). These strata were folded into a NW-SE striking syncline with variable dips ranging from c. 20° to c. 70°. It is difficult to obtain a proper impression of the complexity of the tectonic structure in the absence of a detailed cross section. The only section showing the disposition of coal seams is that published by TEIXEIRA (1944a, 1945a, p. 125) after a sketch diagram provided by the then director of the coal mine.

The other outlier of certain importance, at Vale do Burro (or Vale de Figueiras), is presently covered to a large extent by the reservoir of Pego do Altar. A borehole campaign (GUERREIRO & SANTOS 1955; C. ANDRADE 1955c; C. ANDRADE *et al.* 1951) failed to discover workable coals but showed the presence of shallowly dipping terrestrial strata with similar plant remains to those from Jongeis.

GONÇALVES *in* DOMINGOS *et al.* (1983 — this volume) has recently studied the geology of Santa Susana from temporary exposures in the reservoir and outcrops in its immediate vicinity. He described the stratigraphic succession in general terms as consisting of 150 m of coarse conglomerates, arkoses and shales.

The Remeiras outlier in between those of Jongeis and Vale do Burro presents similar characteristics as follows from the information obtained by a borehole campaign reported by SANTOS (1955).

5. PALAEONTOLOGICAL RECORDS

The Carboniferous strip near Oporto, the Buçaco Basin and the Santa Susana Basin have all yielded a fair abundance of palaeontological material. Most are plant impressions, but insect and other arthropod remains, phyllopods and non-marine bivalves were also described (FLEURY 1936/37, 1937; LAURENTIAUX & TEIXEIRA 1948, 1950, 1957/58, 1958; LIMA 1888/92a; TEIXEIRA 1939, 1941a, b, 1942d, 1943c, 1944b, c, 1945, 1947, 1952a; TEIXEIRA & FONSECA 1953; WATTISON 1926).

The bivalves have been studied again by EAGAR (1983 — this volume), and a revision of the fossil floras has been made by WAGNER & SOUSA (1983 — this volume).

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THE CARBONIFEROUS MEGAFLORAS OF PORTUGAL —A REVISION OF IDENTIFICATIONS AND DISCUSSION OF STRATIGRAPHIC AGES (¹)

R. H. Wagner (²) and M. J. Lemos de Sousa (³)

Key words: Carboniferous plants, Portugal; Douro Basin; Douro Coalfield; Buçaco Basin; Santa Susana Basin; Westphalian C; Westphalian D; Stephanian C; Rotliegend.

Palavras-chave: Plantas fósseis do Carbonífero de Portugal; Bacia do Douro; Bacia Carbonífera do Douro; Bacia do Buçaco; Bacia de Santa Susana; Vestefaliano C; Vestefaliano D; Estefaniano C; 'Rotliegend'.

ABSTRACT

A general revision of fossil plants from the different Carboniferous occurrences in Portugal has been undertaken. Composite lists of species have been established for each area and comments are made on certain identifications. A number of taxa have been illustrated.

The stratigraphic dating of the different occurrences is as follows:

The Criaz-Serra de Rates outcrop north of Oporto has yielded only one species which ranges from mid-Westphalian to lower Stephanian. The Casais-Alvarelhos outcrop in the same strip contains a flora not older than Westphalian B and not later than early Westphalian D. A Westphalian C age has been regarded as most likely. The Douro Coalfield, in the same strip but southeast of Oporto, contains an early Stephanian C flora. The thin thrust slice of Ervedosa, which lies parallel to the northwestern part of the Stephanian C coal-measures of the Douro Coalfield, contains a flora of late Westphalian D age.

The Buçaco Basin, north of Coimbra, possesses a flora of lower Rotliegend type with probably a very late Stephanian C age.

The Santa Susana Basin, in the Alto Alentejo, has a flora of very late Westphalian D aspect. The earliest Cantabrian cannot be excluded however.

RESUMO

Efectua-se uma revisão geral das plantas fósseis dos diferentes terrenos carboníferos portugueses de fácies continental. Para cada uma das áreas estudadas apresenta-se uma lista das espécies identificadas e fazem-se comentários acerca das determinações. Ilustram-se, além disso, alguns dos táxones.

Estabeleceram-se as seguintes datações estratigráficas:

O afloramento de Criaz-Serra de Rates a norte do Porto apenas forneceu uma espécie cuja extensão conhecida vai do Vestefaliano médio ao Estefaniano inferior. O afloramento de Casais-Alvarelhos contém, por seu lado, uma flora não mais antiga que o Vestefaliano B, nem mais moderna que o Vestefaliano D inferior,

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(²) Departement of Geology, the University, Beaumont Building, Brookhill, Sheffield S3 7HF, England, U. K., and E. N. Carbonífera del Sur, Peñarroya—Pueblonuevo (Córdoba), España.

(³) Mineralogia e Geologia, Faculdade de Ciências, Universidade do Porto, Praça Gomes Teixeira, 4000 Porto, Portugal.

havendo elementos que permitem atribuir-lhe, como mais fidedigna, idade correspondente ao Vestefaliano C. A Bacia Carbonífera do Douro, a sudeste do Porto mas pertencente ainda à mesma faixa que os afloramentos antes citados, contém uma flora cuja idade corresponde ao Estefaniano C inferior. A pequena mancha de Ervedosa, localizada paralelamente à Bacia do Douro, patenteia uma flora do Vestefaliano D superior.

A Bacia do Buçaco, a norte de Coimbra, possui uma flora do tipo da do Rotliegendes inferior, com idade provavelmente correspondente ao Estefaniano C mais superior.

A Bacia de Santa Susana no Alto Alentejo contém uma flora de idade que se pode paralelizar com o Vestefaliano D mais superior. Todavia, não é de excluir a presença do Cantabriano mais inferior.

1. HISTORY OF INVESTIGATIONS

The first reference to the Carboniferous plant fossils of Portugal is found in SHARPE (1849) who collected Stephanian plant remains from the 'Vallongo Section' (in fact the São Pedro da Cova region) and had them identified by C. F. J. BUNBURY. The same palaeontologist later identified plant fossils from Buçaco (BUNBURY *in* RIBEIRO 1853). This included one new species, *Pecopteris leptophylla* BUNBURY, which was to create a certain amount of confusion in the literature as the result of a misidentification by ZEILLER (1892). More complete lists of plant fossils from the Douro Basin appeared in RIBEIRO (1861, 1862, 1863). These presented a mixture of Stephanian and Westphalian species, probably as a result of misidentifications. No illustrations were provided. RIBEIRO also distinguished between two sets of strata in the São Pedro da Cova region. These correspond to what has later been called the 'bacia clássica' and the 'bacia oriental' of the coal mine at São Pedro da Cova, i.e. the western and eastern coal-mining areas.

GOMES (1865) provided the first systematic description of the Portuguese Carboniferous floras, dealing with the Douro Basin as well as Buçaco and Santa Susana (Moinho da Ordem) basins. GEINITZ (1867) commented upon GOMES's work and his remarks were duly taken into consideration by GOMES (1870). The significance of GOMES's work was fully discussed by TEIXEIRA (1941/42). The plants studied by GOMES were re-examined by TEIXEIRA, who compiled a catalogue of specimens kept in the Faculdade de Ciências de Lisboa. This collection is still available. Some of GOMES's specimens are in the Serviços Geológicos de Portugal and in the Faculdade de Ciências do Porto.

The records given by SHARPE, RIBEIRO and GOMES were discussed by FEISTMANTEL (1875), who commented on certain identifications. FEISTMANTEL's comments were analysed later by TEIXEIRA (1941c).

GRAND'EURY (1877, p. 433) also commented upon GOMES's work, referring the Portuguese coal deposits to his 'terrain houiller sous-supérieur'.

HEER (1881), in his memoir on the Fossil Floras of Portugal mainly dealt with Mesozoic and Tertiary plants but also included a brief (and incomplete) summary of Palaeozoic records, as well as providing the description of *Distrigophyllum lusitanicum*, a plant later referred to *Dicranophyllum* (LIMA 1888), and *Baiera gomesiana* which was subsequently identified as *Rhacopteris gomesiana* by TEIXEIRA (1943b). General comments on HEER's work were published by LIMA (1883/87).

KIDSTON (1886), in his Catalogue of the Palaeozoic Plants in the British Museum (Natural History), made certain corrections on the identifications given by GOMES and eliminated synonyms.

An important step forward in the knowledge of the Portuguese Carboniferous floras was made by LIMA, who published a monographic description of *Dicranophyllum gallicum* and *D. lusitanicum* in the Douro Basin (LIMA 1888), and also discussed the stratigraphic ages of the Buçaco and Santa Susana basins in a series of notes without illustrations (LIMA 1888/92, 1891, 1894, 1895/98). One of these (1894) included the description of *Neuropteris zeilleri* LIMA for which he referred to *Neuropteris cordata* GÖPPERT (*non* BRONGNIART) as figured in GÖPPERT's 'Permische Formation'. He recorded this species in Portugal from the Buçaco Basin. TEIXEIRA (1940d) finally illustrated *N. zeilleri* LIMA from Buçaco, where it is relatively common, and from the Douro Basin where it is rare. He envisaged an exhaustive description of the relevant floras and prepared a substantial atlas which remained unpublished as a result of other commitments (WENCESLAU DE LIMA became Foreign Minister and eventually Prime Minister of the last monarchist administration in Portugal — COSTA 1958). Nine plates had been printed already and these plates as well as a number of additional ones which had been prepared but not yet printed, were later published by TEIXEIRA (1941e), reduced by one half. TEIXEIRA also commented upon the identifications.

LIMA named several new species in his floral lists but these remained *nomina nuda*. His collection is still in the Serviços Geológicos de Portugal and has been re-examined by the present writers. However, with the exception of '*Desmopteris guimarãesi*' which TEIXEIRA (1944, 1945, p. 40) later identified as a pinnule (leaf) of *Taeniopteris jejunata*, these manuscript names do not appear on the labels of specimens currently in the collections of the Serviços Geológicos de Portugal.

STERZEL (1903) described a new species, *Sphenophyllum costae*, from the Stephanian of São Pedro da Cova. His description and illustration of the type specimen was reproduced again by ZOBEL (1910), *in* POTONIE's 'Abbildungen und Beschreibungen fossiler Pflanzen-Reste'. This species is a possible synonym of *Sphenophyllum alatifolium* RENAULT, but is generally kept separate and is fairly widely used (see TEIXEIRA 1955a).

The most voluminous work on the Carboniferous plant impressions of Portugal is that of TEIXEIRA, who published a long series of generally short papers dealing with all the different areas of terrestrial Carboniferous in the country. From 1937 to 1944 he systematically revised the material found in the older collections to which he added by further collecting. Successively, he published on Lycopodiales (1937a), *Pecopteris feminaeformis* (1939a), *Mariopteris* (1939c — dealing with two species subsequently attributed to *Diplothmema* (TEIXEIRA, 1941d) and currently regarded as belonging to *Pseudomariopteris* DANZÉ-CORSIN), *Sphenophyllum costae* (1939d), *Alethopteris* (1940a), *Neuropteris zeilleri* (1940d), and *Rhacopteris gomesiana* (1943b). He also published three papers discussing the work of FEISTMANTEL, LIMA, and GOMES (TEIXEIRA 1941c, 1941e, 1941/42). In his 1939c paper he introduced a new species, *Mariopteris corsini* TEIXEIRA which he later transferred to *Diplothmema* and which was finally referred to *Pseudomariopteris* by WAGNER (1962).

At the same time he published several papers on the stratigraphic significance of the floras from the various Carboniferous areas in Portugal (TEIXEIRA 1937b, 1938a, b, 1938/40, 1939b, c, 1941a, b, 1942a, b, c, d, e; TEIXEIRA & MEDEIROS 1943). Several new species were described in these papers, viz. *Neuropteris ovata* forma *ervedosensis* (later referred to *Neuropteris ervedosensis* by WAGNER 1963), *Neuropteris machadicostai* (referred to *Neuropteris flexuosa* STERNBERG in the present paper), *Linopteris florini*, *Linopteris* (?) *gomesi* (probably to be referred to *Reticulopteris germari* (GIEBEL) GOTHAN), *Sphenopteris alentejana*, *Sphenopteris mendes-correae*, *Sphenopteris sampaiana* (probably a form of *S. mendescorreae*), *Pecopteris limai* (which he later referred to as *Pecopteris lepidorachis* forma *limai*, but which the present writers prefer to regard as a separate species), *Pecopteris viannai*, *Pecopteris bussacensis*, *Pecopteris hemitelioides* forma *longipinnata* (which has proved to be the lower part of a pinna of *Pecopteris cyathea* (VON SCHLOTHEIM) BRONGNIART — see comments later on in the present paper), *Taeniopteris bertrandiana*, and *Sphenophyllum guerreiroi*.

The papers published in 1942c, 1942d and 1942e are particularly significant in that they proved the presence of Westphalian D strata in the general area of São Pedro da Cova (Douro Basin) where formerly only upper Stephanian deposits had been recorded.

In 1943a TEIXEIRA compared the Portuguese Carboniferous floras with those of the Spanish occurrences. Since very little had been figured from Spain at that time, these comparisons were only partly correct.

A complete summary of all his previous work was given in his major paper 'O Antracolítico Continental Português' (TEIXEIRA 1944, 1945). It is noted that some earlier identifications were modified in this paper, without providing reasons. A notable case is that of *Sphenopteris mendes-correae* which is listed as *Sphenopteris (Crossotheca) mendes-correae*. The earlier description of this species included a diagrammatic representation of the fructification which did not resemble *Crossotheca* and which proved to be quite misleading (see comments later in the present paper). This major paper also contains an exhaustive summary of the earlier records of Carboniferous flora in Portugal, and remains a generally useful guide to the knowledge of these floras up to the mid-nineteen forties.

Afterwards, TEIXEIRA continued to write more sporadic systematic descriptions of certain species (1949, 1951, 1953, 1955a, b, 1956, 1964); two of these are among the most useful records of Portuguese Carboniferous floral remains (1949, 1951). Two new species and one new genus were introduced at this stage, viz. *Sigillariostrobus serreatus* and *Gondomaria alethifolia*. The latter has proved to be a synonym of *'Callipteris' discreta* WEISS, a form that was referred to the undescribed genus *Pseudocallipteris* by GRAND'EURY (1877). *Gondomaria discreta* has been figured and described most recently by WAGNER & SOUSA (1982a). A general record of species found in boreholes in the area of São Pedro da Cova was published by TEIXEIRA in 1946.

TEIXEIRA (1954, 1981), TEIXEIRA & GONÇALVES (1980) and TEIXEIRA & PAIS (1976) repeated the lists of fossils published in TEIXEIRA (1944, 1945) in various syntheses on the Carboniferous and general geology and palaeontology of Portugal. These failed to take into account later revisions of identifications.

The Stephanian conifers of Portugal (partly recorded as lower Permian) were incorporated in the world-wide monograph of Palaeozoic conifers by FLORIN (1938/45), who summarised the Portuguese elements in a note published in 1940. One new species, *Lebachia parvifolia*, was based on a specimen from Buçaco.

Apart from TEIXEIRA's papers, already referred to, little work was published in the nineteen fifties. FONSECA (1954, 1959) recorded cannelate *Sigillaria* from the 'bacia oriental' of São Pedro da Cova and published some general lists of flora from this eastern part of the coalfield. DOUBINGER (1956) compared TEIXEIRA's records of flora from the Douro Basin and Buçaco with similar ones from the Massif Central in France. She regarded the Stephanian flora of the Douro Basin as probably representative of Stephanian C (rather than upper Stephanian B as TEIXEIRA had indicated), but accepted the early Autunian age which TEIXEIRA had ascribed to Buçaco. WAGNER (1959) compared TEIXEIRA's records with those from Northwest Spain, and proposed a few changes in the identifications. He did not query TEIXEIRA's age determinations.

In the seventies a few additional contributions were made by PAULE CORSIN & SOUSA (1972), PAIS (1973), and PAULE CORSIN (in COURBOULEIX 1972). The former dealt with the 'bacia oriental' of São Pedro da Cova, PAIS redescribed *Alethopteris* from the Douro Basin, and the latter recorded a Stephanian plant assemblage from Algeriz in the Buçaco Basin which was supposed to indicate the presence of Stephanian as well as Autunian strata in this area. In fact, the assemblage published by COURBOULEIX came from the same beds as yielded the so-called Autunian floras recorded earlier, the difference being due to habitat and not to stratigraphic age.

In the nineteen eighties commenced the revision reported in the next pages. This is based on the collections in the Serviços Geológicos de Portugal (Lisbon), the Faculty of Science in the University of Oporto (Museu Wenceslau de Lima), the Faculty of Science in the University of Lisbon, the Faculty of Science and Technology in the University of Coimbra, and additional collecting by the present authors. The material in Lisbon and Oporto had already been studied by TEIXEIRA, and included specimens identified by GOMES and LIMA. Among LIMA's material are specimens sent to France by LIMA and possibly identified by ZEILLER. Two papers providing systematic descriptions have already been published (WAGNER & SOUSA 1982a, b).

2. OCCURRENCES NEAR OPORTO

This refers to a 130 km long strip of cleaved Westphalian C (?), upper D, and lower Stephanian C strata in a faulted area which extends from Criaz to Mioma (see map in SOUSA & WAGNER 1983 — this volume, Fig. 1).

2.1. Criaz-Serra de Rates

The only plant fossils identified from this area are:

Linopteris florini TEIXEIRA
Calamites sp.

The present writers have identified the same species as TEIXEIRA (1942c, 1944, 1945, 1954). However, the latter deduced a Westphalian D age from the presence of *Linopteris florini*, which seems unjustified in the light of a recent revision indicating a total range from mid-Westphalian to early Stephanian (WAGNER & SOUSA 1982b).

2.2. Casais-Alvarelhos

A single assemblage has been described from the locality of Cidai-Bougado (WAGNER & SOUSA 1982a):

Linopteris florini TEIXEIRA
Alethopteris cf. *davreuxi* (BRONGNIART) GÖPPERT
Mariopteris nervosa (BRONGNIART) ZEILLER
Eusphenopteris nummularia (VON GUTBIER) NOVIK
Sphenopteris (Alloiopteris ?) cf. *dixoni* KIDSTON
Sphenopteris sp.
Gondomaria discreta (WEISS) WAGNER & LEMOS DE SOUSA
Sphenophyllum majus BRONN
Asterophyllites sp.

It is noted that *Gondomaria* is a synonym of *Pseudocallipteris* GRAND'EURY. However, this genus has never been described (GRAND'EURY 1877, p. 430).

The age is certainly Westphalian, with a possible range from Westphalian B to early Westphalian D. A Westphalian C age has been regarded as most likely (op. cit.).

2.3. Ervedosa

This classical locality of Westphalian D plants has been discussed repeatedly by TEIXEIRA (1942c, d, e, 1944, 1945, 1951, 1954) who also figured and described a number of plant fossils from collecting spots mentioned as Ervedosa, Tanjarro and Montalto.

A revision of material stored in the collections of the Serviços Geológicos de Portugal in Lisbon and at the Faculdade de Ciências do Porto has yielded the species listed in Table 1.

Certain changes have been made with regard to the earlier lists published from Ervedosa. These include a few additional identifications. The lists published by TEIXEIRA (1944, 1945, 1954) for the Tanjarro and Montalto localities in the general area of Ervedosa were already commented upon by WAGNER (1959, p. 405), whose observations have been taken into consideration. Additional comments are as follows:

Neuropteris ervedosensis TEIXEIRA

Neuropteris ovata forma *ervedosensis* TEIXEIRA is a species in its own right (compare WAGNER 1963, p. 27), and probably unrelated to *Neuropteris ovata*. Comparison can be made with *Neuropteris flexuosa* STERNBERG but there is no apparent relationship to this species. It is noted that DE JONG (1974, p. 66) retained *Neuropteris ervedosensis* in the group of *Neuropteris ovata*.

Dicksonites plueckeneti forma *sterzeli* ZEILLER

Two specimens have been figured by TEIXEIRA (1942d, Est. III, figs 5, 6) as *Pecopteris*

Table 1 — Composite list of the upper Westphalian D flora of Ervedosa (Oporto region)

Neuropteris ervedosensis TEIXEIRA
Neuropteris scheuchzeri HOFFMANN
Neuropteris sp.
Linopteris florini TEIXEIRA
Linopteris obliqua (BUNBURY) ZEILLER (pl. I, fig. 3; pl. II, fig. 1)
Callipteridium (Praecallipteridium) jongmansi (P. BERTRAND) WAGNER
Alethopteris corsini BUISINE (pl. I, figs 1-2)
Dicksonites plueckeneti forma *sterzeli* ZEILLER
Eusphenopteris cf. *neuropteroides* (BOULAY) NOVIK
Sphenopteris sp.
Oligocarpia pulcherrima STUR
cf. *Polymorphopteris polymorpha* (BRONGNIART) WAGNER
Lobatopteris micromiltoni (P. BERTRAND) WAGNER
cf. *Lobatopteris vestita* (LESQUEREUX) WAGNER
Pecopteris dentata BRONGNIART
Pecopteris cf. *hucheti* CORSIN
Pecopteris cf. *nyranensis* NĚMEJC
Pecopteris unita BRONGNIART
Pecopteris sp.
Sphenophyllum sp. indet.
Annularia sphenophylloides (ZENKER) VON GUTBIER
Annularia stellata (VON SCHLOTHEIM) WOOD
Asterophyllites sp.
Calamites cf. *gigas* BRONGNIART
Calamites suckowii BRONGNIART
Calamites sp.
Cordaites sp.
Lepidocarpon major (BRONGNIART) HEMINGWAY
Sigillariophyllum sp.

plueckeneti (SCHLOTHEIM). His Fig. 5 depicts a still unidentified species of *Sphenopteris* which the present writers have re-examined in Oporto. His Fig. 6 does represent *Dicksonites plueckeneti.* The latter specimen has been figured in TEIXEIRA (1942c, 1944, 1945).

Sphenophyllum sp. indet.

The specimen figured as *Sphenophyllum* cf. *cuneifolium* (STERNBERG) (TEIXEIRA 1942d, Est. IV, fig. 4) does not seem to be available any more and cannot be judged from the illustration. In the absence of any other specimen corresponding to *S. cuneifolium,* this species has been struck from the list.

Pecopteris cf. *nyranensis* NĚMEJC

The two specimens figured as *Pecopteris crenulata* BRONGNIART (TEIXEIRA 1942d, Est. III, figs 3, 4; the latter refigured in TEIXEIRA 1942c, Est. II, fig. 9, and 1944, 1945, fig. 27) unfortunately seem to have disappeared. Similar material is tentatively attributed to *Pecopteris* cf. *nyranensis* NĚMEJC. It shows simple veins.

Alethopteris corsini BUISINE

The species mentioned as *Alethopteris corsini* BUISINE has been figured and described by TEIXEIRA (1940a, 1941/42, 1942c, 1942d, 1944, 1945; TEIXEIRA & PAIS, 1976) as *Alethopteris* *lonchitifolia* P. BERTRAND. Subsequently, WAGNER (1968, p. 105-110) changed the identification to *Alethopteris missouriensis* WHITE. A re-examination of various specimens from TEIXEIRA's locality of Ervedosa has shown a mixture of simple and once forked lateral veins which apparently excludes *A. missouriensis* which has generally once, rarely twice forked nervules. The vein density also appears to be wrong. The specimens from Ervedosa show c. 25-27 veins per cm on the pinnule border, whereas *A. missouriensis* displays more numerous veinlets, 40-50 per cm. *Alethopteris corsini* BUISINE, however, shows once forked as well as simple veins with a vein density of about 30 per cm (BUISINE, 1961, p. 116).

Discussion on stratigraphic age

There are at least two species in the list which suggest the higher part of Westphalian D. These are *Callipteridium jongmansi* and *Polymorphopteris polymorpha*. The latter has been identified very tentatively and the former ranges down to a level below Tonstein 60 in Lorraine (LAVEINE *et al.* 1977), i.e. below the horizon of the mid-Westphalian D floral change as described by LAVEINE (1977). This species does indeed occur, albeit very rarely, in lower to middle Westphalian D strata of the Central Asturian Coalfield in Northwest Spain where it has been recorded under the name of *Callipteridium armasi* by WAGNER (1971).

Another species, generally regarded as commencing its occurrence in upper Westphalian D, is *Dicksonites plueckeneti*. However, although its range commences at Tonstein 60 in Lorraine (LAVEINE 1977) and at an equivalent horizon in South Wales (CLEAL 1978), this species has been recorded once at what may be a slightly lower horizon in the Central Asturian Coalfield of northern Spain (JONGMANS & WAGNER 1957).

Other elements suggest a position not quite as high as upper Westphalian D. For instance, *Alethopteris corsini* has been described from the Westphalian C and lower Westphalian D of northern France (BUISINE 1961). *Lobatopteris micromiltoni*, in conjunction with remains identified tentatively as *Lobatopteris vestita* would tend to suggest a mid-Westphalian D age corresponding to either the upper part of lower Westphalian D or the lower part of upper Westphalian D.

Linopteris obliqua characterises deposits below the level of upper Westphalian D in Spain, but occurs, abeit very rarely, in higher Westphalian and even lower Stephanian strata. However, it should be noted that this species and variety is quite common in uppermost Westphalian D strata of Saar-Lorraine (LAVEINE 1977) and Britain (CLEAL 1978).

Pecopteris unita ranges upwards from a low level in Westphalian D.

Altogether, it seems that the Ervedosa flora fits an age corresponding to a level which is possibly equivalent to upper Westphalian D but most likely the lower part of this interval.

2.4. **Douro Basin**

This refers to a narrow strip of upper Stephanian deposits immediately west of the Valongo Anticline. It extends in 90 km from S. Pedro Fins to Mioma. Coal mines at São Pedro da Cova, and Pejão have yielded a large quantity of fossil plant impressions, generally with tectonic deformation and showing the characteristic silvery sheen associated with advanced coalification. Additional remains were recovered from several outcrop localities. A large proportion of these plant impressions are poorly preserved with regard to the detail of nervation and the outline of pinnules. Some of the mines, the recently adandoned mine of São Pedro da Cova in particular, have provided large specimens allowing observations of frond structure. The quantity of material recovered has given rise to abundant records going back to SHARPE (1849), RIBEIRO (1861, 1862, 1863), GOMES (1865, 1870), GEINITZ (1867), FEISTMANTEL (1875), HEER (1881), LIMA (1888), STERZEL (1903) and ZOBEL (1910). However, the most detailed records are due to FLORIN (1938/45, 1940) and TEIXEIRA (see references).

Table 2 — Composite list of the lower Stephanian C flora of the Douro Basin (Oporto region)

Ernestiodendron filiciforme (VON SCHLOTHEIM *pars*) FLORIN
cf. *Lebachia frondosa* var. *zeilleri* FLORIN
Lebachia parvifolia FLORIN
Neuropteris cordata BRONGNIART (pl. V, fig. 2)
Neuropteris gallica ZEILLER
Neuropteris ovata var. *pseudovata* GOTHAN & SZE (pl. IV; pl. V, figs 1, 3-4
Neuropteris planchardi ZEILLER
Neuropteris zeilleri LIMA
Neuropteris sp.
Reticulopteris germari (GIEBEL) GOTHAN (pl. X, fig. 3)
Linopteris neuropteroides (VON GUTBIER) POTONIÉ
Odontopteris brardi BRONGNIART
Lescuropteris genuina (GRAND'EURY) REMY
Callipteridium (Eucallipteridium) gigas (VON GUTBIER) WEISS
Callipteridium (Eucallipteridium) zeilleri WAGNER
Alethopteris zeilleri RAGOT (pl. X, fig. 2)
Pseudomariopteris cf. *busqueti* (ZEILLER) DANZÉ-CORSIN
Pseudomariopteris corsini (TEIXEIRA) WAGNER
Pseudomariopteris ribeyroni (ZEILLER) DANZÉ-CORSIN
Dicksonites plueckeneti (VON SCHLOTHEIM) STERZEL
Eusphenopteris rotundiloba (NĚMEJC) VAN AMEROM
Sphenopteris cf. *cremeriana* POTONIÉ
Sphenopteris cf. *germanica* WEISS
Sphenopteris cf. *lenis* ZEILLER
Sphenopteris matheti ZEILLER (pl. XI, figs 2, 4)
Sphenopteris mendescorreae TEIXEIRA (pl. VI, figs 1-4)
Sphenopteris sp. cf. *chaerophylloides* BRONGNIART
Sphenopteris sp.
Oligocarpia leptophylla (BUNBURY) GRAUVOGEL-STAMM & DOUBINGER
Alloiopteris sp.
Gondomaria discreta (WEISS) WAGNER & LEMOS DE SOUSA
Nemejcopteris feminaeformis (VON SCHLOTHEIM) BARTHEL
Polymorphopteris polymorpha (BRONGNIART) WAGNER
Lobatopteris viannae (TEIXEIRA) WAGNER
Pecopteris cf. *ameromi* STOCKMANS & WILLIÈRE (pl. XI, fig. 1)
Pecopteris candolliana BRONGNIART
Pecopteris cyathea (VON SCHLOTHEIM) BRONGNIART
Pecopteris daubreei ZEILLER (pl. VII, fig. 1)
Pecopteris densifolia GÖPPERT
Pecopteris gruneri ZEILLER
Pecopteris limae TEIXEIRA (pl. VIII; pl. IX, figs 1-2)
Pecopteris cf. *melendezi* WAGNER
Pecopteris monyi ZEILLER (pl. VII, fig. 2; pl. X, fig. 4)
Pecopteris unita BRONGNIART
Pecopteris sp. nov. (cf. *hemitelioides* BRONGNIART) (pl. XI, fig. 3)
Pecopteris sp.
Rhacopteris gomesiana (HEER) TEIXEIRA
Taeniopteris bertrandiana TEIXEIRA
Taeniopteris jejunata GRAND'EURY
cf. *Taeniopteris multinervis* WEISS
Sphenophyllum alatifolium RENAULT
Sphenophyllum costae STERZEL (pl. X, fig. 1)
Sphenophyllum longifolium (GERMAR) UNGER

Table 2 *(Cont.)*

Sphenophyllum oblongifolium (GERMAR & KAULFUSS) UNGER (pl. XI, fig. 1)
Sphenophyllum cf. *thoni* var. *minor* STERZEL
Annularia sphenophylloides (ZENKER) VON GUTBIER
Annularia stellata (VON SCHLOTHEIM) WOOD
Asterophyllites equisetiformis (VON SCHLOTHEIM) BRONGNIART
Equisetites zeaformis ANDRAE
Sigillaria brardi BRONGNIART
Sigillariostrobus serreatus TEIXEIRA
Cyperites sp.
Calamites carinatus STERNBERG
Calamites schuetzeiformis forma *waldenburgensis* KIDSTON
Calamites suckowii BRONGNIART
Macrostachya carinata (GERMAR) ZEILLER
Calamostachys tuberculata STERNBERG
Cordaites sp.
Lepidophylloides sp.
Dicranophyllum gallicum GRAND'EURY
Dicranophyllum lusitanicum (HEER) LIMA (pl. XII)
Seeds

Some of the identifications can only be regarded as tentative and this state of affairs is reflected in the revised list given here (Table 2). Certain corrections on TEIXEIRA's work are discussed below:

Neuropteris ovata var. *pseudovata* GOTHAN & SZE

This name is here applied to specimens originally identified as *Neuropteris flexuosa* STERNBERG and *Neuropteris auriculata* BRONGNIART by GOMES (1865) and TEIXEIRA (1944, 1945, p. 88, fig. 40, Est. XIV, fig. 4), and which WAGNER (1959, 1963) has assigned to *Neuropteris ovata* HOFFMANN. The nervation density of these specimens allows comparison with *N. ovata* var. *grandeuryi* WAGNER which has been described from the Stephanian of Northwest Spain (WAGNER 1963). However, the veins are perhaps a little more oblique on the pinnule margin and the pinnules themelves show a more subtriangular shape than is common in the standard lateral pinnules of the var. *grandeuryi*. In these respects it would appear that a closer comparison can be made with *Neuropteris peseudovata* GOTHAN & SZE of the late Stephanian and early Permian of China (compare STOCKMANS & MATHIEU 1939, pl. IX). This species has been incorporated with *Neuropteris* ovata by GU & ZHI (1974, pl. 74 — figures after STOCKMANS & MATHIEU 1939).

The more inclined position of the nervules and the somewhat more tapering pinnule apices show a similarity with the lower Rotliegend form *Neuropteris neuropteroides* (GÖPPERT) BARTHEL (see BARTHEL 1976 and ŠETLÍK 1980). CORSIN (*in* PRUVOST & CORSIN 1949) suggested that *Neuropteris ovata* and *Neuropteris neuropteroides* would represent an evolutionary lineage. The incorporation of *Neuropteris pseudovata* in the *ovata-neuropteroides* lineage would tend to suggest an unbroken stratigraphic range extending from Westphalian D to early Permian. It is known that the top occurrence of *Neuropteris ovata* is extremely variable in different parts of Europe and elsewhere. In Portugal the var. *pseudovata* (ex *N. auriculata* GOMES and TEIXEIRA, *non* BRONGNIART) has been recorded repeatedly from the 'bacia do Douro', i. e. from strata of probable early Stephanian C age. *Neuropteris neuropteroides* (GÖPPERT) has been found in the upper Stephanian C (part of the so-called Stephanian D of BOUROZ & DOUBINGER (1977)) of Buçaco, near Coimbra.

A slightly different form of *Neuropteris ovata*, showing also rather tapering pinnules but possessing a wider venation, has been found most recently in lower Stephanian C strata of El Bierzo, in north-western León (Spain).

The different varieties of the *Neuropteris ovata* complex still have to be described in detail. Some points of synonymy have been dealt with by WAGNER (1963) for material from the Iberian Peninsula.

Reticulopteris germari (GIEBEL) GOTHAN

TEIXEIRA (1940) described a *Linopteris gomesi* TEIXEIRA which he compared with *Linopteris brongniarti* GUTBIER. This means probably *L. brongniarti* sensu ZEILLER which is equivalent to *Linopteris gangamopteroides* (DE STEFANI) (WAGNER & SOUSA 1982b). He also compared with *Linopteris germari* GIEBEL. His illustrations allow a comparaison with *Reticulopteris germari*, and WAGNER (1964a) did, in fact, refer TEIXEIRA's species to the latter. TEIXEIRA (1940b) also described a *Linopteris* cf. *germari* from what may be the same locality (although this is not entirely clear). Both identifications (without using cf.) were mentioned in a later paper (TEIXEIRA 1941b), but only *Linopteris germari* appeared in the composite list published in his thesis (TEIXEIRA 1944, 1945). This suggests that he came to the conclusion that his species *gomesi* should be regarded as a synonym of *Reticulopteris germari* (although no such explanation was given). TEIXEIRA's types have been re-examined in Oporto and do, indeed, appear marginally different to standard *Reticulopteris germari*. The comparison with *Linopteris brongniarti (gangamopteroides ?)* is irrelevant. The present writers support the synonymy with *Reticulopteris germari*.

Callipteridium (Eucallipteridium) zeilleri WAGNER

This name has been applied to material figured by TEIXEIRA (1951, Est. VI) as *Callipteridium pteridium* (VON SCHLOTHEIM). One of the specimens figured (Est. VI, figs 2, 2a) shows the elongate terminal which distinguishes this species from *Callipteridium pteridium*. Additionally, it would appear that VON SCHLOTHEIM's species from Central Europe has generally smaller pinnules than *Callipteridium zeilleri* from France, Spain and Portugal. These differences are quite minor and WENDEL (1980), who has recently studied *C. pteridium* in detail, does not regard them as sufficiently important to merit a specific distinction. It is certainly apparent that the morphological variations within these two species are partly overlapping. Perhaps the Iberian form should be regarded as a variety. However, this decision can only be taken after the morphological variation in *Callipteridium zeilleri* has been re-examined in the light of WENDEL's study.

