LEIDSE GEOLOGISCHE MEDEDELINGEN, Deel 52, Aflevering 1, pp. 23-56, 1-7-1981

FRAMEWORK AND EVOLUTION OF HERCYNIAN MINERALIZATION IN THE IBERIAN MESETA

BY

L. J. G. SCHERMERHORN*

ABSTRACT

The Hercynian cycle, starting in Late Precambrian times and terminated at the end of the Palaeozoic, is associated in the Iberian Peninsula with the deposition of a wide variety of metallic and nonmetallic mineral resources. The most famous of these are the base-metal sulphides of the Iberian Pyrite Belt (Rio Tinto and other deposits), tin and tungsten (Panasqueira), and mercury (Almadén).

The depositional stage of the Hercynian cycle saw the accumulation of syngenetic mineral deposits, resulting from the interplay of palaeogeographical, sedimentary and volcanic controls. During and after the following orogenic stage, epigenetic minerals originated through magmatic activity, mostly as direct deposits from magmatic-derived fluids and also indirectly through thermal activation of existing rock. In both stages felsic magmatism was the dominant agent of mineralization, both for the more important volcanogenic and for the plutonic mineral deposits.

Framework and evolution of Hercynian mineralization are defined by the geotectonic intraplate – not plate-margin – setting of the Meseta and by its palaeogeographical and structural development during the cycle, modified by regional and local factors, foremost among which are volcanic and plutonic heat and mass transfer.

Metallogenetic provinces and epochs are distinguished, metallotects outlined, and possible sources for the introduced ore elements discussed.

CONTENTS

*) Prof. Dr. L. J. G. Schermerhorn, Institut für Mineralogie, Freie Universität Berlin, Takustrasse 6, D-1000 Berlin 33, B.R.D.

INTRODUCTION

The Iberian Peninsula is well endowed with mineral resources, a large part of which were created during the Hercynian cycle, that is, the geological cycle of sedimentation, magmatism and tectonism leading up to the Hercynian orogeny which terminates the cycle at the end of the Palaeozoic (Fig. 1). The Iberian Meseta, or Hesperic Massif, occupying a large part of the Peninsula, was consolidated during the Hercynian orogeny. It is the site of a great variety of metallic and nonmetallic mineral deposits. Three world-famous Iberian ores belonging to the Hercynian cycle spring to mind: the base-metal sulphides of the Iberian Pyrite Belt, with Rio Tinto as their most celebrated representative, the tin-tungsten belt, with Panasqueira as the most important tungsten mine outside the Far East, and the mercury of Almadén.

Two factors dominate the context of mineral deposition in the Iberian Peninsula during the Hercynian cycle. They are, first, the long-drawn evolution of the Hercynian geosyncline, starting in Late Precambrian times, with volcanism occurring throughout, and, second, extensive granitic intrusion.

The depositional stage, when the geosyncline was being filled, provided the setting for the accumulation of syngenetic mineral resources, ranging from sedimentary ironstone to volcanogenic polymetallic sulphides. During the subsequent orogenic stage the earlier deposits were deformed, metamorphosed and intruded. The deformed geosynclinal rocks served as hosts to a wide range of epigenetic mineral deposits, from barite and fluorite to the famed Iberian tin and tungsten lodes. The geosynclinal mineral deposits are the result of the interplay of palaeogeographical, sedimentary and volcanic controls, and the orogenic and postorogenic minerals arose from magmatic activity, directly as plutonic deposits or indirectly through thermal activation of existing materials. Thus were produced a great variety of mineral deposits, some of them on a grand scale indeed: the Iberian Pyrite Belt contains the largest massive sulphide bodies in the world, the Panasqueira mine is the largest tungsten producer in Europe, and Almadén is by far the vastest mercury concentration known.

Hercynian geosynclinal and orogenic evolution in Europe and North Africa is of a different type than the evolution of the Alpine and other orogens related to lithosphere subduction, though this is still an endlessly debated point. This difference however is undeniable as it makes itself felt in a number of ways, one of which is the scarcity of Hercynian ultramafic rocks and in particular the absence of true ophiolite, that is, oceanic crust complexes. As a result, the minerals associated with ultramafic magmatism, such as the ores of nickel and chromium, are rare in the Iberian Meseta.

Geotectonic and structural setting

The sedimentary, volcanic and tectonic history of the Meseta during the Precambrian and Palaeozoic is very complex and has not yet been completely unravelled. There is a clearcut zonation across the orogenic trend, Lotze's well-known subdivision into six zones which has suffered little change in the 34 years since it was first propounded. Julivert et al. (1974) point out that the significance of these zones is essentially palaeogeographical: sedimentary facies and thicknesses change from one zone to another but tend to remain constant within each zone, that is, along the strike. The zones represent not only different palaeogeographical domains and different stratigraphic and sedimentary facies but also more or less different tectonic, metamorphic and magmatic styles together with different mineralizations (Fig. 2 – this figure follows the adaptation by Julivert et al. (1974) of Lotze's original scheme).

The Hercynian geosyncline in Iberia was a multiplex, ceaselessly shifting assemblage of subsiding basins, more or less stable platforms and geanticlinal uplifts, evidencing pronounced vertical mobility of the basement underlying the geosyncline. Positive epeirogenic movements in a region at one time may turn to negative soon after, and so on. In consequence, formation thicknesses can vary greatly between regions. Moreover, volcanism took place in all periods from the Precambrian and Cambrian on (Fig. 1).

LATE PRECAMBRIAN

The Hercynian cycle in the Iberian Peninsula begins when the Hercynian geosyncline formed and sedimentation started, on a floor of older Precambrian rocks, some time during the Late Precambrian.

Upper Precambrian strata crop out in several places in the Meseta. In the south they are known generically as the 'Serie Negra', a slate-greywacke succession containing thin layers and lenses of black quartzites and cherts, and some limestone. Locally intermediate volcanics appear at the top of the sequence (Vegas, 1974). The Serie Negra is overlain by the Lower Cambrian, and the junction is conformable, disconformable or locally even unconformable. Similar slate-greywacke successions appear elsewhere in the Meseta below Cambrian or probable Cambrian beds and may contain felsic volcanics at or near their top. Little is known as yet of the relationship between the Upper Precambrian successions and the older schists and gneisses of the crystalline basement to the Hercynian geosyncline.

No syngenetic mineralizations have been recorded from the Upper Precambrian (meta) sediments, and their slate-greywacke lithology – often called flysch-like – does not favour the occurrence of synsedimentary ores of any importance. Nor are important orebodies known from the volcanic intercalations, with the possible exception of the copper deposits at Arinteiro and Fornás.

Late Precambrian (?) copper sulphides

The Arinteiro and Fornás cupriferous pyrrhotite deposits near Santiago de Compostela in Galicia, in exploitation since 1975, are metamorphosed sulphide lenses interbedded with garnetiferous amphibolites in the polymetamorphic Ordenes complex. The rocks and ores underwent polyphase deformation and were metamorphosed to the hornblende granulite facies, suffering retrogradation to the amphibolite facies (van Zuuren, 1969). The Ordenes complex consists of paragneiss and schist (mostly original greywackes according to van Zuuren (1969)) enclosing amphibolites in the lower levels. Its age is considered to be probably Late Precambrian (E. Den Tex, pers. comm.). The occurrence of thin beds of black quartzite (van Zuuren's graphite-bearing

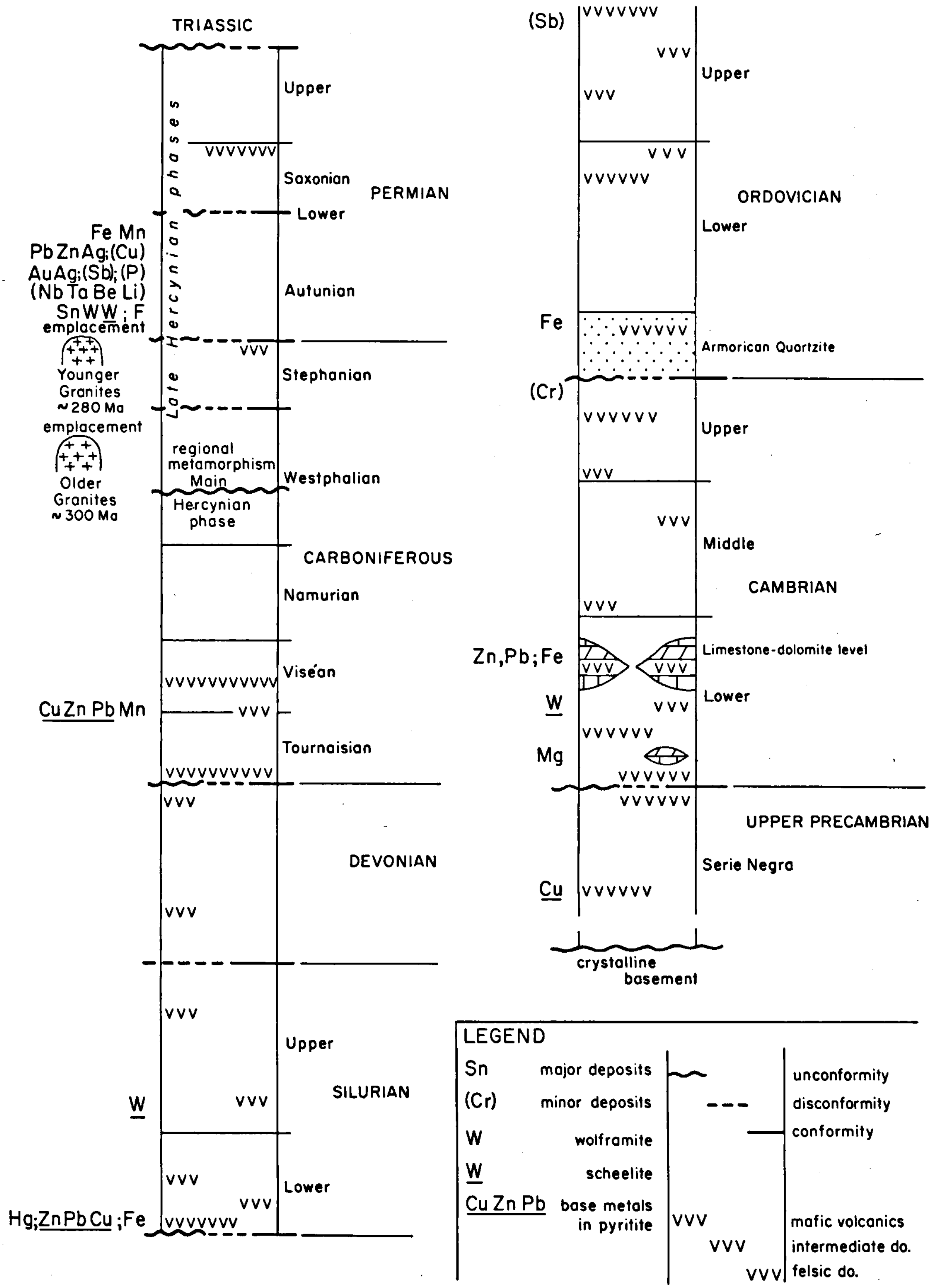

Fig. 1. Generalized section illustrating timing and evolution of mineralization in the Iberian Meseta.

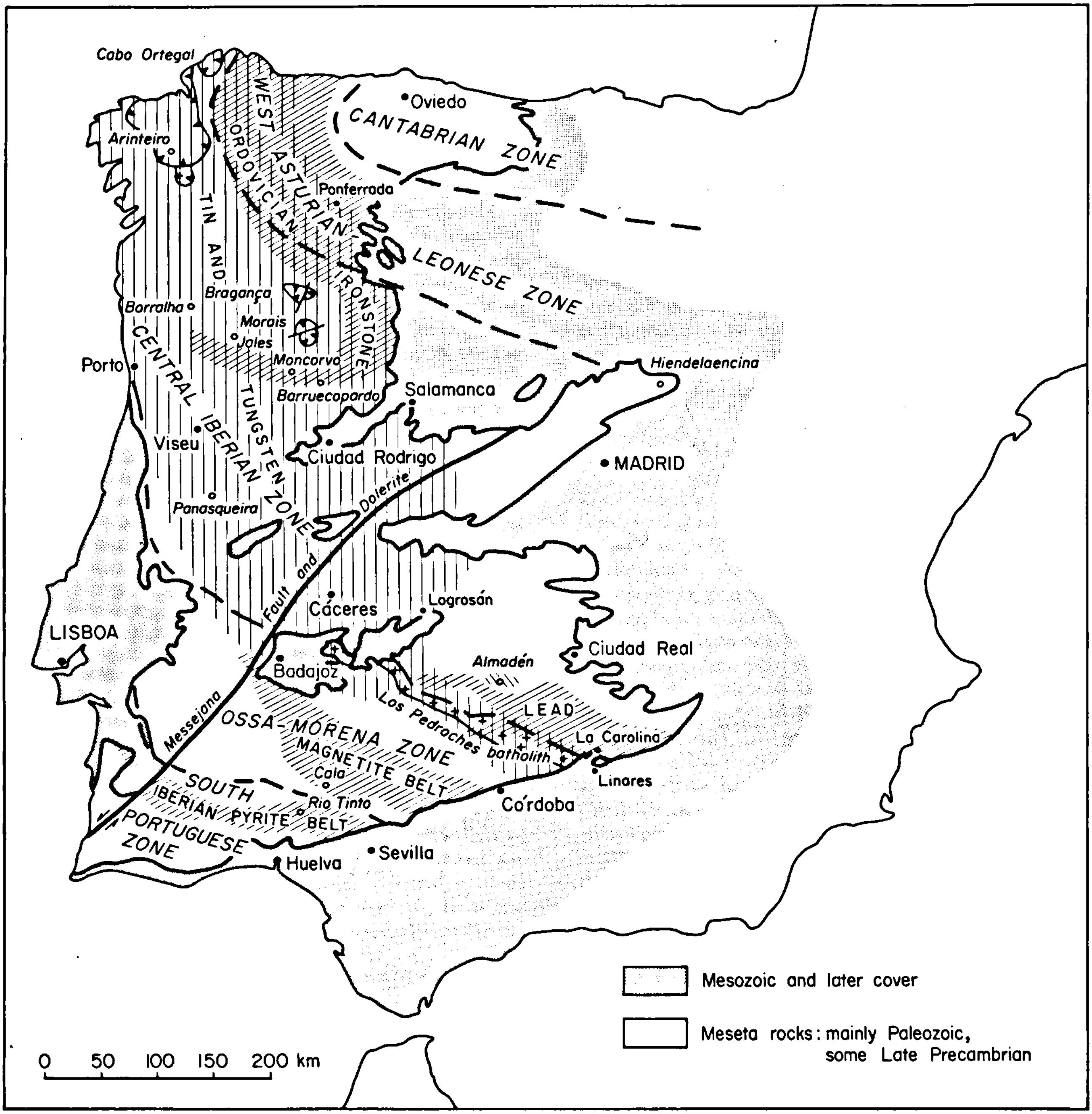

Fig. 2. Principal metallogenetic provinces in the Iberian Meseta.

schists) does in fact recall the black quartzites and cherts of the Serie Negra (unless they are metamorphosed Silurian black cherts).

At Fornás three stacked lenses of disseminated to fairly massive sulphides (pyrrhotite with chalcopyrite and pyrite), together 30 m thick by a strike length of about 200 m, are enclosed with sharp contacts in unmineralized garnet amphibolite. They total 750,000 tonnes grading 1.02% Cu. Arinteiro is a much larger orebody of pyrrhotite with chalcopyrite and pyrite disseminated in garnet amphibolite in the east flank of an anticlinorium. The west flank contains a similar orebody, Bama. Sulphide mineralization is found in other places in this structure as well (Chabod et al., 1976).

At Arinteiro and Bama the mineralization is linked to a particular level of garnet amphibolite 20–80 m thick. The three sulphide deposits add up to almost 25 million tonnes of copper ore reserves grading 0.66% Cu (White, 1977a). The Fornás ore is much more massive than Arinteiro and Bama, grading 20–30% S as against 5–7% S for Arinteiro (N. Rhoden, pers. comm.), but is still moderately diluted by gangue minerals such as anthophyllite or gedrite and other silicates, with little garnet (Ypma, 1966).

From their field aspect and their simple composition (iron sulphide with chalcopyrite and little else), these deposits appear to be original syngenetic stratabound cupriferous pyrite mineralizations in submarine mafic volcanics. Subsequent recrystallization under high-grade conditions caused the coarse grain of the sulphides, destroying any original depositional structures and textures, and the alteration to pyrrhotite of original pyrite.

As the ore-bearing amphibolites of the Ordenes complex are mafic to intermediate metavolcanics interbedded among metasediments, the Fornás, Arinteiro and Bamba deposits may be classified as belonging to the Besshi type of volcanogenic sulphide mineralization (see discussion of sulphide-deposit classification under Lower Carboniferous base-metal sulphides).

CAMBRIAN

The Cambrian of the Iberian Meseta is generally very thick and Cambrian volcanism is well developed, especially in south Portugal and southwest Spain.

The Lower Cambrian (Georgian) is a sandy-slaty sequence with a limestone-dolomite level in its upper part that is thick and continuous in the south, thinning northwards and thickening again in the northern Meseta (Schermerhorn, 1955; Julivert et al., 1974; Schmitz & Walter, 1974; Walter & Schmitz, 1975). This carbonate level is locally interbedded with mafic or felsic volcanics (Guillou, 1971a; Vázquez Guzmán & Fernández Pompa, 1976), and volcanics are also found near the base of the Cambrian in several places. Less important carbonates, the Cándana Limestone, occur in the lower part of the Lower Cambrian in northwest Spain.

The Middle Cambrian (Acadian) consists of slates and sandstones and contains spilites and some felsic volcanics. Where still present, the Upper Cambrian (Potsdamian) shows a sandstone-slate lithology. Fossiliferous Cambrian crops out in south Portugal, south Spain and northwest Spain. In addition, the Beira Schists and the Ollo de Sapo Formation are here regarded, in common with many Spanish and Portuguese geologists, as essentially Cambrian successions.

The Beira Schists of Portugal and equivalent formations in Spain occupy vast tracts in the central and western Meseta. The name Beira Schist Formation, or Beira Schists for short, is to be preferred since it has clear priority (Xistos das Beiras, Formação Xistosa da Beira: Delgado, 1907); this formation has also become known as 'xistograuváquico', an equivocal designation not in accordance with modern lithostratigraphical nomenclature. The Beira Schists and coeval formations in Spain (such as the Valdelacasa Series, the Villalba Series, the Porto Series and the upper Alcudiense) are generally unfossiliferous, but other arguments show their equivalence to, at least, the Middle and Lower Cambrian (Schermerhorn, 1955, 1956; Bard et al., 1972; Martínez García, 1973; Ribeiro, 1974) and locally they pass conformably into the basal Ordovician (A. Ribeiro, pers. comm.). They overlie Serie Negra rocks, and the precise chronostratigraphical age of the junction between Serie Negra and the Beira Schists or the fossiliferous Lower Cambrian is still uncertain.

The Beira Schists (and equivalent), like the Serie Negra, consist mostly of interbedded slates and greywackes – often called flysch-like (as also the fossiliferous Middle Cambrian of southeast-central Portugal and southwest Spain) – and represent the infilling of a large, fairly deep basin. The fossiliferous Cambrian of the northern and southern Meseta was largely deposited in shallow waters and it shows a much greater development of carbonates.

Characteristic for the Beira Schists and coeval formations are thin interbeds of quartz conglomerates made of rounded pebbles of quartz and quartzite, with rare clasts of black quartzite and chert derived from the Serie Negra. Characteristic too, and like the quartz conglomerates useful for correlation purposes, are amphibole schists of Garbenschiefer type. They occur as thin beds of metamorphosed impure calcareous sediments, consisting of quartz, calcic plagioclase and amphibole (often hornblende) poikiloblasts, frequently accompanied by garnet and other silicates. They have been described from the Beira Schists in Portugal (Schermerhorn, 1956) and from similar successions in Spain (Capdevila, 1969 (his fig. 5 shows a rock entirely similar to Schermerhorn's fig. 13); Martínez Carcía, 1973; Martínez García & Nicolau, 1973); these rocks are of interest because they host scheelite mineralizations (see below).

The Ollo de Sapo is a sequence in which coarse augengneisses are dominant, mostly consisting of metamorphosed feldspathic sediments, with interbedded felsic metavolcanics. This formation is exposed in northwest and central Spain. It is a lateral equivalent of Cambrian and Beira Schist rocks, and may extend into the basal Ordovician. Ollo de Sapo-like rocks occur at lower levels too, though their relationship to the Serie Negra is not always clear.

The Ollo de Sapo constitutes a regional lithofacies largely made of the freshly eroded debris of Precambrian porphyritic granites. Its deposition went on from the early Cambrian or earlier into the Ordovician, indicating the continuous existence during this period of a granitic source area within the Hercynian geosyncline.

Thus Hercynian geosynclinal sedimentation starts with a great thickness of terrigenous clastics, mostly sand and clay, deposited during the Late Precambrian and Cambrian. However, basin configuration was not uniform over the extent of the Meseta. One or more geanticlinal ridges exposing older Precambrian basement shed feldspathic detritus into the basin, and shallow-water conditions prevailed during most of the Cambrian along the northern and southern edges of the Meseta. Here carbonate deposition attained great importance during part of the Lower Cambrian.

The Cambrian carbonate rocks and the associated mineralizations are economically important. Though very extensive, the terrigenous clastics making up most of the fossiliferous Cambrian, the Beira Schists and the Ollo de Sapo do not contain syngenetic mineral deposits of any significance. Locally the quartz conglomerates are auriferous: Tamain (1971) refers to grades up to 16 g/t Au in the Valle de Alcudia in southern Spain. The slates, sandstones, schists and gneisses are quarried to some extent to provide local building and roadmaking material, but it is the Lower Cambrian limestones and marbles that are extensively exploited as ornamental stone and for industrial purposes in south Portugal, northwest and southwest Spain.

Local sedimentary differentiation to magnesium-rich rocks within the Lower Cambrian carbonates has led to the formation of several magnesite deposits in northwest Spain. The largest is worked at Pacios near Incio (Lugo province) and is one of the two active magnesite mines in the Peninsula. Here the magnesite is associated with dolomites belonging to the Cándana Formation and is of lagoonal origin (Doval et al., 1977).

Cambrian scheelite

Scheelite has been found disseminated in the thin layers of amphibole schist (sometimes called quartzite of gneiss) in the Beira Schists and equivalent sequences, as south of Salamanca in western Spain (Pellitero, Arribas & Saavedra, 1976). Though formerly ascribed to the epigenetic Sn-W mineralization generated by the Hercynian granites, discussed later, occurrences such as near Castro Daire in northern Portugal where scheelite is found dispersed in hornblende schist (outside the granite contact aureoles) over a strike length exceeding 400 metres (Schermerhorn, 1956, p. 77, 537) lead to reinterpretation as a syngenetic mineralization in marly sediments. This problem will be examined when reviewing the similar scheelite occurrences in Silurian calc-silicate rocks.

Cambrian lead-zinc sulphides

The Lower Cambrian carbonates form an important metallotect for Pb-Zn mineralization. In west and northwest Spain, Upper and Lower Georgian limestones carry syngenetic stratiform base-metal mineralizations sometimes associated with antimony and traces of mercury (Guillou, 1971a; Monseur, 1977). These are considered by Guillou to represent marine concentrations of elements leached and transported from a weathered landmass.

Rubiales in northwest Spain (Piedrafita del Cebrero, Lugo province), a recent discovery, is a zinc-lead mine. The ore is found in silicified limestones alternating with mudstone layers, in the dragfolded east limb of an anticline. Reserves are at least 12 million tonnes grading 8.1% Zn and 1.5% Pb (2.5–3% Pb has also been quoted) with some Cu and Ag (White, 1977b).

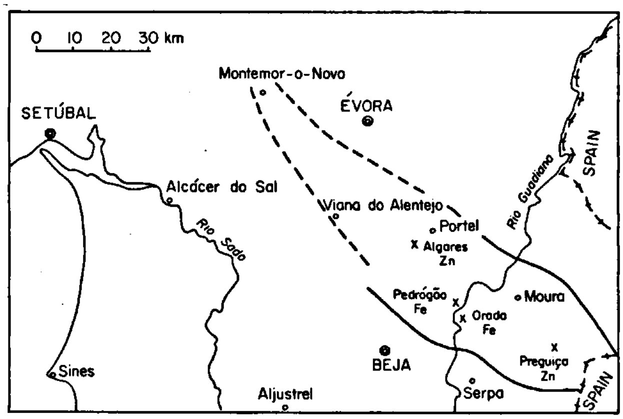

Fig. 3. The South-Portuguese zinc-magnetite belt. After Goinhas (1971).

Pn-Zn sulphides are widespread in the carbonates near the top of the Lower Cambrian in the southwest Meseta. In south Portugal a Lower Palaeozoic lead-zinc-magnetite belt (Faixa Zincífera e Magnetítica Alentejana, also known as Faixa Zincífera, or Zinc Belt, for short – Fig. 3) has been defined and thoroughly explored by the Serviço de Fomento Mineiro (Andrade, 1969, 1972; Goinhas, 1971; Carvalho et al., 1971a, b). This metallogenetic province continues into southwest Spain (Guillou, 1971). Here stratabound sulphide mineralizations occur as disseminations near or at the top of the carbonate horizon in the upper part of the Lower Cambrian (Upper Georgian), and as massive bodies at the base of disconformably overlying volcanics which also contain magnetite orebodies. The age of these volcanics is not yet known with certainty. In the opinion of the Serviço de Fomento Mineiro geologists they are Lower Silurian (or at most Upper Ordovician), and the Middle and Upper Cambrian and the Ordovician are lacking in the Zinc Belt (Goinhas, 1971; Carvalho et al., 1971a, b; Carvalho, 1976a). According to others the volcanism could be Middle Cambrian (Andrade, 1972; Guillou, 1971a). We will follow S. F. M. opinion and regard the Zinc Belt as a metallogenetic province where mineralizations dating from different epochs have become superposed. All rocks and ores have been strongly folded and metamorphosed during the Hercynian orogeny.

The carbonate level, here mostly dolomitic, is unique in the whole Late Precambrian-Palaeozoic succession of the southwest Meseta: it is a marker horizon and by dint of its singularity reliably correlated with the less metamorphosed fossiliferous Upper Georgian carbonates nearby in southwest Spain (Carvalhosa, 1965).

There are two types of sulphide mineralization in this mineral belt. The most frequent type, found at Preguiça, Ficalho, Balsa and other prospects, consists of stratabound accumulations of marmatitic sphalerite, argentiferous galena and pyrite in siliceous metadolomite horizons, with the sulphides mostly forming veinlets along microfractures. The other type, found at Algares, comprises massive orebodies discussed under Silurian metallization. At Vila Ruiva, near Preguiça, and elsewhere, primary sulphides were weathered, leached and redeposited on a karst surface developed in the Cambrian carbonates, possibly during the Tertiary. The ore consists of iron oxides and calamine, that is, zinc oxides, silicates and carbonates, such as smithsonite, hydrozincite $Zn_5(CO_3)_2(OH)_6$ and descloizite Pb (Zn, Cu) $(VO_4)(OH)$.

The disseminated sulphides in the dolomites of the Zinc Belt are thought to be synsedimentary mineralizations unrelated to volcanism. Later remobilization caused redistribution of the sulphides in veinlets following cracks in the dolomites, accompanied by silicification and redolomitization.

In the continuation of the Zinc Belt in southwest Spain the same two types of sulphide mineralization occur. According to Guillou (1971b) Pb and barite occur disseminated in Cambrian dolomites, accompanied by subordinate Zn and little Cu, Ag, Sb, while the overlying volcanics (felsic pyroclastics) contain sulphide lenses (pyrite with pyrrhotite, magnetite, sphalerite, chalcopyrite and galena) associated with carbonate and chert lenses. Guillou concludes to a syngenetic origin of both mineralizations: Cu, Pb, Sb, Hg and Ba were deposited with the carbonates at the top of the Lower Cambrian carbonate level, with the metals deriving from weathered land masses, while Zn – locally abundant, elsewhere almost lacking – is volcanogenic.