Alethopteris zeilleri RAGOT

WAGNER (1968) has figured and described this species which TEIXEIRA (1941b, Est. IV, figs 1-2; 1941/42, Est. IV, figs 1-2; 1944, 1945, fig. 39), following ZEILLER (1888), had attributed to *Alethopteris grandini* (BRONGNIART) GÖPPERT. A recent paper by PAIS (1973) has perpetuated the misidentification of this species in Portugal.

Sphenopteris mendescorreae TEIXEIRA

This species was described together with *Sphenopteris sampaiana* TEIXEIRA from the locality of Alto da Bela (Ermesinde) (TEIXEIRA 1939). Strangely enough, the two species were not compared. From the examination of the type specimens, which are in the Museu Wenceslau de Lima (Oporto), it follows that the pinnules showing more elongate segments were distinguished as *Sphenopteris sampaiana*, whereas the specimens with relatively shorter, more spreading pinnules were attributed to *Sphenopteris mendescorreae*. It seems that this is a matter of variation within the frond of a single species. The name *mendescorreae* is preferred because TEIXEIRA (1939) figured some fertile remains under this name. In the text he referred to both female and male fructifications and he also attempted to illustrate these by diagrammatic drawings which the present writers have no hesitation in declaring both imaginative and misleading. TEIXEIRA (1944, 1945, p. 75) must have come to the same conclusion since he later referred to the fructification as *Crossotheca* (without, however, discussing it again). The fructification is figured here after a specimen not illustrated by TEIXEIRA. It shows two rows of elongate sporangia laterally fused in a synangiate structure which is attached to the tip of a linear pinnule segment. It does, indeed, bear a superficial resemblance to *Crossotheca*. Cl. BROUSMICHE (pers. comm.) has described this fructification from Saar-Lorraine under a special name.

Pecopteris candolliana BRONGNIART and *P. densifolia* GÖPPERT

One of the specimens figured by TEIXEIRA as *Pecopteris lepidorachis* BRONGNIART is here referred to *Pecopteris candolliana* (TEIXEIRA 1940b, Est. VII, fig. 1; 1944, 1945, Est. XVII, fig. 3). This specimen, which is in the collection of the Serviços Geológicos de Portugal, shows ribbon-shaped pinnules, apparently with a very slightly contracted base in the lower and middle parts of pinnae. These pinnules are closely spaced, laterally touching, inserted perpendicularly on a strong rachis, and display a regular venation pattern characterised by once bifurcate nervules; the vein bifurcation taking place at a short distance from the midvein. This is the venation pattern shown on BRONGNIART's (1834 *in* BRONGNIART 1828/1837) pl. 100. *Pecopteris lepidorachis* BRONGNIART (op. cit. 1834, pl. 103) shows both bifurcate and three-pronged veins. It is also noted that the type material of *Pecopteris candolliana* displays contracted pinnule bases. The only difference with the specimen in hand is found in the wider spacing of pinnules in the type material of *P. candolliana*. This may be a matter of habitat, however.

Most of the specimens attributed to *Pecopteris lepidorachis* by TEIXEIRA show ribbon-shaped pinnules with a regular vein pattern characterised by once bifurcate veins, with the vein bifurcation taking place at the midvein. Such specimens are here referred to as *Pecopteris densifolia* GÖPPERT.

Specimens figured and described by TEIXEIRA as *Pecopteris lepidorachis* forma *limai* TEIXEIRA are regarded as a separate species (as TEIXEIRA originally described it), viz. *Pecopteris limae* TEIXEIRA.

Pecopteris gruneri ZEILLER

This species of the *Senftenbergia plumosa-dentata* group has been called *Pecopteris bioti* BRONGNIART in all TEIXEIRA's publications on the 'Bacia do Douro'. It is here preferred to apply the name *Pecopteris gruneri* since this is the species ZEILLER (1888) described from the upper Stephanian of Commentry in south-central France. It is noted that *Pecopteris bioti* shows smaller, more rounded pinnules than are commonly encountered in Stephanian material of this group. Consideration has been given to *Senftenbergia saxonica* BARTHEL, which shows identical pinnule shapes, but ZEILLER's species clearly enjoys priority.

Pecopteris limai TEIXEIRA (recte *limae*)

TEIXEIRA (1941b, pp. 8-9, Est. I, figs 3-6) described this species from Bed 9 of the Valongo Anticline in an outcrop along the road from São Pedro da Cova to Valongo. He later (TEIXEIRA 1944, 1945, Est. XVII, fig. 4 — plate explanation) referred to one of the type specimens as *Pecopteris lepidorachis* forma *limai* TEIXEIRA. There is no doubt that this form is a species in its own right. TEIXEIRA's (1941b, Est. I, fig. 3) is designated here as the lectotype and refigured on pl. VIII. It shows the terminal part of a pinna of the penultimate order in which there is a range of variation in pinnule shape and venation. It displays gradual lobing which is accompanied by a sloping midvein and simple, rather steeply inclined laterals in poorly individualised pinnules, whereas the better developed pinnules show a straight midvein, sometimes very slightly decurrent at the extreme base, and once, more rarely twice bifurcate laterals. The vein bifurcations occur at one third to one half of the distance from the midvein to the pinnule border. A topotype (pl. IX, fig. 2) shows gradually fused terminals to the pinnae of the last order which also display the gradual passage of pinnules with bifurcate veins to pinnules with simple veins in the upper parts of pinnae. These characteristics were not brought out by TEIXEIRA's illustrations of the type specimens. One of the other type specimens figured by TEIXEIRA (1941b, Est. I, figs 4, 4a) is refigured here on pl. IX, fig. 1. It is only provisionally regarded as conspecific with the lectotype, but may represent a pinna lower down the frond, in which case the vein bundles reach a three-pronged stage. The vein density is c. 16-18 per cm on the pinnule border.

TEIXEIRA (1942b) also identified *Pecopteris limai* from Buçaco but this identification appears to have been in error. The plant from Buçaco is here regarded as a new species.

Pecopteris limae TEIXEIRA is an extremely characteristic species which will have to be redescribed in detail. It also appears to be rare. There is a marked similarity between TEIXEIRA's species and *Pecopteris ameromi* STOCKMANS & WILLIÈRE from the Stephanian B of Ciñera-Matallana (Léon, Northwest Spain) (see STOCKMANS & WILLIÈRE 1965, pl. XXIX, figs 3, 3a; WAGNER & ARTIEDA 1970, lám. XI, fig. A; lám. XII, fig. A; lám. XIX, figs A-B), but the latter shows generally smaller pinnules with a somewhat higher vein density. Also the terminals in this species are not quite as extensively fused as in *Pecopteris limae* TEIXEIRA. STOCKMANS & WILLIÈRE (1965) failed to compare with TEIXEIRA's species, partly because of TEIXEIRA's somewhat limited figuration and partly because their type material of *Pecopteris ameromi* was inadequate to show the range of variation in this species.

Pecopteris sp. nov. (cf. *hemitelioides* BRONGNIART)

From the locality of Alto da Bela (Ermesinde), TEIXEIRA (1939b) has figured a well characterised pecopterid which he has called *Pecopteris hemitelioides* BRONGNIART. It compares with the latter in that the pinnules show relatively fine, simple veins. However, typical *Pecopteris hemitelioides* has more perpendicular pinnules even in the uppermost parts of pinnae, and a well individualised, small, rounded apical pinnule, whereas the specimens from Alto da Bela show elongate terminals with a increasingly more oblique insertion of near-terminal pinnules. The length/breadth ratios of pinnules also seem somewhat different in the two species.

Sphenophyllum sp. (cf. *S. thoni* var. *minor* STERZEL)

TEIXEIRA (1946, p. 49, fig. 3) figured some very fragmentary remains of a *Sphenophyllum* with rounded distal borders as *Sphenophyllum* cf. *verticillatum* (VON SCHLOTHEIM). Any attempt at an identification can only be very tentative. The present writers prefer to compare with *S. thoni* var. *minor,* in view of the fact that some of the veins apparently abut onto the lateral borders which grade into the distal border.

Discussion on stratigraphic age

The general composition of the flora is more or less standard for the upper Stephanian (i.e. Stephanian B and C), and TEIXEIRA (1944, 1945, 1954) has indeed referred it to the upper part of the middle Stephanian (which, at that time, meant Stephanian B) and, possibly, the lower part of the upper Stephanian (meaning Stephanian C). TEIXEIRA & PAIS (1976) mentioned it as Stephanian B-C. DOUBINGER (1956, p. 73-74), when comparing the Portuguese records with similar ones in the Massif Central of south-central France, suggested that the lower Stephanian C seemed more likely than upper Stephanian B. WAGNER (1959) mentioned the flora published by TEIXEIRA as Stephanian B.

Two recent identifications, unfortunately both tentative, do indeed suggest that DOUBINGER's appraisal was the correct one and that an early Stephanian C age appears most likely. These refer to *Pseudomariopteris* cf. *busqueti* and cf. *Taeniopteris multinervis*. Both identifications are based on single specimens of rather poor preservation, but the *Pseudomariopteris busqueti* is almost good enough to be identified positively. A single specimen comparable to *Sphenopteris germanica* WEISS provides a further indication for the highest Stephanian. The presence of three different conifer species, as identified by FLORIN (1938/45, 1940), though perhaps more important with respect to habitat and palaeogeographical environment than for the stratigraphic age, also tends to favour a Stephanian C rather than Stephanian B age.

Fairly important for the determination of the stratigraphic age is the presence of a species of the *Neuropteris ovata* complex. Records from the Villablino Coalfield in León, Northwest Spain, show that *Neuropteris ovata* HOFFMANN occurs up to and including the lower Stephanian C. This species has been found most recently and in fair abundance in lower Stephanian C deposits of the El Bierzo area of north-western León. The form encountered in the Douro Basin seems to some extent transitional to *Neuropteris neuropteroides* of the highest Stephanian C and lower Autunian. It also appears to be rare in the Douro Basin. This may be another indication for a Stephanian C age.

The present writers are reasonably confident that the coal-bearing strata of the Douro Basin belong to the lower Stephanian C (and perhaps not even to the very basal part of this stage).

In regard to a comparison with lower Stephanian C deposits in the Cantabrian Zone of Northwest Spain, occurring at El Bierzo, Villablino (upper part of the succession in this coalfield), Rengos, Carballo, Tineo, Tormaleo, etc., it is noted that the Douro Basin contains varied conifer remains (three species) and *Dicranophyllum* (two species). These are special elements which are practically absent from strata of similar age in Northwest Spain, the exceptions being a single illustrated record of *Dicranophyllum gallicum* by STOCKMANS & WILLIÈRE (1965) and the mention of *Walchia* by ZEILLER (1882). These are extrabasinal elements (sensu HAVLENA 1970, PFEFFERKORN 1980).

3. BUÇACO BASIN

This is an elongate basin, aligned north-south along the Oporto - Coimbra - Badajoz - Córdoba Fault, one of the most important long-lasting fractures in the Iberian Peninsula. Its stratigraphy has been studied most recently by the present writers, who distinguish between a lower red beds succession, a short interval of grey beds, and a higher, very substantial succession consisting mainly of fluviatile conglomerates. The plant fossils from Buçaco are almost invariably from the 42 m thick interval of grey beds which alternate with unfossiliferous red beds in the lower part of this interval. It is clear that the floras recorded from Buçaco by LIMA (1888/92, 1891, 1894) and TEIXEIRA (1941a, 1942b, 1944, 1945) as well as by COURBOULEIX (1972) all correspond to a single, rather short time span.

The sedimentary characteristics of the Buçaco Basin show it to be intramontane, and the floral assemblages found in this basin accord well with this contention since they include quite a few extrabasinal elements (WAGNER *et al.* in prep). Table 3 shows the sum total of species recognised at present from Buçaco, after a revision of floras in the Serviços Geológicos de Portugal and in the Museu Wenceslau de Lima in Oporto, and the identification of plants found in different localities along a measured stratigraphic section. A preliminary list of species corresponding to the present revision has been published by the present writers in WAGNER & MARTINEZ (1982).

Comments on the revised identifications are as follows:

Table 3 — Composite list of the upper Stephanian C (or lowermost Autunian) flora of the Buçaco Basin

Lebachia goeppertiana FLORIN
Lebachia laxifolia FLORIN
Lebachia parvifolia FLORIN
Callipteris conferta (STERNBERG) BRONGNIART
Neuropteris neuropteroides (GÖPPERT) BARTHEL (pl. XIII, figs 1-2)
Neuropteris planchardi ZEILLER
Neuropteris praedentata GOTHAN
Neuropteris zeilleri LIMA (pl. XIII, fig. 3; pl. XIV)
Neuropteris sp.
Reticulopteris germari (GIEBEL)
Linopteris gangamopteroides (DE STEFANI) WAGNER (pl. XVIII, fig. 2)
Odontopteris brardi BRONGNIART (pl. XIII, fig. 4)
Mixoneura sp. cf. *Odontopteris osmundaeformis* (VON SCHLOTHEIM) ZEILLER (pl. XVI, figs 1-3)
Lescuropteris genuina (GRAND'EURY) REMY
Callipteridium densinervium WAGNER
Callipteridium gigas (VON GUTBIER) WEISS
Alethopteris schneideri STERZEL
Alethopteris zeilleri RAGOT
Pseudomariopteris busqueti (ZEILLER) DANZÉ-CORSIN
Pseudomariopteris ribeyroni (ZEILLER) DANZÉ-CORSIN
Dicksonites leptophylla (ZEILLER) DOUBINGER (pl. XV, figs 3-4)
Dicksonites plueckeneti (VON SCHLOTHEIM) STERZEL
Eusphenopteris cf. *rotundiloba* (NĚMEJC) VAN AMEROM
Sphenopteris cf. *biturica* ZEILLER
Sphenopteris casteli ZEILLER (pl. XVIII, fig. 3)
Sphenopteris cremeriana POTONIÉ
Sphenopteris elaverica (ZEILLER) ALVAREZ-RAMIS
Sphenopteris cf. *minutisecta* FONTAINE & WHITE, non ALVAREZ-RAMIS
Sphenopteris sp.
Oligocarpia leptophylla (BUNBURY) GRAUVOGEL-STAMM & DOUBINGER (pl. XVIII, fig. 1)
Alloiopteris sp.
Nemejcopteris feminaeformis (VON SCHLOTHEIM) BARTHEL
Polymorphopteris polymorpha (BRONGNIART) WAGNER
Lobatopteris corsini WAGNER
Lobatopteris viannae (TEIXEIRA) WAGNER
Pecopteris arborescens (VON SCHLOTHEIM) BRONGNIART
Pecopteris (Oligocarpia?) cf. *bredovii* GERMAR
Pecopteris bussacensis TEIXEIRA
Pecopteris cf. *candolliana* BRONGNIART
Pecopteris cyathea (VON SCHLOTHEIM) BRONGNIART (pl. XV, figs 1-2)
Pecopteris densifolia GÖPPERT
Pecopteris gruneri ZEILLER
Pecopteris monyi ZEILLER
Pecopteris unita BRONGNIART
Pecopteris sp. nov. (pl. XVII, figs 1-2)
Pecopteris sp.
Taeniopteris jejunata GRAND'EURY
Taeniopteris multinervis WEISS
Sphenophyllum angustifolium (GERMAR) UNGER
Sphenophyllum costae STERZEL
Sphenophyllum oblongifolium (GERMAR & KAULFUSS) UNGER
Sphenophyllum thoni VON MAHR (pl. XVI, fig. 4)

Table 3 *(Cont.)*

Annularia sphenophylloides (ZENKER) VON GUTBIER
Annularia stellata (VON SCHLOTHEIM) WOOD
Asterophyllites equisetiformis (VON SCHLOTHEIM) BRONGNIART
Asterophyllites longifolius STERNBERG
Asolanus camptotaenia WOOD
Calamites suckowii BRONGNIART
Calamostachys tuberculata STERNBERG
Cordaites sp.
Dicranophyllum sp.
Cordaianthus sp.
Hexagonocarpus sp. and other seeds

Neuropteris praedentata GOTHAN

This species replaces the record of *Neuropteris crenulata* BRONGNIART published by TEIXEIRA (1942b, Est. VIII, figs 6, 6a). Only a single pinnule was illustrated but this is reasonably convincing. The species appears very rarely at Buçaco. The specific name refers to GOTHAN's (1909) redescription of the material misidentified by ZEILLER *in* RENAULT & ZEILLER (1888) as *Neuropteris crenulata* BRONGNIART.

Mixoneura sp. cf. *Odontopteris osmundaeformis* (VON SCHLOTHEIM) ZEILLER

A single locality in Buçaco, viz. Fonte do Salgueiro, has yielded a large number of fragments ranging in pinnule morphology from *Neuropteris* to *Odontopteris*. There is no doubt that these represent a single species in the group of *Mixoneura* WEISS. Most of the specimens are odontopteroid and show a marked resemblance to *Odontopteris osmundaeformis* (VON SCHLOTHEIM). The size and shape of the odontopteroid pinnules is identical with those of the latter, and some of the pinnae show an elongate terminal which is also very similar to that shown on the illustrations of SCHLOTHEIM's species from its type locality in Thuringia, East Germany (e.g. REMY & REMY 1977, p. 297, Bild 170). However, the large neuropteroid pinnules found at Fonte do Salgueiro do not seem to have been described for *Odontopteris osmundaeformis*. In this respect, the species from Buçaco shows a similarity to *Odontopteris pseudoschlotheimi* DE MAISTRE from St. Etienne (DE MAISTRE 1960), but there is no specific identity, the latter showing more rounded pinna terminals and a denser nervation. Comparison may also be made with *Odontopteris duponti* ZEILLER, from Commentry in south-central France, but this little known species does not apparently display the large neuropteroid pinnules that have been found in Buçaco.

The closest comparison remains with VON SCHLOTHEIM's species, and it is as *Odontopteris osmundaeformis* that TEIXEIRA (1941a) has figured a single specimen from Fonte do Salgueiro. This specimen, which is in the Serviços Geológicos de Portugal, has also been prepared for illustration by LIMA (TEIXEIRA 1941e, Est. VI). It shows the odontopteroid pinnules which resemble those of *O. osmundaeformis* very closely, but fails to show the pinna terminals or completely fused, neuropteroid pinnules.

Callipteridium densinervium WAGNER

TEIXEIRA (1942b, Est. III, figs 1-2, 1a-2a; 1944, 1945, p. 100, fig. 51) figured and described this species as *Callipteridium regina* ROEMER. It is clear that he compared with ZEILLER's (1890) illustration of this species and not with ROEMER's original figuration. WAGNER (1965) has placed both ZEILLER's and TEIXEIRA's records in the synonymy of *Callipteridium densinervium* as described from the upper Stephanian of Peña Cildá in Palencia, Northwest Spain.

Oligocarpia leptophylla (BUNBURY) GRAUVOGEL-STAMM & DOUBINGER

This species was first described from Buçaco by BUNBURY, *in* RIBEIRO (1853). It was later misidentified in France by ZEILLER (1892) who applied BUNBURY's name to a species of *Dicksonites* which also happens to be commonly present in the Buçaco Basin. Although BUNBURY's figure of the type specimen is diagrammatic, it is clearly recognisable as the form misidentified as *Sphenopteris cristata* BRONGNIART by ZEILLER (1888) *in* RENAULT & ZEILLER's memoir on Commentry in south-central France, and which has later been named *Ovopteris pecopteroides* by LANDESKROENER (*in* POTONIÉ 1906). The species was subsequently identified from Buçaco as *Ovopteris pecopteroides* or *Sphenopteris pecopteroides* by TEIXEIRA (1941a, p. 12-13, Est. II, fig. 1, Est. III, fig. 4; 1942b, Est. II, fig. 6; 1944, 1945, fig. 49), who followed ZEILLER in applying the name '*leptophylla*' to the species of *Dicksonites* recorded from France. The error was spotted by WAGNER (1959, p. 414), who subsequently figured the correct species as *Pecopteris leptophylla* BUNBURY from Northwest Spain (WAGNER 1962). ALVAREZ-RAMIS (1965) transferred this species to *Sphenopteris*, and GRAUVOGEL-STAMM & DOUBINGER (1975) described its fructification, thus proving its affinity to lie with *Oligocarpia*.

The name *Dicksonites leptophylla* (ZEILLER) DOUBINGER has been retained for specimens identical to the remains misidentified as *Pecopteris leptophylla* by ZEILLER (1892). Since DOUBINGER (1956) transferred ZEILLER's species to *Dicksonites*, there is no longer a case of homonymy with *Oligocarpia leptophylla* (BUNBURY).

Pecopteris cyathea (VON SCHLOTHEIM) BRONGNIART

The full range of variation in this species, as figured most recently from its type area by BARTHEL (1980), has been found in the roof of a thin coal which crops out in the metalled road above Algeriz. This may be the same locality as mentioned by TEIXEIRA (1942b, p. 13) and is certainly that described by COURBOULEIX (1972, 1974). A fully characteristic specimen from this locality is figured here on pl. XV, fig. 1, together with the holotype of *Pecopteris hemitelioides* forma *longipinnata* TEIXEIRA (1941a, Est. IX, figs 1-2). The latter represents the lower part of a major pinna of *Pecopteris cyathea*, and shows very elongate pinnules which are very slightly lobate in places. *Pecopteris cyathea* is a common species in the Buçaco Basin.

Pecopteris sp. nov.

This is an extremely variable species which is of common occurrence in the Buçaco Basin. It shows rather strongly vaulted, ribbon-shaped pinnules with a rounded apex; a well developed, straight midvein and generally either simple or once forked laterals, or a mixture of both. Lobing pinnules are uncommon, but when they occur they show the rather flat lobes which are also characteristic of *Pecopteris daubreei* ZEILLER. This is not the place to introduce this species which will be described elsewhere and figured rather more fully so as to illustrate the wide variation observed. It is commented upon here merely as an aid to the understanding of certain records in the literature.

TEIXEIRA (1942b, Est. IV, figs 1-2, 2a), figured two fragments showing a bifurcate vein pattern under the name of *Pecopteris limai* TEIXEIRA. These specimens are clearly different to *Pecopteris limae* as known from the type locality in the Douro Basin, which shows larger pinnules with a wider venation. The pinna terminals are also quite different. Later, TEIXEIRA (1944, 1945, fig. 46-1) refigured one of these specimens as *Pecopteris lepidorachis* forma *limai* TEIXEIRA.

In the list of species from Buçaco, TEIXEIRA (1944, 1945, p. 91) only mentioned *Pecopteris lepidorachis* BRONGNIART.

It seems likely that the specimens illustrated by TEIXEIRA (1941a, Est. VII, figs 1-3) under the name of *Pecopteris daubreei* ZEILLER also belong to the new species discussed here. However, the illustrations show too little of the venation to be able to judge these specimens adequately. They have not been seen in the collections re-examined by the present writers.

Discussion on stratigraphic age

Dating of the Buçaco Basin has varied from Autunian (TEIXEIRA 1944, 1945, 1954; DOUBINGER 1956, p. 74) to Stephanian D (TEIXEIRA & PAIS 1976). After collecting a flora of Stephanian aspect near Algeriz and Vale da Mó, (COURBOULEIX 1972, 1974), postulated that both Stephanian and Autunian would be present in different parts of the basin. However, the stratigraphic work by the present writers indicates that the plant-bearing interval is always the same. COURBOULEIX's conclusion should therefore be modified in the sense of GOTHAN & GIMM's (1930) recognition of coal-forming and allochthonous plant assemblages. COURBOULEIX's flora, identified by PAULE CORSIN, occurs in the roof of a coal seam and shows a coal-forming assemblage similar to those known from the upper Stephanian coal-measures in France. It is not necessarily older than the assemblages recorded by GOMES (1865); LIMA (1888/92, 1891, 1894) and TEIXEIRA (1941a, 1942b, 1944, 1945) which contain elements traditionally regarded as Autunian.

The general aspect of the flora is lower Rotliegend which could mean either late Stephanian or early Autunian. Several Stephanian pecopterids and sphenopterids are present as floodplain assemblages, most notably in association with the single coal horizon found in the Buçaco Basin (e.g. the Algeriz locality of COURBOULEIX 1972, p. 44). Another locality in the Algeriz road section has yielded *Neuropteris neuropteroides* (GÖPPERT) and *Alethopteris schneideri* STERZEL which are typical Rotliegend plants and which may be regarded as extrabasinal elements in the sense of HAVLENA (1970) and PFEFFERKORN (1980).

The presence of *Pseudomariopteris busqueti* (ZEILLER), *Sphenophyllum thoni* VON MAHR, *Sphenophyllum angustifolium* GERMAR, *Taeniopteris multinervis* WEISS and also *Dicksonites leptophylla* (ZEILLER), clearly shows that a level at least as high as Stephanian C is represented. The abundant presence of *Pecopteris cyathea* (VON SCHLOTHEIM) is in agreement with this conclusion.

There is not a single species at Buçaco which can be regarded as restricted to Stephanian C. The floodplain assemblages found all occur in both upper Stephanian and lower Autunian. Among the extrabasinal elements, which are well represented in the Buçaco Basin, the three conifer species identified by FLORIN (1938/45, 1940), are all late Stephanian and lower Autunian taxa. Most significant, perhaps, is the presence of *Callipteris conferta* (STERNBERG) and the absence of other callipterids. This would tend to indicate the 'Stephanian D' level as described by BOUROZ & DOUBINGER (1977) and which is a floral zone straddling the upper Stephanian C and basal Autunian (compare WAGNER, in press b). It follows that both the latest Stephanian (late Stephanian C) and earliest Autunian are possible ages for the plant-bearing grey beds of the Buçaco Basin.

In comparison with the Douro Basin, which is coal-bearing and which has a correspondingly smaller proportion of extrabasinal floral elements (i.e. plants living on the higher ground), it is noted that the floodplain assemblages are virtually identical in the two basins. It may thus be assumed that the age difference is very small. However, there are some elements that are apparently restricted to either one or the other, and the general assumption is that the Douro Basin is slightly older. The latter has been assigned an early (?) Stephanian C age, whereas the Buçaco Basin is apparently at least of late Stephanian C age and possibly a little younger, i. e. early Autunian in the traditional sense. A late Stephanian C age appears most likely to the present writers.

4. SANTA SUSANA BASIN

This basin, consisting of three separate outliers, is aligned along the fault which separates the Ossa-Morena and South Portuguese zones, i.e. two of the most important palaeogeographic and structural areas in the Palaeozoic of the Iberian Peninsula. The two main outliers, at Jongeis and Vale do Burro, have both yielded floral remains as reported by LIMA (1895/98) and TEIXEIRA (1938/40, 1940c, 1944, 1945). The corresponding specimens, in so far as they are still available in the collections of the Serviços Geológicos de Portugal and at the Faculdade de Ciências do Porto (Museu Wenceslau de Lima), have been re-examined by the present writers. An additional collection was made available by F. GONÇALVES from the locality of Monte da Casa Branca. A composite list of species with partly revised identifications is given in Table 4.

Comments on the revised identifications are as follows:

Neuropteris flexuosa STERNBERG

A comparison between material from Somerset, England, the type area of this species, and the published figures and additional specimens of *Neuropteris machadicostae* TEIXEIRA, has shown that the latter is to be referred to *N. flexuosa*. TEIXEIRA (1938/40, p. 14) compared with members of the *Neuropteris ovata* group and also with *Neuropteris macrophylla* BRONGNIART from Somerset, but failed to identify with *Neuropteris flexuosa* STERNBERG, probably because, at that time, this species was confused with a late form of *Neuropteris ovata* which had been misidentified as *Neuropteris flexuosa* by GRAND'EURY (1890).

Linopteris palentina WAGNER

Specimens attributed to this species are rather small detached pinnules identified originally by TEIXEIRA (1940c) as *Linopteris* aff. *subbrongniarti* GRAND'EURY, and afterwards (TEIXEIRA 1944, 1945, 1951) as *Linopteris obliqua* (BUNBURY). They show subparallel margins, a rounded apex, and completely anastomosed veins displaying relatively short but elongate meshes. Similar material was originally identified from uppermost Westphalian D and lower Cantabrian strata in Northwest Spain as *Linopteris brongniarti* (VON GUTBIER) and subsequently as *Linopteris* cf. *subbrongniarti* GRAND'EURY (WAGNER in KANIS 1956, 1960). These remains were later described as a new species, viz. *Linopteris palentina* and *Linopteris linearis* (or *L. neuropteroides* var. *linearis*) (WAGNER 1964b, 1965). This material has since proved to be quite variable and although a full description is still outstanding, it seems likely that a single taxon is involved for which the name *Linopteris palentina* is preferred.

The specimens from Santa Suzana mainly represent the smaller kind of pinnule which was originally associated with the name *linearis*, and the vein meshes are accordingly wider than is common in the larger pinnules of *Linopteris palentina*. Among the various specimens examined in the collections of the Serviços Geológicos de Portugal and the Faculdade de Ciências do Porto (Museu Wenceslau de Lima) there is not a single one which shows the more isodiametric vein meshes found in *Linopteris obliqua* (compare pl. I, fig. 3 and pl. II, fig. 1).

Alethopteris lesquereuxi WAGNER

TEIXEIRA (1938/40, Est. VIII, fig. 5; 1944, 1945, Est. II, fig. 1) has figured two specimens from Vale do Burro under the name of *Alethopteris davreuxi* (BRONGNIART). These specimens

Table 4 — Composite list of the uppermost Westphalian D flora of the Santa Susana Basin

Neuropteris flexuosa STERNBERG
Neuropteris scheuchzeri HOFFMANN
Linopteris palentina WAGNER (pl. III, fig. 9)
Callipteridium (Praecallipteridium) jongmansi (P. BERTRAND) WAGNER
Alethopteris lesquereuxi WAGNER (pl. II, fig. 6)
Dicksonites plueckeneti (VON SCHLOTHEIM) STERZEL
Mariopteris rotundata HUTH
Eusphenopteris nummularia (VON GUTBIER) NOVIK
Eusphenopteris trigonophylla (BEHREND) VAN AMEROM (pl. III, fig. 10)
Sphenopteris alentejana TEIXEIRA
Sphenopteris (Palmatopteris ?) spinosa GÖPPERT
Sphenopteris cf. *pecopteroides* KIDSTON
Sphenopteris sewardi KIDSTON (pl. III, fig. 8)
Sphenopteris sp. nov. ? (cf. *douvillei* ZEILLER)
cf. *Alloiopteris* sp.
Lobatopteris cf. *lamuriana* (HEER) WAGNER
Lobatopteris vestita (LESQUEREUX) WAGNER
cf. *Pecopteris avoldensis* (STUR) CORSIN
Pecopteris daubreei KIDSTON, *non* ZEILLER (pl. II, fig. 2)
Pecopteris dentata BRONGNIART
Pecopteris haussei STERZEL
Pecopteris (Lobatopteris) cf. *camertonensis* KIDSTON
Pecopteris monyi ZEILLER (pl. III, fig. 2)
Pecopteris nyranensis NĚMEJC (pl. III, fig. 7)
Pecopteris cf. *obliquenervis* CORSIN
Pecopteris plumosa (ARTIS) BRONGNIART
Pecopteris raconensis NĚMEJC (pl. II, figs 3-5)
Pecopteris cf. *saraefolia* P. BERTRAND
Pecopteris unita BRONGNIART
Pecopteris sp.
Sphenophyllum emarginatum BRONGNIART
Sphenophyllum emarginatum forma *truncatum* SCHIMPER
Sphenophyllum guerreiroi TEIXEIRA (pl. III, figs 3-6)
Annularia sphenophylloides (ZENKER) VON GUTBIER
Annularia stellata (VON SCHLOTHEIM) WOOD
Asterophyllites equisetiformis (VON SCHLOTHEIM) BRONGNIART
Asterophyllites longifolius STERNBERG
Macrostachya carinata GERMAR (pl. III, fig. 1)
Calamites carinatus STERNBERG
Calamites suckowi BRONGNIART
Lepidodendron cf. *aculeatum* STERNBERG
Lepidophloios ? sp. nov. ?
Lepidostrobophyllum hastatum (LESQUEREUX) CHALONER
Lepidocarpon major (BRONGNIART) HEMINGWAY
Lycopodites sp.
Sigillaria cf. *tessellata* BRONGNIART
Cordaites sp.

do not show the widely spaced, irregular nervules of this species and clearly belong to the *Alethopteris grandini — Alethopteris lesquereuxi* complex. Since the pinnules of these fragments are not barrel-shaped but relatively straight-sided, the second of these species is preferred.

Mariopteris rotundata HUTH

This refers to a specimen figured by TEIXEIRA (1933c) as *Mariopteris nervosa* (BRONGNIART) and which he later refigured as *Mariopteris* aff. *nervosa* (TEIXEIRA 1951). This specimen has not been found again but the plant collection of the Serviços Geológicos de Portugal has one, possibly two fragments of the same plant which is compared with *Mariopteris* cf. *rotundata* HUTH as referred to in WAGNER *et al.* (1969, p. 119). These remains as well as the more fragmentary specimens from Santa Susana show a good resemblance to the types from the upper Westphalian of Saarland in Germany, and also to those of *Mariopteris witieri* CORSIN, from the same basin. The latter would appear to be a synonym of HUTH's species. Two specimens figured by CORSIN (1932) under the name of *Mariopteris rotundata* HUTH are not wholly convincing. BOERSMA (1972) has transferred this species to *Karinopteris* but failed to provide reasons for removing it from *Mariopteris*. Since its frond structure is apparently unknown, and in view of its vague resemblance to *Mariopteris nervosa*, there seems to be no good reason for accepting the transfer to *Karinopteris*.

Eusphenopteris nummularia (VON GUTBIER) NOVIK

TEIXEIRA (1938/40, Est. V, figs 3, 4; Est. VI, figs 1-3) has figured this species under the name of *Sphenopteris obtusiloba* BRONGNIART. Several of the figured specimen have been seen by the present writers. ALVAREZ-RAMIS (1965, p. 43) tentatively referred this material to *Sphenopteris rotundiloba* NĚMEJC. This is a Stephanian species which is very similar to *Eusphenopteris nummularia*, and which possibly grades into the latter (compare VAN AMEROM 1975, p. 83).

Sphenopteris sp. nov.? (cf. *douvillei* ZEILLER)

TEIXEIRA (1938/40, Est. V, figs 1-2) has figured and described a well preserved *Sphenopteris* from Santa Susana under the name of *Sphenopteris* aff. *chaerophylloides* BRONGNIART. The present writers prefer to compare with *Sphenopteris douvillei* ZEILLER and also with *Sphenopteris hadrophylla* KNIGHT, but agree that no exact identification with a published species appears possible.

Pecopteris daubreei KIDSTON, *non* ZEILLER

TEIXEIRA (1938/40, Est. VII, figs 4-6) has figured some apparently well preserved remains as *Pecopteris daubreei* ZEILLER and, subsequently (TEIXEIRA 1940c, fig. 4), as *Pecopteris* cf. *daubreei* ZEILLER. Although the actual specimens have not been recovered, the illustration in TEIXEIRA (1940c) is accompanied by a nervation diagram which show the nervation pattern. This conforms to material collected most recently by F. GONÇALVES from exposures in the reservoir covering Vale do Burro, and which is comparable to *Asterotheca daubreei* as recorded by KIDSTON (1925) from the upper Westphalian D of Radstock, Somerset, England. Although similar to *Pecopteris daubreei* ZEILLER, it is not identical to ZEILLER's species and should be regarded as a different taxon, still to be described.