All in all, the age difference between the Cambrian dolomites and the overlying volcanics, whether Silurian, Ordovician or even Middle Cambrian, seems sufficiently significant that, taken together with their different modes of occurrence, the two types of mineralization must belong to two different periods of deposition and represent different environments. Cambrian carbonates are a metallotect for lead-zinc ore, and it seems likely that the end of the period of Lower Cambrian carbonate sedimentation became favourable for the deposition of terrigenous-sedimentary lead and zinc sulphides. Concentration instead of dissipation of Pb and Zn in the Cambrian sea could have been brought about by suitable sedimentary and palaeogeographical configurations.

Cambrian (?) mercury

At Usagre in southwest Spain (near Burguillos del Cerro) was a mercury mine working stratiform cinnabar impregnations along three horizons in Lower Cambrian limestones, accompanied by cinnabar veinlets; other minerals included pyrite, galena, chalcopyrite, barite and quartz. It is not certain whether this mineralization is syngenetic or epigenetic, postdating the Cambrian. Vázquez Guzmán and Fernández Pompa (1976) think it may well be of pre-orogenic age and related to volcanism, with subsequent remobilization due to Hercynian tectonism.

Cambrian iron

No important iron mineralization is found in the Cambrian or supposed Cambrian sequences of the northern and central Meseta. However, iron assumes considerable economic significance in the south.

In southwest Spain a magnetite belt extends from near Córdoba to south of Badajoz (Vázquez Guzmán & Fernández Pompa, 1976), continuing in Portugal in the Elvas region (Fig. 2). In this metallogenetical province magnetite deposits of skarn type occur in Lower Cambrian carbonates within the contact aureoles of Hercynian granites to quartz diorites.

Thus, in the Elvas region of central Portugal, near Spain, the Alagada deposit of stratiform magnetite bodies, totalling one million tonnes at 40% Fe, 12% SiO_2 and 0.73% S, is enclosed in Lower Cambrian marbles, calcsilicate hornfelses and skarns in contact with a Hercynian granite (Carvalho et al., 1971a; Carvalho, 1976a).

In Spain several iron mines are found in this belt (San Guillermo near Jerez de los Caballeros, Monchi near Burguillos del Cerro, Cala, Teuler, etc.). The ore is magnetite, sometimes hematite and rarely siderite, and averages 48–52% Fe, with sulphur and silica as principal impurities (1% S or over); the phosphorus content is negligible (Doetsch, 1967).

The most important deposit is Cala, with over a hundred million tonnes of reserves. Here magnetite associated with copper sulphides is worked by opencast and underground mining, producing Fe and Cu. The magnetite ore forms conformable levels and lenses and is banded (5–30 cm), with layers and zones rich in pyrite and chalcopyrite. The deposit occurs interbedded in limestone-quartzite-slate country rock and is associated with much skarn, near granite.

The origin of the mineralization in this belt is uncertain. Many authors favour a pyrometasomatic mode of formation, with the iron and the sulphides deriving from intruding magma, but a syngenetic sedimentary or volcanic origin modified by later metamorphism caused by granite intrusion has also been defended (Doetsch, 1967). In favour of syngenetic deposition is the occurrence of minor iron ore as folded layers older than the granites. Elsewhere there are strong signs of epigenetic mineralization, such as the formation of mineralized skarns in a carbonate horizon that outside the contact aureole consists of rather pure limestone only. More study is needed.

Early Palaeozoic chromium

Ultramafic rocks, the source of chromium, are rare in the Meseta. They can be divided into two groups: 1) the Early Palaeozoic ultramafics – peridotites and serpentinites – that occur associated with mafic rocks and paragneisses in a number of high-grade massifs in the northwest corner of the Meseta (Fig. 2), and 2) ultramafics occurring associated with mafic intrusives and extrusives of ages varying from Early to Late Palaeozoic; these are discussed on a later page.

In the northwest Meseta are found the Cabo Ortegal, Santiago de Compostela (only mafics exposed), Sobrado, Mellid and Lalín mafic-ultramafic complexes in Galicia, and the Bragança (or Bragança-Vinhais) and Morais mafic-ultramafic complexes in Trás-os-Montes (northeast Portugal). The best known is the Cabo Ortegal complex which consists of amphibolite, granulite and eclogite-facies ultramafics, mafics and paragneisses. These complexes are mushroom-shaped intrusions and they are seen as mantle diapirs whose emplacement was attended by high-grade metamorphism and various other effects (van Calsteren et al., 1979). The ultramafic rocks (lherzolite, harzburgite and dunite) are derived from the mantle. Rb-Sr whole-rock isochron dating of Cabo Ortegal lherzolites indicates that intrusion took place about 500 ± 100 Ma ago (van Calsteren et al., 1979).

Little chromite occurs in these complexes, most perhaps in northeast Portugal where the Abessedo mine in the Bragança massif has produced some thousands of tonnes of ore during the last world war. Extraction was on a small scale and no chrome mines are working any more, for the chromitite deposits are small, irregular and mostly rather low-grade.

The chromite occurrences in the Bragança and Morais complexes in Portugal have been described by Cotelo Neiva (1947a). Chromitite is present as vaguely bounded pockets, lenses and layers in peridotite and serpentinite. The Abessedo mine worked chromitite pockets grading 40–48% Cr_2O_3; some pockets yielded up to 1500 t chromite. In these ores chromite is associated with serpentinized olivine.

Platinum occurs as minute inclusions in chromite (Cotelo Neiva, 1947a, b).

ORDOVICIAN

During the Upper Cambrian a phase of folding, tilting, uplift and erosion, often called the Sardic phase, affected the Cambrian over a considerable part of the Meseta, after which the Ordovician transgressed (Schermerhorn, 1955, 1956). Thus the Cambrian-Ordovician junction is conformable, disconformable or unconformable in different localities.

The Lower Ordovician generally begins with quartzites that reach a few hundred metres in thickness, locally underlain by conglomerate. This level is the so-called Armorican Quartzite, well developed in central and west Iberia but thinning out in southern Spain and Portugal, where Silurian overlies Cambrian carbonates, and in north Portugal and northwest Spain. In a few places the quartzites enclose interbedded volcanics.

The quartzites are followed by a sequence of dark trilobite slates, and the Upper Ordovician contains slates, sandstones and locally limestones. Felsic and mafic volcanism took place during the deposition of these sequences. Towards the end of the Ordovician tectonic movements caused local uplift and emergence, and the Ordovician-Silurian junction, though mostly conformable, can be locally disconformable or even unconformable. During the late Ordovician some granites and peralkaline massifs were emplaced in east-central Portugal and western Galicia (Priem et al., 1970).

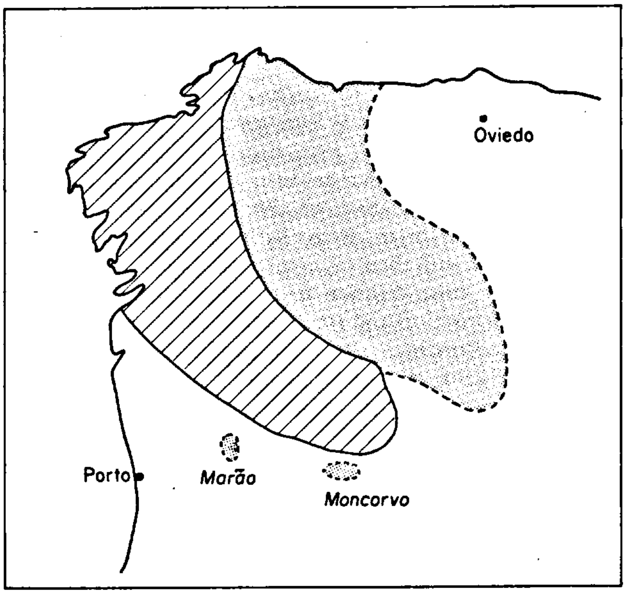

Fig. 4. Distribution of Lower Ordovician ironstone deposits around an emerged landmass in the northwest Meseta. Stippled: zone in which ironstone occurs; hatched: probable land during Ordovician. After Ribeiro (1974).

Ordovician iron

The Lower Ordovician quartzites of the northwest Meseta often enclose stratabound marine-sedimentary iron-ore deposits, especially near or at their top (Ribeiro & Rebelo, 1971; Armengot de Pedro & Campos Juliá, 1971; Ribeiro, 1974). The largest deposits are near Ponferrada in northwest Spain (Coto Vivaldi, Coto Wagner mines) and Moncorvo and Marão in north Portugal. The ore is mostly oxidic, sometimes sideritic or chamositic. Guadramil in northeast Portugal is a deposit of oolitic siderite in this horizon. The Marão mine exploited magnetite interbedded at the top of the quartzites and in the overlying slates; the ore is metamorphic (Priem, 1962).

By far the largest occurrence is M o n c o r v o in northeast Portugal, which has been intermittently worked since the 13th century and is to be developed into a 1.5 million tpy producer in the near future. The reserves exceed 550 million tonnes and may be as high as one billion tonnes (J. A. Rebelo, pers. comm.) though the average grade is low (36% Fe) and the phosphorus content (0.4–0.5%) troublesome. The main deposit is 8 km long by 1 km wide and occurs in the north limb of the Moncorvo synclinorium. A smaller deposit, 0.5 by 1 km in size, is found in a syncline nearby. These deposits have been folded and suffered regional and contact metamorphism during the Hercynian orogeny. The ore is mainly hematitic, including specularite and martite, and rather siliceous. Below a depth of 80–100 m magnetite predominates (Rebelo, pers. comm). Banded ore/quartzite alternations are frequent. The orebodies occur as distinct levels in the upper part of the Armorican quartzite formation and pass laterally into quartzites. They were deposited in a shallow sea, apparently as crossbedded iron oxide sands (Thadeu, 1952, 1965a). The Ordovician iron ores of northwest Spain and north Portugal are low to medium grade (up to 55% Fe), contain varying proportions of siliceous and aluminous gangue, and are characterized by phosphorus contents, in apatite and collophane, that may exceed 0.8% (Doetsch, 1967; Carvalho, 1976a).

They show a sedimentary facies indicating deposition in a shallow epicontinental sea and owe their existence to the palaeogeographical configuration that arose as a result of the Upper Cambrian tectonic movements (Sardic phase). The northwest corner of the Meseta remained as an emergent landmass during the Ordovician while around it the Lower Ordovician quartzite transgressed over more or less eroded Cambrian strata. The iron-ore deposits in the Lower Ordovician surround this landmass, according to Ribeiro and Rebelo (1971) and Ribeiro (1974), the source area for the iron which they think may have been leached from weathered mafic rocks (Fig. 4).

The iron precipitated in coastal waters, mostly as oxides, in places as silicate or carbonate, dependent on local physicochemical conditions. After some current transport, the ore became concentrated in local accumulations.

Spain has no large iron ore production, unlike for instance France, Sweden or Russia. The largest mine, the Marquesado open pit in the southeast, produces 3 million tonnes per year of hematite-goethite ore in Triassic limestones, from reserves estimated at 100 million tonnes grading 55–56% Fe. Spanish production figures indicate a decline in iron ore output: 7,684,000 t ore (50% Fe) was produced in 1976, 10.8% below 1975 production, itself 4.3% below 1974 (Mining Mag., August 1977). Portugal produced 51,989 t magnetite, hematite and ferromanganese ore in 1977 (Bol. Minas (Lisboa), December 1977). These figures illustrate the potential importance within the Iberian context of the Moncorvo orebody with its enormous reserves.

Ordovician antimony

In northwest Spain antimony has been mined from several small deposits associated with Upper Ordovician (Ashgill) carbonate lenses. This is interpreted as a synsedimentary mineralization in a shallow-marine environment, remobilized and redistributed during the Hercynian orogeny. The origin of the Sb could be sought in earlier Ordovician volcanism, with pedological concentration during subsequent emergence, and redeposition along the coast following erosion (Guillou, 1969, 1971a).

SILURIAN

The Silurian of the Meseta usually presents a monotonous facies of dark graptolite slates with frequent interbeds of sandstone, limestone and black chert (lydite). The Upper Silurian often is sandy. In south Portugal the Lower Silurian is highly volcanic, with abundant levels of mafic and felsic extrusives. In northeast Portugal and northwest Spain the Silurian is developed as a thick eugeosynclinal facies of flysch-type turbidites, greywackes and slates containing black chert layers, with plentiful felsic and mafic volcanics.

Near the base of the Silurian occur important mercury, iron and base-metal mineralizations. At higher levels stratabound scheelite is found.

The Silurian-Devonian junction is generally a little disturbed transition.

Silurian mercury: Almadén

The celebrated mercury deposit of Almadén (i.e. 'the mine' in Arabic) is by far the largest and richest known Hg accumulation on earth. It is a stratiform mineralization consisting of three cinnabar-impregnated quartzite layers, each about 10 m thick, within the quartzite formation known as 'cuarcita del criadero' (ore-deposit quartzite). This quartzite horizon is the host of the ore at Almadén and several much smaller occurrences in the area.

The grade at Almadén reached 20% Hg (cinnabar and native mercury) in some places in the richest lodes. Even richer is the recently (1975) discovered cinnabar deposit of El Entredicho, 17 km from Almadén, where ore consisting of 50% cinnabar occurs (Bol. Geol. Minero (Madrid), vol. 89, 1978, p. 188). The average grade of the Almadén ore worked during the first quarter of this century was 8% Hg, around 1940 it was still 6–7% Hg and it is now around 1.5%. No other mercury deposit anywhere has ever approached these extremely rich grades. Almadén accounts for 18% of world mercury production.

The cuarcita del criadero is by analogy considered to be of Early Silurian (Valentian) age. In the Almadén area, mafic volcanism started before the deposition of this quartzite formation and continued afterwards during the Silurian and Devonian.

The cinnabar, accompanied by native quicksilver and some pyrite, shows sedimentary features indicating its early introduction, before diagenesis (Saupé, 1976). Saupé suggests a direct genetic relationship between the Hg mineralization and the Silurian volcanism which reached its greatest development in the mineralized area, in the sense that the Lower Silurian volcanic activity set into motion the convection of solutions releasing mercury preconcentrated by adsorption in Ordovician carbonaceous muds and transporting it into overlying Silurian sands where the mercury was trapped and redeposited by sulphidization. Maucher (1976) considers such an origin unlikely and proposes a volcanic source for the Hg.

Silurian base-metal sulphides and magnetite

In the lead-zinc-magnetite belt of south Portugal (Fig. 3) and southwest Spain, as discussed, massive sulphide and magnetite orebodies occur associated with volcanics overlying the Lower Cambrian carbonate level. These volcanics, in south Portugal, are greenstones or greenschists (intermediate to mafic metavolcanics) enclosing lenses of marbles and calcschists. This formation passes up into felsic metavolcanics, likewise associated with calcareous metasediments but not containing orebodies.

At the Algares zinc prospect in south Portugal the two types of sulphide mineralization characteristic of this metallogenetic belt (see above) are met with: in depth, beneath a gossan, a massive pyritic mineralization overlies a disseminated Pb-Zn mineralization. The first consists of banded to almost massive orebodies made mostly of pyrite, with subordinate pyrrhotite, magnetite, sphalerite, barite, and some chalcopyrite and galena. These occur at the base of presumably Silurian greenstones (intermediate to mafic metatuffs) which rest on Lower Cambrian dolomites containing sphalerite, pyrite and galena veinlets (Andrade, 1969; Goinhas, 1971; Carvalho et al., 1971a, b).

Magnetite occurs disseminated and as massive lenses in the intermediate to mafic metavolcanics and has been worked in several small mines in south Portugal (Nogueirinha and Monges near Montemor-o-Novo, Alvito, Pedrógão, Orada and others) the most important of which was Orada (Fig. 3) where mining ceased in 1971 (its original reserves were two million tonnes – Carvalho, 1971, 1976a). Magnetite may be accompanied by pyrite, locally plentiful, and pyrrhotite (Carvalho, 1971; Carvalho et al., 1971a, b). The largest orebody known reached 30 m thickness by a length of about 250 m.

The polymetallic pyritic lenses at the base of the greenstones and the magnetite-pyrite-pyrrhotite lenses within the greenstones are now mostly regarded as syngenetic exhalative-sedimentary deposits (Andrade, 1969, 1972; Carvalho, 1971, 1976a; Carvalho et al., 1971a, b).

The María Luisa mine near La Nava in the Sierra Morena is an old cupriferous pyrite exploitation now worked for copper with gold and silver as byproducts. The ore consists of pyrite with chalcopyrite and sphalerite, and in places magnetite and pyrrhotite. The minerals are concentrated in bands parallel to the metamorphic layering in the host rocks (greenstone, jasper and schist). Detailed studies of the complex paragenetical relationships by Vázquez Guzmán (1972, 1974) revealed the existence of syngenetic sulphide and magnetite mineralizations on which an epigenetic sulphides-magnetite assemblage has been superposed through skarn-type (re) crystallization (with diopside, actinolite and epidote) caused by diorite intrusions.

María Luisa and other occurrences in the same region originated as submarine exhalative-sedimentary sulphide accumulations in volcanic horizons and associated jaspers; locally too the jaspers contain manganese ore. The age of these ores is uncertain since their host rocks have not yet been dated. They were regarded as Middle to Upper Devonian by Vázquez Guzmán and Fernández Pompa (1976). Other authors, such as Monseur (1977), favour a Cambrian age. Still, since the syngenetic stratiform ores of these occurrences are clearly a volcanogenic (or at least volcanic-associated) sulphide-oxide mineralization, of the type found in the lead-zinc-magnetite belt, a Lower Silurian age seems possible.

In Galicia, the metasediments and metavolcanics surrounding the Cabo Ortegal mafic-ultramafic complex, formerly thought to be Precambrian, have recently been shown to be of Silurian and, possibly, Devonian age (Martínez García et al., 1975; van der Meer Mohr, 1975). In the Moeche area and elsewhere in the same zone occur small stratiform massive to disseminated orebodies containing pyrite, hematite and chalcopyrite, interbedded among Silurian metasediments and metavolcanics.

A s b e s t o s (tremolite-actinolite) has been worked on a small scale in the Arado do Castanheiro mine near Portel in south Portugal, in a small mass of serpentinized ultramafics associated with metaspilites belonging to the presumably Silurian greenstone formation overlying the Cambrian carbonates (Gaspar, 1971; Carvalho et al., 1971b).

Silurian scheelite
As mentioned above, scheelite disseminations are found in calc-silicate interbeds (amphibole schists: originally impure marls) in the greywacke-slate sequence of the Beira Schists in northern Portugal and western Spain. Similar tungsten mineralization occurs in the calc-silicate intercalations in the Silurian greywacke-slate facies of northeast Portugal and adjacent Spain (L. Ribeiro, 1971).

However, the Silurian metacalcareous interbeds containing this mineralization are unlike the amphibole schists of the Beira Schists in aspect: they range from fine-grained, thinly banded rocks consisting of quartz, intermediate plagioclase and varying amounts of garnet, actinolite-tremolite, epidote-group minerals, sulphides and other minerals such as diopside (Sousa, 1975; Noronha, 1976) to skarns (L. Ribeiro, 1971).

Scheelite may be of wider distribution in Cambrian and Silurian metamarly intercalations than was previously supposed. Carvalho (1971) refers to the occurrence of some dispersed scheelite in Silurian calc-silicate hornfelses near the old Orada iron mine in south Portugal.

In northeast Portugal the C r a v e z e s scheelite prospect is being actively explored. Mineralization extends over three calc-silicate horizons up to 34 m thick and up to 2.5 km long (Sousa, 1975; Viegas et al., 1976). Provisional tenors obtained by surface sampling before drilling started are as follows: 1.43–3.2 kg/t WO_3, that is 1134–2538 ppm W, and 420–770 ppm Sn.

It seems likely that most of the tungsten in these marls is syngenetic. It may have been deposited as scheelite or, as Noronha (1976) proposes, originally dispersed tungsten may have been remobilized by metamorphism to produce scheelite. In either case the syngenetic tungsten and accompanying tin may be of sedimentary or volcanic origin. However, though the Silurian contains volcanics in the vicinity of the scheelite beds, this is not so for the Cambrian. It is moreover difficult to explain the preference of syngenetic tungsten for marly sediments. More study is needed.

DEVONIAN

Over most of the Meseta the Devonian presents a variable epicontinental facies of interbedded sandstones, shales and carbonates not attaining any great thickness.

In northeast Portugal and part of northwest Spain the Devonian is a flysch-type greywacke-slate sequence and in the southwest Meseta (the Iberian Pyrite Belt) it consists of monotonous shales with quartzite and rare limestone beds. Devonian volcanism, mostly mafic, is rare. No important mineral deposits are associated with the Devonian.

EARLY TO MIDDLE CARBONIFEROUS

After the Devonian lull, the Carboniferous gave rise to extensive mineral resources of some importance in the Iberian economy. At the Devonian-Carboniferous boundary the Meseta became differentiated into a large central block which emerged and two flanking troughs in which subsidence and sedimentation continued. The central block englobes the West Asturian-Leonese, the Central Iberian and the Ossa-Morena Zones. The flanking troughs are the Cantabrian Zone in northern Spain and the South Portuguese Zone in south Portugal and southwest Spain (Fig. 2). These external basins are secondary geosynclines that were to a large extent fed by detritus from the central block, a geanticlinal source area. This block is composite, containing internal basins, though the Upper Palaeozoic sediments here deposited do not reach the great thicknesses attained by contemporary strata in the flanking troughs in northern and southwest Iberia.

As a result, marine Carboniferous up to and including Westphalian strata is well represented in the Cantabrian Chain and the South Portuguese Zone. In the Cantabrian Chain the Lower Carboniferous is very thin, a condensed marine succession of shales and limestones not bearing mineral resources. The South Portuguese Zone, on the other hand, is the site of the Pyrite Belt geosyncline and here a thick eugeosynclinal sequence was laid down, divided into three conformable groups. Devonian shales and quartzites, the Phyllite-Quartzite Group, are conformably overlain by Lower Carboniferous volcanics and sediments of Tournaisian and Early Viséan age, the Volcanic-Siliceous Complex which hosts the Pyrite Belt mineralization. It is covered conformably by a thick barren flysch sequence of turbidite greywackes and shales, the Culm Group which ranges from the Upper Viséan into the Westphalian A in south Portugal. Geosynclinal deposition was then brought to an end by orogeny (Schermerhorn, 1971a; Carvalho et al., 1971b, 1976).

Early Carboniferous base-metal sulphides: the Iberian Pyrite Belt

The Iberian Pyrite Belt (Fig. 5) is a metallogenetic province of polymetallic pyrite deposits, copper and copper-zinc disseminations, some copper veins and manganese orebodies, oc-

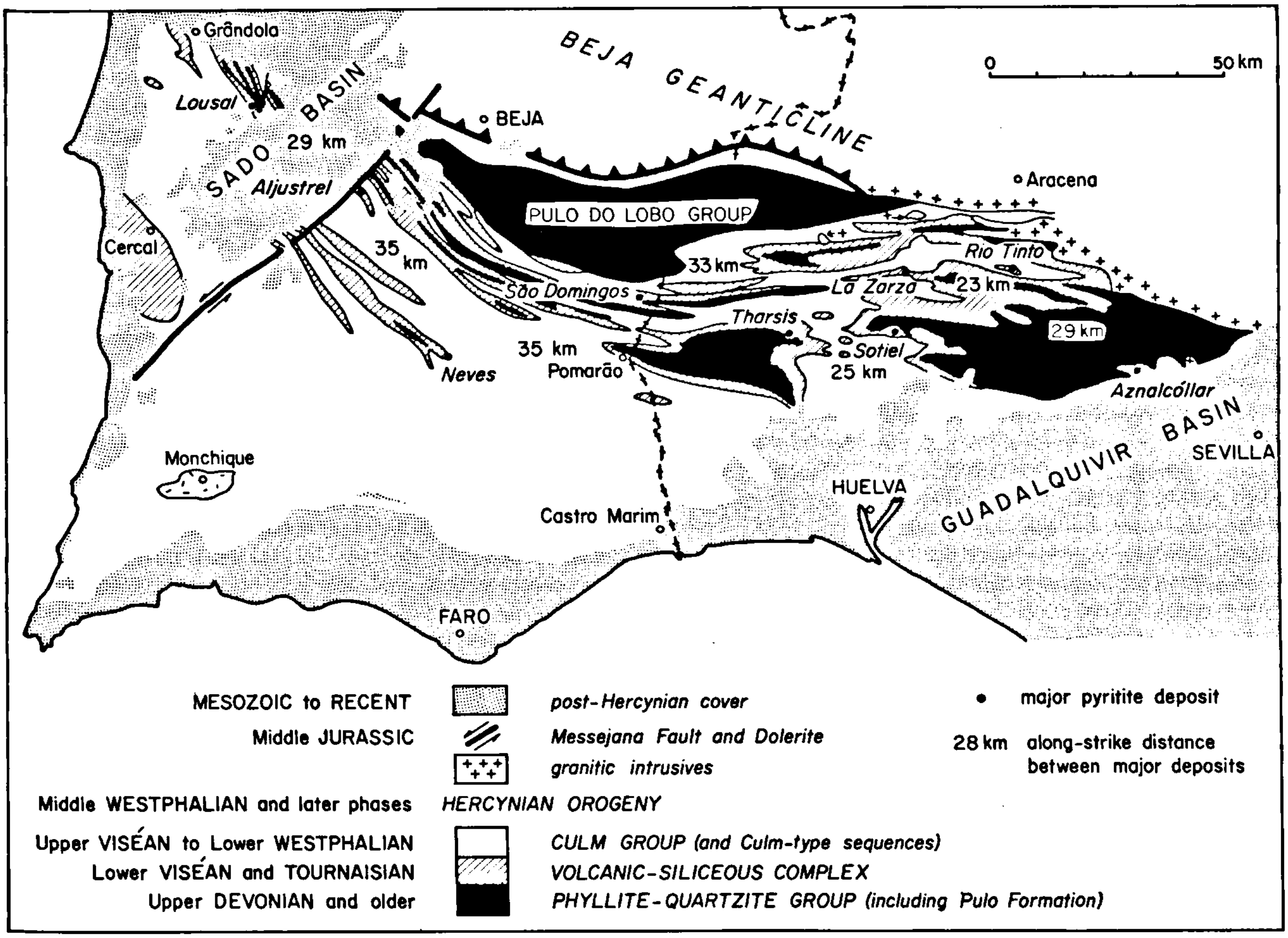

Fig. 5. Geology and major pyritite deposits in the Iberian Pyrite Belt. Note regular spacing of the large deposits. Adapted from Carvalho et al. (1976).

cupying a zone 230 km long by up to 30 km wide. Mining here goes back for at least 30 centuries, at first for copper and noble metals, since 1866 also for sulphur, with iron and other metals a byproduct. Annual production of pyrite ore is 2.5–3 million tonnes but new projects in Spain and Portugal aim to double this figure. The Spanish Pyrite Belt contains five large pyrite occurrences of which three are large mines, Tharsis, La Zarza and Rio Tinto, while Aznalcóllar will soon enter production and Sotiel Coronada is still under study. In addition there are several smaller workings. In the Portuguese Pyrite Belt three important occurrences, together with some smaller ones, are, or have been, worked: Aljustrel, São Domingos and Lousal. The annual production of Aljustrel is to be raised from 300,000 to 2 million tonnes.

The Iberian Pyrite Belt is the type area of the Iberian type of volcanogenic polymetallic pyritic deposits (as discussed at the end of this section). These are stratiform, associated with felsic volcanism and are composed of pyrite with small proportions of chalcopyrite, sphalerite, galena and other minerals. Massive pyritic deposits of this and other types consist of p y r i t i t e (Schermerhorn, 1970).

The Volcanic-Siliceous Complex consists of sediments (including prominent jaspers), felsic volcanics (largely quartz-keratophyric tuffs), mafic volcanics (spilite lavas and tuffs and diabase sills), sulphide deposits and manganese ore deposits. The sulphide mineralization and most of the manganese is genetically linked to the felsic volcanism in the geosyncline (Schermerhorn, 1971a; Strauss & Madel, 1974; Carvalho et al., 1976; Strauss et al., 1977).