Pecopteris nyranensis NĚMEJC

This name is applied here to the material identified as *Pecopteris cyathea* (VON SCHLOTHEIM) by TEIXEIRA (1938/40, Est. V, figs 6-7). The figured specimens have not been recovered but a similar fragment has been found in the collection at Oporto (Museu Wenceslau de Lima). This conforms to *Asterotheca hemitelioides* as figured by KIDSTON (1924) and which, together with *Asterotheca cyathea* sensu KIDSTON (1924), has been referred to *Pecopteris nyranensis* by NĚMEJC (1940).

Pecopteris raconensis NĚMEJC

This is the species introduced by NĚMEJC (1940) for *Asterotheca (Pecopteris) lepidorachis*

BRONGNIART as figured by KIDSTON (1925). Material attributable to this species is in the Serviços Geológicos de Portugal as well as in the Museu Wenceslau de Lima in Oporto. It has apparently gone unrecorded by TEIXEIRA (1938/40), although it is possible that it was incorporated with the *Pecopteris daubreei* ZEILLER as figured by TEIXEIRA (1938/40, 1940c) and which refers to *Pecopteris daubreei* KIDSTON, *non* ZEILLER.

Lepidophloios ? sp. nov.?

The lycophyte figured by TEIXEIRA (1937a, 1938/40, 1944, 1945) as *Lepidodendron dichotomum* STERNBERG is characterised by lepidodendroid leaf cushions and scars with leaf trace and parichnos, but no infrafoliar parichnos and no apparent ligule. NĚMEJC (1946) has pointed out that *Lepidodendron dichotomum* as figured by STERNBERG belongs to two different types, one of which apparently represents young shoots of *Lepidodendron obovatum* STERNBERG whilst the other one comes close to *Lepidophloios* and is to be identified with *L. longifolium* PRESL. He therefore recommends that the species name *'dichotomum'* be abandoned. The Portuguese material, which is reasonably well preserved, probably should be described as a new species. It is still an open question whether it may be described under *Lepidophloios*.

Discussion on stratigraphic age

The composition of the fossil flora of Santa Susana points at an age which is either very late Westphalian D or earliest Cantabrian. Comparison should be made with the late Westphalian D floras of South Wales and the Bristol-Somerset district of England on the one hand, and the late Westphalian D/early Cantabrian floras of the post-Leonian basin in Northwest Spain on the other. *Neuropteris flexuosa, N. scheuchzeri, Alethopteris lesquereuxi, Dicksonites plueckeneti, Lobatopteris vestita, Pecopteris nyranensis, P. raconensis* and *P. daubreei* (sensu KIDSTON) are all common elements of the British area (compare KIDSTON 1924, 1925; CROOKALL 1959; CLEAL 1978). With the exception of *Neuropteris flexuosa* and *Pecopteris daubreei* (sensu KIDSTON), these species are also of common occurrence in the upper Westphalian D and lower Cantabrian of Northwest Spain (compare, particularly, WAGNER in press a). *Linopteris palentina* is another common species for the stated interval in the Spanish area. *Callipteridium jongmansi*, which is very rare in the British area, occurs commonly in Northwest Spain.

Particularly important is the presence of *Lobatopteris vestita* in association with specimens tending towards *Lobatopteris lamuriana* since the transitional forms between these two species apparently characterise the lower Cantabrian in Spain. *Pecopteris monyi* is a species which commences in the highest Westphalian D but which is more common in Stephanian strata. *Sphenophyllum emarginatum* forma *truncatum* occurs preferentially in upper Westphalian D but continues into the lower Cantabrian (and beyond, albeit very rarely). *Pecopteris haussei* is a rare species found in Stephanian and lower Permian strata, and which is closely similar to *Pecopteris monyi*.

There are some Westphalian elements in the Santa Susana flora that have also been recorded in the lower Cantabrian of Northwest Spain. These are *Mariopteris rotundata* and *Eusphenopteris nummularia*, both elements found in the lower Cantabrian flora of Tejerina (WAGNER *et al.* 1969). *Eusphenopteris trigonophylla* is a Westphalian species that has not yet been found in the Cantabrian of Spain. However, CLEAL (1978) shows its range as going right up into what he regards as basal Cantabrian in Britain.

Although there are no species in the flora from Santa Susana that would be restricted to the early Cantabrian, it seems unwise to regard this age as impossible. Further collecting will be necessary to either eliminate or prove this possibility. However, for the time being, the age of the Santa Susana flora is regarded as probably very late Westphalian D.

TEIXEIRA (1944, 1945, 1954) regarded the flora of Santa Susana as being of the same age as that of Ervedosa in the Oporto region. There are certain differences however. The *Alethopteris corsini* identified from Ervedosa is at the top of its range in the upper Westphalian D, whereas *Alethopteris lesquereuxi* (as found at Santa Susana) ranges well into the lower Cantabrian. *Linopteris palentina* is found commonly at a higher horizon in Northwest Spain than *Linopteris obliqua* (although both ranges overlap). Also, the pecopterids at Santa Susana seem more varied than those at Ervedosa. One may ascribe these differences to some extent to limited collecting, particularly at Ervedosa. However, they do create the impression that the Santa Susana flora may be a little later in age than that of Ervedosa.

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PLATES

The specimens illustrated are in the collections of the different institutions as follows:

SGP — Serviços Geológicos de Portugal, Lisboa
FCUL — Faculdade de Ciências, Universidade de Lisboa
FCUP — Faculdade de Ciências, Universidade do Porto (Museu Wenceslau de Lima)

Plate I

Ervedosa (Oporto region)
upper Westphalian D

1. *Alethopteris corsini* BUISINE (x1); la (x3).
   Montalto — Coll. «Bernardino António Gomes» cat. n.º 2 (5) 7 (FCUL).

2. *Alethopteris corsini* BUISINE (x1).
   Montalto — Coll. SGP.

3. *Linopteris obliqua* (BUNBURY) ZEILLER (x3).
   Montalto — Coll. SGP.
   Specimen figured by TEIXEIRA (1944, 1945, Est. V, 4; 1951, Est. XV, 1, 1a).

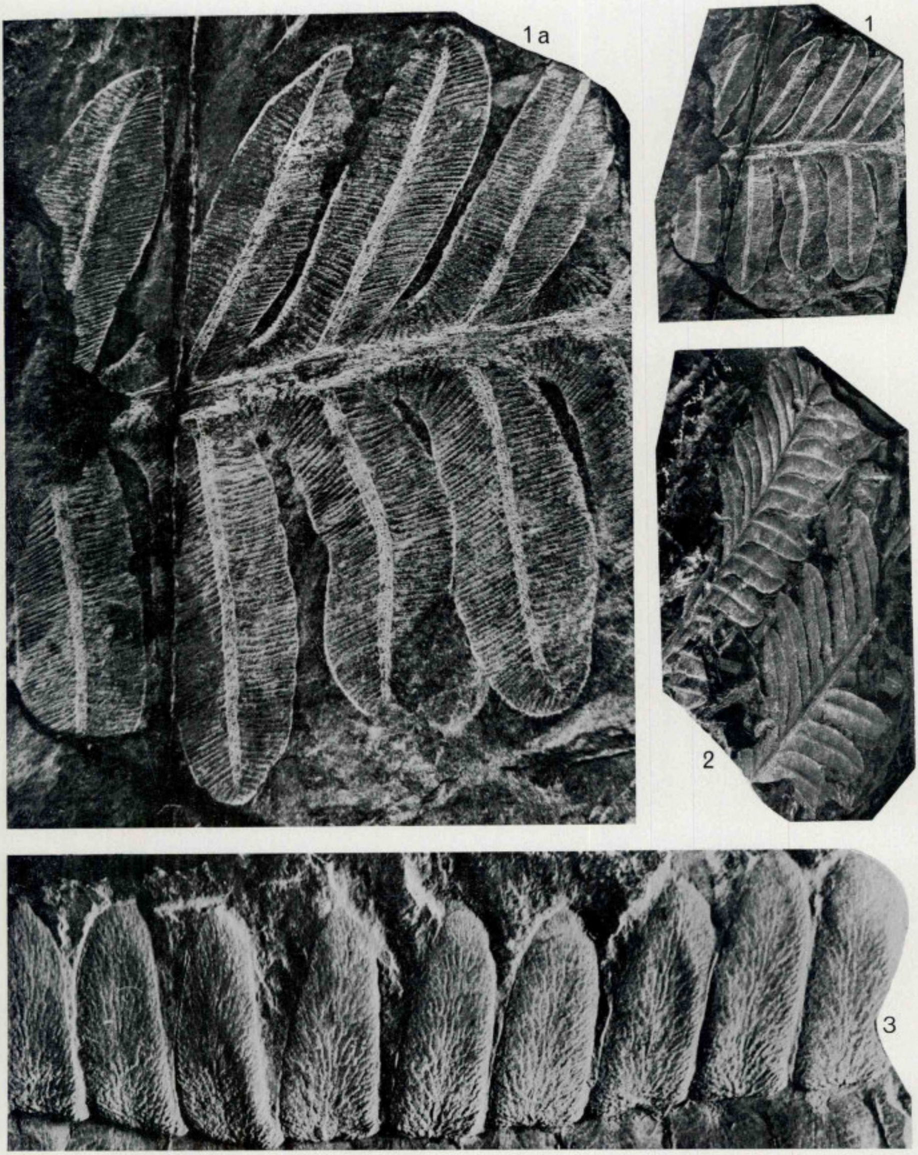
1a
1
2
3

Plate II

Ervedosa (Oporto region)
upper Westphalian D

1. *Linopteris obliqua* (BUNBURY) ZEILLER (x1).
Same as figured on pl. I, fig. 3.

Santa Susana Basin
uppermost Westphalian D

2. *Pecopteris daubreei sensu* KIDSTON, non ZEILLER (x3).
1600 m south of Monte da Casa Branca — Coll. «F. Gonçalves» (SGP).

3-5. *Pecopteris raconensis* NĚMEJC (x3).
   3. Jongeis — Coll. FCUP.
   4. 1600 m south of Monte da Casa Branca — Coll. «F. Gonçalves» (SGP).
   5. Jongeis — Coll. SGP.

6. *Alethopteris lesquereuxi* WAGNER (x3).
1250 m S 20°W of Moinho da Casa Branca — Coll. SGP.

Plate II

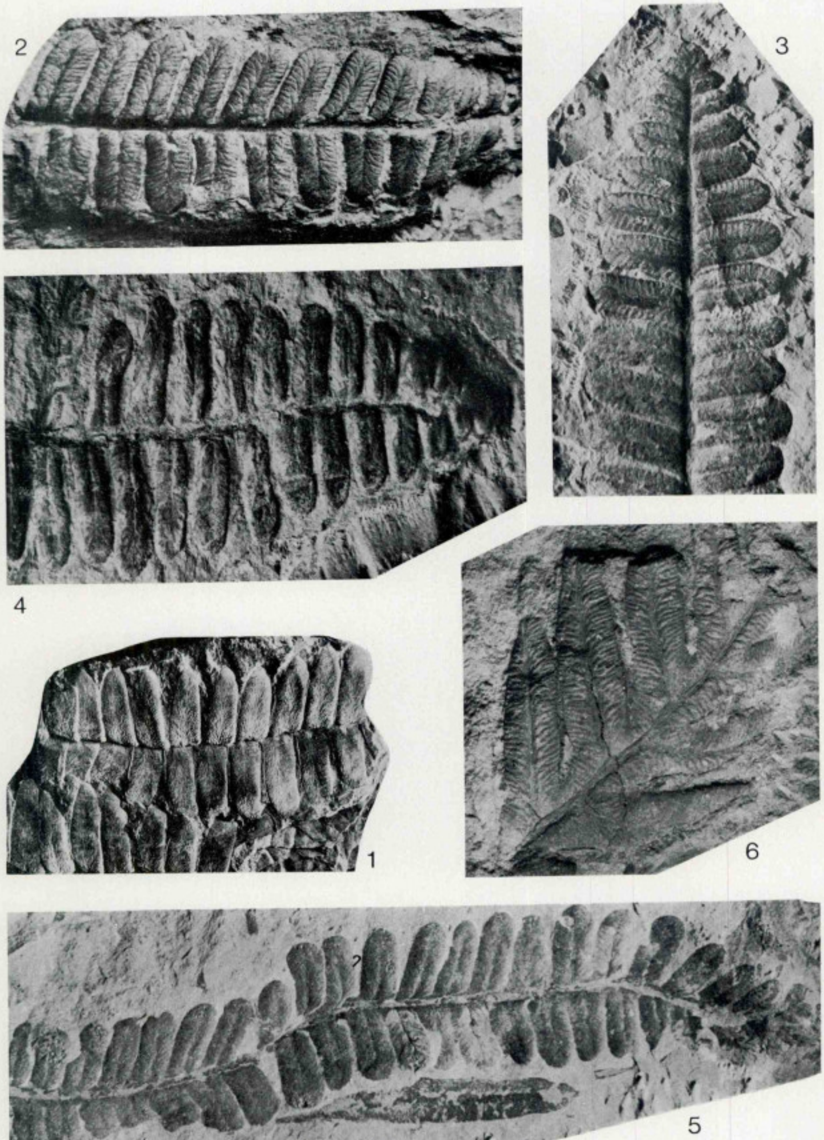

Plate III

Santa Susana Basin
uppermost Westphalian D

1. *Macrostachya carinata* GERMAR (x1).
Vale de Figueira de Baixo — Coll. SGP.

2. *Pecopteris monyi* ZEILLER (x3).
Exact locality unknown — Coll. SGP.

3-6. *Sphenophyllum guerreiroi* TEIXEIRA (x3).
Exact locality unknown — Coll. SGP.

7. *Pecopteris nyranensis* NĚMEJC (x3).
Vale de Figueira de Baixo, São Cristovão — Coll. FCUP.

8. *Sphenopteris sewardi* KIDSTON (x3).
1600 m south of Monte da Casa Branca — Coll. «F. Gonçalves» (SGP).

9. *Linopteris palentina* WAGNER (x3).
200 m N 58°E of Monte da Casa Branca — Coll. SGP.

10. *Eusphenopteris trigonophylla* (BEHREND) VAN AMEROM (x3).
Exact locality unknown — Coll. SGP.

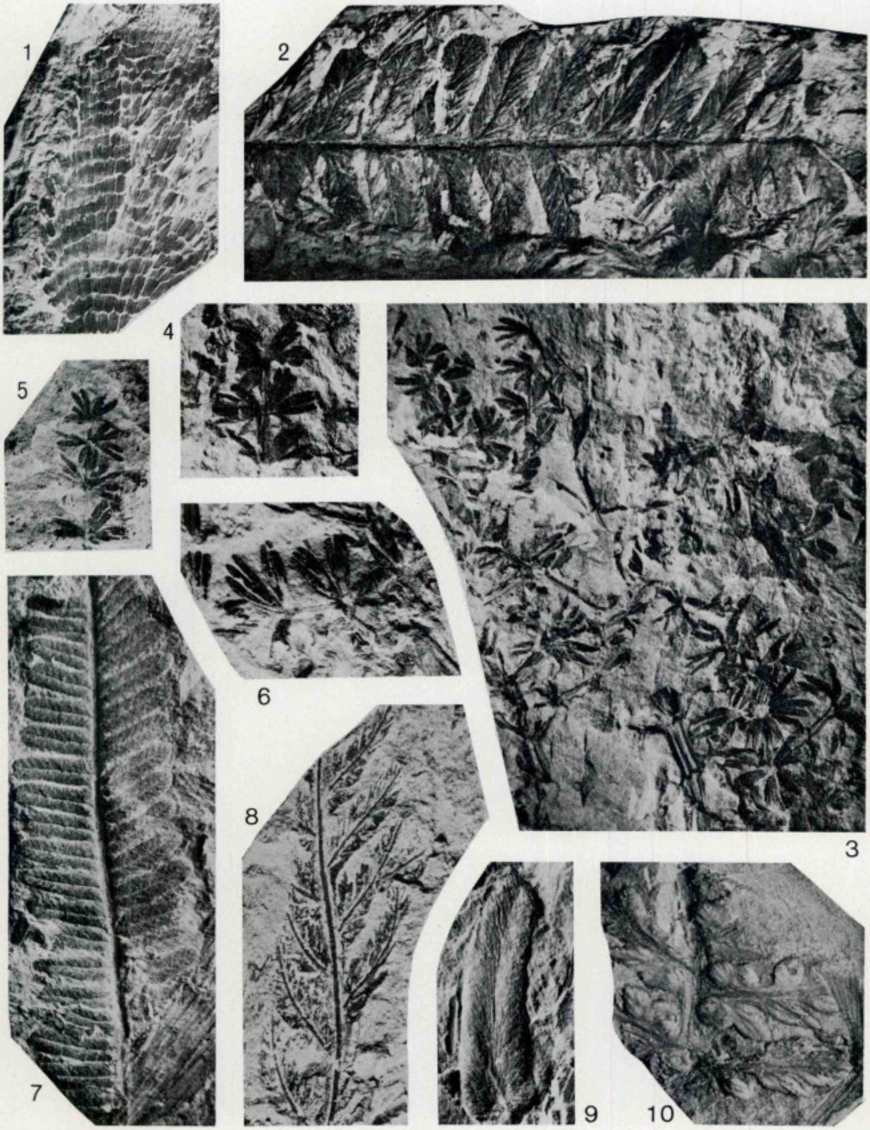
1
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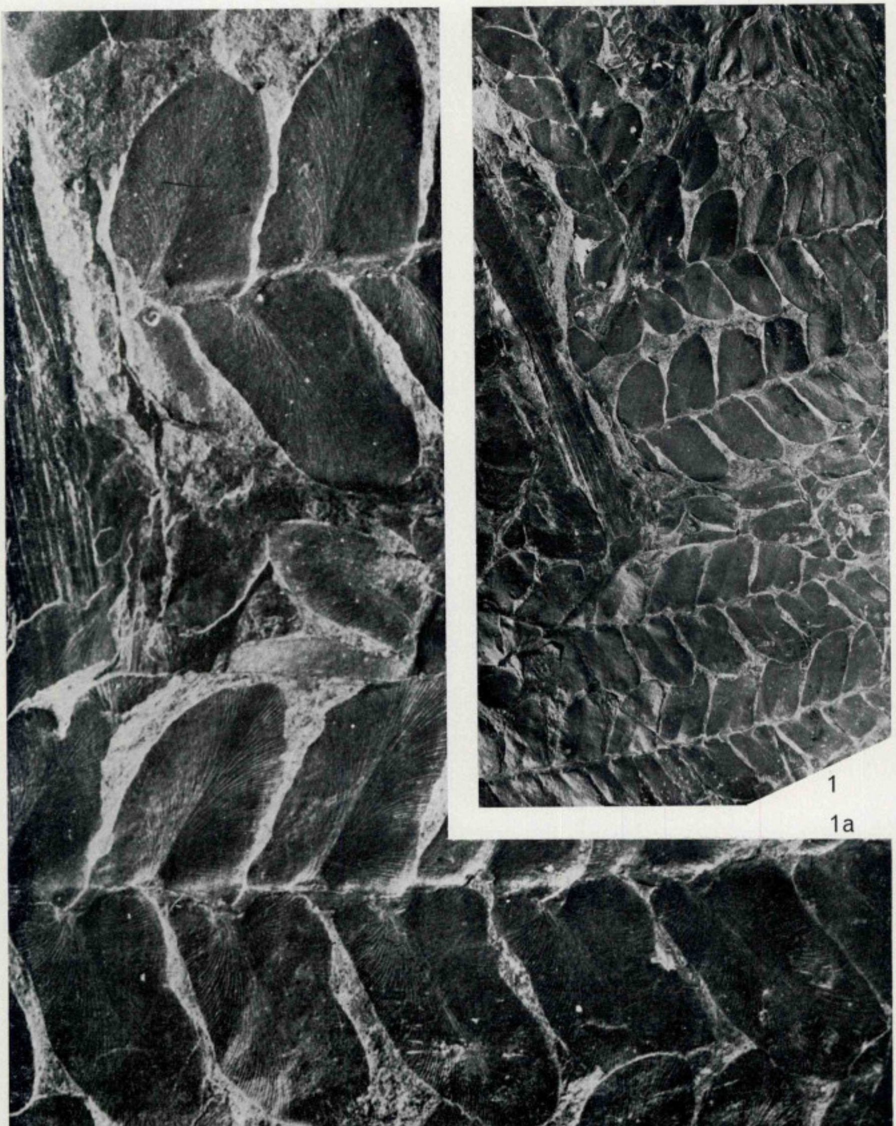
Plate IV

Douro Basin (Oporto region)
lower Stephanian C

1. *Neuropteris ovata* var. *pseudovata* GOTHAN & SZE (x1); 1a (x3).
Between the Esperança and Farrobo collieries, Vila Verde, São Pedro da Cova — Coll. «Bernardino António Gomes» cat. n.° 18 (3) 9 (FCUL).



1

1a

Plate V

Douro Basin (Oporto region)
lower Stephanian C

1. *Neuropteris ovata* var. *pseudovata* GOTHAN & SZE (x1).
Between the Esperança and Farrobo collieries, Vila Verde, São Pedro da Cova — Coll. «Bernardino António Gomes» cat. n.º 9 (4) 4 (FCUL).

2. *Neuropteris cordata* BRONGNIART (x1).
Alto da Bela, Ermesinde.
Specimen figured by TEIXEIRA (1939b, Est. IV, 3).

3-4. *Neuropteris ovata* var. *pseudovata* GOTHAN & SZE: 3 (x1); 3a (x3); 4 (x3).
Between the Esperança and Farrobo collieries, Vila Verde, São Pedro da Cova — Coll. «Bernardino António Gomes» cat. n.º 18 (2) 9 (FCUL).

Plate V

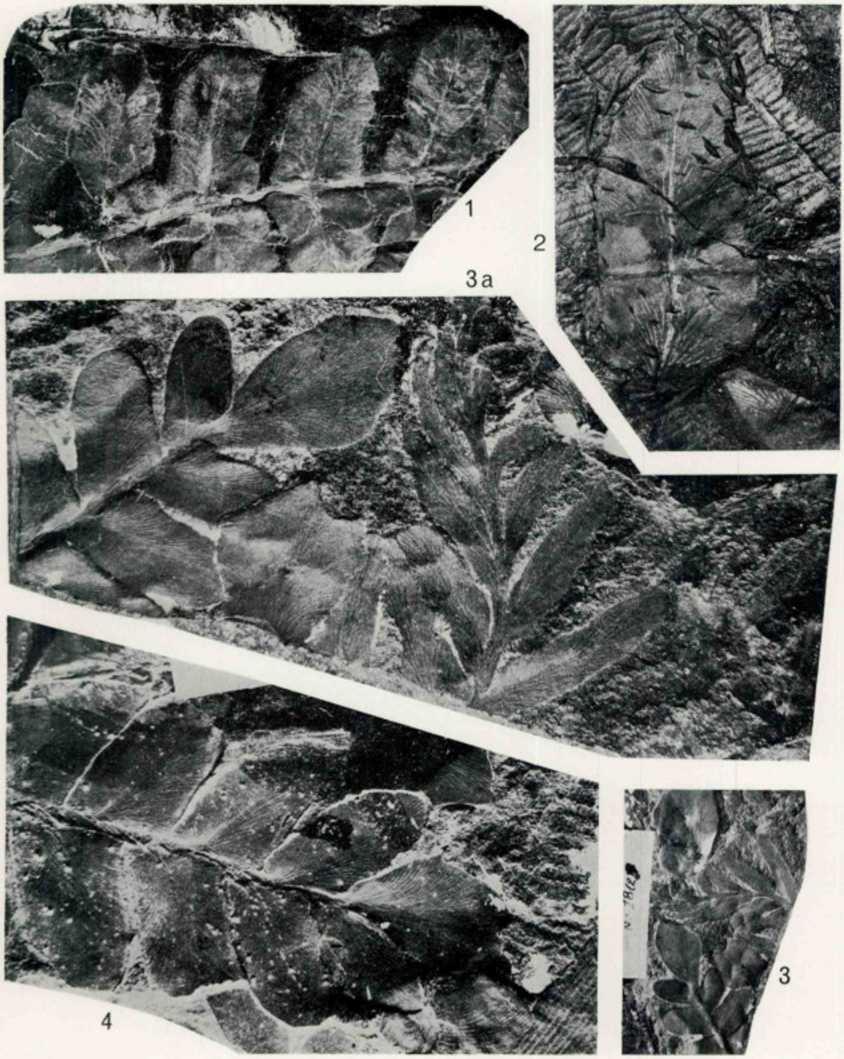


Plate VI

Douro Basin (Oporto region)
lower Stephanian C

1-2. *Sphenopteris mendescorreae* TEIXEIRA (x3).
Fig. 1 shows immature synangia at the tip of pinnule segments.
Alto da Bela, Ermesinde — Coll. FCUP.

3. *Sphenopteris mendescorreae* TEIXEIRA (x3).
50 m west from Alto da Bela, Ermesinde — Coll. FCUP.
Specimen figured by TEIXEIRA (1939b, Est. VIII, 4) as one of the types of *Sphenopteris sampaiana* TEIXEIRA.

4. Fructification of *Sphenopteris mendescorreae* TEIXEIRA (x15).
Alto da Bela, Ermesinde — Coll. FCUP.

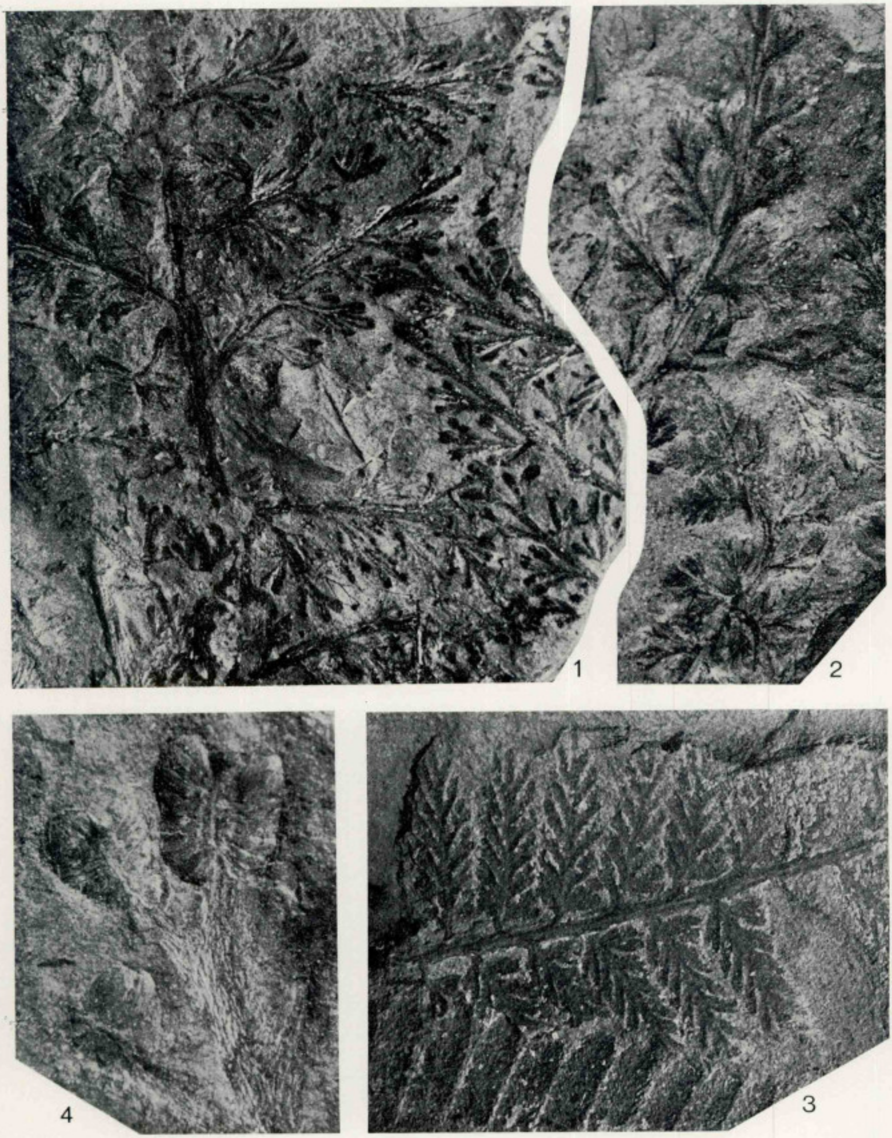

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Plate VII

Douro Basin (Oporto region)
lower Stephanian C

1. *Pecopteris daubreei* ZEILLER (x3).
   São Pedro da Cova — Coll. FCUP.

2. *Pecopteris monyi* ZEILLER (x3).
   São Pedro da Cova — Coll. FCUP.
   Specimen figured by TEIXEIRA (1940b, Est. X, 6).

1

2

Plate VIII

Douro Basin (Oporto region)
lower Stephanian C

Pecopteris limae TEIXEIRA (x3). Lectotype (by designation in the present paper).
Road from São Pedro da Cova to Valongo — Coll. SGP.
Specimen figured by TEIXEIRA (1941b, Est. I, 3).

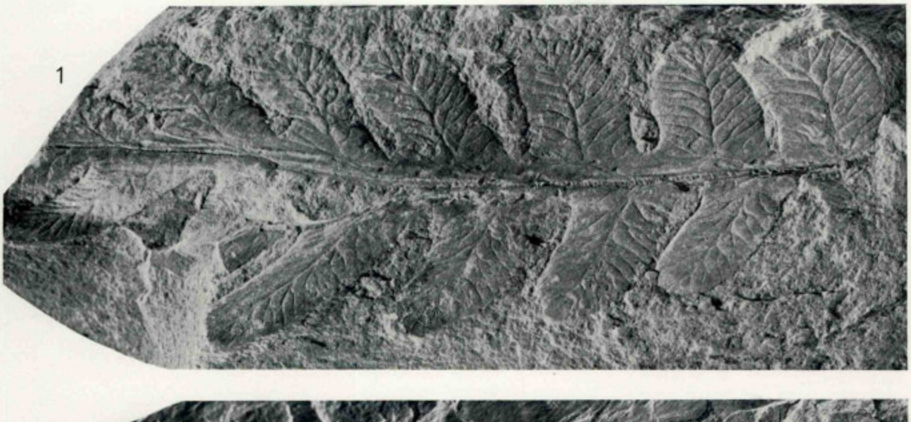
Plate IX

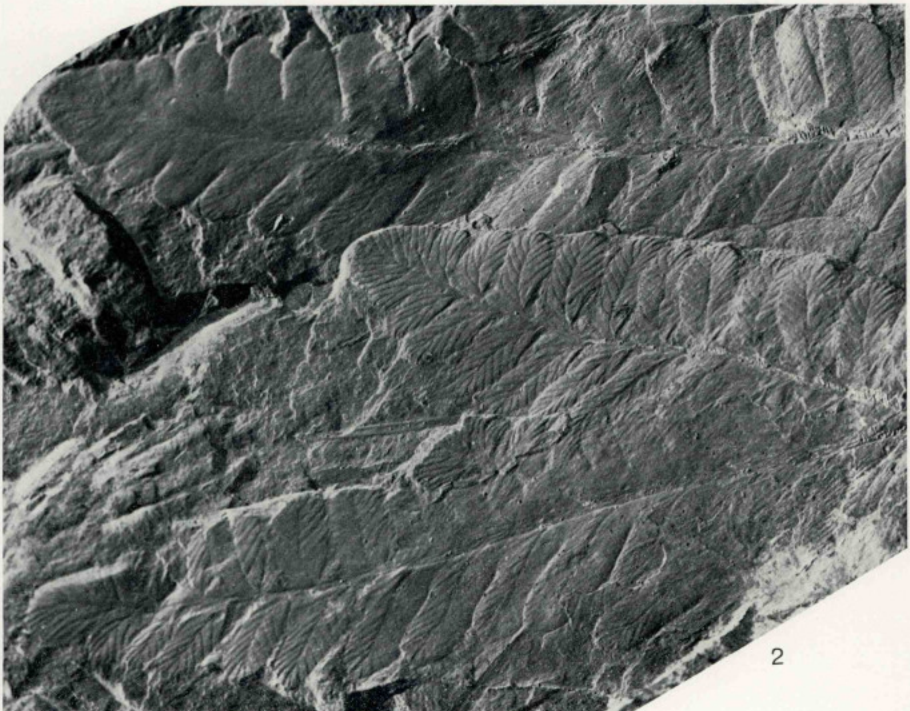
Douro Basin (Oporto region)
lower Stephanian C

1. *Pecopteris limae* TEIXEIRA (x3).
   Road from São Pedro da Cova to Valongo — Coll. FCUP.
   Specimen figured by TEIXEIRA (1941b, Est. I, 4, 4a) as one of the types of *Pecopteris limai.*

2. *Pecopteris limae* TEIXEIRA (x3).
   Road from São Pedro da Cova to Valongo — Coll. Geologisch Bureau, Heerlen, cat. n.° 41 674 (Specimen identified by TEIXEIRA as *P. lepidorachis*).
   This specimen occupies a position in the frond slightly below that of the lectotype, and is regarded as entirely typical.

1





2

Plate X

Douro Basin (Oporto region)
lower Stephanian C

1. *Sphenophyllum costae* STERZEL (x1).
   900 m S 54°W from the church of Pedorido, Melres — Coll. SGP.
   Specimen figured by TEIXEIRA (1955a, Pl. VII, 1, 1a).

2. *Alethopteris zeilleri* RAGOT (x3).
   Covelo, São Pedro da Cova — Coll. «Bernardino António Gomes» cat. n.° 1 (1) 7 (FCUL).
   Specimen figured by TEIXEIRA (1941/42, Est. IV, 2) as *Alethopteris grandini* BRONGNIART.

3. *Reticulopteris germari* (GIEBEL) GOTHAN (= *Linopteris gomesi* TEIXEIRA) (x3).
   500 m south from Alto de Sete Casais — Coll. FCUP.

4. *Pecopteris monyi* ZEILLER (x3).
   1000 m S 68°E from the church of São Pedro da Cova — Coll. SGP.

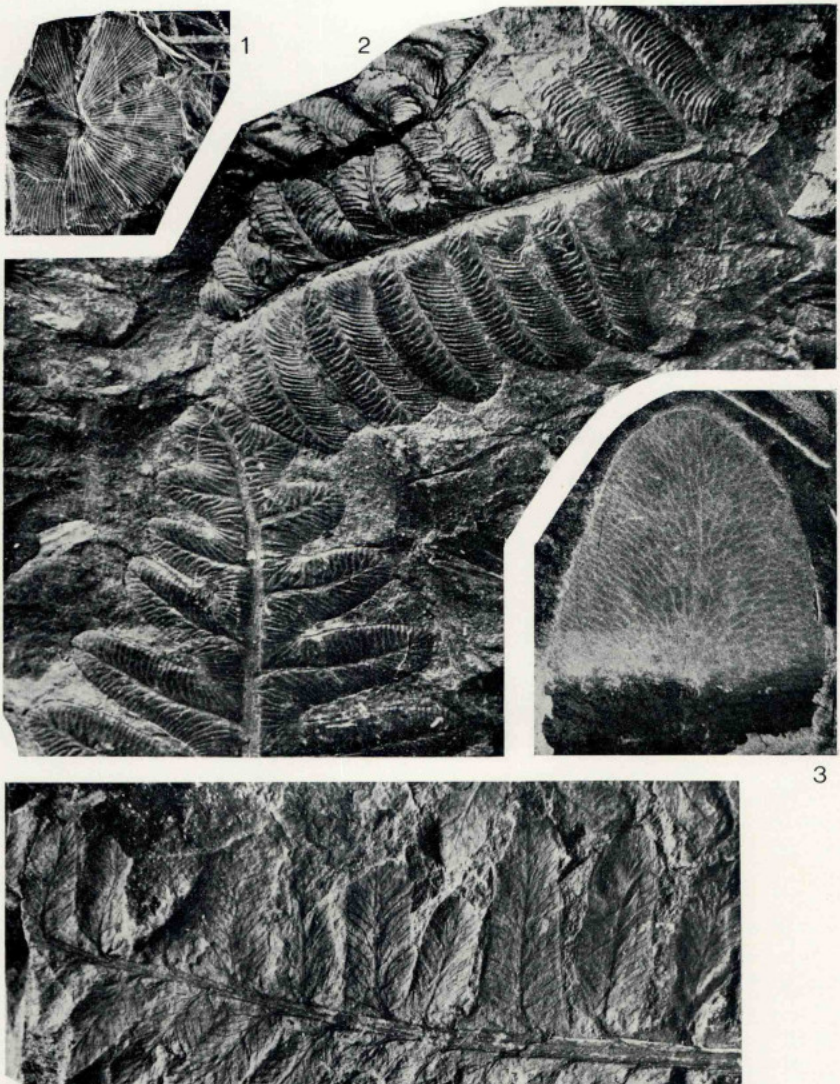

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Plate XI

Douro Basin (Oporto region)
lower Stephanian C

1. *Pecopteris* cf. *ameromi* Stockmans & Willière; *Sphenophyllum oblongifolium* (Germar & Kaulfuss) Unger (x3).
São Pedro da Cova — Coll. FCUL.

2. *Sphenopteris matheti* Zeiller (x3).
Road from São Pedro da Cova to Valongo — Coll. FCUP.

3. *Pecopteris* sp. nov. (cf. *hemitelioides* Brongniart) (x3).
50 m west from Alto da Bela, Ermesinde — Coll. FCUP.
Specimen identified by Teixeira as *Pecopteris hemitelioides* (as stated on the label and as follows from Teixeira 1939b).

4. *Sphenopteris* cf. *matheti* Zeiller (x3).
São Pedro da Cova — Coll. FCUP.

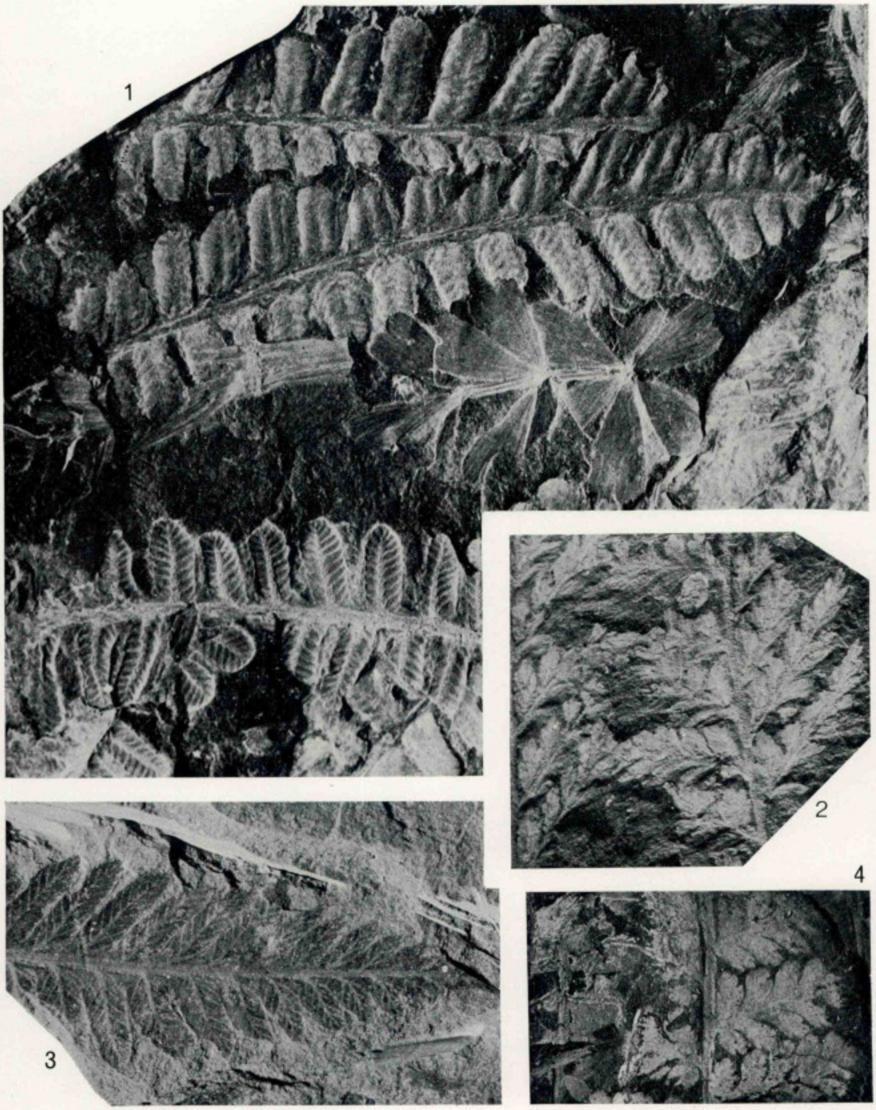
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Plate XII

Douro Basin (Oporto region)
lower Stephanian C

Dicranophyllum lusitanicum (HEER) LIMA (x1).
Specimen figured by GOMES (1865, Tab. 1, 1) as *Cyperites* sp.?, and subsequently under its present name by LIMA (1888, Pl. II), TEIXEIRA (1944, 1945, fig. 4), and TEIXEIRA & PAIS (1976, fig. 38).

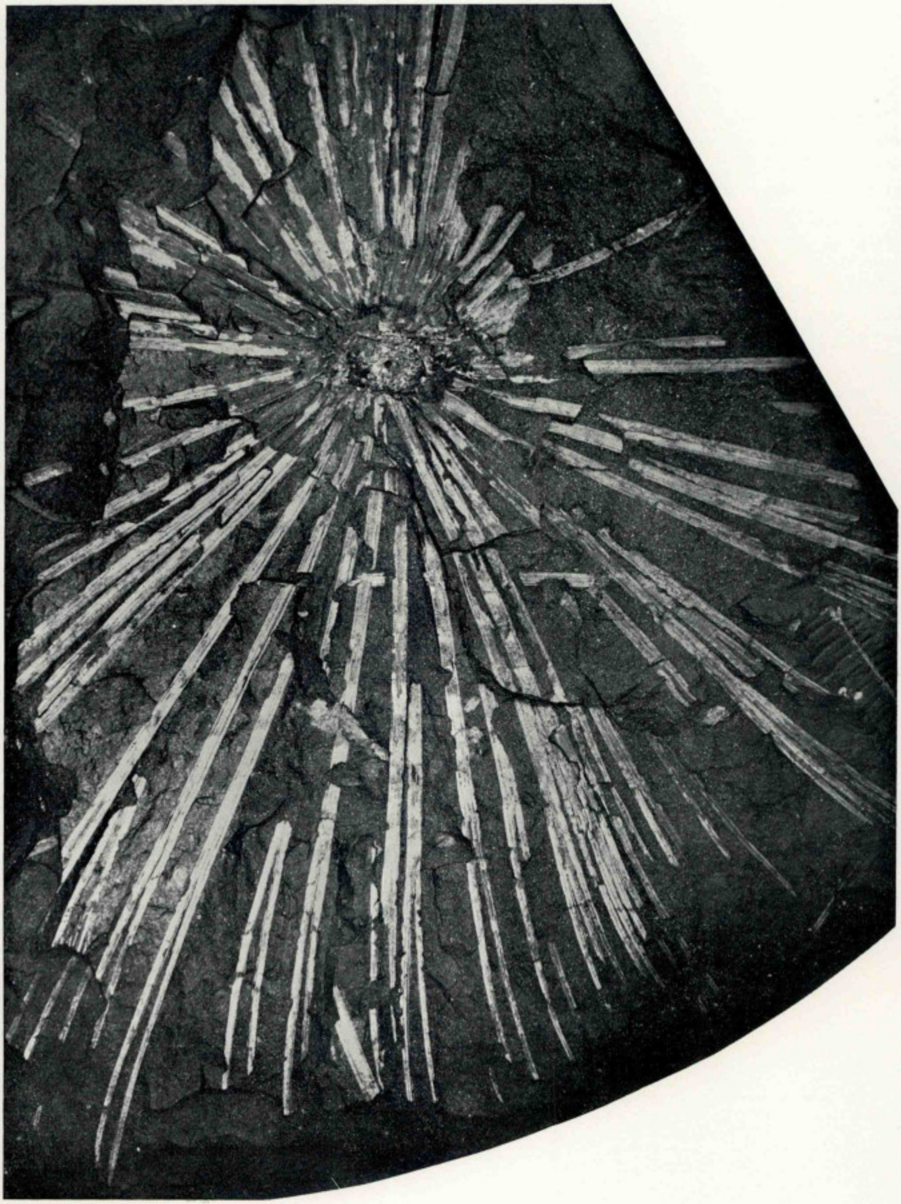

Plate XIII

Buçaco Basin
upper Stephanian C (or lowermost Autunian)

1. *Neuropteris neuropteroides* (GÖPPERT) BARTHEL (x3).
Junqueira — Algeriz road section, Loc. 3905 *in* WAGNER *et al.* (in prep.) — Coll. «Wagner & Lemos de Sousa» (FCUP).

2. *Neuropteris neuropteroides* (GÖPPERT) BARTHEL (x3).
Lapa do Solhal, Salgueiral — Coll. SGP.

3. *Neuropteris zeilleri* LIMA (x1).
1000 m N 10°W from Salgueiral — Coll. SGP.

4. *Odontopteris brardi* BRONGNIART (x3).
Road section near Salgueiral, Loc. 3402 *in* WAGNER *et al.* (in prep.) — Coll. «Wagner & Lemos de Sousa» (FCUP).

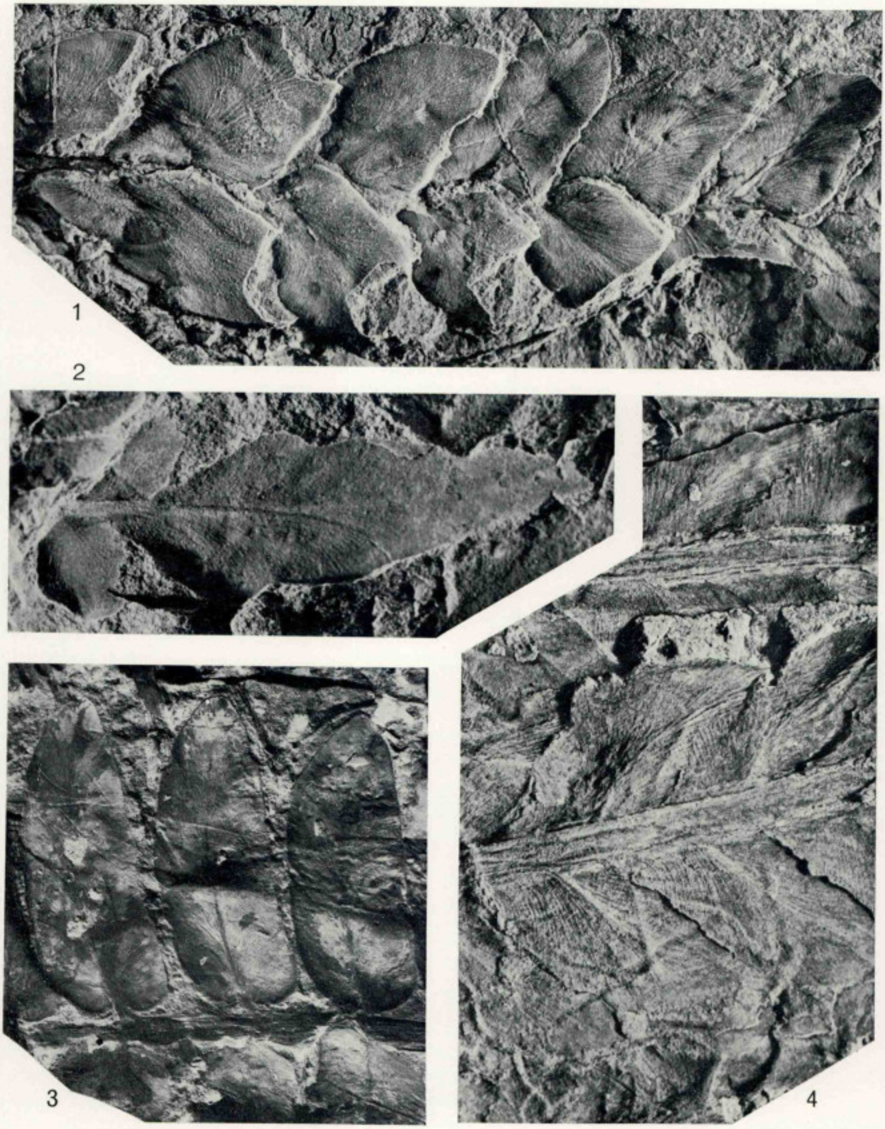
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Plate XIV

Buçaco Basin
upper Stephanian C (or lowermost Autunian)

Neuropteris zeilleri LIMA (x3).
Same as figured natural size on pl. XIII, fig. 3.

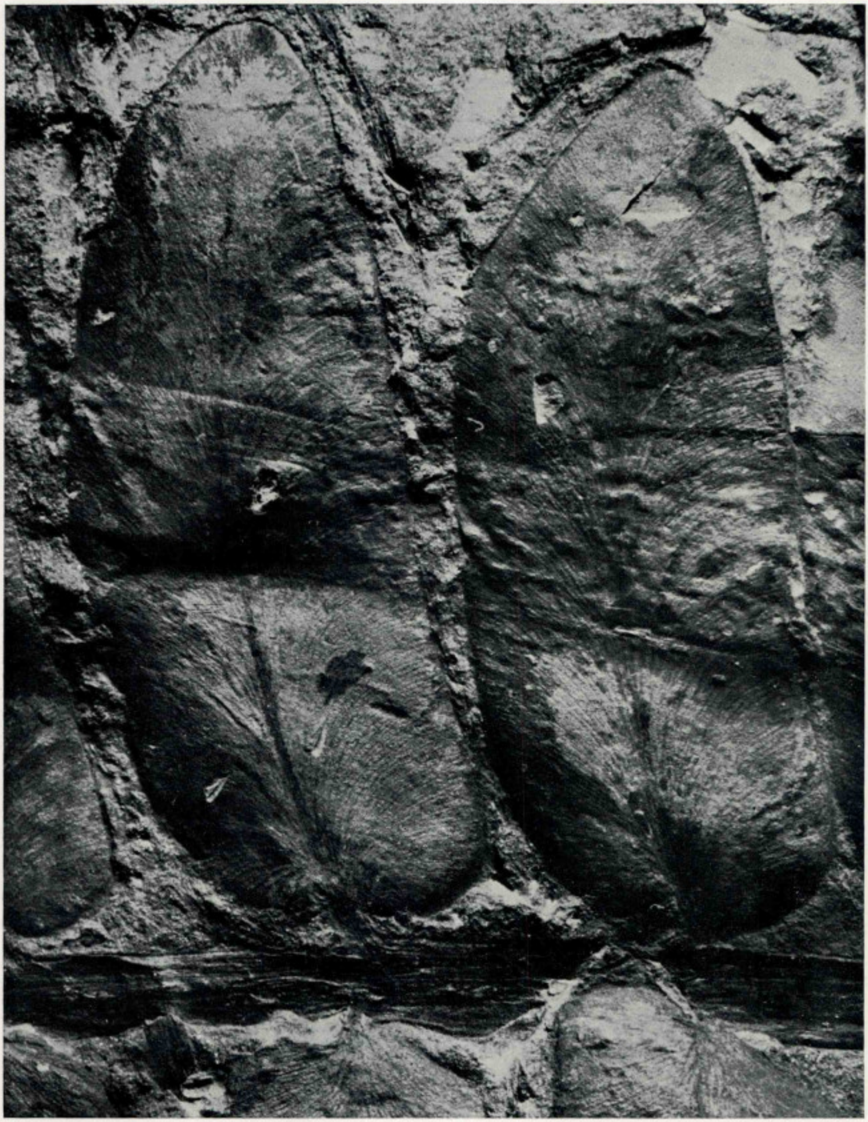

Plate XV

Buçaco Basin
upper Stephanian C (or lowermost Autunian)

1. *Pecopteris cyathea* (VON SCHLOTHEIM) BRONGNIART (x3).
   Junqueira — Algeriz road section, roof shales of coal seam above the village of Algeriz, Loc. 3401 *in* WAGNER *et al.* (in prep.) — Coll. «Wagner & Lemos de Sousa» (FCUP).

2. *Pecopteris cyathea* (VON SCHLOTHEIM) BRONGNIART (x3).
   Santa Cristina, Mealhada — Coll. FCUP.
   Type of *Pecopteris hemitelioides* forma *longipinnata* (TEIXEIRA 1941a, Est. IX, 1, 2).

3-4. *Dicksonites leptophylla* (ZEILLER) DOUBINGER
   Junqueira — Algeriz road section, roof shales of coal seam above the village of Algeriz, Loc. 3401 *in* WAGNER *et al.* (in prep.) — Coll. «Wagner & Lemos de Sousa» (FCUP).

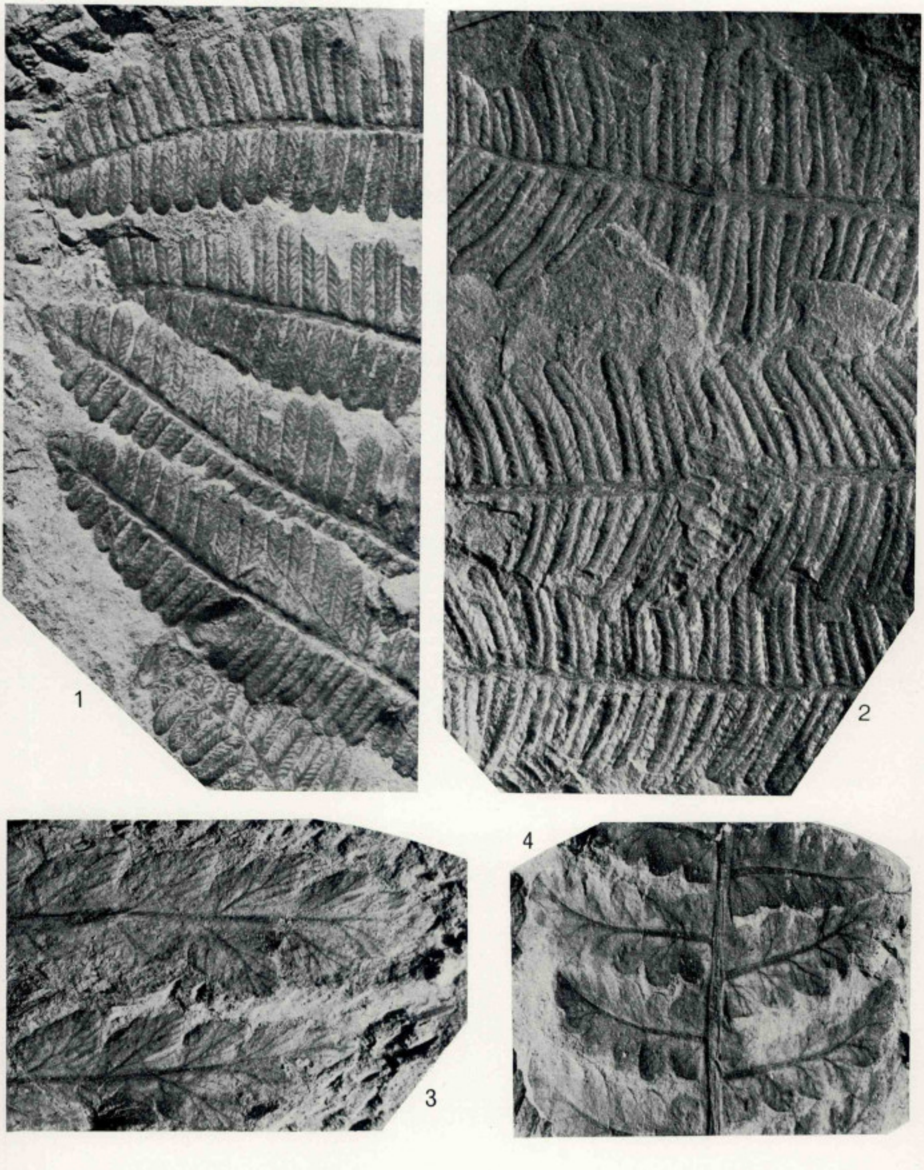
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Plate XVI

Buçaco Basin
upper Stephanian C (or lowermost Autunian)

1-3. *Mixoneura* sp. cf. *Odontopteris osmundaeformis* (VON SCHLOTHEIM) ZEILLER.
80 m southwest of Fonte do Salgueiro, Luso.
1. odontopteroid pinna (x3) — Coll. FCUL.
2. mixoneuroid pinna (x1); 2a (x3) — Coll. SGP, Specimen figured by TEIXEIRA (1941e, Est. VI).
3. neuropteroid pinna (x3) — Coll. SGP.

4. *Sphenophyllum thoni* VON MAHR (x3).
Exact locality unknown — Coll. FCUP.

Plate XVI

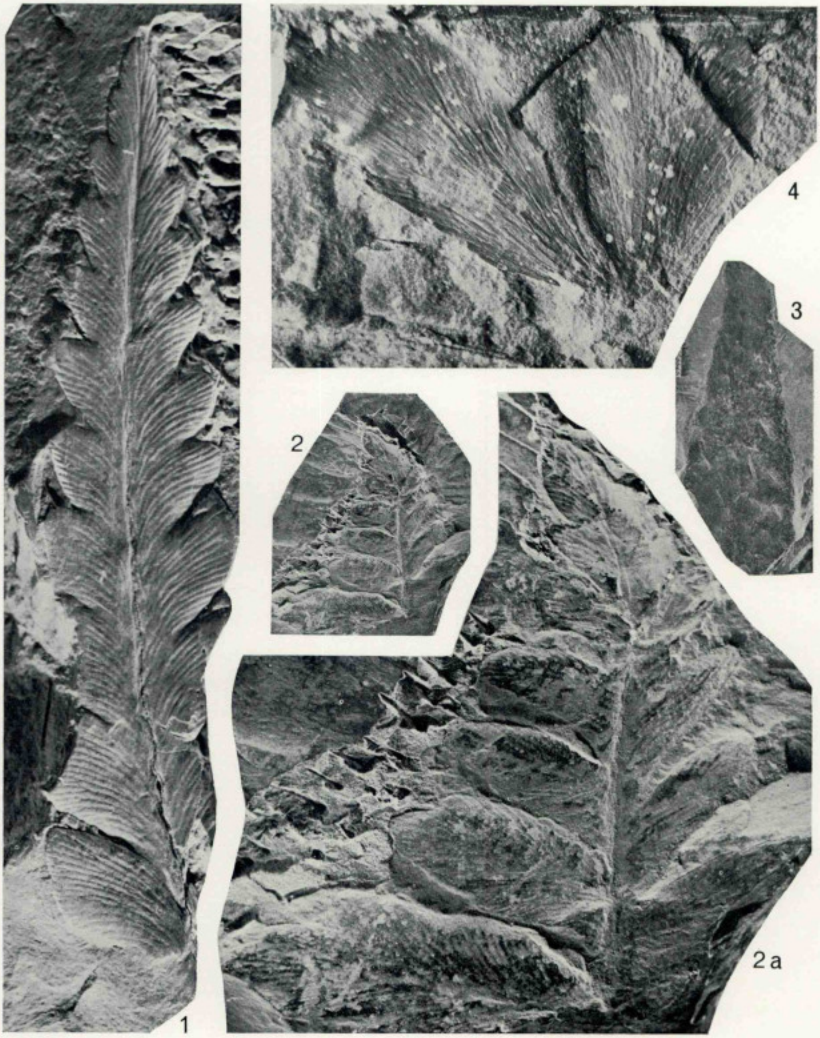
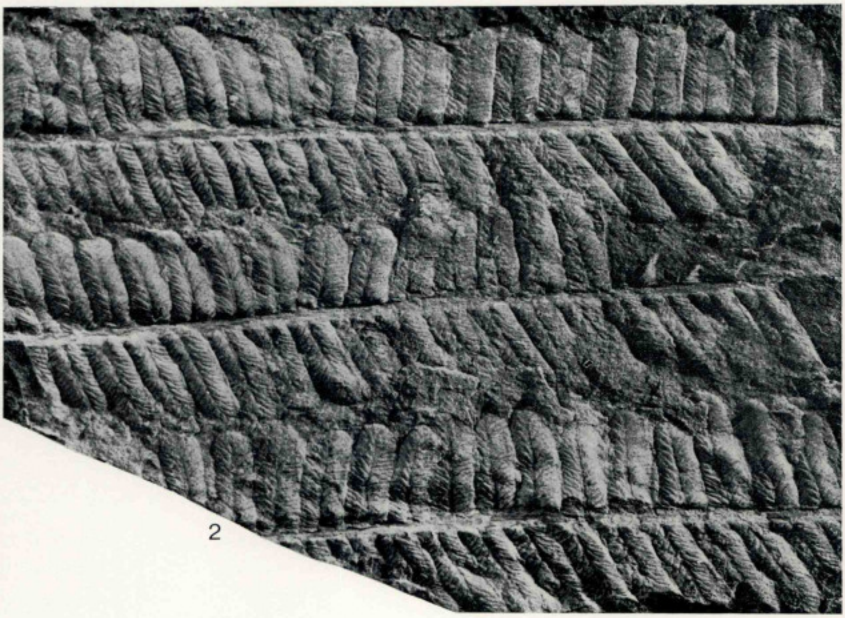


Plate XVII

Buçaco Basin
upper Stephanian C (or lowermost Autunian)

1-2. *Pecopteris* sp. nov. (x3).
Salgueiral — Coll. SGP.

1



2

Plate XVIII

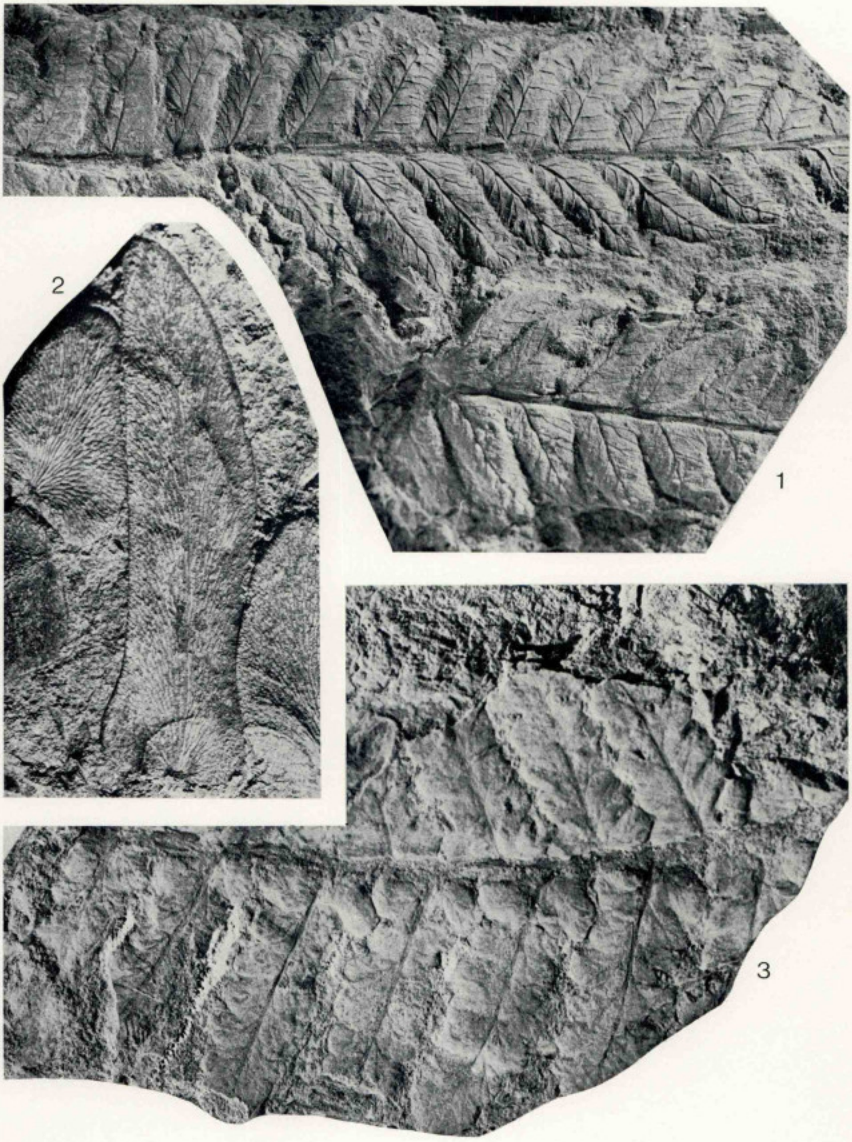
Buçaco Basin
upper Stephanian C (or lowermost Autunian)

1. *Oligocarpia leptophylla* (BUNBURY) GRAUVOGEL-STAMM & DOUBINGER (x3).
   550 m south of Algeriz — Coll. SGP.

2. *Linopteris gangamopteroides* (DE STEFANI) WAGNER (x3).
   Porta das Lapas — Coll. SGP.

3. *Sphenopteris casteli* ZEILLER (x3).
   Between Ferreiros and Pirâmide da Gralheira — Coll. SGP.

Plate XVIII



THE PALAEOGEOGRAPHICAL AND AGE RELATIONSHIPS OF THE PORTUGUESE CARBONIFEROUS FLORAS WITH THOSE OF OTHER PARTS OF THE WESTERN IBERIAN PENINSULA [1]

R. H. Wagner [2]

Key words: Carboniferous plants Iberian Peninsula; Intramontane basins; Coastal basins; Westphalian B; Westphalian C; Westphalian D; Cantabrian; Stephanian B; Stephanian C; Autunian.

Palavras-chave: Plantas fósseis do Carbonífero da Península Ibérica; Bacias intramontanhosas; Bacias costeiras; Vestefaliano B; Vestefaliano C; Vestefaliano D; Cantabriano; Estefaniano B; Estefaniano C; Autuniano.

ABSTRACT

The floral assemblages found in the elongate strip of Carboniferous deposits near Oporto, in the Buçaco Basin and in the Santa Susana Basin are compared with those of similar ages in the adjacent parts of Spain. It is noted that the Portuguese occurrences are all devoid of marine influences. They all occur along important fracture zones and are apparently all intramontane basins (although this cannot be proved for the highly sheared Westphalian C (?) and D deposits in the region near Oporto). The composition of the early Stephanian C flora of the Douro Basin (near Oporto) and the very late Stephanian C flora of Buçaco shows the common presence of extrabasinal elements, as can be expected in intramontane basins. Both occur in the Central Iberian Zone and comparison is made with the late Stephanian B flora of Puertollano in Ciudad Real, La Mancha. There is a clearly marked compositional difference with the early Stephanian C floras of the Cantabrian Zone in Northwest Spain, which occur in a coastal basin.

The floral assemblage recorded from Santa Susana is perhaps most similar to that of upper Westphalian D strata in the South Wales, Forest of Dean and Bristol-Somerset areas of Britain. There are no deposits of comparable age in the Ossa-Morena Zone in Spain.

RESUMO

Comparam-se as associações florísticas encontradas na faixa de idade Carbonífera existente nas proximidades do Porto e nas Bacias do Buçaco e de Santa Susana com as associações congéneres conhecidas nos terrenos de idade similar em áreas adjacentes de Espanha. Faz-se notar que as formações portuguesas não sofreram influência marinha. Com efeito, as referidas formações situam-se, todas, ao longo de importantes zonas e fractura e, bem assim, em todos os casos, parecem corresponder a bacias intramontanhosas. Isto, muito embora, o carácter intramontanhoso não se possa provar no caso das formações fortemente cisalhadas do Vestefaliano C (?) e D que ocorrem na região do Porto.

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(1) The Portuguese Carboniferous floras were studied with the aid of NATO Research Grant n.º 85.80/D1/D2, which is gratefully acknowledged. The author is indebted to his colleague, M. J. Lemos de Sousa, for suggesting this paper and contributing materially to its completion.

(2) Department of Geology, The University, Beaumont Building, Brookhill, Sheffield S3 7HF, England, U. K., and E. N. Carbonífera del Sur, Peñarroya-Pueblonuevo (Córdoba), España.

Tal como seria de esperar em bacias intramontanhosas, as floras da Bacia do Douro (Estefaniano C inferior) e do Buçaco (Estefaniano C superior) patenteiam, ambas, elementos provenientes das margens da bacia («extra-basinal elements»). Como tanto a Bacia do Douro como a do Buçaco se situam na Zona Centro-Ibérica efectuaram-se comparações com as floras do Estefaniano B superior de Puertollano, Província de Ciudad Real (La Mancha). Por outro lado, faz-se notar que existe uma assinalável diferença na composição das floras intramontanhosas que ocorrem no Estefaniano C inferior da Zona Centro-Ibérica e na bacia costeira da mesma idade na Zona Cantábrica (Noroeste de Espanha).

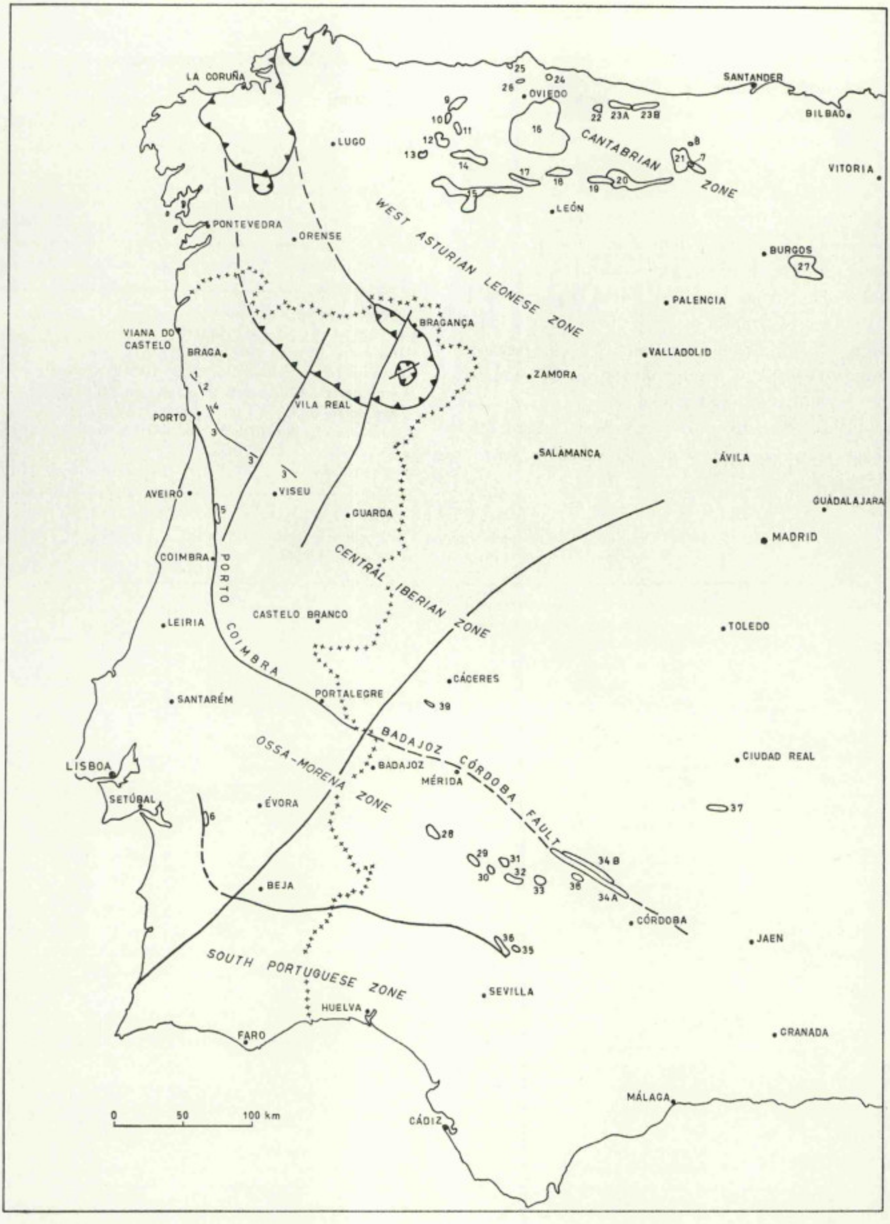
A associação florística que se encontra em Santa Susana é, talvez, mais similar às que ocorrem no Vestefaliano D superior da parte sudeste do País de Gales e nas áreas de Forest of Dean e de Bristol-Somerset em Inglaterra. Em Espanha não existem, contudo, depósitos desta idade na Zona de Ossa-Morena.