The pyrite deposits of this belt are lenticular or sheetlike bodies conformably interbedded among volcanics and/or sediments: they overlie felsic volcanics or overstep onto shales and siliceous slates (dust tuffs) or may be wholly enclosed in shales, as at Tharsis.

The felsic volcanism was explosive, along submarine strike fissures (Schermerhorn, 1976), and generated mostly subaqueous flow tuffs, with few lavas and breccias. Most pyrite occurs at or near the top of felsic piles: it was produced at the end of an eruptive cycle when the explosive activity waned. In each mining centre the mineralization is related to an eruptive centre and occurs at one, or rarely more, stratigraphic levels. When two felsic cycles are present, each may have its pyritic mineralization, so there will be two ore levels (Strauss & Madel, 1974; Ramírez Copeiro del Villar, 1976). However, the timing of sulphide deposition varies between centres: it is relatively early, Tournaisian, in some mines, and rather late, Lower Viséan, in others where microfossils in associated limestones permit dating, as at Sotiel Coronada. As a result, the local volcanic and ore stratigraphic columns cannot be correlated from mine to mine: detailed stratigraphy at Rio Tinto is quite different from Tharsis to the west or Aznalcóllar to the east. This fact has sometimes been ignored.

The orebodies may be divided into autochthonous and allochthonous deposits. Some of the pyritite bodies enclosed in felsic volcanics are autochthonous, as at Rio Tinto, which is to say that they overlie feeder zones, now stockwork pipes strongly altered by chloritization and silicification (Strauss & Madel, 1974; Williams et al., 1975). This is relatively rare in the Pyrite Belt where most pyritite has suffered some transport during or shortly after its original precipitation on the seafloor as a very fine-grained pyritic mud, subsequently consolidating elsewhere as allochthonous stratiform deposits. The pyritite is usually thinly and rhythmically bedded, except in the autochthonous deposits overlying their own feeders. The frequent occurrence of internal structures denoting primary sediment movement, such as small slump structures, cut-and-fill, low-angle crossbedding and graded bedding, together with the varying lithological environment of the orebodies, from volcanic to sedimentary, led to the recognition of allochthonous pyritic masses mobilized and redeposited by gravity flow at some distance downslope from the feeder vents (Schermerhorn, 1970, 1971b). The pyritite is first precipitated on the seafloor around exhalative orifices as a sulphide gel containing pore water (seawater). Thixotropic gel-sol transformation causing liquefaction of sulphide muds is the mobilization mechanism, and the gravity gradient existing on the slopes of submarine volcanoes sets the instable mass in motion. This process, many times repeated, produces the well-bedded allochthonous pyritite lenses. Farthest travelled, away from the volcanic feeder vents, are the orebodies overlying sediments on the basin floor. Thus there is a gradation from proximal to distal pyritite deposits. All Pyrite Belt pyritite is exhalative-sedimentary in the sense that it was originally laid down as a chemical sediment, namely a hydrothermal surface precipitate. The transported allochthonous pyritite may then be termed exhalative-resedimented since it has been redeposited at some distance from the site of primary precipitation.

Usually the autochthonous pyritite bodies are large single lenses or sheets, as at Rio Tinto (Williams et al., 1975) or La Zarza (Strauss & Madel, 1974) while the allochthonous deposits often occur in clusters: five pyritite lenses at Tharsis North and eighteen at Lousal (Strauss & Madel, 1974).

The Iberian Pyrite Belt contains several pyritite deposits of giant size, the largest sulphide accumulations in the world. Most important was Rio Tinto where owing to folding and subsequent uplift and erosion most of the original deposit has been removed. It had been folded into an anticline, with pyritite left in the north and south limbs and as a gossan on its crest. Reconstitution by unfolding yields an original pyritite sheet measuring approximately 4.5 km long by 1.5 km wide and up to 80 m thick (Schermerhorn, 1971b). This leads to an original tonnage of roughly three-quarter billion tonnes.

Present reserves in the Pyrite Belt total 620 million tonnes (Carvalho et al., 1976). Largest are the reserves of Aljustrel (250 million tonnes) and Tharsis (130 million tonnes) while Rio Tinto still has 55 million tonnes. Taking into account the amount of pyritite removed by erosion, extracted by mining and still to be discovered, and applying Zipf's law, the total amount of sulphide deposited in the Iberian Pyrite Belt will have been somewhat over 2 billion tonnes.

The distribution of this enormous amount over the 230 km length of the Pyrite Belt presents a regular rather than a random pattern. The large pyritite deposits are isolated: where other orebodies are found in the same area they are very much smaller. A plot of the large deposits shows them to lie at roughly similar intervals (Fig. 5). The Neves-Corvo prospect, a large blind orebody was only discovered in 1977, and the apparent lack of significant mineralization in the interval between Aljustrel and São Domingos, which seemed twice as large as the normal strike distance between comparable pyritite bodies elsewhere in the Pyrite Belt, was an incentive to prospecting in that area. It thus seems that large accumulations of pyritite, 20 million tonnes and over, were only produced at distances between 23 and 35 km. La Zarza and Sotiel are separated by 12 km only but this is across the strike in a strongly folded area and the original distance must have been at least twice that amount. Between the major deposits numerous small deposits were generated in the Spanish Pyrite Belt but not – curiously enough – in the Portuguese half of this orefield where only a few small occurrences are known beside the four large deposits.

There are four main alignments of sulphide mineralization along the strike. The most northerly alignment is in Spain and comprises several small to medium-sized deposits such as Concepción. The next alignment joins Rio Tinto, La Zarza and São Domingos. The third alignment has Aznalcóllar, Sotiel, Tharsis and Aljustrel. The southernmost alignment is in Portugal and runs through Lousal and Neves.

Very approximately, the known pyritite reserves of 620 million tonnes present the following average composition:

sulphur	46 %
iron	40 %
copper	0.7%
zinc	2.9%
lead	1.1%
arsenic	0.6%
gold	0.8 g/t
silver	30 g/t

The remainder is largely siliceous, chloritic and carbonate gangue: as the high sulphur content shows, most pyritite in this province is massive sulphide ore of high purity.

In the very large deposits the copper tenor is low, 0.3–0.7%, though zinc can be relatively high, up to 4–5%. It appears to be a geochemical law in the Pyrite Belt that pyritite bodies exceeding 30 million tonnes, or over 40 m thickness, do not contain over 0.7% Cu. Higher copper tenors occur in smaller orebodies (up to 2% Cu) and locally as oreshoots in larger pyritite masses.

Vázquez Guzmán (1976) has established that primary magnetite and hematite, together with some siderite, appear in small quantities in pyritite, especially in the northern part of the Spanish Pyrite Belt.

The massive sulphides may be associated with disseminated sulphides constituting syngenetic (pre-folding) or epigenetic (post-folding) stockwork ores underlying pyritite deposits, or syngenetic stratabound ores in hanging wall or footwall tuffs and slates. These may contain workable proportions of chalcopyrite, sphalerite and galena, and reach large tonnages. This type of mineralization – sulphides diluted by volcanic material at the source – is common though not often economic.

Of considerable metallogenetic and explorational interest are two lithofacies that accompany pyritite: chloritite and jasper. Chloritite, first studied and named at Aljustrel, consists mostly of a fine-grained chlorite aggregate generally containing disseminated sulphides and sometimes carbonate; it may grade into chloritic tuffs. When cleaved it could be, and has been, mistaken for black slate. Chloritite bodies may be found in the footwall or sometimes in the hanging wall of pyritite deposits, and may reach 50 m thickness. At Aznalcóllar a massive pyritite orebody is overlain by tuffaceous chloritite carrying copper-zinc sulphides. Chloritite mostly occurs near or at the periphery of an orebody. In a similar peripheral position, generally beyond the ore boundary and slightly higher stratigraphically, are jasper lenses, silica rocks containing some chlorite and pyrite but mostly poor or lacking in iron oxides, unlike the red or black jaspers unrelated to sulphide mineralization.

Pyrite, chlorite and silica are intimately associated, as bodies of pyritite, chloritite and jasper, or intergrown and even interlaminated. This association is widespread the world over, constituting a volcanogenic sulphide-chlorite-silica association.

Chlorite and silica accompany sulphide precipitation both as subsurface replacements in the feeder zones underneath autochthonous pyritite and as surface deposits accompanying autochthonous and allochthonous pyritite. Chloritization and silicification of footwall rock in alteration pipes is caused by syngenetic hydrothermal activity around the feeder channels below the seafloor on which pyritite was deposited. Such replacement bodies show irregular, discordant shapes. Chloritite and jasper deposited on the seafloor, as volcanogenic muds (they may grade into tuff) and transported downslope along with allochthonous pyritite masses, are stratiform, concordant bodies. Chloritite deposits were formed immediately preceding, and sometimes succeeding, the phase of sulphide precipitation, while the jasper represents siliceous precipitation occurring slightly later in sequence, and travelling farther out.

Pyritite originates through an interaction between volcanogenic emanations and seawater that gives rise to the precipitation of colloidal sulphides on the seafloor (pyritite may show original colloform gel textures). There is a whole range of possibilities here, between the extremes of volcanic heat only plus seawater supplying all the components of pyritite, and of seawater only serving to chill upwelling hot ore solutions wholly of volcanic origin, to precipitate pyritite.

However, it appears likely, in the light of data on the isotopic constitution of sulphur at Rio Tinto (Williams et al., 1975) and elsewhere, that the sulphur in pyritite represents a mixture of juvenile sulphur associated with the felsic volcanism and of sulphur derived from seawater sulphate. Seawater circulating as pore water in the tuff piles building the submarine volcanoes and participating in convective hydrothermal systems driven by magmatic heat, would supply the sulphate sulphur component. The Iberian ore carries an extremely wide range of metals (Schermerhorn, 1971b): tin, for instance, is widespread, occurring as cassiterite, sometimes accompanied by stannite (Aye & Picot, 1976). The source of the metals would be volcanic too, either directly from the magmatic residue or by leaching from volcanics.

As the geology shows, pyritite is deposited at the end of a volcanic cycle, from hydrothermal solutions discharging on the seafloor and building up large volumes of very pure sulphides. During active volcanism the hydrothermal process cannot take place without interruptions and the sulphides produced are scattered. Moreover, the mineralizing fluids bringing the sulphides were accompanied by solutions from which chlorite and silica were deposited, within the vents (chloritization and silicification of footwall tuffs) or on the surface (primary chloritite and jasper). The Mg, Fe and Si needed for these processes are likely to be of volcanic origin too, either directly or indirectly.

The mafic volcanism of the Pyrite Belt – mafic to intermediate spilites and diabases – is completely distinct from the felsic volcanism, showing no transitions. Though the mafics are likely to be mantle-derived, such an origin seems less probable for the felsic volcanics, and the few known initial $^{87}Sr/^{86}Sr$ ratios favour a crustal origin (Schermerhorn, 1976). If such is the case, then the sulphur and metals brought up by the magmas either have an origin in remelted Precambrian rocks, or have been transferred from mantle-derived mafic magma to ascending deep-crustal felsic magmas, a problem awaiting resolution. Either type of source involves a pre-existing sulphur-metals concentration, to explain the vast tonnage of Pyrite Belt deposits.

According to their volcanic environment there can be distinguished three types of volcanogenic polymetallic pyritic deposits, grading from massive (pyritite) to disseminated.

1. The Cyprus type is found in the mafic (spilitic) lavas at the top of ophiolite complexes. The best known example is Cyprus and similar deposits have been described from Newfoundland and elsewhere. This type represents polymetallic sulphide mineralization in an oceanic environment, outside continental plates.

2. The Iberian type is linked to submarine felsic volcanics (often quartz keratophyres) and it comprises most of the important pyritite deposits through time and space, including the Iberian Pyrite Belt. This type may be found along or within continental plates.

3. The Besshi type is associated with non-ophiolite mafic and intermediate volcanics, along or within continental plates.

Though the Iberian type is sometimes called the Kuroko type, the kuroko ores form a subtype of infrequent occurrence and small tonnage. Briefly, the differences between the Kuroko and Iberian types are as follows:

1. Kuroko-type mineralization is restricted in time and space, (Miocene in Japan), unlike the widespread Iberian type (from the Archean on, all over the world);
2. Kuroko deposits are small, exceptionally over 10 million tonnes;
3. Kuroko is black ore, consisting mostly of sphalerite, galena and barite, with little pyrite and chalcopyrite: this is not base-metal-carrying pyritite such as makes up the bulk of the Iberian type deposits;
4. Oko, the yellow pyrite-chalcopyrite ore (pyritite) underlying most kuroko ore, is less well developed than kuroko, is often sharply bounded against kuroko, and is coarser-grained than most pyritite. In pyritite bodies of the Iberian type the transition from a copper-rich lower to a zinc-lead-rich upper part of the deposit takes place gradually;
5. Abundant barite and gypsum accompany kuroko and may even be more important than the sulphides but are rare elsewhere;
6. Kuroko is associated with lava domes and tuff breccias formed by vent explosions; true tuff is extremely rare. In the Iberian Pyrite Belt tuff is the normal felsic rock, tuff breccias are infrequent and lavas are rare;
7. Every lava dome in the Kuroko province has its kuroko deposit, and allochthonous kuroko has not been transported over great distances; kuroko is very rarely found in a black shale environment, unlike the Pyrite Belt where even very large deposits may be sited in shales.

Copper veins have been worked at many places in and near the Pyrite Belt: they are crosscutting epigenetic quartz veins containing chalcopyrite and some other minerals, and are post-orogenic since they cut folded, cleaved and metamorphosed Culm strata, the youngest Pyrite Belt group. This mineralization could well have originated by remobilization of cupriferous pyritite in depth.

Early Carboniferous manganese

Manganese ore deposits are far more numerous than sulphide occurrences in the Iberian Pyrite Belt though always much smaller: manganese mineralization was much more extensive but less intensive than pyritite mineralization. The total amount of manganese ore produced in this belt is about 5 million tonnes, from a few hundred occurrences. The last mines were abandoned some years ago.

Where found in the neighbourhood of pyritite, as at Aljustrel, manganese deposits are slightly younger than the sulphide mineralization. Most of them have been mined for secondary manganese oxides, derived from rhodochrosite and rhodonite in depth. They often are associated with jasper lenses.