1. CARBONIFEROUS OCCURRENCES NEAR OPORTO

These belong to three different ages, viz. Westphalian C (?), late Westphalian D, and early Stephanian C (WAGNER & SOUSA 1983 — this volume). According to DOMINGOS *et al.* (1983 — this volume), the corresponding strata were probably deposited in a single basin formed by subsidence along a shear zone west of the Valongo Anticline in Lower Palaeozoic rocks. Although this may be the case, it is also possible that this shear zone merely preserved slices of a number of successive Carboniferous successions of more or less independent origin. The fact is that the various Carboniferous successions in the narrow strip which extends in 130 km from Criaz in the northwest to Mioma (northeast of Viseu) in the southeast (Fig. 1), all appear with a similar NW-SE strike and in an identical steeply inclined position. In the northwestern area only some rather poorly dated Westphalian successions are found (including the probable Westphalian C of Casais-Alvarelhos — WAGNER & SOUSA 1982), whilst the southeastern part contains the lower Stephanian C strata of the Douro Coalfield. In the central part the upper Westphalian D strata of Ervedosa occur in a thrust slice between Lower Palaeozoics and parallel to lower Stephanian C deposits.

Fig. 1 — TERRESTRIAL CARBONIFEROUS AND LOWER PERMIAN OCCURRENCES IN THE WESTERN PART OF THE IBERIAN PENINSULA

1 — Criaz — Serra de Rates (Westphalian ?)
2 — Casais — Alvarelhos (Westphalian C ?)
3 — Douro (lower Stephanian C)
4 — Ervedosa (upper Westphalian D)
5 — Buçaco (uppermost Stephanian C or lower Autunian)
6 — Santa Susana (uppermost Westphalian D)
7 — Peña Cildá (upper Stephanian B or lower Stephanian C)
8 — Pico Cordel (lower Stephanian C)
9 — Tineo (lower Stephanian C)
10 — Cangas del Narcea (lower Stephanian C)
11 — Carballo (lower Stephanian C)
12 — Rengos (lower Stephanian C)
13 — Tormaleo (lower Stephanian C)
14 — Villablino (Stephanian B + C)
15 — El Bierzo (lower Stephanian C)
16 — Central Asturian Coalfield (Westphalian C + D)
17 — La Magdalena (Stephanian B)
18 — Ciñera-Matallana (Stephanian B)
19 — Sabero (upper Stephanian A + Stephanian B)
20 — Guardo (upper Westphalian D + lower Cantabrian)
21 — S. Cebrían, La Pernía, Barruelo (Westphalian D, Cantabrian + lower Stephanian A)
22 — Sebarga (Stephanian B or C)
23A — Gamonedo — Inguanzo (Cantabrian)
23B — Cabrales — Panes (Stephanian B)
24 — La Camocha (Namurian, Westphalian A + lower B)
25 — Arnao (lower Stephanian C)
26 — Ferroñes (lower Stephanian C)
27 — La Demanda (Westphalian C + D)
28 — Los Santos de Maimona (Viséan)
29 — Bienvenida (Viséan)
30 — Casas de Reina (Viséan)
31 — Berlanga (Viséan)
32 — Guadalcanal (Autunian)
33 — Valdeinfierno (upper Tournaisian)
34A — Guadiato (lower Namurian)
34B — Guadiato (Westphalian B of Peñarroya-Belmez)
35 — Villanueva del Rio (Westphalian A)
36 — Viar (Autunian)
37 — Puertollano (upper Stephanian B or lower Stephanian C)
38 — Benajarafe (Viséan)
39 — Sierra de S. Pedro (lower? Westphalian)


LA CORUÑA
LUGO
OVIEDO
SANTANDER
BILBAO
VITORIA
CANTABRIAN ZONE
LEÓN
PONTEVEDRA
ORENSE
WEST ASTURIAN LEONESE ZONE
BURGOS
PALENCIA
BRAGANÇA
VIANA DO CASTELO
BRAGA
VALLADOLID
ZAMORA
VILA REAL
PORTO
SALAMANCA
ÁVILA
AVEIRO
VISEU
GUARDA
GUADALAJARA
MADRID
COIMBRA
PORTO COIMBRA
CENTRAL IBERIAN ZONE
LEIRIA
CASTELO BRANCO
TOLEDO
CÁCERES
SANTARÉM
PORTALEGRE
BADAJOZ CÓRDOBA FAULT
OSSA-MORENA ZONE
LISBOA
BADAJOZ
MÉRIDA
CIUDAD REAL
SETÚBAL
ÉVORA
BEJA
CÓRDOBA
JAEN
SOUTH PORTUGUESE ZONE
SEVILLA
HUELVA
FARO
GRANADA
MÁLAGA
CÁDIZ
0
50
100 km

Since no basin margins have been demonstrated, and the lower Stephanian C deposits of the Douro Coalfield show an unconformable contact with Lower Palaeozoic rocks and a Westphalian granite (Oporto Granite) (PRIEM *et al.* 1967, 1970), it is preferred here to regard the Westphalian and upper Stephanian units as remnants of mutually independent successions which have been preserved as thrust slices in a major shear zone. It is also noted that boulders of the Oporto Granite occur in the basal conglomerate of the lower Stephanian C succession (SOUSA 1978). The similar structural position of the Westphalian and upper Stephanian units would be due to post-sedimentary tectonics. There is certainly a wide difference in stratigraphic age.

1.1. Westphalian of Oporto region

Assuming that the Westphalian C (?) of Casais-Alvarelhos (Table 1) and the upper Westphalian D of Ervedosa (Table 2) are differ-

Table 1 — List of Westphalian C (?) plant fossils from Cidai (Bougado) in the Casais-Alvarelhos occurrence, north of Oporto, Portugal (after WAGNER & SOUSA 1982)

Linopteris florini TEIXEIRA
Alethopteris cf. *davreuxi* (BRONGNIART) GÖPPERT
Mariopteris nervosa (BRONGNIART) ZEILLER
Eusphenopteris nummularia (VON GUTBIER) NOVIK
Sphenopteris (Alloiopteris ?) cf. *dixoni* KIDSTON
Sphenopteris sp.
Gondomaria discreta (WEISS) WAGNER & LEMOS DE SOUSA
Sphenophyllum majus BRONN
Asterophyllites sp.

Table 2 — Composite list of the upper Westphalian D flora of Ervedosa, Oporto region, Portugal (after WAGNER & SOUSA 1983)

Neuropteris ervedosensis TEIXEIRA
Neuropteris scheuchzeri HOFFMANN
Neuropteris sp.
Linopteris florini TEIXEIRA
Linopteris obliqua (BUNBURY) ZEILLER
Callipteridium (Praecallipteridium) jongmansi (P. BERTRAND) WAGNER
Alethopteris corsini BUISINE
Dicksonites plueckeneti forma *sterzeli* ZEILLER
Eusphenopteris cf. *neuropteroides* (BOULAY) NOVIK
Sphenopteris sp.
Oligocarpia pulcherrima STUR
cf. *Polymorphopteris polymorpha* (BRONGNIART) WAGNER
Lobatopteris micromiltoni (P. BERTRAND) WAGNER
cf. *Lobatopteris vestita* (LESQUEREUX) WAGNER
Pecopteris dentata BRONGNIART
Pecopteris cf. *hucheti* CORSIN
Pecopteris cf. *nyranensis* NĚMEJC
Pecopteris unita BRONGNIART
Pecopteris sp.
Sphenophyllum sp. indet.
Annularia sphenophylloides (ZENKER) VON GUTBIER
Annularia stellata (VON SCHLOTHEIM) WOOD
Asterophyllites sp.
Calamites cf. *gigas* BRONGNIART
Calamites suckowii BRONGNIART
Calamites sp.
Cordaites sp.
Lepidocarpon major (BRONGNIART) HEMINGWAY
Sigillariophyllum sp.

ent remnants of what may have been a single upper Westphalian succession, there is no immediate comparison in regard to stratigraphic age with other Westphalian successions in the Central Iberian Zone. In the province of Cáceres, in Spain (Fig. 1), some conglomeratic, lower (?) Westphalian strata have been recorded in the Sierra de San Pedro by BOCHMANN (1977) *in* WALTER, but these rocks are of Westphalian B age. A more complete Westphalian B succession, also non-marine, is known from the Peñarroya-Belmez Coalfield in the province of Córdoba. This terrestrial succession, close to 2,000 m thick, possibly commences with uppermost Westphalian A and corresponds almost entirely to the Westphalian B. The higher levels, as found in the Oporto region, have been eroded in this area. It is quite striking, however, that the Westphalian occurrences near Oporto, south of Cáceres, and at Peñarroya-Belmez, are all devoid of marine influences. This makes a striking difference with the largely marine Westphalian successions of the Cantabrian Zone in Northwest Spain.

The Westphalian successions of the Central Iberian Zone are therefore likely to correspond to local depressions in a general hinterland area. Of course, the very fragmentary records of Westphalian strata in the area near Oporto can teach us very little about the basin development in this area. In this respect, it is only the Peñarroya-Belmez Coalfield which is instructive. This coalfield shows clear evidence of a faulted northeastern basin margin with a moderate but discernible palaeotopography which was filled in by alluvial fan deposits. These were followed by fluviatile and floodplain deposits laid down by streams orientated in the long axis of the NW-SE striking basin. Important lacustrine deposits followed later.

The Westphalian B floras of the Peñarroya-Belmez Coalfield (Table 3) and the Westphalian C (?) and D floras of Casais-Alvarelhos and Ervedosa are all comparable to the standard Westphalian floras of the paralic coal belt of northern Europe and also with those of the non-marine area of Saar-Lorraine. There are minor differences however. At Peñarroya-Belmez, one of the most common floral elements is a new species of *Neuropteris*, whilst Ervedosa shows the presence of another endemic form, *Neuropteris ervedosensis* TEIXEIRA. It is also noted that some of the common forms in Peñarroya-Belmez, e.g. *Sphenophyllum kidstoni* HEMINGWAY, are rare elements in the North European floras of the paralic coal belt. On the whole, however, the same species occur, thus confirming the general homogeneity of Westphalian floras all over the Amerosinian palaeoequatorial belt of mid-Carboniferous times.

There is not a great deal of difference between the upper Westphalian D flora of Ervedosa and the higher Westphalian D floras of the Central Asturian Coalfield in the Cantabrian

Table 3 — Composite list of Westphalian B flora from the Peñarroya-Belmez Coalfield (Guadiato region), Province of Córdoba, Spain

Neuropteris sp. nov. (2 species)
Neurodontopteris sp. nov. ?
Paripteris gigantea (STERNBERG) GOTHAN
Paripteris linguaefolia (P. BERTRAND) LAVEINE
Linopteris cf. *neuropteroides* (VON GUTBIER) POTONIÉ
Alethopteris davreuxi (BRONGNIART) GÖPPERT
Alethopteris lonchitica (VON SCHLOTHEIM) ZEILLER
Alethopteris urophylla (BRONGNIART) VON ROEHL
Lonchopteris rugosa BRONGNIART
Mariopteris muricata (VON SCHLOTHEIM) ZEILLER
Mariopteris nervosa (BRONGNIART) ZEILLER
Mariopteris cf. *sauveuri* forma *stricta* LUTZ
cf. *Karinopteris daviesi* (KIDSTON) BOERSMA

Table 3 — *(Cont.)*

Karinopteris soubeirani (ZEILLER) BOERSMA
Eusphenopteris hollandica (GOTHAN & JONGMANS) NOVIK
Eusphenopteris pelloi (WAGNER) comb. nov.
Eusphenopteris polyphylla (LINDLEY & HUTTON) NĚMEJC
Eusphenopteris sauveuri (CRÉPIN) SIMSON-SCHAROLD
Eusphenopteris cf. *schumanni* (STUR) VAN AMEROM
Eusphenopteris cf. *trifoliolata* (ARTIS) NOVIK
Palmatopteris furcata (BRONGNIART) POTONIÉ
Diplotmema sp. nov.
Rhodeopteridium souichi (ZEILLER) WAGNER
Rhodeites gutbieri (VON ETTINGSHAUSEN) NĚMEJC
Sphenopteris (Zeilleria) frenzli (STUR) ZEILLER
Sphenopteris (Zeilleria) hymenophylloides KIDSTON
Sphenopteris (Urnatopteris) herbacea (BOULAY) CARPENTIER
Sphenopteris (Discopteris ?) cf. *goldenbergi* ANDRAE
Sphenopteris (Renaultia) cf. *chaerophylloides* (BRONGNIART) ZEILLER
Sphenopteris (Renaultia) schwerini STUR
Sphenopteris (Oligocarpia) brongniarti STUR
Sphenopteris (Oligocarpia) gutbieri GÖPPERT
Sphenopteris artemisiaefolioides CRÉPIN
Sphenopteris dufayi DANZÉ
Sphenopteris cf. *latinervis* DANZÉ
Sphenopteris magnifica DANZÉ
Sphenopteris obtuse-dentata GOTHAN
Sphenopteris spinosa GÖPPERT
Sphenopteris vuellersi STUR
Alloiopteris coralloides (VON GUTBIER) POTONIÉ
Alloiopteris tenuissima (PRESL) NĚMEJC
Corynepteris sp.
Pecopteris (Lobatopteris) lignyi STOCKMANS & WILLIÈRE
Pecopteris (Lobatopteris) precursor STOCKMANS & WILLIÈRE
Pecopteris pennaeformis BRONGNIART
Pecopteris plumosa (ARTIS) BRONGNIART
Sphenophyllum cuneifolium STERNBERG
Sphenophyllum kidstoni HEMINGWAY
Sphenophyllum lineare WAGNER
Sphenophyllum myriophyllum CRÉPIN
Sphenophyllum trichomatosum STUR
Sphenophyllum wingfieldense HEMINGWAY
Annularia microphylla SAUVEUR
Annularia radiata BRONGNIART
Asterophyllites grandis (STERNBERG) GEINITZ
Asterophyllites lycopodioides ZEILLER
Bothrodendron minutifolium (BOULAY) ZEILLER
Lepidodendron aculeatum STERNBERG
Lepidodendron simile KIDSTON
Lepidophloios laricinus STERNBERG
Lepidostrobophyllum cf. *lanceolatum* (LINDLEY & HUTTON) CHALONER
Sigillaria cf. *boblayi* BRONGNIART
Sigillaria elegans (STERNBERG) BRONGNIART
Cordaites palmaeformis GÖPPERT
Stigmaria ficoides STERNBERG

Zone. Unfortunately, an exact comparison with the Central Asturian floras is made difficult by problems of correlation within the Central Asturian Coalfield and by the doubts existing about a number of identifications. Modern work on the fossil floras of the Asturian Coalfield commenced with the pictorial documentation provided by JONGMANS (1952). Unfortunately, a number of his identifications could not be sustained, particularly those referring to Stephanian species (WAGNER 1959). The same doubts exist with regard to later identifications by CARIDE *et al.* (1973), whose record of Stephanian species in the two highest divisions of the succession in the northeastern part of the Central Asturian Coalfield (Nalón Valley area) is clearly questionable. For example, the specimen figured as *Alethopteris bohemica* FRANKE (op. cit., pl. II, figs 4, 4a) should probably be attributed to *Callipteridium (Praecallipteridium) jongmansi* (P. BERTRAND) WAGNER. Their *Pecopteris subelegans* (POTONIÉ) WAGNER (op. cit., pl. IV, figs 1, 1a) is certainly not that species, but resembles *Pecopteris simoni* ZEILLER, and their *Pecopteris lepidorachis* BRONGNIART and *Pecopteris cyathea* (SCHLOTHEIM) BRONGNIART (op. cit., pl. IV, figs 2, 2a and 3, 3a, respectively) are probably to be compared with *Pecopteris saraefolia* P. BERTRAND and *Pecopteris (Lobatopteris) micromiltoni* P. BERTRAND. It is also possible that their *Dicksonites plueckeneti* (SCHLOTHEIM) STERZEL (op. cit., pl. III, fig. 1) may be referred to *Dicksonites potieri* (ZEILLER) P. BERTRAND. Another clear misidentification in their paper is the specimen figured as *Lonchopteridium* cf. *alethopteroides* GOTHAN (op. cit., pl. III, figs 2, 2a) which is a beautiful example of *'Alethopteris' riosensis* WAGNER. With this number of certain misidentifications, the lists presented in CARIDE *et al.* 's paper

Table 4 — Composite revised list of mid-Westphalian D flora from Ablanedo in the area of Riosa, northwestern part of the Central Asturian Coalfield, Cantabrian Mts, Northwest Spain

Neuropteris sp. nov. ?
Neurodontopteris peyerimhoffi (P. BERTRAND) comb. nov.
Linopteris obliqua (BUNBURY) ZEILLER
Lonchopteris longepinnata JONGMANS (nomen nudum)
Dicksonites potieri (ZEILLER) P. BERTRAND
Eusphenopteris sp. (cf. *leonardi* P. BERTRAND)
Sphenopteris cf. *goldenbergi* ANDRAE
Sphenopteris spp.
Alloiopteris cristata (VON GUTBIER) NĚMEJC
Pecopteris (Zeilleria) avoldensis (STUR) SIMSON-SCHAROLD
Pecopteris acuta BRONGNIART
Pecopteris cf. *bioti* BRONGNIART
Pecopteris dentata BRONGNIART
Pecopteris cf. *saraefolia* P. BERTRAND
Pecopteris unita BRONGNIART
Pecopteris sp.
Sphenophyllum sp. (cf. *kidstoni* HEMINGWAY)
Annularia radiata BRONGNIART
Annularia sphenophylloides (ZENKER) VON GUTBIER
Annularia stellata (VON SCHLOTHEIM) WOOD
Asterophyllites equisetiformis (VON SCHLOTHEIM) BRONGNIART
Sigillaria brardi BRONGNIART
Calamites cf. *schuetzei* STUR
Lepidostrobophyllum cf. *lanceolatum* (LINDLEY & HUTTON) CHALONER
Cordaianthus jongmansi FLORIN
Cordaites sp.

cannot be taken at face value. The only species of Stephanian or late Westphalian D age which seems to have been figured correctly, is *Pseudomariopteris ribeyroni* (ZEILLER) DANZÉ-CORSIN (op. cit., pl. I, figs 6, 6a), and even this specimen is not wholly typical. The present writer has repeatedly expressed his reservations about the Stephanian age of the highest strata in the Central Asturian Coalfield (compare WAGNER 1971a, p. 468 — footnote), and these doubts were confirmed after collecting floral remains from the very highest measures at Pajomal (WAGNER & LAVEINE *in* LAVEINE 1977, p. 81). The assemblage from these highest measures was compared with the floral composition of mid-Westphalian D strata between tonsteins 60 and 40 in the succession at Lorraine (LAVEINE 1977).

In the northwestern part of the Central Asturian Coalfield, a probably less complete succession of Westphalian D coal-measures exist. The highest beds in this area are the Olloniego, Ablanedo and Loredo formations (PELLO & CORRALES 1971). The fossil flora of the Olloniego Formation has been figured and described by WAGNER (1971b) and various assemblages from the Ablanedo Formation have been listed by JONGMANS & WAGNER (1957). The detailed description of the Ablanedo flora is still outstanding, but a revised list of most of the floral elements present in the Ablanedo Formation is given here (Table 4). The age is mid-Westpha-

Table 5 — Composite list of upper Westphalian D flora from the post-Leonian basin in northern Palencia, Cantabrian Mts, Northwest Spain (after WAGNER & VARKER 1971 and WAGNER in press a)

Neuropteris ovata HOFFMANN
Neuropteris planchardi ZEILLER
Neuropteris scheuchzeri HOFFMANN
Odontopteris cantabrica WAGNER (very rare)
Linopteris obliqua (BUNBURY) ZEILLER (rare)
Linopteris palentina WAGNER
Alethopteris ambigua LESQUEREUX
Alethopteris grandinioides KESSLER var. *grandinioides*
Alethopteris lesquereuxi WAGNER
Alethopteris missouriensis WHITE
Alethopteris palentina WAGNER (very rare)
Alethopteris robusta LESQUEREUX
Callipteridium (Praecallipteridium) jongmansi (P. BERTRAND) WAGNER
Mariopteris nervosa (BRONGNIART) ZEILLER
Pseudomariopteris ribeyroni (ZEILLER) DANZÉ-CORSIN (rare)
Dicksonites plueckeneti (VON SCHLOTHEIM) STERZEL
Eusphenopteris neuropteroides (BOULAY) NOVIK
Oligocarpia gutbieri GÖPPERT
Pecopteris (Oligocarpia?) bredovi GERMAR
Polymorphopteris polymorpha (BRONGNIART) WAGNER
Lobatopteris vestita (LESQUEREUX) WAGNER
Pecopteris hemitelioides BRONGNIART (rare)
Pecopteris monyi ZEILLER (rare)
Pecopteris ocejensis WAGNER (= *P. nyranensis* NĚMEJC ?)
Pecopteris raconensis NĚMEJC
Pecopteris unita BRONGNIART
Sphenophyllum emarginatum BRONGNIART forma *truncatum* SCHIMPER
Annularia sphenophylloides (ZENKER) VON GUTBIER
Annularia stellata (VON SCHLOTHEIM) WOOD
Asterophyllites equisetiformis (VON SCHLOTHEIM) BRONGNIART

lian D, corresponding most likely to a position at or just below the mid-Westphalian D floral change which LAVEINE (1977) puts at Tonstein 60 in Lorraine. The Ablanedo flora is probably marginally older than the flora from Ervedosa in the Oporto region. Seen in conjunction with the Olloniego flora, there are certain resemblances in the presence of *Linopteris obliqua* (which is very common in the Central Asturian Coalfield), *Callipteridium jongmansi* (recorded as *C. armasi* in WAGNER 1971b — see comments in LAVEINE *et al.* 1977) and *Lobatopteris micromiltoni* (which is a common species in the Esperanza Formation which precedes the Olloniego Formation). *Pecopteris unita* also occurs in the Ablanedo flora as well as in Ervedosa. However, *Dicksonites plueckeneti* and *Polymorphopteris polymorpha* which are apparently present at Ervedosa but not a Albanedo, impart a slightly younger aspect to the Ervedosa flora. Unfortunately, the former is not well figured (TEIXEIRA 1942) and the latter is identified only with cf. The similar species, *Dicksonites potieri*, occurs at Ablanedo and has in fact been recorded as *Dicksonites plueckeneti* by JONGMANS & WAGNER (1957).

The flora from Ervedosa should also be compared with those found in the upper Westphalian D strata in the lower part of the succession of the post-Leonian basin in the southeastern corner of the Cantabrian Mts, Northwest Spain (Table 5). The latter are clearly later in age than those recorded from the Olloniego and Ablanedo formations (though not necessarily later than the flora of Pajomal referred to earlier), and undoubtedly correspond to a level above the mid-Westphalian D floral change (compare LAVEINE 1977). A reasonably good comparison exists with the flora from Ervedosa, but it is noted that *Linopteris obliqua* which is common at Olloniego-Ablanedo and which also exists at Ervedosa, is virtually replaced by *Linopteris palentina* in the late Westphalian D floras of the post-Leonian basin. *Linopteris obliqua* still exists but is rarely encountered. The latter apparently persists into the very highest Westphalian D in the British Isles (CLEAL 1978) and in Lorraine (LAVEINE 1977). *Polymorphopteris polymorpha* is already common in the upper Westphalian D strata of the post-Leonian basin, and the very occasional presence of such Stephanian species as *Odontopteris cantabrica* and *Neuropteris planchardi* is also noted. The latter probably does not possess great value as a stratigraphic indicator since it is a mesophilous element, but its occurrence in the late Westphalian D floras of the post-Leonian basin is the lowest on record. *Odontopteris cantabrica* has also been recorded from the highest Westphalian D in Lorraine (LAVEINE, *in* Discussion of WAGNER 1972). Altogether, it appears that the flora from Ervedosa is a little older than the late Westphalian D floras of the post-Leonian basin in the Cantabrian Zone of Northwest Spain.

1.2. **Stephanian of Oporto region**

The lower Stephanian C strata of the Douro Coalfield (Table 6) are also non-marine. Its detailed stratigraphy is not well known, but the coal-bearing strata are floored by a coarse conglomerate and also contain a laterally continuous conglomerate inside the sucession. A basin of comparable age in the Central Iberian Zone is the Puertollano Coalfield in the province of Ciudad Real (La Mancha). The latter contains essentially lacustrine deposits, including oil shales, and it also shows the influence of acid volcanism, with agglomerates and tuffs as well as a strong tuffaceous element in the mudstones and siltstones of the lower part of the succession. The Puertollano Basin is entirely nonmarine. Its age has been determined as late Stephanian B rather than early Stephanian C (WAGNER & UTTING 1967) (Table 7).

The floral assemblages of the Douro and Puertollano basins are comparable in that conifer remains have been reported from both areas. This may be regarded as indicative of non-depositional areas in the near vicinity. Only one conifer species has been seen at Puertollano, viz. *Lebachia piniformis* (VON SCHLOTHEIM), whereas FLORIN (1940) reported *Lebachia parvifolia* FLORIN, cf. *Lebachia frondosa* var. *zeilleri* FLORIN and *Ernestiodendron filiciforme* (VON

Table 6 — Composite list of the lower Stephanian C flora of the Douro Basin, Oporto region, Portugal (after WAGNER & SOUSA 1983)

Ernestiodendron filiciforme (VON SCHLOTHEIM *pars*) FLORIN
cf. *Lebachia frondosa* var. *zeilleri* FLORIN
Lebachia parvifolia FLORIN
Neuropteris cordata BRONGNIART
Neuropteris gallica ZEILLER
Neuropteris ovata var. *pseudovata* GOTHAN & SZE
Neuropteris planchardi ZEILLER
Neuropteris zeilleri LIMA
Neuropteris sp.
Reticulopteris germari (GIEBEL) GOTHAN
Linopteris neuropteroides (VON GUTBIER) POTONIÉ
Odontopteris brardi BRONGNIART
Lescuropteris genuina (GRAND'EURY) REMY
Callipteridium (Eucallipteridium) gigas (VON GUTBIER) WEISS
Callipteridium (Eucallipteridium) zeilleri WAGNER
Alethopteris zeilleri RAGOT
Pseudomariopteris cf. *busqueti* (ZEILLER) DANZÉ-CORSIN
Pseudomariopteris corsini (TEIXEIRA) WAGNER
Pseudomariopteris ribeyroni (ZEILLER) DANZÉ-CORSIN
Dicksonites plueckeneti (VON SCHLOTHEIM) STERZEL
Eusphenopteris rotundiloba (NĚMEJC) VAN AMEROM
Sphenopteris cf. *cremeriana* POTONIÉ
Sphenopteris cf. *germanica* WEISS
Sphenopteris cf. *lenis* ZEILLER
Sphenopteris matheti ZEILLER
Sphenopteris mendescorreae TEIXEIRA
Sphenopteris sp. cf. *chaerophylloides* BRONGNIART
Sphenopteris sp.
Oligocarpia leptophylla (BUNBURY) GRAUVOGEL-STAMM & DOUBINGER
Alloiopteris sp.
Gondomaria discreta (WEISS) WAGNER & LEMOS DE SOUSA
Nemejcopteris feminaeformis (VON SCHLOTHEIM) BARTHEL
Polymophopteris polymorpha (BRONGNIART) WAGNER
Lobatopteris viannae (TEIXEIRA) WAGNER
Pecopteris cf. *ameromi* STOCKMANS & WILLIÈRE
Pecopteris candolliana BRONGNIART
Pecopteris cyathea (VON SCHLOTHEIM) BRONGNIART
Pecopteris daubreei ZEILLER
Pecopteris densifolia GÖPPERT
Pecopteris gruneri ZEILLER
Pecopteris limae TEIXEIRA
Pecopteris cf. *melendezi* WAGNER
Pecopteris monyi ZEILLER
Pecopteris unita BRONGNIART
Pecopteris sp. nov. (cf. *hemitelioides* BRONGNIART)
Pecopteris sp.
Rhacopteris gomesiana (HEER) TEIXEIRA
Taeniopteris bertrandiana TEIXEIRA
Taeniopteris jejunata GRAND'EURY
cf. *Taeniopteris multinervis* WEISS
Sphenophyllum alatifolium RENAULT
Sphenophyllum costae STERZEL

Table 6 — *(Cont.)*

Sphenophyllum longifolium (GERMAR) UNGER
Sphenophyllum oblongifolium (GERMAR & KAULFUSS) UNGER
Sphenophyllum cf. *thoni* var. *minor* STERZEL
Annularia sphenophylloides (ZENKER) VON GUTBIER
Annularia stellata (VON SCHLOTHEIM) WOOD
Asterophyllites equisetiformis (VON SCHLOTHEIM) BRONGNIART
Equisetites zeaformis ANDRAE
Sigillaria brardi BRONGNIART
Sigillariostrobus serreatus TEIXEIRA
Cyperites sp.
Calamites carinatus STERNBERG
Calamites schuetzeiformis forma *waldenburgensis* KIDSTON
Calamites suckowii BRONGNIART
Macrostachya carinata (GERMAR) ZEILLER
Calamostachys tuberculata STERNBERG
Cordaites sp.
Lepidophylloides sp.
Dicranophyllum gallicum GRAND'EURY
Dicranophyllum lusitanicum (HEER) LIMA
Seeds

SCHLOTHEIM *pars* FLORIN from the Douro Coalfield. On the other hand, there is a major difference in the relative absence of medullosan pteridosperm taxa in the Puertollano Coalfield which shows only *Neuropteris ovata* in moderate abundance, and *Neuropteris* cf. *auriculata* and *Alethopteris zeilleri* as single specimens. *Callipteridium* is also rare, and *Linopteris* and *Odontopteris* have not been seen in Puertollano. The medullosan seed ferns are strongly represented in the Douro Coalfield. Other seed ferns are also better represented in the Douro Basin. True ferns are probably just as well represented in both areas, and so are the sphenophytes. However, the Douro Basin shows the presence of *Taeniopteris*, which has not been seen in Puertollano, and the rather special lycophytes *Dicranophyllum gallicum* and *D. lusitanicum*. The comparison is therefore not particularly close. This may be due to the intramontane character of both basins which are likely to have had local restrictions on the floral composition as a result of limited migration potential. Local endemism can also be expected and the two species of *Dicranophyllum* in the Douro Basin may be a case in point.

Comparison should also be made with the floral composition of lower Stephanian C deposits in the various post-Asturian coalfields of the Cantabrian Zone. These are the areas of El Bierzo (Table 8), Villablino (Table 9), Tormaleo, Rengos, Carballo, Cangas de Narcea, Tineo (Table 10), Arnao, Ferroñes and Pico Cordel. These coalfields all belong to a coastal basin on the edge of a hinterland area which corresponds more or less to the West Asturian — Leonese Zone (Fig. 1). The seaward side of this basin lies in the concavity of the Asturian arcuate fold belt, the evidence being found in the area of the Picos de Europa (MARTINEZ & WAGNER 1982). The composition of the early Stephanian C floras in these coalfields is always very similar. Examples are given in Tables 8-10. There are obvious differences with the composition of the flora in the Douro Basin, differences which are independent of stratigraphic age. Three different species of *Lebachia* and *Ernestiodendron* have been found in the Douro Basin. These extrabasinal elements are notably absent in the coastal basin of the Cantabrian Zone which has yielded no conifers apart from a single record from Tineo by ZEILLER (1882), who mentioned *Walchia* (now *Lebachia*) *pini-*

Table 7 — Composite list of upper Stephanian B flora from Puertollano, Province of Ciudad Real (La Mancha), Central Spain

Lebachia piniformis (Von Schlotheim) Florin
Neuropteris cf. *auriculata* Brongniart
Neuropteris ovata Hoffmann
Callipteridium zeilleri Wagner
Alethopteris zeilleri Ragot
Eusphenopteris rotundiloba (Němejc) Van Amerom
Sphenopteris biturica Zeiller
Oligocarpia grigorievi (Zalessky & Tchirkova) Wagner
Oligocarpia leptophylla (Bunbury) Grauvogel-Stamm & Doubinger
Nemejcopteris feminaeformis (Von Schlotheim) Barthel
Polymorphopteris polymorpha (Brongniart) Wagner
Polymorphopteris villablinensis Wagner
Lobatopteris corsini Wagner
Lobatopteris lamuriana (Heer) Wagner
Pecopteris ameromi Stockmans & Willière
Pecopteris arborescens (Von Schlotheim) Brongniart
Pecopteris cf. *densifolia* Göppert
Pecopteris gruneri Zeiller
Pecopteris jongmansi Wagner
Pecopteris monyi Zeiller
Pecopteris cf. *paleacea* Zeiller
Pecopteris unita Brongniart
Pecopteris sp.
Sphenophyllum costae Serzel
Sphenophyllum oblongifolium (Germar & Kaulfuss) Unger
Sphenophyllum thoni var. *minor* Sterzel
Sphenophyllum verticillatum (Von Schlotheim) Zeiller
Annularia sphenophylloides (Zenker) Von Gutbier
Annularia stellata (Von Schlotheim) Wood
Asterophyllites equisetiformis (Von Schlotheim) Brongniart
Lepidostrobophyllum hastatum (Lesquereux) Chaloner
Macrostachya carinata (Germar) Zeiller
Sigillaria brardi Brongniart
Sigillariostrobus sp.

formis. The absence of additional records of conifers in the Cantabrian Zone is particularly noticeable because of the exhaustive collections made in this area over the last thirty years (tens of thousands of specimens). Furthermore, although all the medullosan pteridosperms of the Douro Basin also occur in the Stephanian B-C coastal basins of the Cantabrian Zone (with the exception of *Neuropteris zeilleri*), there are certain species which are excessively rare in the latter. These are *Neuropteris cordata, Neuropteris gallica, Neuropteris planchardi, Reticulopteris germari,* and *Lescuropteris genuina.* All these species are mentioned as common in the composite list presented by Teixeira (1944, 1945). It is no coincidence that these are precisely the species regarded as not belonging to floodpain assemblages (compare Bouroz & Wagner 1972). A further difference is the presence of *Dicranophyllum gallicum* and *D. lusitanicum* in the Douro Basin. The latter is marked as common by Teixeira (1944, 1945). Only *Dicranophyllum gallicum* has been observed in the upper Stephanian of Northwest Spain, and that only once (Stockmans & Willière 1965). *Taeniopteris jejunata* is marked as common in the Douro Basin (Teixeira 1944, 1945), whereas it is rather uncommon in Northwest Spain. It is thus fairly obvious that extrabasinal (xerophilous and mesophilous) elements are more common in the Douro Basin than they are in the upper Stephanian of Northwest Spain. This suggests higher ground with better drained soils in the vicinity of the depositional area of the Douro Basin, which is consistent with the absence of marine influences and the probability that this represents an intramontane basin.

Table 8 — Composite list of lower Stephanian C plant fossils from El Bierzo, northwestern León, Northwest Spain

Neuropteris cf. *eveni* LESQUEREUX
Neuropteris ovata HOFFMANN
Odontopteris brardi BRONGNIART
Mixoneura matallanae WAGNER
Callipteridium striatum-zeilleri WAGNER
Alethopteris bohemica FRANKE
Alethopteris leonensis WAGNER (= *A. virginiana* FONTAINE & WHITE ?)
Alethopteris zeilleri RAGOT
Pseudomariopteris busqueti (ZEILLER) DANZÉ-CORSIN
Pseudomariopteris corsini (TEIXEIRA) WAGNER
Pseudomariopteris ribeyroni (ZEILLER) DANZÉ-CORSIN
cf. *Karinopteris paleaui* (ZEILLER) BOERSMA
Dicksonites leptophylla (ZEILLER) DOUBINGER
Dicksonites plueckeneti (VON SCHLOTHEIM) STERZEL
Eusphenopteris rotundiloba (NĚMEJC) VAN AMEROM
Palmatopteris cf. *furcata* (BRONGNIART) POTONIÉ
Sphenopteris biturica ZEILLER
Sphenopteris biturica forma *densipennata* ALVAREZ-RAMIS
Sphenopteris cristata BRONGNIART
Sphenopteris doubingeri ALVAREZ-RAMIS
Sphenopteris durbanensis CORSIN & VILA
Sphenopteris elaverica (ZEILLER) ALVAREZ-RAMIS
Sphenopteris cf. *lenis* ZEILLER
Sphenopteris cf. *matheti* ZEILLER
Sphenopteris melendezi ALVAREZ-RAMIS
Sphenopteris mendescorreae TEIXEIRA (with fructifications)
Sphenopteris minutisecta FONTAINE & WHITE
Sphenopteris cf. *ovalis* ANDRAE
Sphenopteris picandeti ZEILLER
Sphenopteris cf. *wagneri* ALVAREZ-RAMIS
Oligocarpia gutbieri GÖPPERT
Oligocarpia leptophylla (BUNBURY) GRAUVOGEL-STAMM & DOUBINGER
Oligocarpia cf. *pulcherrima* STUR
Nemejcopteris feminaeformis (VON SCHLOTHEIM) BARTHEL
Polymorphopteris polymorpha (BRONGNIART) WAGNER
Polymorphopteris subelegans (POTONIÉ) sensu WAGNER
Polymorphopteris villablinensis WAGNER
Lobatopteris corsini WAGNER
Pecopteris ameromi STOCKMANS & WILLIÈRE
Pecopteris cf. *apicalis* KNIGHT
Pecopteris cf. *candolliana* BRONGNIART
Pecopteris daubreei ZEILLER
Pecopteris densifolia GÖPPERT
Pecopteris gruneri ZEILLER
Pecopteris hemitelioides BRONGNIART

Table 8 *(Cont.)*

Pecopteris integra ANDRAE
Pecopteris jongmansi WAGNER
Pecopteris cf. *lepidorachis* BRONGNIART
Pecopteris monyi ZEILLER
Pecopteris robustissima WAGNER
Fascipteris sp.
Taeniopteris jejunata GRAND'EURY
Sphenophyllum alatifolium RENAULT
Sphenophyllum angustifolium (GERMAR) UNGER
Sphenophyllum costae STERZEL
Sphenophyllum crenulatum KNIGHT
Sphenophyllum longifolium GERMAR
Sphenophyllum oblongifolium (GERMAR & KAULFUSS) UNGER
Annularia sphenophylloides (ZENKER) VON GUTBIER
Annularia stellata (VON SCHLOTHEIM) WOOD
Asterophyllites equisetiformis (VON SCHLOTHEIM) BRONGNIART
Lepidostrobophyllum hastatum (LESQUEREUX) CHALONER
Sigillaria brardi BRONGNIART
Sigillaria cf. *ovata* SAUVEUR
cf. *Asolanus camptotaenia* WOOD

Table 9 — Composite list of lower Stephanian C plant fossils from the Calderón, Paulina and María Bolsada formations, Villablino Coalfield, northwestern León, Cantabrian Mts, Northwest Spain (after WAGNER 1964, and ALVAREZ-RAMIS 1965, with some modifications)

Neuropteris ovata var. *grandeuryi* WAGNER
Mixoneura neuropteroides sensu ZEILLER
Linopteris neuropteroides (VON GUTBIER) POTONIÉ
Odontopteris brardi BRONGNIART
Callipteridium (Eucallipteridium) gigas (VON GUTBIER) WEISS
Callipteridium (Eucallipteridium) zeilleri WAGNER
Alethopteris leonensis WAGNER (= *A. virginiana* FONTAINE & WHITE ?)
Pseudomariopteris busqueti (ZEILLER) DANZÉ-CORSIN
Pseudomariopteris corsini (TEIXEIRA) WAGNER
Pseudomariopteris ribeyroni (ZEILLER) DANZÉ-CORSIN
Dicksonites plueckeneti (VON SCHLOTHEIM) STERZEL
Sphenopteris beyschlagi POTONIÉ
Sphenopteris biturica ZEILLER
Sphenopteris biturica densipennata ALVAREZ-RAMIS
Sphenopteris cremeriana POTONIÉ
Sphenopteris cristata BRONGNIART
Sphenopteris decheni WEISS
Sphenopteris dimorpha (LESQUEREUX) BARTHEL
Sphenopteris doubingeri ALVAREZ-RAMIS
Sphenopteris fossorum ZEILLER
Sphenopteris lebachensis WEISS
Sphenopteris melendezi ALVAREZ-RAMIS
Sphenopteris rotundiloba NĚMEJC

Table 9 *(Cont.)*

Sphenopteris wagneri ALVAREZ-RAMIS
Sphenopteris aff. *weissi* POTONIÉ
Sphenopteris sp.
Oligocarpia grigorievi (ZALESSKY & TCHIRKOVA) WAGNER
Oligocarpia leptophylla (BUNBURY) GRAUVOGEL-STAMM & DOUBINGER
Nemejcopteris feminaeformis (VON SCHLOTHEIM) BARTHEL
Polymorphopteris polymorpha (BRONGNIART) WAGNER
Polymorphopteris subelegans (POTONIÉ) sensu WAGNER
Lobatopteris corsini WAGNER
Lobatopteris serpentigera WAGNER
Pecopteris acuta-dentata BRONGNIART
Pecopteris arborescens (VON SCHLOTHEIM) BRONGNIART
Pecopteris integra ANDRAE
Pecopteris jongmansi WAGNER
Pecopteris subcrenulata LESQUEREUX
Pecopteris unita BRONGNIART
Pecopteris spp.
Fascipteris hispanica (WAGNER) comb. nov.
Taeniopteris jejunata GRAND'EURY
Taeniopteris multinervis WEISS
Sphenophyllum angustifolium (GERMAR) UNGER
Sphenophyllum longifolium (GERMAR) UNGER
Sphenophyllum oblongifolium (GERMAR & KAULFUSS) UNGER
Sphenophyllum thoni VON MAHR
Annularia stellata (VON SCHLOTHEIM) WOOD
Asterophyllites equisetiformis (VON SCHLOTHEIM) BRONGNIART
Sigillaria brardi BRONGNIART
Pholidophloios sp.

Table 10 — Composite list of lower Stephanian C plant fossils from Tineo (Asturias), Cantabrian Mts, Northwest Spain (after WAGNER 1965, ALVAREZ-RAMIS 1965, WAGNER & ALVAREZ-RAMIS 1967, and ALVAREZ-RAMIS & DOUBINGER 1970)

Neuropteris ovata HOFFMANN
Neuropteris praedentata GOTHAN
Linopteris neuropteroides (VON GUTBIER) POTONIÉ
Odontopteris brardi BRONGNIART
Callipteridium (Eucallipteridium) zeilleri WAGNER
Alethopteris leonensis WAGNER (= *A. virginiana* FONTAINE & WHITE ?)
Alethopteris zeilleri RAGOT
Mariopteris cantabrica WAGNER & ALVAREZ-RAMIS
Mariopteris melendezi WAGNER & ALVAREZ-RAMIS (= *Karinopteris paleaui* (ZEILLER) BOERSMA ?)
Pseudomariopteris busqueti (ZEILLER) DANZÉ-CORSIN
Pseudomariopteris ribeyroni (ZEILLER) DANZÉ-CORSIN
Karinopteris paleaui (ZEILLER) BOERSMA
Dicksonites leptophylla (ZEILLER) DOUBINGER
Dicksonites plueckeneti (VON SCHLOTHEIM) STERZEL
Sphenopteris cf. *alabamensis* LESQUEREUX

Table 10 *(Cont.)*

Sphenopteris cf. *beyrichi* (WEISS) ALVAREZ-RAMIS
Sphenopteris biturica ZEILLER
Sphenopteris casteli ZEILLER
Sphenopteris cristata BRONGNIART
Sphenopteris cf. *devians* GOTHAN
Sphenopteris elaverica (ZEILLER) ALVAREZ-RAMIS
Sphenopteris goldenbergi ANDRAE
Sphenopteris kidstoni ZEILLER
Sphenopteris matheti ZEILLER
Sphenopteris minutisecta FONTAINE & WHITE
Sphenopteris ovalis GUTBIER
Sphenopteris rotundiloba NĚMEJC
Oligocarpia leptophylla (BUNBURY) GRAUVOGEL-STAMM & DOUBINGER
Nemejcopteris feminaeformis (VON SCHLOTHEIM) BARTHEL
Polymorphopteris polymorpha (BRONGNIART) WAGNER
Polymorphopteris subelegans (POTONIÉ) sensu WAGNER
Lobatopteris corsini WAGNER
Lobatopteris viannae (TEIXEIRA) WAGNER
Pecopteris acuta-dentata BRONGNIART
Pecopteris arborescens (VON SCHLOTHEIM) BRONGNIART
Pecopteris daubreei ZEILLER
Pecopteris cf. *dawsoniana* (FONTAINE & WHITE) WAGNER
Pecopteris dentata BRONGNIART
Pecopteris hemitelioides BRONGNIART
Pecopteris cf. *hucheti* CORSIN
Pecopteris jongmansi WAGNER
Pecopteris longiphylla CORSIN
Pecopteris cf. *odontopteroides* (FONTAINE & WHITE) FRANKE
Pecopteris oreopteridea (VON SCHLOTHEIM) BRONGNIART
Pecopteris paleacea ZEILLER
Pecopteris cf. *pectinata* P. BERTRAND
Pecopteris unita BRONGNIART
Pecopteris spp.
Fascipteris sp.
Desmopteris cf. *robusta* DOUBINGER
Taeniopteris jejunata GRAND'EURY
Sphenophyllum angustifolium (GERMAR) UNGER
Sphenophyllum oblongifolium (GERMAR & KAULFUSS) UNGER
Sphenophyllum thoni VON MAHR
Sphenophyllum sp.
Annularia stellata (VON SCHLOTHEIM) WOOD
Pholidophloios sp. nov. ?
Lepidostrobophyllum sp.
Calamites gigas BRONGNIART
Calamites multiramis WEISS
Calamites sp.
Calamostachys tuberculata STERNBERG
Equisetites sp.
Cordaites palmaeformis GÖPPERT
Cordaites sp.
Seeds

The various species of *Sphenophyllum* are virtually identical in the Douro Basin and in the Douro Basin and in the Cantabrian Zone of Northwest Spain. Minor differences probably relate to the larger number of collecting sites in Northwest Spain, but it is also possible that these floodplain elements are slightly more varied in this area. *Pseudomariopteris, Dicksonites* and *Eusphenopteris* are represented almost identically in the two different areas. A major difference however appears in the much more varied composition of sphenopterid and pecopterid ferns in the upper Stephanian of Northwest Spain. Sphenopterids do not occur very commonly in either the Douro Basin or the Cantabrian Zone, but they are notably more diversified in the latter area. This cannot be explained entirely by the larger number of collecting sites in Northwest Spain. The extremely common pecopterids, which form the bulk of the flora in both areas, show a much larger number of species in the Cantabrian Zone. Most likely, this reflects better migration routes and the consequent intermingling of species in the latter area (in other words, a lack of endemism).

The floral composition of the lower Stephanian C flora of the coastal basin in the Cantabrian Zone is therefore markedly different to that of the Douro Basin which is likely to have been intramontane.

The same differences have been observed between the 'classical' (late) Stephanian floras of the Massif Central in France and the Stephanian B-C floras of the Cantabrian Zone in Northwest Spain (WAGNER 1971a). The basins of the Massif Central are mainly intramontane depressions, including some graben fills.

2. BUÇACO BASIN

This basin which occurs alongside the Porto--Coimbra-Badajoz-Córdoba Fault (Fig. 1), probably the most important early fracture zone in the Iberian Peninsula, shows all the hallmarks of rapid deposition and erosion of a nearby topographic relief. It seems likely that the fracture zone experienced vertical movements at the time of deposition of the very late Stephanian C strata of the Buçaco Basin and that the subsequent deformation into a tight, almost isoclinal syncline with a sheared western flank is also related to such vertical movements (WAGNER *et al.* in prep.; DOMINGOS *et al.* 1983 — this volume). The presence of a nearby topographic relief and tectonic mobility are both evident from the nature of the sediments laid down in the Buçaco Basin and their stratigraphic succession. A fanglomerate linked to palaeotopography is followed by almost 200 m of mudflow deposits and debris flow conglomerates in red beds facies, about 40 m of grey measures representing floodplain deposition (and forming a gradual transition with the red beds), and a long succession (in excess of 600 m) of fluviatile conglomeratic measures which probably represent braided streams in a very high energy regime. Plant fossils (Table 11) are virtually restricted to the relatively thin interval of grey measures.

Table 11 — Composite list of the upper Stephanian C (or lowermost Autunian) flora of the Buçaco Basin, Portugal (after WAGNER & SOUSA 1983)

Lebachia goeppertiana FLORIN
Lebachia laxifolia FLORIN
Lebachia parvifolia FLORIN
Callipteris conferta (STERNBERG) BRONGNIART
Neuropteris neuropteroides (GÖPPERT) BARTHEL
Neuropteris planchardi ZEILLER
Neuropteris praedentata GOTHAN
Neuropteris zeilleri LIMA
Neuropteris sp.

Table 11 *(Cont.)*

Reticulopteris germari (GIEBEL)
Linopteris gangamopteroides (DE STEFANI) WAGNER
Odontopteris brardi BRONGNIART
Mixoneura sp. cf. *Odontopteris osmundaeformis* (VON SCHLOTHEIM) ZEILLER
Lescuropteris genuina (GRAND'EURY) REMY
Callipteridium densinervium WAGNER
Callipteridium gigas (VON GUTBIER) WEISS
Alethopteris schneideri STERZEL
Alethopteris zeilleri RAGOT
Pseudomariopteris busqueti (ZEILLER) DANZÉ-CORSIN
Pseudomariopteris ribeyroni (ZEILLER) DANZÉ-CORSIN
Dicksonites leptophylla (ZEILLER) DOUBINGER
Dicksonites plueckeneti (VON SCHLOTHEIM) STERZEL
Eusphenopteris cf. *rotundiloba* (NĚMEJC) VAN AMEROM
Sphenopteris cf. *biturica* ZEILLER
Sphenopteris casteli ZEILLER
Sphenopteris cremeriana POTONIÉ
Sphenopteris elaverica (ZEILLER) ALVAREZ-RAMIS
Sphenopteris cf. *minutisecta* FONTAINE & WHITE, non ALVAREZ-RAMIS
Sphenopteris sp.
Oligocarpia leptophylla (BUNBURY) GRAUVOGEL-STAMM & DOUBINGER
Alloiopteris sp.
Nemejcopteris feminaeformis (VON SCHLOTHEIM) BARTHEL
Polymorphopteris polymorpha (BRONGNIART) WAGNER
Lobatopteris corsini WAGNER
Lobatopteris viannae (TEIXEIRA) WAGNER
Pecopteris arborescens (VON SCHLOTHEIM) BRONGNIART
Pecopteris (Oligocarpia?) cf. *bredovii* GERMAR
Pecopteris bussacensis TEIXEIRA
Pecopteris cf. *candolliana* BRONGNIART
Pecopteris cyathea (VON SCHLOTHEIM) BRONGNIART
Pecopteris densifolia GÖPPERT
Pecopteris gruneri ZEILLER
Pecopteris monyi ZEILLER
Pecopteris unita BRONGNIART
Pecopteris sp. nov.
Pecopteris sp.
Taeniopteris jejunata GRAND'EURY
Taeniopteris multinervis WEISS
Sphenophyllum angustifolium (GERMAR) UNGER
Sphenophyllum costae STERZEL
Sphenophyllum oblongifolium (GERMAR & KAULFUSS) UNGER
Sphenophyllum thoni VON MAHR
Annularia sphenophylloides (ZENKER) VON GUTBIER
Annularia stellata (VON SCHLOTHEIM) WOOD
Asterophyllites equisetiformis (VON SCHLOTHEIM) BRONGNIART
Asterophyllites longifolius STERNBERG
Asolanus camptotaenia WOOD
Calamites suckowii BRONGNIART
Calamostachys tuberculata STERNBERG
Cordaites sp.
Dicranophyllum sp.
Cordaianthus sp.
Hexagonocarpus sp. and other seeds

Classical 'flöznah' and 'flözfremd' assemblages have been recorded from this interval (WAGNER *et al.* in prep.; WAGNER & SOUSA 1983 — this volume), very much on the same lines as recorded from the Rotliegend of Saxony and Thuringia (compare GOTHAN & GIMM 1930; BARTHEL 1976). Extrabasinal elements (HAVLENA 1970; PFEFFERKORN 1980) derived from better drained soils (probably hill slopes) in the immediate vicinity of the basin of sedimentation, are the conifers (three species of *Lebachia*), *Callipteris conferta,* several species of *Neuropteris, Reticulopteris germari, Mixoneura, Alethopteris schneideri, Taeniopteris* and *Dicranophyllum.* As is the case for the Douro Basin, there is a relatively low diversity of floodplain elements such as the sphenopterid and pecopterid ferns. The floral composition thus agrees with the sedimentary characteristics of the Buçaco Basin in supposing this to have been intramontane. DOUBINGER (1956) had no difficulty in comparing with the floras of Igornay (Autun Basin), Bert and Brive of the Massif Central, and attributed the assemblage to a 'Stephanian D' which BOUROZ & DOUBINGER (1977) later defined in chronostratigraphic terms (as against the original biostratigraphic concept) as comprising strata formerly belonging to the highest Stephanian C and the lowermost Autunian. This stage has not been sanctioned by the IUGS Subcommission on Carboniferous Stratigraphy (it impinges on both Stephanian C and Autunian in their general acceptance) and the concept should probably remain biostratigraphical. WAGNER (in press b) has proposed to recognise DOUBINGER's Stephanian D as the *Callipteris conferta* Zone (which is characterised by *C. conferta* without additional species of *Callipteris* and in the general context of a Stephanian flora).

The only comparable floral assemblage in the western part of the Iberian Peninsula is at the locality of San Tirso, near Mieres, in the Asturias (GERVILLA *et al.* 1978). This locality shows unconformable upper Stephanian C strata at the base of a Permian succession with a strong volcanic component, and dates the lowermost part of the post-Hercynian cover in the Cantabrian Zone. The floral remains are being studied by WAGNER and LAVEINE who published a provisional list (Table 12) in WAGNER & MARTINEZ (1982, p. 276). They constitute a small assemblage, mainly consisting of apparent

Table 12 — List of plant fossils from San Tirso, near Mieres, central Asturias, Cantabrian Mts, Northwest Spain (after WAGNER and LAVEINE *in* WAGNER & MARTINEZ 1982)

Neuropteris neuropteroides (GÖPPERT) BARTHEL
Neuropteris praedentata GOTHAN
Neuropteris cf. *zeilleri* LIMA
Odontopteris brardi BRONGNIART
Callipteridium cf. *gigas* (VON GUTBIER) WEISS
Pseudomariopteris busqueti (ZEILLER) DANZÉ-CORSIN
Pseudomariopteris ribeyroni (ZEILLER) DANZÉ-CORSIN
Sphenopteris cf. *lescuriana* FONTAINE & WHITE
Sphenopteris pachypteroides FONTAINE & WHITE
Sphenopteris sp. nov. ?
pecopteris spp.
Sphenophyllum oblongifolium (GERMAR & KAULFUSS) UNGER
Annularia sphenophylloides (ZENKER) VON GUTBIER
Annularia stellata (VON SCHLOTHEIM) WOOD
Cordaites sp.

floodplain elements but containing also *Neuropteris neuropteroides* and *Neuropteris* cf. *zeilleri* which are regarded as mesophilous.

There is no real comparison between the palaeogeographic relationships of the Buçaco Basin and those of the San Tirso occurrence

Table 13 — Composite list of the uppermost Westphalian D flora of the Santa Susana Basin, Portugal (after WAGNER & SOUSA 1983)

Neuropteris flexuosa STERNBERG
Neuropteris scheuchzeri HOFFMANN
Linopteris palentina WAGNER
Callipteridium (Praecallipteridium) jongmansi (P. BERTRAND) WAGNER
Alethopteris lesquereuxi WAGNER
Dicksonites plueckeneti (VON SCHLOTHEIM) STERZEL
Mariopteris rotundata HUTH
Eusphenopteris nummularia (VON GUTBIER) NOVIK
Eusphenopteris trigonophylla (BEHREND) VAN AMEROM
Sphenopteris alentejana TEIXEIRA
Sphenopteris (Palmatopteris ?) spinosa GÖPPERT
Sphenopteris cf. *pecopteroides* KIDSTON
Sphenopteris sewardi KIDSTON
Sphenopteris sp. nov. ? (cf. *douvillei* ZEILLER)
cf. *Alloiopteris* sp.
Lobatopteris cf. *lamuriana* (HEER) WAGNER
Lobatopteris vestita (LESQUEREUX) WAGNER
cf. *Pecopteris avoldensis* (STUR) CORSIN
Pecopteris daubreei KIDSTON, *non* ZEILLER
Pecopteris dentata BRONGNIART
Pecopteris haussei STERZEL
Pecopteris (Lobatopteris) cf. *camertonensis* KIDSTON
Pecopteris monyi ZEILLER
Pecopteris nyranensis NĚMEJC
Pecopteris cf. *obliquenervis* CORSIN
Pecopteris plumosa (ARTIS) BRONGNIART
Pecopteris raconensis NĚMEJC
Pecopteris cf. *saraefolia* P. BERTRAND
Pecopteris unita BRONGNIART
Pecopteris sp.
Sphenophyllum emarginatum BRONGNIART
Sphenophyllum emarginatum forma *truncatum* SCHIMPER
Sphenophyllum guerreiroi TEIXEIRA
Annularia sphenophylloides (ZENKER) VON GUTBIER
Annularia stellata (VON SCHLOTHEIM) WOOD
Asterophyllites equisetiformis (VON SCHLOTHEIM) BRONGNIART
Asterophyllites longifolius STERNBERG
Macrostachya carinata GERMAR
Calamites carinatus STERNBERG
Calamites suckowi BRONGNIART
Lepidodendron cf. *aculeatum* STERNBERG
Lepidophloios ? sp. nov. ?
Lepidostrobophyllum hastatum (LESQUEREUX) CHALONER
Lepidocarpon major (BRONGNIART) HEMINGWAY
Lycopodites sp.
Sigillaria cf. *tessellata* BRONGNIART
Cordaites sp.

in Asturias. The regional extent of the San Tirso beds is still poorly known (op. cit.). However, these beds are wholly non-marine as are those at Buçaco.

The Autunian flora of Guadalcanal (BROUTIN 1981), in the province of Sevilla, and occurring well down south in the Ossa-Morena Zone (Fig. 1), is comparable in most respects but different because of the presence of additional Cathaysian type elements corresponding to more humid climatic conditions than seem to have prevailed in most of Europe. The progressive aridisation of Europe and eastern North America in Autunian times has given rise to floral assemblages of the Euramerian or Atlantean Province (CHALONER & MEYEN 1973). The flora of Buçaco clearly belongs to that province, and does not show any trace of the humid, tropical elements which constitute the characteristic feature of the generally more diversified Cathaysian type floral assemblages.

Table 14 — Composite list of upper Westphalian D and lower Cantabrian plant fossils from the Guardo Coalfield (northwestern Palencia and northeastern León), Cantabrian Mts, Northwest Spain (after WAGNER in press a). Species marked with an asterisk occur from lower Cantabrian onwards

* *Neuropteris cordata* BRONGNIART
Neuropteris ovata HOFFMANN
Neuropteris planchardi ZEILLER
* *Neuropteris praedentata* GOTHAN
Neuropteris scheuchzeri HOFFMANN
* *Neurodontopteris raymondi* (ZEILLER) WAGNER
* *Odontopteris brardi* BRONGNIART
Odontopteris cantabrica WAGNER
Odontopteris cf. *robusta* ZALESSKY
* *Mixoneura subcrenulata* (ROST) ZEILLER
Linopteris cf. *brongniarti* (VON GUTBIER) POTONIÉ
Linopteris elongata ZEILLER
Linopteris florini TEIXEIRA
Linopteris cf. *obliqua* (BUNBURY) ZEILLER
Linopteris palentina WAGNER
* *Callipteridium (Eucallipteridium) striatum* WAGNER
Callipteridium (Praecallipteridium) jongmansi (P. BERTRAND) WAGNER
Alethopteris ambigua LESQUEREUX
* *Alethopteris barruelensis* WAGNER
* *Alethopteris bohemica* FRANKE
Alethopteris grandinioides KESSLER var. *grandinioides*
* *Alethopteris grandinioides* var. *subzeilleri* WAGNER
* *Alethopteris leonensis* WAGNER (= *A. virginiana* FONTAINE & WHITE ?)
Alethopteris lesquereuxi WAGNER (including
A. lesquereuxi var. *cerverae* WAGNER and *A. kanisi* WAGNER)
Alethopteris missouriensis WHITE
Alethopteris robusta LESQUEREUX
* *Alethopteris zeilleri* RAGOT
Mariopteris nervosa (BRONGNIART) ZEILLER
Mariopteris rotundata HUTH
* *Pseudomariopteris corsini* (TEIXEIRA) WAGNER
Pseudomariopteris ribeyroni (ZEILLER) DANZÉ-CORSIN
Dicksonites plueckeneti (VON SCHLOTHEIM) STERZEL
Eusphenopteris neuropteroides (BOULAY) NOVIK
Eusphenopteris nummularia (VON GUTBIER) NOVIK
* *Eusphenopteris rotundiloba* (NĚMEJC) VAN AMEROM

Table 14 — *(Cont.)*

Eusphenopteris trigonophylla (BEHREND) VAN AMEROM
Palmatopteris sturi GOTHAN
Radstockia sphenopteroides KIDSTON
Sphenopteris (Renaultia) chaerophylloides (BRONGNIART) ZEILLER
Sphenopteris (Renaultia) schatzlarensis (STUR) ZEILLER
Sphenopteris artemisiaefolioides CRÉPIN
* *Sphenopteris biturica densipennata* ALVAREZ-RAMIS
Sphenopteris dimorpha (LESQUEREUX) WAGNER
Sphenopteris mixta SCHIMPER
Sphenopteris ovalis VON GUTBIER
Alloiopteris angustissima (STERNBERG) STOCKMANS & WILLIÈRE
Alloiopteris cristata (VON GUTBIER) NĚMEJC
Oligocarpia bredovii (GERMAR) WAGNER
Oligocarpia gutbieri GÖPPERT
* *Oligocarpia leptophylla* (BUNBURY) GRAUVOGEL-STAMM & DOUBINGER
Oligocarpia pulcherrima STUR
Polymorphopteris cf. *cistii* (BRONGNIART) WAGNER
* *Polymorphopteris multifurcata* WAGNER
Polymorphopteris polymorpha (BRONGNIART) WAGNER
Lobatopteris camertonensis (KIDSTON) WAGNER
* *Lobatopteris lamuriana* (HEER) WAGNER
Lobatopteris vestita (LESQUEREUX) WAGNER
* *Lobatopteris viannae* (TEIXEIRA) WAGNER
Lobatopteris cf. *waltoni* (CORSIN) WAGNER
Pecopteris cf. *candolliana* BRONGNIART
Pecopteris dentata BRONGNIART
Pecopteris hemitelioides BRONGNIART
* *Pecopteris melendezi* WAGNER
Pecopteris monyi ZEILLER
Pecopteris ocejensis WAGNER
(= *P. martinezi* STOCKMANS & WILLIÈRE = *P. nyranensis* NĚMEJC ?)
Pecopteris raconensis NĚMEJC
Pecopteris unita BRONGNIART
Pecopteris villaverdenensis STOCKMANS & WILLIÈRE
Sphenophyllum emarginatum BRONGNIART
Sphenophyllum emarginatum forma *truncatum* SCHIMPER
Sphenophyllum cf. *nageli* GRAND'EURY
* *Sphenophyllum oblongifolium* (GERMAR & KAULFUSS) UNGER
Annularia sphenophylloides (ZENKER) VON GUTBIER
Annularia stellata (VON SCHLOTHEIM) WOOD
Asterophyllites equisetiformis (VON SCHLOTHEIM) BRONGNIART
Lepidodendron wortheni LESQUEREUX
Lepidostrobus variabilis LINDLEY & HUTTON
Lepidostrobophyllum lanceolatum (LINDLEY & HUTTON) CHALONER
Sigillaria brardi BRONGNIART
Sigillaria candollei BRONGNIART
Sigillaria scutellata BRONGNIART
Sigillaria tessellata BRONGNIART
Bothrodendron minutifolium BOULAY
* *Pholidophloios* sp.
Calamites carinatus STERNBERG
Calamites suckowi BRONGNIART
Cordaites sp.
Cordaianthus sp.

3. SANTA SUSANA BASIN

This basin is found alongside the major fault which separates the Ossa-Morena and South Portuguese zones (Fig. 1). Its sedimentary characteristics are poorly known, but it contains partly conglomeratic deposits of non-marine facies.

The floral assemblage of Santa Susana has been discussed most recently by WAGNER & SOUSA (1983 — this volume) and is apparently of latest Westphalian D age (Table 13). There are no deposits of similar age elsewhere in the Ossa-Morena Zone. The floral composition is of standard floodplain type and the assemblage invites comparison with the upper Westphalian D floras of the British Isles (South Wales, Forest of Dean and Bristol-Somerset district) and, to some extent, also with the upper Westphalian D and lower Cantabrian floras of the post-Leonian basin in the Cantabrian Zone of Northwest Spain (Table 14). The British comparison is the most interesting one. There are two elements in the Santa Susana flora which it has in common with the British flora and which are apparently absent elsewhere on the Continent of Europe. *Neuropteris flexuosa* STERNBERG (recorded originally as *Neuropteris machadicostai* by TEIXEIRA 1938/40) is a common species in the upper Westphalian D (and basal Cantabrian) of Britain (CROOKALL 1959; CLEAL 1978), and it also occurs abundantly in Nova Scotia (ZODROW & MCCANDLISH 1980) and in the Illinois Basin of North America (compare LANGFORD 1958). This species is apparently absent in other parts of Europe. *Pecopteris daubreei* sensu KID:TON has also been recorded from the British area. This form may also be present in Northwest Spain, but has still to be redescribed.

With the exception of the local elements, *Sphenopteris alentejana* TEIXEIRA and *Sphenophyllum guerreiroi* TEIXEIRA, and perhaps also *Lepidophloios*? sp. nov., all the species mentioned in the list from Santa Susana occur in comparable strata elsewhere in Europe.

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THE NON-MARINE BIVALVE FAUNA OF THE STEPHANIAN C OF NORTH PORTUGAL

R. M. C. Eagar (1)

Key words: Limnic biospecies; *Anthraconaia lusitanica* (TEIXEIRA); *Anthraconaia* (?) *altissima* sp. nov.; Wettin Shales; Saale Basin.

Palavras-chave: Espécies límnicas; *Antraconaia lusitanica* (TEIXEIRA); *Anthraconaia* (?) *altissima* sp. nov.; 'Wettin Shales'; Bacia de Saale.

ABSTRACT

Amongst exceptionally large, presumed limnic shells two biospecies have been recognized on biometric criteria. The first and most abundant is attributed to *Anthraconaia lusitanica* (TEIXEIRA) and the second to *Anthraconaia* (?) *altissima* sp. nov., which is described. Both biospecies have several varieties and morphic elements strikingly similar to those of faunas in the upper part of the Wettin Shales of the Saale Basin of East Germany. The upper Wettin Shales have been ascribed by EAGAR to lower Stephanian C on the basis of the sucession of northern European non-marine bivalves. The same age is therefore suggested for the North Portuguese fauna.

RESUMO

Com base em critérios biométricos identificaram-se duas espécies de bivalves límnicos de dimensões excepcionais. A mais abundante foi atribuída a *Anthraconaia lusitanica* (TEIXEIRA) e a outra a *Anthraconaia* (?) *altissima* sp. nov. que se descreve nesta publicação. Ambas as espécies patenteiam variações e elementos morfológicos muito similares com os das faunas da parte superior dos 'Wettin Shales' da Bacia de Saale na República Democrática Alemã. A parte superior dos 'Wettin Shales' foi anteriormente atribuída por EAGAR ao Estefaniano C inferior com base na sucessão de bivalves não marinhos do norte da Europa. Em consequência, admite-se a mesma idade para a fauna portuguesa.

1. INTRODUCTION

WATTISON (1926) found shells near São Pedro da Cova which WHEELTON HIND determined as *Anthracomya* (now *Anthraconaia*) *wardi* (ETHERIDGE), *Anthracomya* (now *Anthraconauta*) *phillipsii* (TRUEMAN & WEIR 1946) and *Carbonicola* sp. nov. FLEURY (1937) referred to the probable existence of four species of *Anthracomya* and at least one of *Carbonicola;* but TEIXEIRA (1942) was unable to find WATTISON's original material, which is presumed lost. The following summarized account is based on several of the specimens illustrated by TEIXEIRA (1942, 1943, 1945, 1952), on illustrations by TEIXEIRA & FONSECA (1953), and on the study of some

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(1) The Manchester Museum, The University, Manchester M13 9PL, England, U.K.

seventy variably complete shells in the Department of Geology, Oporto University (Museu Wenceslau de Lima) and in the collections of the Serviços Geológicos de Portugal at Lisbon. I am much indebted to Professor Dr M. J. Lemos de Sousa, to Dr Delfim de Carvalho and Professor Dr M. Magalhães Ramalho for the loan of this material and for permission to develop it. The fauna is herein treated as a whole, all localities known to me being given in the descriptions of Figure 2 (left half) and of Plates I, II.

Fig. 1 — Height and length measurements of shells from the Stephanian of São Pedro da Cova, Varziela, Mina do Pejão (Limoeira), Mina da Lomba and Mina da Varziela (Castelo de Paiva). Measurements have been taken as defined in the description of Table 1. Those of the two paratypes of *Anthraconaia* (?) *altissima* sp. nov. are indicated by the addition of oblique dashes. The fitted lines have been calculated by the method of IMBRIE (1956).