The manganese deposits originated through seafloor precipitation of volcanogenic manganese oxides, hydroxides and carbonates, probably as a gel. Later metamorphic recrystallization formed rhodonite, braunite, spessartite and other minerals (Thadeu, 1965; Carvalho et al., 1971b, 1976; Cramer, 1976).

The manganese orebodies are small in comparison not only to the pyritite masses associated with the same volcanism but also to manganese deposits elsewhere. Their deposition as chemical sediments fed from a volcanic-hydrothermal source took place in an oxidizing environment, after pyritite deposition – under reducing conditions – had ceased or else at some distance from the centres of sulphide accumulation. Because the available Mn, an exhalative product, was not absorbed by sulphide precipitation, as was iron, it entered seawater to be dispersed, presumably as a colloidal hydrosol. Transport away from the volcanic sources under the influence of gravity and sea currents did not lead to complete dispersal by dilution with seawater, for no thin extensive beds of manganese ore are known. Rather, the manganese deposits form lenses strung out along a favourable horizon. Accumulation of manganese mineral gels, together with silica gels producing jasper, must therefore have depended on mass transport and settling in situ.

MIDDLE CARBONIFEROUS

Within the Westphalian the break takes place that separates geosynclinal from orogenic evolution in the Hercynian sequence, and mainly syngenetic from mainly epigenetic mineralization episodes.

Tectonic movements largely of epeirogenic character made their influence felt at various times during the development of the Hercynian geosyncline, leaving their traces as more or less widespread disconformities and unconformities. They did not involve Meseta-wide orogenic deformation. As we have seen, one of these tectonic phases, acting at the end of the Devonian, caused emergence of the central Meseta block and subsidence of the Cantabrian and South Portuguese external geosynclines. This phase is sometimes equated with the Bretonian orogeny and considered to be the first Hercynian orogenic phase. Succeeding phases are largely of the nature of epeirogenic movements, as for instance the uplift during the Viséan of the source area in the southwest part of the central block that gave rise to the deposition of the thick widespread Culm group in the South Portuguese Zone, the second most extensive sedimentary unit, after the Beira Schists, in the Meseta.

When full-scale orogeny set in with the Hercynian main phase, orogenic development superseded geosynclinal sedimentation in the two external remnants of the Hercynian geosyncline, the Cantabrian and South Portuguese Zones.

Hercynian orogeny: the main phase

The main Hercynian folding took place during the Westphalian (Schermerhorn, 1956, 1971a). The evidence for this is principally of two kinds: firstly, everywhere in the Meseta the older geosynclinal formations up to and including the Viséan and Namurian (where present) were folded before the deposition of the continental coal-bearing Middle to Upper Carboniferous strata (mostly Upper Westphalian and Stephanian), themselves deformed by later Hercynian phases. Secondly, this folding was accompanied by the development of cleavage and closely followed in many places by regional metamorphism, migmatization and the emplacement of the widespread Older Granites. Many of these have been radiometrically dated and their average age is around 300 Ma, that is approximately Middle Westphalian (Mendes, 1968; Priem et al., 1970; Leutwein et al., 1970; van Calsteren et al., 1979).

The succession of Hercynian phases and how they affected unfolded or already consolidated regions is of importance in the genesis of the Hercynian epigenetic mineral deposits.

Without entering into detail, it may be mentioned that the structural history of the Hercynian folding phases is still far from having been satisfactorily worked out. This is to a large extent due to the prevalence of coaxial refolding and the scarcity of time horizons that can be used to distinguish between discrete phases. It is therefore difficult to assess the extent of pre-main phase deformational episodes and whether they involved epeirogenic or true orogenic movements. So far, it appears that the large-scale fold structures of the Meseta were laid out during the main phase, possibly starting during some local earlier folding. Later tectonic phases consisted of block movements affecting crust that had already been consolidated by main-phase folding, cleavage, metamorphism and granite intrusion. These later phases only caused true folding in post-main phase sediments, and their effect on the earlier formations was tightening of main phase folds and differential vertical movements giving rise to faulting and horst-and-graben tectonics.

Pervasive folding and shearing, with the development of cleavage, was only associated with the main phase, possibly in several subphases. This is illustrated by the deformation of the pre-main phase granites of Lower Carboniferous and Upper Ordovician age which have become gneissose (Priem et al., 1970). The Older Granites, associated with the main phase, have not suffered similar tectonization, though they were subjected to later Hercynian phases. Thus only the main-phase deformation was forceful enough to shear and gneissify granite massifs.

Westphalian talc

At Puebla de Lillo in León, the main talc producer in Spain, Namurian limestone, the 'caliza de montaña', has been locally transformed into talc by dolomitization and later silica metasomatism during low-grade regional metamorphism when silica derived from adjacent detrital sediments was redistributed; most talc is found near the contact with overlying Westphalian shales (Galán-Huertos & Rodas, 1973).

As such, this important talc deposit may be classified as one of the few metamorphic mineral deposits in the Iberian Peninsula. Moreover, since the regional metamorphism is of Westphalian age, this is the age of the talc too.

Hercynian Older Granites

Granite plutonism took place in various episodes during the Hercynian cycle (Mendes, 1968; Priem et al., 1970):

1. Late Ordovician	460–430 Ma
2. Early Tournaisian	ca. 350 Ma
3. Westphalian	ca. 300 Ma
4. Earliest Autunian	ca. 280 Ma

The Westphalian and Autunian granites are distinguished as Older and Younger Granites (Schermerhorn, 1956), a distinction originally based on structural relationships and later confirmed by radiometric dating (Priem et al., 1970). These two suites present different characteristics. The Ordovician and Tournaisian granites are much rarer and are not associated with mineralization.

The designation of Older and Younger Granites serves to situate the two suites in time: it must be realized that either group contains units varying in texture, structure, mineralogy, composition and environment. The distinction between the two suites, based on crosscutting relationships, is a valid one as radiometric dating all over the Meseta has shown. The Older Granites have been equated with an alkaline suite, as leucogranites, in contrast to the Younger Granites as a calcalkaline suite (Capdevila, 1969; Oen Ing Soen, 1970; Capdevila et al., 1973) but this chemical subdivision only applies locally in the northwestern Meseta (Floor, 1970). Elsewhere, as in south Portugal and south Spain, the Older Granites often are calcalkaline rocks. Moreover, in the northwestern Meseta also many Older Granites are rich in biotite and show calcalkaline tendencies, with plagioclase as calcic as andesine and even labradorite (Schermerhorn, 1956). Also, many Younger Granites show an alkaline, leucogranite trend, containing abundant albite and muscovite.

Granites were emplaced over the width of the Hercynian orogen in the Iberian Peninsula, from the Cantabrian Zone in the north to the South Portuguese Zone in the southwest (according to Capdevila et al. (1973) granites are lacking in the latter zone but this is erroneous: granites (granodiorite to quartz diorite) occur in the Spanish Pyrite Belt). They are most abundant in the central block.

The Older Granites are late-syntectonic to post-tectonic with reference to the Hercynian main phase, show concordant to crosscutting relationships with their country rock (mostly schists), often occur in areas of medium to high-grade regional metamorphism, may be associated with extensive migmatization, and range from gabbroic to granitic with quartz diorite and granodiorite most common. They often show an alkaline character, with albitic plagioclase, and they contain biotite and muscovite, the latter frequently abundant. Yet calcalkaline rocks with oligoclase to andesine and little or no muscovite are also known, especially in the southern Meseta. In the northwest Meseta fine to mediumgrained two-mica granodiorites and granites are common but sparsely to fully porphyritic rocks are found too. These granites often are leucocratic, containing rather less dark constituents than the Younger Granites, but the Older Granites also include rather dark varieties, not only leucogranites. In south Portugal and south Spain occur much more mafic Older Granites ranging from gabbros and diorites to granites; most frequent are quartz diorites (tonalites) and granodiorites. Radiometric ages determined by Mendes (1968) are mostly around 303 Ma but a few older (about 360 Ma) and younger (280 Ma) rocks are present as well.

The initial $^{87}Sr/^{86}Sr$ ratios in the Older Granites of the Meseta range mainly between 0.709 and 0.711; lower values are rare but higher values are not infrequent.

MIDDLE TO UPPER CARBONIFEROUS

Geosynclinal sedimentation and the deposition of syngenetic mineral resources came to an end with the onset of Hercynian orogeny. After the main Hercynian deformation had run its course, with widespread metamorphism and the emplacement of the Older Granites about 300 Ma ago, the Meseta had become an emerged consolidated massif and the Upper Carboniferous is characterized by continental deposits and coal measures.

Over most of the Meseta Upper Carboniferous and Permian strata were laid down in isolated post-orogenic basins formed by block movements. The sequences in these intramontane troughs characteristically start with orogenic conglomerates derived from adjacent rising blocks and further contain shales and sandstones in continental to paralic facies, with coal seams. Some volcanism occurred locally. The Puertollano coalfield in south-central Spain and some other basins also contain oil shales.

The ages of these sequences mostly range from Westphalian D to Upper Stephanian and Autunian; some are Westphalian B (Ortuño, 1970; Wagner, 1971).

These basins are numerous in Spain and Portugal and some of them have been or are still being worked for coal. Most important, with many mines, is the large central Asturian Coalfield where the Upper Carboniferous is very thick. Spanish coal and anthracite output (excluding lignite) was 11.9 million tonnes in 1977. Portugal produced 195,000 tonnes of anthracite in 1977.

Hercynian orogeny: later phases

The ages of the later Hercynian deformation can only be worked out in areas where dated Carboniferous and Permian formations are present. An unconformity usually separates Stephanian B/C from earlier formations, due to tectonic movements of Early Stephanian age. Similarly, the Autunian in Portugal lies on folded and cleaved older Palaeozoic rocks but was itself folded and deformed, indicating a terminal Hercynian phase of post-Autunian age. As the Autunian consists largely of orogenic conglomerates, tectonic movements must have taken place shortly before their deposition, that is, during the Late Stephanian. Folded Upper Westphalian, Stephanian and Autunian conglomerates may carry clasts from metamorphic and granitic rocks but have not themselves been regionally metamorphosed; the Younger Granites cut the Westphalian and Stephanian, but not the Autunian.

PERMIAN

The Permian of the Meseta continues the deposition of continental facies in intramontane basins, though coal is much rarer. Most of it is Lower Permian, the Upper Permian being unknown in Portugal and very rare in Spain. The Permian basins were formed by late-Hercynian epeirogenic block movements and have themselves been deformed and folded by the latest Hercynian movements. As a result, the Permian is covered disconformably to unconformably by Triassic sediments, at the close of the Hercynian cycle.

Autunian stratabound fluorite

In northern Asturia a fluorite mineralization is found in a Permotriassic basin developed on folded, eroded and planated Carboniferous and older strata. This mineralization is worked in several mines, together with vein deposits, and most of the Spanish fluorite production (275,000 tonnes in 1976 or well over 6% of world fluorite output) derives from this district.

The Lower Permian starts with a basal breccia on Middle Carboniferous limestones and other rocks, on which sandy limestones, dolomites and sandstones follow. The breccia and overlying sediments are impregnated by fluorite. According to Forster (1974) this is a stratabound syngenetic mineralization of Lower Permian age, originated on a shallow-marine shelf along a shoreline. The clastic material and the fluorite derive from the emerged Cantabrian hinterland in the south, where the Palaeozoic sediments already contained relatively much fluorite. However, García Iglesias (1978) admits an epigenetic mode of origin for these fluorite deposits.

Hercynian Younger Granites

The Younger Granites are late-syntectonic to post-tectonic with respect to the Upper Stephanian deformational phase (this phase caused compression of Stephanian and older synclines and took place just before the deposition of the Lower Autunian). They intruded all kinds of earlier rocks, including Older Granites, and form mostly crosscutting massifs. The Younger Granites are largely calcalkaline rocks, ranging in composition from gabbroic to granodioritic and granitic, but they are commonly granitic. Their plagioclase is generally oligoclase to andesine and the mica is mostly biotite with less muscovite. These granites usually are rich in radioactive accessories (zircon, monazite, xenotime, orthite) unlike most Older Granites (Schermerhorn, 1956) and may be associated with abundant pegmatites and aplites. Coarse porphyritic biotite granite with abundant potash feldspar megacrysts is common and forms large batholiths.

A widespread phenomenon is the development of potash feldspar megacrysts in equigranular Older Granites or earlier Younger Granites in the vicinity of porphyritic Younger Granites, a process described as potash-feldspathization of earlier by later granites (Schermerhorn, 1956). This is a secondary process evidenced by the irregular distribution of newly crystallized potash feldspar megacrysts as an inhomogeneous fabric element superimposed on the homogeneous original consolidation texture of the affected older granites. Moreover, megacryst density varies with the distance from the adjacent Younger Granite. The megacrysts are few and far between in the distal facies of potash-feldspathization, giving rise to sparsely porphyritic granite. They increase in abundance towards the adjacent or underlying Younger Granite next to which the proximal facies of highly porphyritic granite is developed. Apparently both metasomatic introduction of potash (as deduced from modal analyses) and metamorphic rearrangement of already present potash feldspar (as deduced from chemical analyses: Brink, 1960) play a role. Potash-feldspathization and the accompanying processes of decalcification of plagioclase to albite and muscovitization are contact-metamorphic effects exerted by large intruding masses of Younger Granites.

Two aspects of the megacryst development in older granites need to be considered here. First, the process ultimately generates porphyritic granites which must be distinguished from primary, magmatic porphyritic granites. Second, this type of megacryst production in Older Granites or earlier Younger Granites is a contact effect of neighbouring Younger

Granites and as such often coincides with Sn-W mineralization. Thus Brink (1960) mapped a large area of equigranular Older Granite in which two zones of megacryst development also are sites of tungsten mineralization which he ascribed to a buried Younger Granite.

Initial $^{87}Sr/^{86}Sr$ ratios in the Younger Granites vary from about 0.705 to higher values. Their abundant and varied mineralization – in contrast to the Older Granites – seems to indicate a different source in deeper levels or a different mode of generation. The Older Granites often are more aluminous than the Younger Granites as is brought out by their higher normative corundum contents (Priem, 1962), indicative of sialic assimilation. In western Spain, on the other hand, occur cordierite-bearing granites assigned to the Younger Granites whose high alumosilicate content is ascribed to assimilation in middle crustal levels (Ugidos & Bea, 1976).