2. GENERAL DESCRIPTION

On the evidence of shell height and length measurements (Fig. 1, H, L) two groups of shells were present, the bulk of each one probably representing a biospecies. The first, *Anthraconaia lusitanica* (TEIXEIRA) has been found to be the most abundant, particularly in the finer grained, darker shales, being fairly variable with a wide range in size ($r_{HL} = 0.97$, Fig. 1). It is illustrated in Figure 2 a-h, o and in Plate I A-H. It includes the unusual shell of Figure 2 k and Plate II H2. It apparently includes shells formerly referred to *Anthracomya prolifera* (WATERLOT) by TEIXEIRA (1952, pl. 1, figs 1-8) and by TEIXEIRA & FONSECA (1953, pl. 1). The elongate forms of Figure 2 i, j and Plate 1 G may also belong to it. The second group (Fig. 1, small circles, Fig. 2 1-n, p, Pl. II, A-H1) appears to be very variable ($r_{HL} = 0.87$). Its varieties commonly have *Anthracosia*-like profile and occasionally resemble the lateral profile of *Anthracosphaerium* (Pl. II G, H1). The bulk of it is herein referred to the new species *Anthraconaia* (?) *altissima*. Whether some of the larger shells in the group, such as that of Plate II I, belong to this biospecies cannot be determined with certainty on available evidence. External measurements of the two groups of Figure 1 are summarized in Table I.

Table 1

	Length		Height/Length		Anterior end/Length	
	O. R.	Mean	O. R.	Mean	O. R.	Mean
A. lusitanica	7.0-50.7	25.41	34.8-50.0	43.55	15.8-27.4	21.54
Numbers measured	32		29		16	
A. (?) *altissima*	10.4-37.3	21.48	53.1-79.0	61.20	16.4-25.1	21.56
Numbers measured	20		20		11	

Length is the maximum length of the shell measured parallel to the line of the hinge. Height is the greatest height measured at right angles to length. Anterior end is the greatest length of the shell measured anterior to the umbo and parallel to length. H/L and A/L ratios are expressed as percentages. O. R. Observed range.

3. SYSTEMATIC DESCRIPTION

The genera: No internal features of the shells have been seen. Comparisons of shell outlines with those of the Stephanian of northern Europe (see below) and with a few shells from eastern North America of probable Stephanian age and showing internal features (EAGAR 1975, pl. 1 F-H) suggests that the genus *Anthraconaia* should be retained for *A. lusitanica*. The generic name is followed by a query in the case of *altissima* sp. nov. because of its external morphological similarity to small-shell anthracosiid-like forms, also probably of Stephanian age, which usually separate clearly from *Anthraconaia* on external morphology and occasionally reveal small internal differences of the shell (EAGAR 1975, fig. 3, pl. 1 A-D). These small shells have been placed provisionally in the category *Anthraconaia* (?). It is noteworthy that no anthracosiids have been proved anywhere in strata younger than those of Westplalian B age (CALVER 1969; EAGAR 1970, 1973, 1975, in press [1979]).

Family MYALINIDAE FRECH 1891, emend. NEWELL 1942

Genus *Anthraconaia* TRUEMAN & WEIR 1946

Anthraconaia lusitanica (TEIXEIRA)

Fig. 2 a, a′

Anthracosiidae sp. TEIXEIRA 1942, est. II 1, 1a (bis),
Anthracomya lusitanica TEIXEIRA 1943, pl. III, fig. 1.
Anthracomya lusitanica TEIXEIRA 1945, est. VIII, figs 4, 5.
Non Anthracomya lusitanica TEIXEIRA 1952, pl. I, fig. 9.

The holotype (Fig. 2 a) consists of two conjoined closed valves which have been compressed laterally but retain low relief. The shell itself is preserved as a thin rind on which growth lines indicate slight ventral curvature at a length of 10.5 mm, which was followed by straightening of the ventral margin. The posterior end of the dorsal margin is not perfectly preserved and may be interpreted as slightly more curved than is indicated in Figure 2 a, so the it may approach that of Figure 2 b, b′. Curvature of the dorsal margin is also indicated in the varieties shown as TEIXEIRA (1942, est. II 1) right hand shell, and in this paper Figure 2 o. In his original description TEIXEIRA (1943) stated that height/length ratio of the species remained constant between lengths of 12 and 30 mm, but it will be seen from Figure 1 that growth led to a reduction in H/L ratio over this range of length. A further reduction in this ratio is seen in the larger shells which TEIXEIRA (1952, pl. 1, figs 1-6, 8) and TEIXEIRA & FONSECA (1958, pl. 1, figs 1-3) attributed to *Anthracomya prolifera* (WATERLOT). The available material from Portugal is insufficient to provide evidence that these large shells belonged to a different biospecies. Moreover similar forms have been shown to be inseparable from small shells with greater H/L ratios where abundant material has been available, as in the Wettin Shales of East Germany (EAGAR *in prep.* [1979]). They are therefore tentatively included within the biospecies *A. lusitanica* pending the collection of further material.

It should be noted that the shell referred to as the holotype of *Anthracomya lusitanica* by TEIXEIRA (1952, pl. I, fig. 9) is not the specimen he designated as holotype in 1943. Its shape and measurements link it with *Anthraconaia* (?) *altissima* sp. nov.

Genus *Anthraconaia* (?) EAGAR 1975

Anthraconaia (?) altissima sp. nov.

Figure 2 1-n, p; Plate II A-H1, I

Diagnosis: Shell subrhomboidal in lateral outline, with a variably arched dorsum which merges into the hinge line by gradual curvature and to the posterior forms a widely rounded angle with the continuously curved posterior margin. Postero-ventral angle rounded. Ventral margin gently curved. Umbones are low and

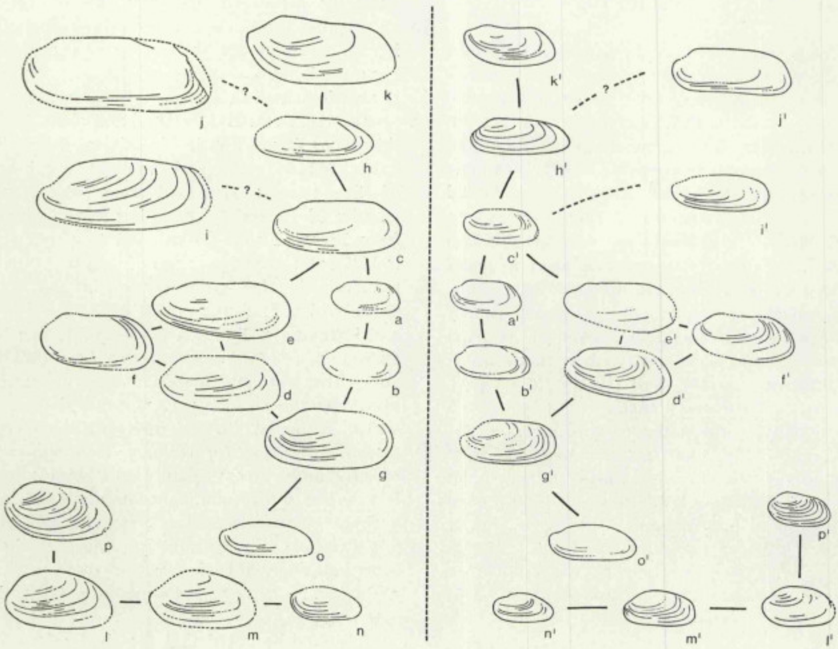

Fig. 2 — a-p (left). Non-marine shells from the Stephanian of north Portugal compared with a′-p′ (right), shells from Stephanian C strata, mostly from the Wettin Shales of East Germany. Lines joining the shells are shown broken when morphological continuity is probable but not directly demonstrable. All shells are reproduced at approximately natural size.

a. *Anthraconaia lusitanica* (TEIXEIRA). Holotype. Minas da Lomba (Castelo de Paiva). Coll. S.G.P.

a′. *A. lusitanica.* A variety having very slightly curved ventral margin. Wettin Shales (probably their upper part), Saale Valley. Z.G.I. 80103/3.

b. *A. lusitanica,* a variety showing slight arching of the hinge line. Varziela, Douro. S.G.P. 10 B.

b′. *A. lusitanica,* with the same trend as that of (b) but a little more strongly expressed. Stephanian C. From a 0.73-metre band of shells lying below 10 m of kuselite in a road cutting at Oberlinxweiler, near St Wendel and 45 km N.E. of Saarbrücken, West Germany. M.M. Ob.G. 7A.

c. *A. lusitanica,* shown also as Plate I B. As for (b). S.G.P. 7A.

c′. *A. lusitanica,* having trend comparable with that of (c). Provenance as for (a′) above, at a depth of 134 m. M.N.E.B. Rie. 134.

d. *A. lusitanica* with low anterior lobe. Shales above the 3rd Coal, São Pedro da Cova. O.U. 21b G.

d′. *Anthraconaia* sp., cf. *A. lusitanica.* From the so-called 'Lower *Anthracosia* Shales' of Rybno, Poland, probably of upper Stephanian age. M.N.E.B. unregistered slab, /D.

e. *Anthraconaia lusitanica* var. nov., cf. *Anthraconaia modiolaris* (J. DE C. SOWERBY). As for (b) above. S.G.P. 8A.

e′. *A. lusitanica* var. nov., cf. (e) above. Wettin Shales. Z.G.I. 80086.

directed forwardly by oblique growth. Height/length ratio between 60 and 80 per cent. Anterior end/length ratio between 13 and 20 per cent.

Dimensions:

	Length mm	Height mm	Anterior end mm
Pl. II C, holotype	18.5	12.5 (67.5 %)	3.3 (17.8 %)
Pl. II A, paratype	20.0	15.8 (79.0 %)	4.2 (15.8 %)
Pl. II D, paratype	20.7	13.3 (64.0 %)	4.5 (13.3 %)

Further description and distribution: Growth lines are clearly impressed on nearly all the specimens attributed to this species, but preservation of the shell as a greenish, probably kaolinitic film, renders the hinge and hinge line obscure. The new species merits a name however by reason of the uniqueness of the material from localities no longer readily accessible. *A.* (?) *altissima*, named from the extremely high H/L ratio of a number of its varieties, is at present known only from the Stephanian of north Portugal, above Seam 3 of the Mine at

f. *A. lusitanica*, cf. *A. modiolaris*, shown also as Plate I D. As for (d), above. O.U. 21b F.

f′. *Anthraconaia* sp., cf. *A. modiolaris*, probably a variety of *lusitanica*. Wettin Shales. Z.G.I. 80311.

g. *A. lusitanica*, cf. *Anthraconaia thuringensis* (GEINITZ *non* LUDWIG). As for (b). S.G.P. 7B. Shown also as Plate II H2.

g′. *Anthraconaia* sp., cf. *A. thuringensis*. Wettin Shales. Löbejün, East Germany. M.N.E.B. Unregistered slab, 1/A.

h. *A. lusitanica*, with a trend towards *Anthraconaia protracta* EAGAR (cf. also EAGAR in press [1979], fig. 6 1, 1′), shown also as Plate I A. Mina da Lomba. S.G.P. 19.

h′. *Anthraconaia* sp., with a trend towards *A. protracta*, with the same comparisons as for (h). Wettin Shales. Z.G.I. 80086A.

i. *Anthraconaia* sp., tentatively regarded as an extreme variety of *A. lusitanica*. Shown also as Plate I G. Mina de Varziela, Castelo de Paiva. S.G.P. 30.

i′. *Anthraconaia* sp. Wettin Shales, East Germany. Z.G.I. 80465.

j. *A.* sp., cf. *A. weissiana* (GEINITZ). As for (b). The restoration is tentative. S.G.P. 30.

j′. *A.* sp., cf. *A. weissiana*. Wettin Shales. Z.G.I. 80443/1/28.

k. *Anthraconaia lusitanica*, with a trend towards *A. protracta*. As for (b). S.G.P. 4A.

k′. *A.* sp., cf. *A. protracta*. Wettin Shales, Adolph Mine, Thuringia, East Germany. M.N.E.B. U. 10. 1094G.

l. *Anthraconaia* (?) *altissima* sp. nov., an elongate variety with strong dorsal arching, comparable with *Anthracosia phrygiana* WRIGHT. External mould. São Pedro da Cova. O.U. 21b A.

l′. *Anthraconaia* (?) *altissima*, elongate variety with slight dorsal arching. Wettin Shales. Z.G.I. 80095/3A.

m. *A.* (?) *altissima*, elongate subrhomboidal variety comparable with *Anthracosia caledonica* TRUEMAN & WEIR. As for (b). S.G.P. 9C.

m′. *Anthraconaia* (?) sp., cf. *A.* (?) *altissima*. Wettin Shales. Z.G.I. 80086C.

n. *A.* (?) sp., aff. *A.* (?) *altissima*, cf. *Anthracosia fulva* DAVIES & TRUEMAN. As for (b). S.G.P. 8B.

n′. *A.* (?) sp., cf. *C.* (?) *altissima*, an elongate variety with arched dorsum. Z.G.I. 332.75.

o. *Anthraconaia* sp., cf. *A. lusitanica*, and compare also *Anthracosia aquilinoides* (TCHERNYCHEV). As for (b). S.G.P. 2A.

o′. *Anthraconaia* sp., and compare also *Anthracosia aquilinoides*. Upper part of Wettin Shales in Riesigk Borehole No. 1 (see (a′) above) at 132.2 m, Saale valley. M.N.E.B. Rie. 132.2/1A.

p. *Anthraconaia* (?) sp., aff. *A.* (?) *altissima*, a variety with clearly defined beak. Shown also as Plate II E. São Pedro da Cova. O.U. 21b H.

p′. *Anthraconaia* (?) sp., cf. (p) above. Wettin Shales, Löbejün, East Germany. M.N.E.B. Unregistered slab, /D.

The shells (c), (e), (d), (g), (j), (l), (p) and (b′), (g′), (i′) and (p′) are produced as mirror images. Abbreviations for collections are given in the description of the plates.

São Pedro da Cova and at the same or similar horizons in its vicinity, the types coming from Varziela, Douro. Shells near the types in shape and unquestionably members of the same biospecies range between 10 and 27 mm in length. Larger apparent varieties, for example the shell of Plate II I, have slightly lower H/L ratios and, in the limited material available, tend to characterize darker, finer grained shales than the typical examples. There appears to be no justification for separating these from *A.* (?) *altissima*. However still more elongate shells having the general characteristics of the latter are known from the Wettin Shales of East Germany (see below) and are clearly separable in this district from a group of much smaller, short, oval to subcircular forms having several varieties comparable to but smaller than *A.* (?) *altissima* (EAGAR in press [1979], fig. 7 52-73); compare also Fig. 2 1-n, p with l′-n′, p′. More and better preserved material from north Portugal is therefore desirable.

4. THE AGE OF THE FAUNA

In Figure 2 tracings of selected shells, mostly from the Wettin Shales of the Saale area, East Germany, have been matched or compared with those of the main varietal trends in the fauna of north Portugal: comparisons should be made in mirror image across the central broken line. Correspondence, particularly within the species *Anthraconaia lusitanica,* appears striking, although it is noticeable that the Portuguese shells, which tend to be found in lighter coloured, more silty sediments than those of northern Europe, are the larger; also the straight ventral margin of *Anthraconaia lusitanica* is rare in the Wettin Shales. However the existence of a distinct small-shell fauna in the Wettin Shales, markedly comparable with that of *Anthraconaia* (?) *altissima* sp. nov. (see the preceding paragraph and Fig. 2 l′-n′ p′, Pl. II G, H1), enhances the comparison. Moreover, with the exception of the shells shown in Figures 2 b′, d′, no comparisons could be found on other horizons among northern European non-marine shells ranging in age from Stephanian A to the German lower 'Autunian', regarded as Asselian and late Carboniferous by KOZUR (1980).

The bulk of the non-marine shells from the Wettin Shales illustrated in Figure 2 come from the collections of the Zentrales Geologisches Institut, East Berlin. Their detailed documentations, although known, could not be released by the Director, but from comparable documented material in the Museum für Naturkunde, East Berlin, it is likely that these shells came from the upper part of the Wettin Shales. The Wettin Shales group have been placed in the higher Stephanian by KAMPE & REMY (1962), and in Stephanian C by KOZUR (1980). The lower part of the Wettin Shales was tentatively placed in Stephanian B and the upper part in Stephanian C on the basis of a correlation of a non-marine shell band in the underlying Mansfeld Shales with WATERLOT's type assemblage of *Anthraconaia prolifera* in the Saar (EAGAR in press [1979], Fig. 2). Thus evidence from plants and that of pioneer work on non-marine bivalve shells together lead to the same broad conclusion, that the limnic fauna from north Portugal lies in Stephanian C. Finally, the Portuguese shells resemble those of the Wettin Shales much more nearly than those from a band in the middle of the Breitenbach Shales of mid-Stephanian C age at Oberlinxweiler, Saar district (see the description of Fig. 2 b′). The latter share varieties primarily with bands of shells from the overlying Kusel Shales, which were placed in 'Stephanian D' by KOZUR (1980). It seems therefore probable that a lower Stephanian C horizon, rather than an upper one, is represented in the fauna from north Portugal.

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PLATES

The following abbreviations are used for collections referred to in the Plates: — O.U. Oporto University, Department of Geology (Museu Wenceslau de Lima); S.G.P. Serviços Geológicos de Portugal, Lisbon; M.N.E.B. Museum für Naturkunde, East Berlin; Z.G.I. Zentrales Geologisches Institute, East Berlin; M.M. The Manchester Museum, University of Manchester, Great Britain.

Plate I

All shells X 2

A. *Anthraconaia lusitanica* (TEIXEIRA), with a trend towards *A. protracta* EAGAR. Also shown as Fig. 2 h. Mina da Lomba, Castelo de Paiva. S.G.P. 19.

B. *A. lusitanica*, also shown as Fig. 2 c (in mirror image). Outline drawn on the photograph. Varziela, Douro. S.G.P. 7A.

C. *A. lusitanica*, cf. Fig. 2 e. Mina da Varziela, Castelo de Paiva. S.G.P. 31D.

D. *A. lusitanica*, with expanded posterior end, also shown as Fig. 2 f. O.U. 21b F.

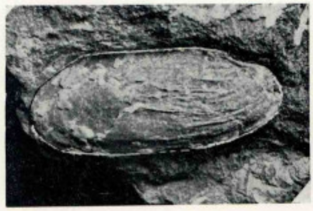
E. *A. lusitanica*, cf. *A. thuringensis* (GEINITZ *non* LUDWIG). Outline drawn on the photograph. São Pedro da Cova. O.U. 21b C.

F. *A. lusitanica*, a variety intermediate between the shells of Figs 2 c and g. Outline drawn on the photograph. As for (B). S.G.P. 3A.

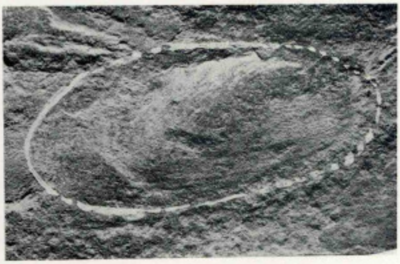
G. *Anthraconaia* sp., tentatively regarded as an extreme variety of *A. lusitanica*. Shown also as Fig. 2 i (in mirror image). Mina da Varziela, Castelo de Paiva. S.G.P. 30. The dorsal margin is partly concealed.

H. *Anthraconaia lusitanica*, with a trend towards *A. protracta*. Outline drawn on the photograph. Also shown as Fig. 2 k. As for (B). S.G.P. 4A.

A



B



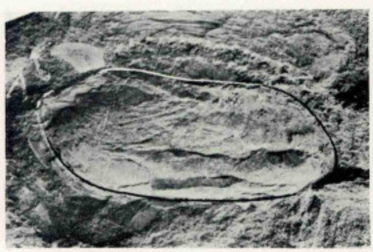
C

D

E

F

G



H

Plate II

All shells X 2

A. *A. Anthraconaia* (?) *altissima* sp. nov. Paratype. A variety with high height/length ratio. The outline is drawn in black on the photograph. Varziela, Douro. S.G.P. 2B.

B. *A.* (?) *altissima*, an unusually large variety, comparable in shape with (D) below. Varziela (50 m from Douro). S.G.P. 38B.

C. *A.* (?) *altissima* sp. nov. Holotype. Provenance as for (A) above. S.G.P. 2C.

D. *A.* (?) *altissima* sp. nov. Paratype. Variety with strongly arcuate dorsum. Outline drawn on the photograph. As for (A) above. S.G.P. 6A.
above. S.G.P. 6B.

E. *A.* (?) sp., aff. *A.* (?) *altissima*, an elongate variety with outline shown as Figure 2 p (in mirror image). São Pedro da Cova. O.U. 21b H.

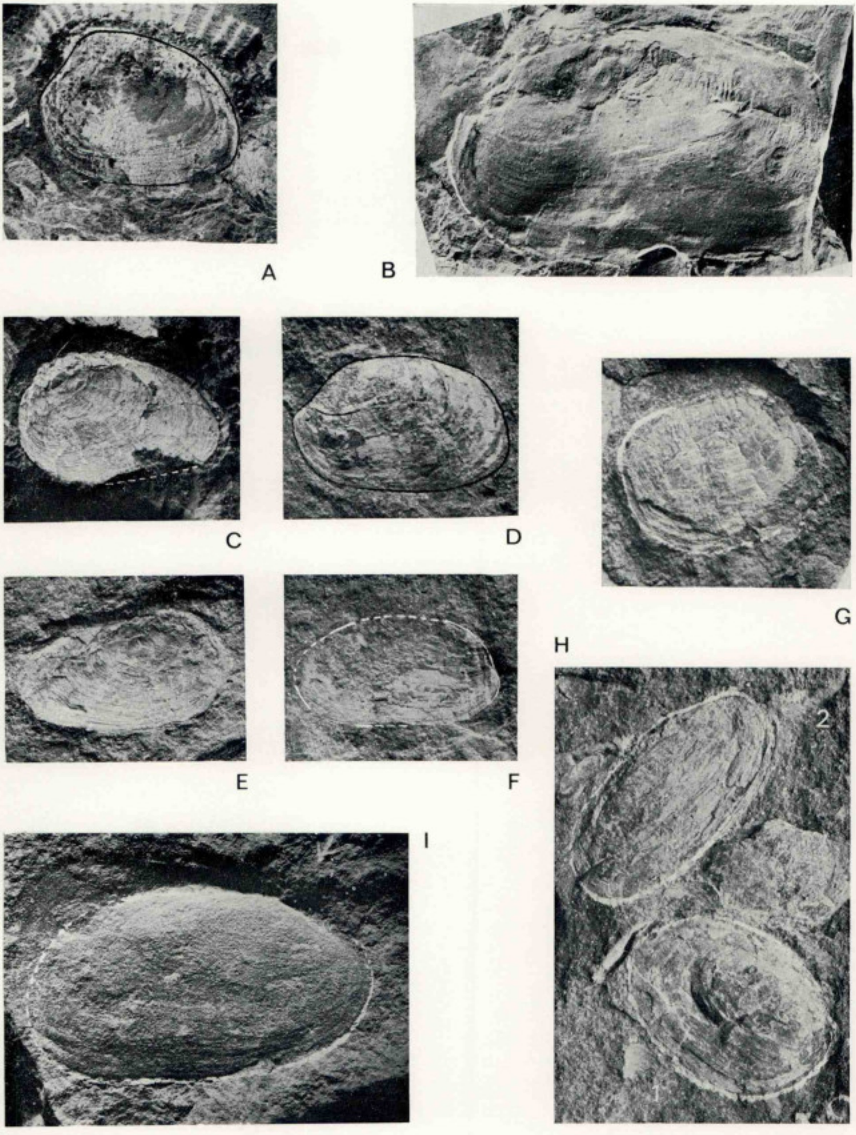
F. *A.* (?) *altissima* sp. nov. Outline drawn on the photograph. Compare (B) and (D) above. S.G.P. 6B.

G. *A.* (?) sp., aff. *A.* (?) *altissima* sp. nov., cf. *Anthracosphaerium pentagonum* TRUEMAN & WEIR. As for (A). S.G.P. 4B.

H1. *A.* (?) sp., aff. *A.* (?) *altissima* sp. nov., a short suboval variety. S.G.P. 7A.

H.2 *Anthraconaia lusitanica* (TEIXEIRA), also shown as Figure 2 g. in mirror image S.G.P. 7B.

I. *A.* (?) *altissima*, an unusually large variety with relatively low height/length ratio. Mina da Varziela, Castelo de Paiva. S.G.P. 25.


A
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THE STRUCTURE OF THE INTRAMONTANE UPPER CARBONIFEROUS BASINS IN PORTUGAL

L. C. G. Domingos (1), J. L. S. Freire (2), F. Gomes da Silva (1), F. Gonçalves (3, 4), E. Pereira (4) and A. Ribeiro (3, 4)

Key words: Buçaco Basin; Douro-Beira though; Granitoid emplacement; Intramontane Basins; Santa Susana Basin; Sedimentation; Shear zone; Stephanian; Structure; Thrusting; Upper Carboniferous; Westphalian.

Palavras-chave: Bacia do Buçaco; Bacias intramontanhosas; Bacia de Santa Susana; Carbónico superior; Cavalgamentos; Estrutura; Estefaniano; Instalação de granitóides; Sedimentação; Sulco Dúrico-Beirão; Vestefaliano; Zona de cisalhamento.

ABSTRACT

The intramontane Upper Carboniferous basins occur along deep major fracture zones with persistent activity during the Variscan Orogeny. These characteristics control the sedimentary environment and internal structure of these basins and its relationship to deformation, granitic emplacement and plutonometamorphism of country rocks.

RESUMO

As bacias intramontanhosas do Carbónico superior ocorrem ao longo de zonas de fractura profundas com actividade tectónica persistente durante a orogenia hercínica. Estes caracteres influenciam a sedimentação, a estrutura interna destas bacias e as relações com a deformação, instalação de rochas graníticas e o plutono-metamorfismo das rochas encaixantes.

1. INTRODUCTION

In the Internal Zones of the Variscan Fold Belt the Upper Carboniferous is mainly of continental character. The individual basins occur along generally narrow depressions where strong subsidence occasionally allowed the deposition of thick coals. These subsiding tracts developed in the later orogenic phases of the Variscan Orogeny, when the internal zones were subjected to uplift induced by crustal thickening due to earlier deformation phases, giving an intramontane character to these basins.

This situation contrast with the prevailing marine conditions in the External Zones (South Portuguese and Cantabrian zones).

The geological conditions are clearly related to the economic importance of the coal basins,

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(1) Museu e Laboratório Mineralógico e Geológico, Faculdade de Ciências e Tecnologia, Universidade de Coimbra, 3049 Coimbra Codex, Portugal.

(2) Serviço de Fomento Mineiro, Rua da Amieira, 4465 São Mamede de Infesta, Portugal.

(3) Departamento de Geologia, Faculdade de Ciências, Universidade de Lisboa, Rua da Escola Politécnica 58, 1294 Lisboa Codex, Portugal.

(4) Serviços Geológicos de Portugal, Rua da Academia das Ciências 19-2.°, 1200 Lisboa, Portugal.

limited and several of them exhausted in the intramontane basins of the Internal Zones but still with large reserves in the much more extensive marine influenced coal basins of the External Zones and the genetically linked accumulations of natural gas.

In Portugal the terrestrial Upper Carboniferous is restricted to the Internal Zones (Central Iberian and Ossa-Morena zones) where intramontane basins are found along major shear zones with a long history of tectonic activity during the Variscan Orogeny, thus suggesting that these are deep crustal structures. In fact, the Santa Susana Basin is located along the major thrust fault between the Ossa-Morena and South Portuguese zones, the Buçaco Basin is located along the major Porto-Tomar shear zone, separating the Centro-Iberian and Ossa-Morena zones, and the Douro-Beira Carboniferous trough is situated along the Douro-Beira shear zone, a major fault within the Centro-Iberian Zone.

In all cases the Upper Carboniferous beds are less deformed than the underlying Devonian and older sediments. This is due to the post-orogenic character of the molasse-type Upper Carboniferous deposits relative to the early Variscan fold phases, which are of interplate type. This post-orogenic character is evident in almost all the lithologies present in these basins and from their unconformable position (see SOUSA & WAGNER 1983 — this volume and WAGNER & SOUSA 1983 — this volume, for all stratigraphic data concerning the Upper Carboniferous). But the intramontane basins were nevertheless affected by late Carboniferous tectonic phases (Santa Susana, Douro-Beira trough) of intracontinental type that controlled the sedimentation, and that rejuvenated faults (Buçaco). The structure in these basins can therefore be quite complex locally.

The late Carboniferous deformation belongs to a regionally extensive tectonic event accompanied by granite pluton emplacement and regional plutonometamorphism and a more localised ductile shear deformation. We thus have to examine the relationships between sedimentation in the intramontane basins and the deformation, granite emplacement and regional metamorphism in the surrounding area.

2. DOURO-BEIRA CARBONIFEROUS TROUGH (J. F., E. P. and A. R.)

The Douro - Beira Carboniferous trough (FREIRE 1981) forms a narrow strip, 0.1 km wide and 90 km long, parallel to the Hercynian structural trend between São Felix de Laundos to the NW and Queiriga-Mioma to the SW. This strip is interrupted northeast of Porto by a fault contact with the older two mica granites from the Porto region, and to the southeast, around Castro Daire, by a younger granitoid central complex.

The age of the sediments (WAGNER & SOUSA 1983 — this volume) seems older to the northwestern part and younger towards the southeast. Westphalian C? and D ages have been determined from floral remains in Bougado and Ervedosa, and early Stephanian C in São Pedro da Cova, Germunde, Pejão. This points to a gradual infilling of an intramontane basin from northwest to southeast with onlap to the southeast. However, the absence of Cantabrian and Stephanian A/B suggests an unconformity within the basin or at least a disconformity.

The sediments that infill the trough are composed of conglomerates and breccias, arkoses, sandstones, shales and coal seams. In some places there is a clear increase in the granitic component of the rock clasts in upward succession, thus suggesting the progressive unroofing of granite batholiths. The sedimentary structures and internal organisation of sequences show an unstable environment of deposition, pointing to syntectonic sedimentation.

In order to determine the relationship of the Upper Carboniferous strata to the underlying rocks and to understand the structure of the trough, the general tectonic situation needs to be explained. The F_1 Variscan phase generated folds with subvertical axial planes in the study area; on the northeastern side of the trough occurs the Valongo Anticline and the trough is preserved in the syncline immediately to the southwest. An important shear zone (JULIVERT *et al.* 1974; ROMANO & DIGGINS 1974) cuts the southwestern limb of the Valongo Anticline and runs across the synclinal axis of the trough. The dip-slip component of the movement that affected the Upper Carboniferous shows always a downthrow on the southwestern side. In the northwestern area (São Felix de Laundos region) and in the southeast (SE of Janarde) the Upper Carboniferous sediments rest on Ordovician to Lower Devonian strata in the southwestern limb of the trough, but in the central tract between Serra de Rates and Janarde the Upper Carboniferous rests on the 'Xisto-grauváquico complex' of Cambrian (and late Precambrian?) age. The base of the Upper Carboniferous succession is everywhere a clear unconformity, marked by basal conglomerates and breccias. The underlying beds show an early S_1 slaty cleavage parallel to the axial planes of F_1 folds, which are refolded by F_2 folds with an axial planar crenulation cleavage. A palaeotopography is fossilised by the Upper Carboniferous sediments. The tightly folded underlying strata are probably a remnant of the southwestern limb of the syncline.

The shear zone is generally steep; in the northwestern part it dips to the northeast but in the southeastern part the dip turns gradually southwestwards, in accordance with the vergence of the F_1 folds in the pre-Upper Carboniferous beds. This suggests that the movement on the shear zone was initiated during the F_1 Variscan folding phase and explains why the shear zone is parallel to the S_1 axial planar cleavage. But the shear zone has also an important strike-slip sinistral component, shown by the attitude of spaced fracture cleavage that disappears away from the shear zone (ROMANO & DIGGINS 1974) and by the presence of sinistral folds with subvertical axes (JULIVERT *et al.* 1974).

Around São Pedro da Cova (Fig. 1) the structure becomes more complex (SOUSA 1977). The thrust plane cutting the southwestern limb of the Valongo Anticline is folded and cut by late high-angle reverse faults, defining a klippe of down-faulted Silurian slices above the lower Stephanian C coal measures worked in the presently abandoned mine of São Pedro da Cova. Since the fold axes plung northwestwards, it shows that the steep reverse fault in the southeast flattens upwards, which is a common phenomenon in this tectonic style. The gently dipping segment of the thrust is easily

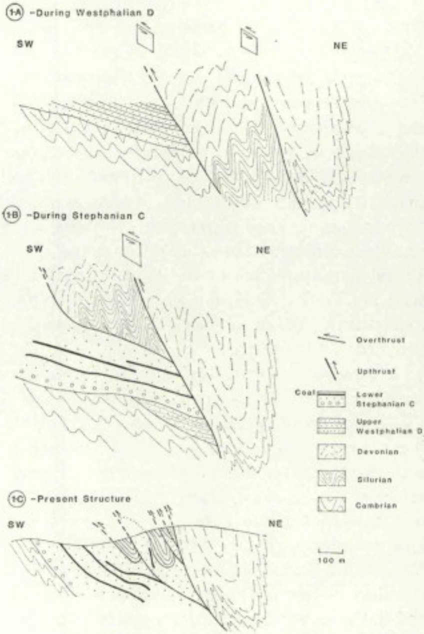

Fig. 1 — Structural evolution of São Pedro da Cova Sector

refolded by continued compression. This has also dragged down the usually steep S_1 cleavage of the overturned southwestern limb of the Valongo Anticline. A further complication in this area is the presence of a narrow strip of Westphalian D sediments immediately northeast of the lower Stephanian C strata. Exposure is poor and without trenching the relationships between the two sets of strata cannot be determined exactly. However, a possible explanation, based on the general tectonic interpretation is that a single trough existed with Stephanian C beds overlapping Westphalian D southwestwards, with only a minor disconformity existing between both successions. This has in fact been claimed by earlier authors (e.g. SCHERMERHORN 1956). This suggests syn-sedimentary tectonism.

The internal deformation of the sedimentary fill is quite variable. Near the shear zone all the beds from Westphalian B to Stephanian C are strongly deformed with subvertical axial planar slaty cleavage parallel to the shear zone and a downdip stretching lineation; but where the trough is wider, as at São Pedro da Cova, the penetrative deformation of Upper Carboniferous strata dies out rapidly southwestwards. This fact is in good agreement with the ideas already expressed (SOUSA 1978) that the mylonitised peranthracites of the 'eastern coal-mine area' are of the same age as the less disturbed coals of the 'western main coal-mine area' of São Pedro da Cova.

We may thus conclude that the Upper Carboniferous was subjected to pressure in a NW-SE direction with a component of sinistral shear in the same sense. This is compatible with the stress field deduced from ductile shear zones (IGLESIAS & RIBEIRO 1981) imposed on regional folds of the Late Variscan fold phase. The complex structure of São Pedro da Cova can be integrated in the more simple general structure if the steep shear zone was curved and flattened out towards the surface, as is usual in areas of block downfaulting.

3. **BUÇACO BASIN** (L. D. and F. G. S.)

The Buçaco Basin constitutes a narrow (0.5-2 km wide) strip, 30 km long, in the general area of the Serra de Buçaco; detailed mapping by the authors is still in progress. The general strike is NNW-SSE.

The stratigraphy (WAGNER *et al.* in prep.) is from top to bottom:

(3) Upper conglomeratic unit (> 600 m thick), consisting of quartzite conglomerates with well rounded boulders, and interbedded shales, arkoses and impure sandstones.

(2) Grey beds (40 m thick) consisting of mudstones, siltstones and sandstones with rootlet beds; well preserved floral remains occur as well as comminuted plant debris ; a thin coal, occasionally up to 1 m thick, forms part of this succession which has been dated as latest Stephanian C.

(1) Lower conglomeratic unit (180-200 m thick) consisting of fanglomerate followed by mass-transported conglomerates and mud flow deposits; these are all red beds.

The basal contact is an angular unconformity with slates of the 'Xisto-Grauváquico complex' (of Cambrian or late Precambrian age).

The upper Stephanian C rocks of the Buçaco Basin have been folded into a tight, almost isoclinal syncline with a slightly overturned western flank which is limited by upthrust Precambrian phyllites of the Ossa-Morena zone (Fig. 2). There is a marked contrast between the relatively undeformed regularly westward dipping eastern flank and the rather heavily sheared western flank which is in contact with the Precambrian phyllites. The mapping has even showed the presence of a small klippe of the lower, red bed unit, on top of the upper conglomeratic unit. This klippe has been folded into a small (200 m wide) syncline, the eastern limb of which is affected by a subsidiary upthrust.

These data suggest that the Porto-Tomar shear zone was reactivated in post-Stephanian C times by E-W compressive movements which thrust the Precambrian phyllites of the Ossa-Morena Zone over the Carboniferous deposits, and producing a substantial amount of shearing on the grey, plant-bearing strata of unit 2. This is probably due to the higher ductility of those beds in comparison with the more competent overlying and underlying conglomeratic units. Continued compression caused refolding of shear planes and thrusting on deeper planes.

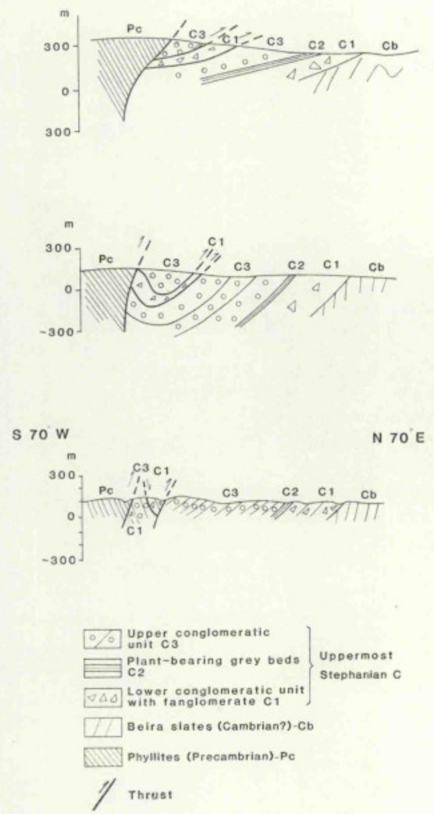

Fig. 2 — Structure of Buçaco Basin

The internal deformation of the Buçaco beds is very low; no fracture cleavage is found, but only brittle faults and slickenslides on bedding planes which confirm the E-W compression deduced from the macrostructure

4. SANTA SUSANA BASIN (F. G.)

The Santa Susana Basin forms a N-S trending narrow strip (0.1-5 km wide, and 12 km long) in the Santa Susana region (northeast of Alcácer do Sal). Floral remains have proved its late Westphalian D age.

The Westphalian D beds of Santa Susana consist of coarse conglomerates, arkoses and shales, 150 m thick. They rest on shales and intercalated intermediate volcanics (Toca da Moura Volcano-sedimentary complex) of unknown, but possibly Namurian age. This volcano-sedimentary complex passes eastwards into acid and intermediate intrusive porphyries. All these igneous rocks are present as clasts in the Westphalian D conglomerates.

The Westphalian D beds of Santa Susana and the underlying Toca da Moura Volcano-sedimentary complex are thrust westwards over phyllites (of Late Devonian age?) of the South Portuguese Zone (Fig. 3). The cleavage of the phyllites predates the thrusting. The thrust plane, the cleavage in the phyllites and the

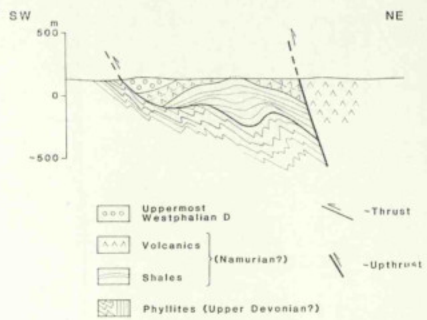

Fig. 3 — Structure of Santa Susana Basin

bedding planes of the upper block are all refolded by N-S striking open folds. The thrust is cut on the eastern side by a steep dextral N-S shear zone (IGLESIAS & RIBEIRO 1981) which shows a westward directed component of

Table 1 — Relationships between Westphalian and Stephanian sedimentary successions, tectonic deformation and granitoid emplacement in northern and central Portugal

GEOCHRONOLOGICAL SCALE	SERIES	STAGES	SEDIMENTARY SUCCESSIONS (after WAGNER & SOUSA 1983 — this volume)	DEFORMATION		GRANITOID EMPLACEMENT	
230							
240	THURINGIAN			BRITTLE FAULTING	(E-W COMPRESSION)		
250							
	SAXONIAN						
260						POST TECTONIC (Central Porphiritic Granite, Castro Daire complex)	YOUNGER Granodiorites and Granites
270	AUTUNIAN				(N-S COMPRESSION)		
280			Buçaco				
	STEPHANIAN	C	São Pedro da Cova	REGIONAL FOLDING ←→ DUCTILE SHEARING AND THRUSTING	LATE VARISCAN FOLD PHASE		
		B				LATE TECTONIC (Regoufe, Alvarenga)	TWO MICA GRANITES AND GRANO-DIORITES
		A					
290	CANTABRIAN		S. Susana				
		D	Ervedosa				
		C	Bougado				
300	WESTPHALIAN	B				SYNTECTONIC (Serra da Freita)	
		A					
310							
	NAMURIAN					EARLY GRANODIORITES	
320							
315							

upthrusting. This shear zone is regarded as the boundary between the Ossa-Morena and South Portuguese zones.

The Westphalian D beds are folded and faulted but are not cleaved.

5. CONCLUSIONS

The Upper Carboniferous intramontane basins of Portugal are important witnesses of the Variscan Orogeny. From the sedimentary record of their infilling important conclusions can be reached regarding the tectonic, magmatic and metamorphic evolution during the Late Variscan phases of the Internal Zones of the Iberian Fold Belt.

With regard to the Beira-Douro trough the first problem is the relationship between the ages of sedimentary successions and the Variscan phases of deformation. The lower Stephanian C basal conglomerate and breccia shows pebbles of metamorphic rocks containing refolded S_1 with axial planar S_2 crenulation cleavage and also pebbles of deformed granites with tectonic fabric (of probably F_2 age). It may be concluded that F_2 penetrative deformation and associated granite emplacement and plutonometamorphism are of pre-Stephanian C age. But the lower Stephanian C beds are themselves deformed by a shear zone wholly compatible with the F_2 stress field. Two possible explanations can be put forward: Either (1) the regional F_2 Variscan folding phase is a post-Westphalian D and pre-Stephanian episode of short duration (the so-called 'Asturian Phase' of earlier authors); or (2) the deformation of the Westphalian C?/D beds is related to the shear zone, as is the one affecting Stephanian C, in which case the more regional penetrative deformation is of pre-Westphalian C age and the deformation of the Upper Carboniferous strata is entirely due to a localised shearing episode. This would be continuous with the F_2 regional fold phase but representing a later phase, when the cooling of the orogenic pile did not allow a penetrative regional deformation but merely ductile shearing along narrow shear zones. This evolution is clearly shown in the structure of granitoids affected by F_2, where the penetrative S_1 gneiss fabric is gradually obliterated by late 'c' shear planes (IGLESIAS & RIBEIRO 1981). The discrimination between these two hypotheses is made difficult by the fact that during the deposition of the basal conglomerate of Westphalian C?/D age the orogenic belt was less eroded than during early Stephanian C times and the refolding of F_1 fabric by F_2 is less clearly expressed in the upper structural level. Nevertheless, we favour the second possibility because a coherent picture emerges from isotopic age dating of granitoids, chronostratigraphic data and structural data (Table 1).

The Buçaco Basin is affected by a late Brittle Faulting Episode with maximum compressive stress in a nearly E-W direction, which also affects the Berlenga post-tectonic granite (RIBEIRO *et al.* 1979) sometime in the late Permian.

The Santa Susana Basin is located on the major boundary between the Ossa-Morena and South Portuguese zones (RIBEIRO *et al.* 1979). The major folding and regional metamorphism in the Ossa-Morena Zone are of Acadian age (post-Eifelian and pre-Frasnian). The Santa Susana beds are clearly post-orogenic in respect to that event. In the South Portuguese Zone the major folding and regional metamorphism are of Viséan to Westphalian ages, and these Carboniferous phases of deformation did not affect the Ossa-Morena Zone penetratively. The boundary between both zones is a deep major fault along which magmas of various types intruded (gabbros, porphyries of the Beja complex) and which even contains highly deformed serpentinites. The movement on that fault zone was mainly westward thrusting from the Middle Devonian onwards and continuing into the Westphalian D. Later on the movement changed to dextral shearing. As in the case of the Douro-Beira trough the steep shear zone was curved and flattened out towards the surface.

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PETROLOGICAL CHARACTERISTICS OF THE UPPER STEPHANIAN COALS IN NORTH PORTUGAL (DOURO COALFIELD) AND THEIR RELEVANCE TO COALIFICATION STUDIES

M. J. Lemos de Sousa [1]

Key words: Douro Basin; Douro Coalfield; Upper Carboniferous; Meta-anthracite; Coalification; Rank; Rank parameters (physical and chemical); Reflectance; Density.

Palavras-chave: Bacia do Douro; Bacia Carbonífera do Douro; Carbonífero superior; Metantracite; Incarbonização (carbonificação); Grau de incarbonização ou carbonificação; Parâmetros de grau (físicos e químicos); Poder reflector; Densidade.

RESUMO

A Bacia do Douro constitui uma estreita faixa que se estende por cerca de 90 km entre São Pedro Fins (a leste do Porto) e Mioma (a nordeste de Viseu). Na parte noroeste, a Bacia do Douro contém camadas de carvão exploradas na chamada Bacia Carbonífera do Douro. Àparte pequenas antigas explorações, duas áreas mineiras principais podem ser definidas: São Pedro da Cova, a Norte do rio Douro (hoje abandonada) e Pejão, a Sul do mesmo rio (única área em actividade hoje em dia).

Estima-se que entre 1894 e a actualidade foram exploradas no conjunto da bacia cerca de 21×10^6 t de carvão. Hoje em dia, na falta de estudos estratigráficos e estruturais de pormenor, apenas se pode falar com seriedade de cerca de 5×10^6 t de reservas *in situ*, das quais se estima poder recuperar $3{,}7 \times 10^6$ t. Na continuidade do jazigo, em profundidade, podem ainda admitir-se recursos adicionais da ordem dos $3{,}5 \times 10^6$ t.

Resumem-se os estudos levados a efeito sobre o grau de incarbonização dos carvões da Bacia Carbonífera do Douro, os quais permitiram demonstrar que se trata de carvões altamente incarbonizados.

Discute-se, quer a classificação dos carvões durienses como metantracites, quer a importância do seu estudo para o conhecimento dos graus mais elevados da carbonificação através das relações entre parâmetros físicos e químicos de grau. Estabelecem-se curvas relacionando os valores médios do poder reflector máximo, mínimo e médio com o teor em carbono (s.s.c.). Também se estuda a relação entre a densidade e o teor em carbono (s.s.c.).

ABSTRACT

The Douro Basin occurs in a 90 km long, narrow strip extending from São Pedro Fins (east of Oporto) and Mioma (northeast of Viseu). In its northwestern part the Douro Basin contains coal-bearing strata worked in the so-called Douro Coalfield. Basides little mines, there are two main mining districts, viz. São Pedro da Cova area (now abandoned), and Pejão area (the

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(1) Mineralogia e Geologia, Faculdade de Ciências, Universidade do Porto, Praça Gomes Teixeira, 4000 Porto, Portugal.

unique mine still being worked), separated by the Douro river.

One can estimate that c. 21×10^6 t of coal were mined out in the whole basin since 1894. Nowadays, and in the absence of a detailed stratigraphical and structural investigation, one can only speak of c. 5×10^6 t of 'proved reserves in place', from which one may expect to recover c. 3.7×10^6 t. 'Additional resources in proven coal province' are c. 3.5×10^6 t.

We summarize the investigations dealing with the rank of the Douro Coalfield coals, which proved to be high coalifield ones.

We discuss the classification of these coals as meta-anthracites, and the relevance of their study to the knowledge of the highest levels of coalification, by means of the relations between physical and chemical rank parameters. Curves relating mean maximum, minimum and random reflectances with carbon content (d.a.f.) are established. The relation between density and carbon content (d.a.f.) is also studied.

1. INTRODUCTION

All the three main Terrestrial Carboniferous occurrences in Portugal, viz., (i) the 130 km long, narrow strip in the general vicinity of Oporto, (ii) the Buçaco Basin, and (iii) the Santa Susana Basin, contain coal-bearing strata which yield the main economic interest of those formations. Owing to the peculiar stratigraphic and structural conditions of those occurrences (Sousa & Wagner 1983 — this volume; Domingos *et al.* 1983 — this volume), the coal deposits are small.

The most important Carboniferous coal deposit in Portugal is the so-called Douro Coalfield. It forms the northern part of the Douro Basin (Viana 1928; Vianna 1952; Thadeu 1965; *Carta mineira de Portugal na escala 1/500.000* 1960) which extends for 90 km along a strip between São Pedro Fins, Ermesinde (just east of Oporto) and Mioma (northeast of Viseu) (Sousa & Wagner 1983 — this volume) and contains coal-bearing strata of early Stephanian C age (Wagner & Sousa 1983 — this volume; Eagar 1983 — this volume).

Besides little mines, the Douro Coalfield includes two main mining areas or sectors, viz. São Pedro da Cova, and Pejão, separated by the Douro river. Pejão is the only area still being mined. Small mines were abandoned in 1957 and those in São Pedro da Cova area in 1972.

Although mining in the Douro Coalfield goes back to the late eighteenth century, the regular exploitation in São Pedro da Cova started in 1804 only (Silva 1814; Schmitz 1852; Azevedo 1858; Ribeiro 1861; Monteiro & Barata 1889; A. M. Carvalho 1891; Lima 1892; Diniz 1941).

The production between 1950 and 1982 is summarized in Table 1. It is estimated that since 1894 c. 21×10^6 t of coal have been mined out in the Douro Coalfield (J. L. S. Freire — personal commun.).

Coal from Douro Coalfield has been used as an alternative fuel, both in domestic and industrial applications, especially during the First and the Second World Wars. Almeida (1929) recommended its use for generating electricity. Actually, it is still its present utilization since 1960. It should also be noted that the utilization of Douro coals as a reducing agent in steelmaking was experimented in the 1950's and 1960's (Solla & Santos 1960, no publication date a, b; Rodrigues *et al.* no publication date).

2. HISTORY OF INVESTIGATIONS

General references about the properties of the coals produced in the Douro Coalfield are found in papers dealing with the coal industry in Portugal (F. Vasconcellos 1877; Cruz 1922, 1923a, b; Pego 1925; Viana 1928; C. Vasconcellos 1929; Almeida no publication date; Machado 1971; Queirós no publication date).

The problem of washability of coals from Douro Coalfield was investigated by Almeida (1931, 1936).

Detailed chemical analyses of the coals run of mine, and of the different industrial types commercially available were published in *Carvões portugueses* (1946).

Brito (no publication date) describes the results of semi-quantative analyses of trace elements present in the ash of Douro coals,

Table 1 — Production of Douro Coalfield mines between 1950 and 1982 (in 10^3 t)

	1950	1951	1952	1953	1954	1955	1956	1957	1958	1959	1960	1961	1962	1963	1964	1965
São Pedro da Cova area	171.1	167.9	169.7	174.5	162.0	140.8	138.5	165.6	189.0	159.8	142.6	143.3	134.1	134.9	137.2	123.4
Pejão area	247.1	248.3	271.8	303.3	269.5	264.4	274.9	333.2	378.1	367.4	291.7	326.7	271.0	281.3	307.1	304.1
Others	0.6	0.7	0.6	0.5	0.6	0.3	0.1	—	—	—	—	—	—	—	—	—
Total	418.8	416.9	442.1	478.3	432.1	405.5	413.5	498.8	567.1	527.2	434.3	470.0	405.1	416.2	444.3	427.5

	1966	1967	1968	1969	1970	1971	1972	1973	1974	1975	1976	1977	1978	1979	1980	1981	1982
São Pedro da Cova area	108.8	109.4	94.1	83.5	27.4	16.3	7.9	—	—	—	—	—	—	—	—	—	—
Pejão area	311.4	333.4	302.8	333.1	243.5	237.0	243.8	220.8	230.2	221.6	193.4	195.3	180.1	179.1	177.5	183.8	178.5
Others	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
Total	420.2	442.8	396.9	416.6	270.9	253.3	251.7	220.8	230.2	221.6	193.4	195.3	180.1	179.1	177.5	183.8	178.5

and A. H. Carvalho & Moura (1960) determined the uranium in the ash of the same fuels.

Almeida (1958) discussed the position of those coals with respect to the principal coal classifications.

The first reference to the petrography of the coals from Douro Coalfield is found in Oliveira (1945). Another more detailed record provided by the same author deals with coals from the Pejão area (Oliveira 1956, 1958), Duparque (1949), who studied a set of samples supplied by Oliveira, classified the coals from Douro coalfield as 'type III anthracites', and criticized Oliveira's work, particularly the etching method used.

Further detailed petrological studies of these coals have been carried out by the present author. We have successively studied the petrography and maceral evolution (Alpern & Sousa 1970; Sousa 1974, 1978a), and several rank parameters, viz., reflectance (Sousa 1971, 1978b, 1979a), microhardness (Sousa 1976), and density (1977a). Papers dealing with properties of oxidized coals in outcrop (Sousa 1978c) and with mineral matter (Sousa 1979b, c) have also been published. One must also mention coalification studies in relation both with the genesis of the basin, and the general geology of the area (Sousa 1977b, 1978b).

3. RESERVES/RESOURCES OF COAL IN THE DOURO COALFIELD

The estimation of the reserves/resources of coal in Douro Coalfield is a crucial question, and a wide range of estimates have been advanced in the literature (Cruz 1922; Santos 1953a, b; Carneiro 1971; Pereira 1981; Solla no publication date) and in unpublished reports. Various reasons may explain this situation. In several cases, the calculations were carried out at different times and, in consequence, the authors had to account for both new findings and the amount of coal mined out in the meantime. Furthermore, rather diverse criteria were employed. Still, in our opinion, the main reason is that, despite borehole campaigns, of which the results were recently synthetized (see FREIRE 1981), the necessary detailed stratigraphic and structural investigations have not been carried out yet. Accordingly, the only estimation that we can safely advance, now, has to be based on the known reserves in place plus the additional resources with a certain degree of geological certainty, only. The estimates, using the terminology of the World Energy Conference (FETTWEIS 1979; KOCH 1980) are as follows:

Proved reserves	*in place* (Category II)......	5×10^6 t
	recoverable (Category IIa)	3.7×10^6 t
Additional Resources (in Proven Coal Provinces) (Category III)		3.5×10^6 t

Cautiously, the Portuguese Energetic Plan, 1982 version *(Plano Energético Nacional — Versão 1982)*, estimates for the Douro Coalfield a recoverable reserve of 3.5×10^6 t, with the recommendation that additional amounts must be confirmed by further research.

Finally, we must mention that in a recent publication (*Listing of Deposits to accompany Worldwide Coal Deposits Color Map* 1978) the São Pedro da Cova and the Pejão areas were classified as SIZE III (Coal deposits $< 10^7$ SCE ([1])) and SIZE II (Coal deposits 10^7-10^9 SCE), respectively.

4. RANK STUDIES ON THE DOURO COALFIELD COALS — THEIR RELEVANCE TO THE KNOWLEDGE OF THE HIGHEST LEVELS OF COALIFICATION

The coalification pattern of Douro Coalfield has been established by means of rank investigation, based mainly on the reflectance of vitrinite (telocolinite) and on chemical analyses of whole coals (SOUSA 1978b). Moreover, owing to the geological applications of coalification studies (ALPERN 1969; M. & R. TEICHMÜLLER 1981; STACH *et al.* 1982) many genetic geological, and economic aspects have been elucidated, viz.:

(i) Coal-bearing strata are always in normal succession;

(ii) Coalification results from an increase of temperature due to regional granitoid emplacement ([2]);

(iii) Coalification is entirely a pre-tectonic event;

(iv) In São Pedro da Cova — structurally, the more complex area — the so-called 'bacia oriental' (i.e., the eastern coal-mining area) corresponds to a dislocated slice of the so-called 'bacia ocidental ou clássica' (i.e., the western, main, coal-mining area);

(v) Coalification increases along the Douro Coalfield from the NW (São Pedro da Cova area) to the SE (Pejão area).

Furthermore, coals from the Douro Coalfield proved to be highly coalified. The main scope of the present paper is to discuss their classification and the relationships between physical and chemical rank parameters at high coalification levels.

4.1. Classification of the Douro Coals

From detailed analytical results already published (SOUSA 1978a, b) and further data, the following range of rank parameters may be established for the whole Douro Coalfield:

	Maximum value (%)	Minimum value (%)
C (d.a.f.) ([3])	96.75	94.23
VM (d.a.f.) ([3])	5.27	2.04
H (d.a.f.) ([3])	2.07	1.12
FC (d.a.f.) ([3])	93.36	84.50
$\bar{R}_{max}$ ([4])	7.30	5.20
$\bar{R}_e$ ([5])	5.30	4.10

([1]) SCE = Standard Coal Equivalent (one SCE represents one metric ton of hard coal with Heat Value of 6,879 k cal/kg equalling 8,000 kWh of electric energy).

([2]) ROQUETTE (1887) was the first author to reach the same conclusion in a general study on metamorphism.

([3]) It was demonstrated (SOUSA 1978b) that for Douro coals, the values of chemical rank parameters calculated on a d.a.f. basis are, in practice, equal to those calculated on a d.m.m.f. (Parr) basis.

([4]) $\bar{R}_{max}$ = Mean maximum reflectance.

([5]) $\bar{R}_e$ = Mean random (or average) reflectance.

Accordingly, and with respect to the two best known classification schemes, our coals are placed as follows:

a) In the International Classification of Hard Coals by Type — Code number 100 A.

b) In the ASTM *'Standard specification for classification of coals by rank', ASTM Designation D 388-66* — As anthracites.

Nevertheless, since PATTEISKY & M. TEICHMÜLLER (1958) (see also SOUSA 1978b; STACH *et al.* 1982; *International Handbook of Coal Petrography 2nd Ed.* 1963), it is known that the rank parameters utilized in those classifications are not appropriate at high coalification levels.

Indeed, if one considers the German (DIN) classification, which calls for the right parameters (M. TEICHMÜLLER *et al.* 1979, (see also STACH *et al.* 1982), all coals of the Douro Coalfield are coherently classified as meta-anthracites. This is so according to either carbon or hydrogen content or reflectance ($\bar{R}_{max}$ and $\bar{R}_e$). A disagreement still remains with respect to the values of volatile matter content. However, besides the little relevance of this parameter at this level, that may be imputed to the mineral matter composition of the ash content in our coals.

Two open questions are the use of the mean minimum reflectance ($\bar{R}_{min}$) as a graphitization index, and the relation between anisotropy ($\bar{R}_{max}$ - $\bar{R}_{min}$) and shearing in Douro Coalfield.

The results obtained for $\bar{R}_{min}$ in Douro Coalfield range between 3.18 and 1.96 %, and the anisotropy varies from 4.68 to 2.27 %. These values agree with those published by M. TEICHMÜLLER *et al.* (1979) (see also STACH *et al.* 1982) for meta-anthracite level. Regrettably those parameters have not been systematically measured in the basin.

The German (DIN) Norm, which was based on relations found between $\bar{R}_{max}$, $\bar{R}_{min}$ and $\bar{R}_e$, sets the boundary between anthracites and meta-anthracites at $\bar{R}_{max} = 4$ % and $\bar{R}_e = 3.5$. It is important to note that this agrees quite well with the conclusions based on the evolution of the reflectance of macerals, according to ALPERN & SOUSA (1970) (Fig. 1) who stated that the

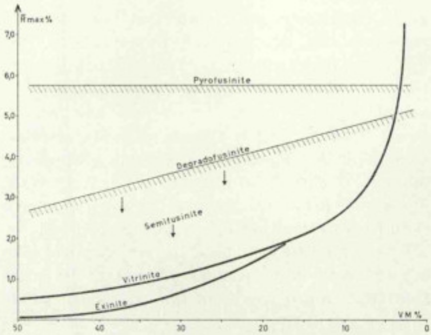

Fig. 1 — Evolution of the reflectance of macerals with the coalification. (Adapted from ALPERN & SOUSA 1970).

anthracites and meta-anthracites ('peranthracites') boundary might be fixed optically when $\bar{R}_{max}$ degradofusinite $\geqslant \bar{R}_{max}$ vitrinite i.e., c. 4.5 % of $\bar{R}_{max}$. This corresponds to $\bar{R}_e$ values c. 4 %, and to less than 2 % hydrogen.

The question arises as to decide if one can retain $\bar{R}_{max}$ or $\bar{R}_e$ as a rank parameter at the meta-anthracite level of coalification.

Since several meta-anthracites display optical biaxiality (see COOK *et al.* 1972), HEVIA (1977) (see also HEVIA & VIRGOS 1977) states — and this is an important point that must be taken in account — that one should use $\bar{R}_e$.

Nevertheless, M. TEICHMÜLLER *et al.* 1979 (see also M. & R. TEICHMÜLLER 1981) relate the $\bar{R}_{min}$ and, consequently, the optical anisotropy of meta-anthracites with shear zones, i.e., with tectonic events occurring after coalification ([1]). If that is the case, $\bar{R}_{max}$ remains

([1]) Biaxiality may also occur in medium volatile coals, and also as a consequence of tectonic events during coalification, as explained by STONE & COOK (1979) in the Southern Coalfield of the Sydney Basin, N. S. W., Australia.

as the only parameter capable of denoting the degree of coalification as stated in the *International Handbook of Coal Petrography Suppl. 2nd Ed.* (1971). However, the question is still open to debate, and necessitates further research.

In his Project of a Universal Classification of Solid Fuels, ALPERN (1981) retains $\bar{R}_e$ as the main rank parameter at all levels of coalification. On the other hand, in that classification the boundary meso- meta-anthracites was set at $\bar{R}_e = 5\ \%$. According to the above considerations, that value for $\bar{R}_e$ is not justifiable and, in our opinion, it should be changed to 4 %. Besides, this would place our coals in the same category with respect to either $\bar{R}_e$ or hydrogen content.

As a matter of fact, if we consider the present version of ALPERN's Project, the coals from the Douro Coalfield are characterized as follows:

Rank: Meso- to Meta-anthracites ...... $\bar{R}_e$ 40-70
Type: Vitric .... $V_{I > L}$
Facies { Ashy coals C30
Washability (< 10 % of ash) 50

We must note that some of our coals fall under the meso-anthracites class. But, if we change the boundary between meso- and meta-anthracites to $\bar{R}_e = 4\ \%$, all the Douro coals, for which $\bar{R}_e$ is always greater than 4 and $2 > H\ \% > 1$, will be coherently classified as meta-anthracites.

4.2. Relationships between physical and chemical rank parameters

Many general coalification curves relating physical and chemical rank parameters have been published, but in most cases they do not cover all the range of coalification, mainly, the highest levels.

As it has been proved that Douro meta-anthracites are among those of the highest coalification in the North-Atlantic Province (ALPERN & SOUSA 1970; SOUSA 1978b) we think that it is important to know how they complete the general coalification curves.

According to the investigations carried out in Douro coals, the relation of carbon content (d.a.f.) to two of the main physical parameters were considered as follows:

(i) Reflectance

Curves relating reflectance to chemical rank parameters have been published by KÖTTER 1960; VAN KREVELEN 1961; LENSCH 1963; NOËL 1966; COPPENS 1967; DE VRIES & BOKHOVEN 1968; DE VRIES *et al.* 1968; ALPERN 1969; MACKOWSKY & SIMONIS 1969; ALPERN & SOUSA 1970; M. TEICHMÜLLER 1970, 1971; MCCARTNEY & M. TEICHMÜLLER 1972; WOLF 1972; HEVIA 1977; DAVIS 1978, and in the *International Handbook of Coal Petrography, 2nd Ed.* 1963. However,

Fig. 2 — Maximum ($\bar{R}_{max}$/closed circles) and minimum ($\bar{R}_{min}$/open circles) reflectances of vitrinite versus carbon content d.a.f. (C″). (Adapted from SOUSA 1979a).

only some authors considered the carbon content as a chemical parameter.

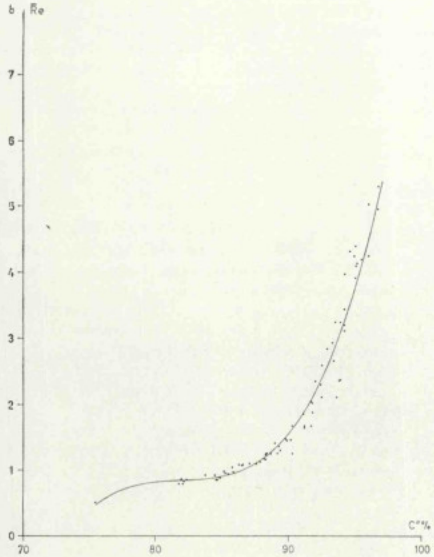

Fig. 3 — Random reflectance ($\bar{R}_e$) of vitrine versus carbon content d.a.f. (C''). (Adapted from SOUSA 1979a).

Fig. 4 — Density versus carbon content d.a.f. (C''). Encircled points correspond to Douro Coalfield coals. (Adapted from SOUSA 1977a).

To establish our curves, the data published by HUNTJENS & VAN KREVELEN (1954), BROADBENT & SHAW (1958), MURCHISON (1958), and ALPERN & SOUSA (1970) were joined with the data from the Douro Coalfield (SOUSA 1979a). The curves for $\bar{R}_{max}$ and $\bar{R}_{min}$ against carbon content (d.a.f.) are shown in Fig. 2, and the curve for $\bar{R}_e$ against carbon content (d.a.f.) is shown in Fig. 3. The equations of the approximating curves are, respectively:

$$\bar{R}_{max} = -541.79 + 9.5751\ (C'') + 0.12768\ (C'')^2 - 0.0034442\ (C'')^3 + 0.000017654\ (C'')^4$$

$$\bar{R}_{min} = -253.63 + 12.994\ (C'') - 0.24245\ (C'')^2 + 0.0019545\ (C'')^3 - 0.0000057207\ (C'')^4$$

$$\bar{R}_e = -398.15 + 11.311\ (C'') - 0.057958\ (C'')^2 - 0.00073962\ (C'')^3 + 0.0000059495\ (C'')^4$$

One may also mention that curves relating reflectance with hydrogen content at the highest levels of coalification were published by MCCARTNEY & M. TEICHMÜLLER (1972) and SOUSA (1978b).

(ii) Density

Curves relating density to carbon content were published by DULHUNTY & PENROSE (1951) and by VAN KREVELEN (1961).

Now, we considered the values of the density (d.a.f.) — carbon content (d.a.f.) relationship published by FRANKLIN (1949), SHERLOCK (1950, 1951), DULHUNTY & PENROSE (1951), and SUN & CAMPBELL (1966) together with those from Douro Coalfield (SOUSA 1977a). The curve obtained is shown in Fig. 4. The equation of the approximating curve is:

$$\text{Density} = 75.97 - 4.290\ (C'') + 0.09164\ (C'')^2 - 0.0008604\ (C'')^3 + 0.000002993\ (C'')^4$$

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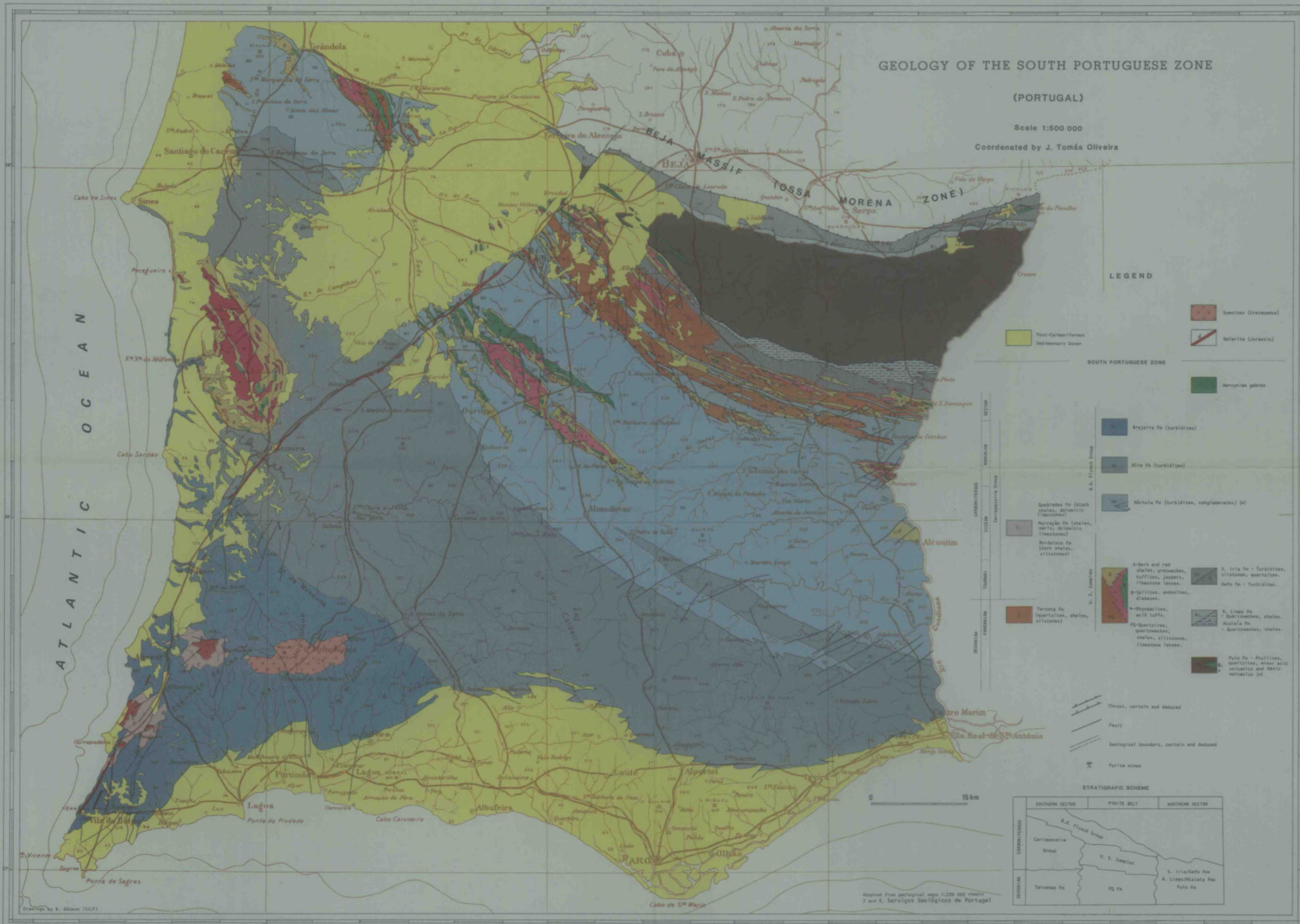

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