Autunian tin and tungsten

Historically the Iberian Peninsula has since pre-Roman times been an important source of tin from lodes and placers in Galicia and northern Portugal. Spain and Portugal are very minor tin producers now: in 1977 Spain's output was 800 t tin concentrate (containing 427 t tin metal) and Portugal produced 369 t concentrate (at 70% Sn), together about a quarter of British mine production and a half percent of world output.

However, Portugal is an important tungsten producer, with 1010 t W (metal contained in concentrates) in 1977 the first in Europe and no. 10 in the world, contributing 2.25% to world output. In 1974 the Portuguese tungsten reserves were put at 10,000 t W, or 0.53% of world reserves, by the U.S. Geological Survey. Spain produced 350 t W (metal content) in 1977.

Modern study of the tin and tungsten ore deposits in the Meseta starts with Cotelo Neiva (1944). Tin is found in Hercynian granites, pegmatites, aplites, greisens and hydrothermal quartz veins and stockworks, and derived from these are post-Hercynian eluvial and alluvial deposits. The main tin mineral is cassiterite; though stannite and other rarer minerals occur, they are not worked. Tungsten deposits are hydrothermal quartz veins, stockworks and replacement bodies mostly, with wolframite and scheelite as economic minerals, and furthermore stratabound scheelite disseminations which as we have seen may well be of syngenetic origin. There also occur post-Hercynian eluvial and alluvial deposits of little significance.

The Sn and W lodes are found in a zone stretching from Galicia over northern Portugal and western Spain into southern Spain (Fig. 2), which is to say, mostly the Central Iberian Zone of the new tectonic subdivision of the Iberian Massif or Meseta by Julivert et al. (1974). Working mines are found in Galicia, Asturia, northern Portugal and western Spain. Most of the rocks outcropping in this zone are either Hercynian granites or Beira Schists and their equivalents, with a lesser proportion of Ordovician and Silurian rocks, and some strata of other ages.

Except for the tin and tungsten placers and the stratabound scheelite, the tin and tungsten deposits are closely associated in space and time with Hercynian granites and are endogenic ores.

The age of this epigenetic granite-bound tin-tungsten mineralization has been in dispute. Studies in northern Portugal indicated a close spatial relationship with the coarse porphyritic and related granites of the Younger Granite suite. Hence a genetic link was deduced and the mineralization was dated as belonging to the consolidation period of the parent granites, which is to say Autunian.

Later investigations in Galicia (Ypma, 1966; Capdevila, 1969) indicated a relationship with two-mica granites regarded as Older Granites. For the northeast corner of Portugal, in eastern Trás-os-Montes, Ribeiro (1968) claimed that tin-tungsten deposits were associated with two-mica granites (belonging to the Older Granite suite) emplaced into the cores of antiforms originated during Westphalian orogeny. These deposits are quartz veins and some aplite veins filling tension fractures normal to the antiform axis. However, the supposed association of tin-tungsten lodes with the Older Granites, indicating an earlier episode of mineralization, has been criticized (Conde et al., 1971; Thadeu, 1973, 1977) and to date it has not been confirmed.

Could there have been two episodes of epigenetic tin-tungsten mineralization in the northwest Meseta, producing deposits of identical aspect? To answer this question care must be taken not to confuse nature and age of the mineralizing granites.

In the first place, too much has been made of a supposed contrast between (older) two-mica granites and (younger) biotite granites, leading in some cases to incorrect assignment of muscovite-bearing granites to the group of Older Granites. Muscovite is a mineral occurring abundantly in the Younger Granites, and muscovitization, even greisenization of Older by Younger granites took place (Schermerhorn, 1956, p. 244).

Secondly, the two-mica granites associated with Sn-W deposits in northern Portugal and Galicia show medium-grained, mostly non-porphyritic textures, but "similar textures do not signify similar ages" (Schermerhorn, 1956, p. 249). Medium-grained muscovite-bearing granites occur among both the Older and Younger Granites, and the same goes for porphyritic granites. Such was confirmed by a study of an area in northernmost Portugal: Farinha Ramos et al. (1972) showed the existence of two granitic cycles separated by a tectonic phase which caused deformation in the older granites. Either cycle displays an identical succession from coarse porphyritic granites over medium-grained non-porphyritic granites to fine-grained granites, and either cycle contains calcalkaline as well as alkaline rocks. Mineralization is related to the younger granites.

Firm evidence as to age is provided only by the Younger Granites where they intrude folded Stephanian strata; the associated pegmatites, aplites, greisens and quartz veins bearing Sn and W minerals occur either within these granites or in their country rock which consists of contact-metamorphosed sediments and potash-feldspathized and muscovitized Older Granites (Schermerhorn, 1956). The radiometric dating of these Younger Granites (Priem et al., 1970) equally defines the Autunian age of these slightly younger mineralizations. For the Older Granites this kind of evidence is not available; unfortunately no supposedly stanniferous Older Granites have yet been dated by a Rb-Sr isochron.

Conde et al. (1971) point out that the porphyritic granites characteristic of Sn-W mineralization in most of northern

Portugal almost disappear in the northernmost regions where the ore deposits are associated with medium-grained two-mica alkaline granites surrounded by contact-metamorphic aureoles. Furthermore, de Boorder (1965) has shown that the coarse porphyritic granite forming large batholiths in northern Portugal may grade into medium-grained two-mica granite.

Whether Sn-W veins associated with two-mica granites supposedly belonging to the Older Granites are really of Westphalian age depends on two conditions. First there must be evidence that the spatially associated granite is truly the parent granite of the ore, for Older Granites often are hosts to ore veins that in reality emanate from an underlying Younger Granite. Second, the spatially associated granite must be an Older Granite, that is, its age must be radiometrically ascertained.

As long as this evidence is not forthcoming the hypothesis of an older, Westphalian episode of Sn-W mineralization remains unconfirmed and, in the light of other evidence, doubtful. Incidental mineralization may have occurred but no important ore formation of economic significance is known as yet. This evidence is the following. In the first place, as Conde et al. (1971) and Thadeu (1973) point out, no Sn-W veins, whether or not related to the Older Granites, are cut by later Younger Granites or their pegmatites, aplites and other veins. This is strong negative evidence since the very widespread Younger Granites very frequently intrude Older Granites and their country rock. Secondly, even though there are mineralogical, petrographical and structural differences, as group characteristics, between the Older and Younger Granites (Schermerhorn, 1956; Floor, 1970; Oen Ing Soen, 1970; Capdevila et al., 1973) the Sn-W deposits cannot be differentiated on this basis. Thus Conde et al. (1971) and Thadeu (1973) conclude in favour of a single period of mineralization. In addition, another argument supporting this view is that the Westphalian and Stephanian conglomerates carrying clasts of Older Granites and metamorphic rocks have not been found to contain cassiterite or other detritus indicative of an important mineralization associated with these granites.

Tungsten is more abundant than tin in this Sn-W province. The two metals mostly occur jointly, forming Sn-W deposits, but there also are some deposits in which only one of these metals is present, with the other absent or appearing in trace amounts. Byproducts of the Sn-W mines include copper, silver (both from chalcopyrite), niobium and tantalum (in columbite-tantalite) and formerly bismuth and lithium as well. I l m e n i t e is recovered together with cassiterite from alluvial tin deposits in Portugal (which produced 367 t ilmenite concentrate (50% TiO_2) in 1976 and 175 t in 1977); however, this mineral derives not from the tin lodes but from the Hercynian granites in which it is almost the only oxidic ore mineral (Schermerhorn, 1956).

The Sn-W lodes are associated especially with the Younger Granites, notably the coarse porphyritic and related granites, as is discussed below. The Sn-W-bearing pegmatites, aplites, greisens and quartz veins are found in the vicinity of the contacts of these granite massifs.

The deposits may be grouped according to type as follows:

I. D i s s e m i n a t i o n s
cassiterite disseminated in muscovite granite or greisen
II. P e g m a t i t e s
a. cassiterite pegmatites with little or no tungsten
b. wolframite pegmatites with little or no cassiterite
III. A p l i t e s
a. cassiterite aplites
b. wolframite aplites
IV. Q u a r t z v e i n s (w i t h s t o c k w o r k s a n d b r e c c i a p i p e s)
a. cassiterite veins with little or no tungsten
b. cassiterite-wolframite (and/or scheelite) veins
c. wolframite veins with little or no cassiterite
d. scheelite veins with little or no wolframite or cassiterite
V. M e t a s o m a t i c s k a r n s
a. scheelite skarns
b. wolframite-scheelite skarns

Excluded are the Cambrian and Silurian metamorphic scheelite skarns (reaction skarns) since their metallization appears to be much older.

T i n d i s s e m i n a t i o n s in greisens are of frequent occurrence but do not often constitute large masses. Normally tin has been concentrated in late-stage fluids which became segregated in pegmatites and vein deposits and only rarely has it impregnated entire rock masses.

Late-stage fine-grained granites or aplogranites sometimes carry sufficient dispersed cassiterite to warrant economic extraction. Columbite-tantalite also occurs. The Sn and Nb-Ta minerals are in some cases found finely disseminated through the rock, in other cases they are associated with quartz veinlets. The grade of this kind of tin ore is variable between 0.3–2 kg cassiterite per tonne (A. Arribas, pers. comm.). Thus, at Penouta (Orense district, Galicia) a kaolinized greisen granite (Gumiel, 1978) is worked which yields 0.4 kg/t cassiterite. At Golpejas near Salamanca a similar granite is mined and several other occurrences of disseminated tin mineralizations are known in Spain (Arribas, 1978).

T i n p e g m a t i t e s contain albite, quartz, potash feldspar and muscovite as main minerals with cassiterite, tourmaline, beryl, apatite and sulphides in accessory amounts. Wolframite is rare; however, a few pegmatites carry wolframite but no cassiterite.

Tin pegmatites, as distinct from the barren pegmatites of simple mineralogy, show very complex mineral assemblages, with the minerals listed above accompanied by plentiful rarer minerals in trace amounts. Many of these pegmatites contain fine-grained aplitic portions (pegmaplites of some authors).

The tin pegmatites have in the past been worked extensively for cassiterite, with columbite-tantâlite, wolframite and other byproducts, but exploitation has mostly ceased, as the grade in pegmatites is generally lower than in veins.

In some regions pegmatites are abundant, as in the Serra de Arga (northern Portugal) where cassiterite and columbite-tantalite-bearing veins were studied by Cotelo Neiva (1954). The tin pegmatite fields of the Amarante region in northern Portugal have been described by Maijer (1965); these pegmatites contained 0.1–0.3 wt % cassiterite.

Some pegmatites attain considerable dimensions: the Lagares tin pegmatite (Cotelo Neiva, 1944; Schermerhorn, 1956) is one kilometre long and up to 12 m thick. The grade was 0.09–0.1 wt % cassiterite; the concentrate produced by the Lagares mine contained in addition 2% columbite-tantalite and 0.1% wolframite.

T i n a n d t u n g s t e n a p l i t e s are much less frequent than tin pegmatites and do not attain similar sizes: they have been worked on a small scale only. These rocks are muscovite aplites generally bearing some tourmaline, and they contain finely disseminated cassiterite occasionally accompanied by wolframite and/or sulphides; some aplites carry wolframite only; some aplites show greisenization (Schermerhorn, 1956).

Where both pegmatites and mineralized quartz veins are developed there exists a zonation from tin pegmatites and/or tin-bearing quartz veins inside or just outside the granite contact to tungsten veins at some distance outside the granite. Though the different types of ore veins grade into each other no transition is known between pegmatites or aplites and quartz veins (Thadeu, 1965b, 1973).

Q u a r t z v e i n s carrying cassiterite and/or wolframite and scheelite constitute the ore exploited by most mines. As a rule cassiterite-quartz veins occurring inside granite pass to cassiterite-wolframite veins across the contact and these to wolframite veins outside the granite. Sulphides accompany this mineralization, especially in the tin-tungsten and tungsten zones. Thadeu (1973) points out that marmatite and arsenopyrite tend to be concentrated in the zone where cassiterite and wolframite occur jointly. Pyrite, pyrrhotite and chalcopyrite are present too but become quantitatively more important in the outer zone of wolframite veins where marmatite and arsenopyrite decrease.

The veins often follow existing joints, especially cross-joints (*ac* joints) and also cleavage planes in schistose country rock. Often too they are somewhat later than the jointing and exhibit diverging strikes and dips. Subhorizontal veins as at Panasqueira are very rare: the veins generally show moderate to steep dips. Stockworks as at Bejanca (Sn and W) in northern Portugal usually occur in greisens (Cotelo Neiva, 1944).

Veining was in many cases accompanied or preceded by greisenization of the host rocks in generally narrow zones along the vein contacts. In contrast, the development of greisen borders in the wall rocks of pegmatites and aplites is rarely seen, as distinguished from internal greisens in these bodies. Greisen borders, often attended by some mineralization, are best developed in granites, much less so in sedimentary host rocks where tourmalinization and other alterations are more common (Schermerhorn, 1956; Thadeu, 1965b).

Metasomatic replacement ore (cassiterite and wolframite) occurs as small bodies locally associated with vein mineralization (Thadeu & Aires Barros, 1973).

The three most important vein deposits in the Iberian Peninsula are the tungsten producers Panasqueira and Borralha in Portugal and Barruecopardo in western Spain. They are not at all alike: Panasqueira presents the peculiarity of a subhorizontal vein system, Borralha has two breccia pipes and is exceptionally rich in molybdenum, and Barruecopardo is a scheelite vein system, not wolframite like the other two.

The P a n a s q u e i r a mine in central Portugal produces wolframite and cassiterite concentrates and since 1962 also argentiferous chalcopyrite concentrates; the recovery of sphalerite is being studied. Mine statistics show that between 1934 and 1970 56,960 t wolframite concentrate, 3,984 t cassiterite concentrate and 2,848 t chalcopyrite concentrate have been produced (Reis, 1971). Although the overall ratio is cassiterite 7% of wolframite output, the tin percentage has since 1960 dropped to 1–2%, again rising to 4–5% during the last few years. Production in 1977 was 1287 t wolframite concentrate (75% of Portuguese output; however, in 1975 it was 1742 t), 1176 t chalcopyrite concentrate and 58 t cassiterite concentrate (16% of Portuguese output).

Panasqueira mineralization consists of subhorizontal wolframite-cassiterite-sulphide-bearing quartz veins. The mineralized area is large: the underground workings cover 3 by 3.4 km. It lies in Beira Schist country, largely within a zone of spotted phyllites. This contact metamorphism is due to a buried granite whose roof has been intersected underground and in boreholes. A dome of greisen (greisenized granite) roughly 150 m in diameter rises about 120 m above the irregular roof of a much larger granite massif (Fig. 6). The roof rocks range from more or less strongly greisenized granite to muscovite-quartz greisen and belong to an undeformed post-tectonic massif (Clark, 1964, 1970; Conde et al., 1971). The greisen carries arsenopyrite, chalcopyrite, sphalerite and a little cassiterite and is itself cut by the mineralized quartz veins. The greisen dome is capped by about 15 m of massive quartz which was deposited in the space left by sagging due to contraction of the dome rock.

The ore veins are up to 200 m long and average 30 cm thick, rarely exceeding one metre. They mostly dip gently at 5–20° SW, SE or NW: moderate to steep SE dips occur locally. The most common minerals are wolframite, arsenopyrite, chalcopyrite, pyrite, pyrrhotite, sphalerite (marmatite) and siderite. Less abundant are cassiterite, apatite, marcasite, muscovite, fluorite, topaz, tourmaline, galena and calcite. Minerals present in small quantities include stannite, molybdenite, bismuth, bismuthinite, cubanite, freibergite and other silver minerals, beryl and others. The most abundant mineral after quartz is arsenopyrite (Reis, 1971). The cassiterite content of the veins increases in depth and towards the granite. The veins display muscovite-tourmaline selvages and greisenized borders.

Greisenization preceded mineralization. Based on four K-Ar age determinations of muscovites from Panasqueira greisens and ore veins which yielded identical results, Clark (1970) concluded that greisenization and hydrothermal mineralization took place in a brief period, possibly less than one million years.

The veins follow existing flat-lying joints (Fig. 6) which is unusual for tin-tungsten vein mineralization. These joints may have been produced as subhorizontal fractures resulting from subvertical pressure release caused by slight sagging of the underlying large granite mass when it contracted during consolidation. As has been noted, shrinking of the greisen dome, a small protuberance on the roof of the main granite

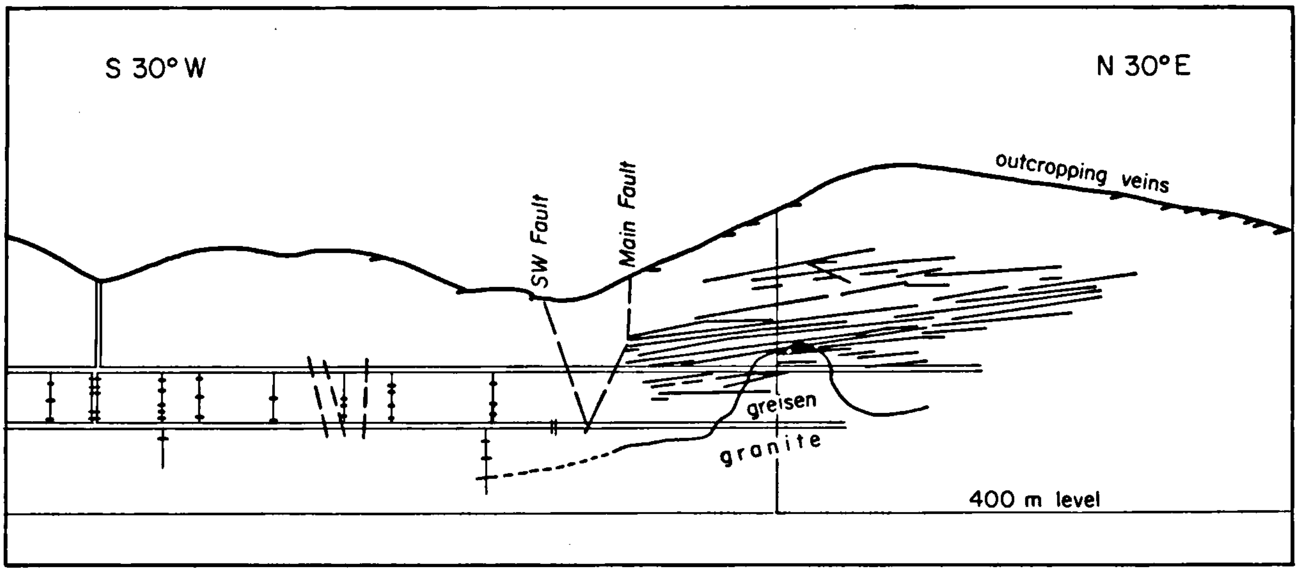

Fig. 6. Section through the Panasqueira vein system and the greisen-granite dome with its quartz cap (black). After Beralt Tin and Wolfram Portugal, S.A.R.L.

mass, left a space that was filled by hydrothermal quartz. The joints formed due to pressure release as here suggested, possibly remained latent until forced open by overpressured hydrothermal solutions from which the vein minerals were deposited.

Hydrothermal mineralization took place in various stages (Orey, 1967). According to Kelly (1974), study of the fluid inclusions in the vein quartz shows that the ore fluids were hot (80–325 °C) and fairly saline (3–10 wt % equivalent NaCl) while mineralization took place within 1.5 km of the contemporaneous ground surface. Stable isotope data indicate a juvenile origin for the sulphur in the sulphides formed during the main stage of mineralization but a shallow source for late-stage sulphur; the carbonates contain a substantial proportion of organic carbon; and the hydrothermal fluids appear to have been predominantly magmatic during the deposition of some minerals but mixtures of meteoric and magmatic during the deposition of others (Rye & Kelly, 1974).

Fluid inclusions in the apatite accompanying the mineralization indicate that it was deposited at temperatures in or slightly above the range of 230–315 °C (peaking at 280 °C) from fluids showing a salinity of about 3–11 wt % equivalent NaCl. The fluid inclusions in later vein minerals (siderite, fluorite, calcite and others) reveal a systematic decline of temperature to below 70 °C.

Kelly's assessment of the depth of mineralization entails that the underlying magma must have risen to less than 1.7 km below the surface. This in turn implies, first, a very high level of emplacement but that is not unusual for the Younger granites and, second, strong denudation. The granite intruded into folded and cleaved Beira Schists. Before erosion, this sequence was overlain by Ordovician, Silurian and Devonian strata to a combined thickness certainly exceeding 500 m. As mentioned in a preceding section, tectonic differentiation at the beginning of the Carboniferous raised the central Iberian block above sea level and erosion started. After the main orogenic phase during the Westphalian, at least two tectonic phases of an epeirogenic, block-faulting nature ensued, during the Early and the Late Stephanian (Schermerhorn, 1956), causing elevation of areas outside the basins in which the Stephanian deposits were laid down, hence strong erosion and denudation. The Stephanian conglomerates in Portugal contain plentiful detritus derived from the Older Granites intruded shortly before and from regional-metamorphic schists likewise formed shortly before, during the Westphalian. This indicates a high rate of erosion and denudation in the uplifted source areas of the Stephanian. It thus is perfectly possible for an Upper Westphalian-Stephanian erosion surface to truncate Beira Schists at a deep level after removal of several kilometres of overlying strata. The shallowness of Panasqueira mineralization hence is not inconsistent with a deep stratigraphic-structural level of emplacement.

The Borralha mine in northernmost Portugal, the second largest tungsten deposit in the Peninsula, produces wolframite, scheelite and argentiferous chalcopyrite concentrates. Together with Panasqueira it accounts for about 90% of Portuguese tungsten output. An extensive zone of tin and tungsten veins adjoins a massif of porphyritic granite regarded as the parent magma; cassiterite veins near this granite pass outwards to wolframite veins (with some scheelite) that are worked at Borralha. The country rock is granite and Silurian schists (Conde et al., 1971).

The lodes, up to 1.5 m thick, form an extensive, very regular system of E to ESE trending mineralized quartz veins dipping at 25–30° N, 45–60° N and 50° N (Noronha, 1974; Noronha & Saavedra, 1975). Two large breccia pipes, one of them outcropping, the other blind, are cemented by ore-bearing quartz. The recently discovered blind pipe contains 415 ppm W, 980 ppm Mo (in molybdenite) and 1300 ppm Cu (in chalcopyrite) (F. Noronha, pers. comm.).

The veins carry wolframite and sulphides (mainly chalcopyrite and pyrite) as principal minerals, accompanied by molybdenite, scheelite (and rare powellite), marmatite, arsenopyrite, other sulphides, tourmaline, white mica, apatite and carbonates (Conde et al., 1971; Noronha & Saavedra, 1975). Study of fluid inclusions in the vein quartz led Noronha (1974) to admit a temperature of formation in the order of 280–350 °C under a pressure around 1000 at.

Noronha estimated this pressure value from figures published by Russian workers. However, Amossé (1976) has evolved a new method of measuring the pressure of crystallization: lattice variation in wolframite before and after annealing at high temperature to remove strain is measured by X-ray methods, and from this value and the compressibility coefficient of wolframite the pressure of crystallization can be derived using the solid state equation. Thus Amossé arrived at a pressure of 1500 bar for wolframite formation at Borralha. Using this value to correct the homogenization temperature of the fluid inclusions in the quartz host of the ore, he estimated that the Borralha wolframite was deposited at temperatures between 325 and 430 °C, the temperature gradient being correlated with distance from a breccia pipe as 'hot centre' of mineralization.

Noronha's and Amossé's figures for the pressure, if taken as load pressure, would indicate that mineralization took place at a depth of roughly 3.8 km, respectively 5.7 km. Further, according to Noronha and Saavedra (1975) the Silurian host rocks of the ore deposits were metamorphosed at a temperature in the order of 650 °C under a pressure of 4.5–5 kbar. This is equivalent to a depth of 17–19 km. Given a Middle Westphalian age of the metamorphism, it means that during the 20–30 million years separating the time of metamorphism and the time of mineralization some 12–14 km of overburden was removed by erosion, at a rate of 400–700 m per million years or less than a millimetre a year. Such does not appear impossible in a tectonically active environment of block-faulting. Still, the depth of mineralization deduced from the above pressure figures seems rather high. But if it is admitted, in view of the presence of breccia pipes, that the pressure under which wolframite formed was a hydraulic overpressure exceeding the lithostatic pressure, hence not correlatable with depth, then the depth of mineralization was even less, say one or two km as at Panasqueira, and this entails an even greater rate of erosion, almost reaching a millimetre a year.

These calculations incidentally tend to confirm the view expressed by Conde et al. (1971) that the mineralization emanated from a porphyritic granite belonging to the Younger Granite suite. An Older Granite would have intruded not long after the regional metamorphism had run its course, when little erosion could have taken place at the surface, but the Borralha ore veins could not have formed at a depth of 10 km or over. Hence this mineralization too is related to the Younger Granites and of Autunian age.

In Spain the largest tungsten producer is the scheelite stockwork of the Barruecopardo open-cast mine northwest of Salamanca, near the frontier with Portugal. This occurrence is a system of parallel NNE-trending subvertical veins averaging 0.5–15 cm thick, developed in granite (Pellitero et al., 1975). This stockwork is over 3 km long by 200 m wide and 100 m deep (Arribas, 1978). The veins are mineralized by scheelite and arsenopyrite with subordinate ferberitic wolframite, pyrite and some chalcopyrite; cassiterite and molybdenite are present in trace amounts.

The stockwork occurs in a granite described as a two-mica adamellite in which pronounced microclinization of biotite and plagioclase took place, and Pellitero et al. (1975, 1976) relate the tungsten mineralization to this process. They note a positive correlation between potash feldspar content and ore, hence they infer that the tungsten was extracted from the surrounding granite itself during microclinization: scheelite precipitated instead of wolframite because the replacement of plagioclase by microcline released calcium. However, it also seems possible that potash-feldspathization of an older by an underlying younger granite took place and that shear zones in the older granite provided pathways for mineralizing solutions related to the younger granite. The presence of predominant scheelite is most unusual in this type of mineralization and difficult to explain on either hypothesis. Because microclinization of plagioclase is widespread in the Meseta granites, scheelite ought to be rather more abundant than wolframite, which is not the case. In fact, Barruecopardo is unique, the more so because of its large size.

Skarn-type scheelite-wolframite mineralization is linked to the occurrence of calcareous beds in the Beira Schists and Silurian country rock of the Younger Granites. As we have seen, formerly such mineralization was uniformly regarded as due to epigenetic introduction of tungsten. It has recently become recognized that syngenetic tungsten exists in Cambrian and Silurian calc-silicate beds, especially where they are found outside the granite contact aureoles, such as the occurrence cited under the heading Cambrian scheelite.

Within the contact aureoles such occurrences have been metamorphosed to reaction skarns, and these are hard to distinguish from contact-metasomatic skarns (or tactites) in which the scheelite has been produced from introduced tungsten and sedimentary lime. Normally the syngenetic scheelite deposits are poor in sulphides and lack wolframite while the metasomatic scheelite deposits contain both sulphides and wolframite.

In the Covas mining district situated in the extreme northwest corner of Portugal, several scheelite deposits, the most important among them being Valdarcas, are associated with a very extensive skarn horizon developed on top of Ordovician quartzite in a large anticline near two intrusive granites; many barren aplitic pegmatites (pegmaplites) occur in the area. The skarn consists of calcite, grossular-rich garnet, diopside, vesuvianite, actinolite-tremolite, hornblende and other minerals. The mineralized skarn zones contain in addition scheelite, ferberite (pseudomorphosing scheelite), wolframite, plagioclase (albite to labradorite), apatite, fluorite, a little quartz and much pyrrhotite, with arsenopyrite, pyrite and other sulphides. Wolframite prefers the zones rich in sulphides where scheelite is rare and scheelite prefers the zones in which calc-silicates prevail (Bayer, 1968; Conde et al., 1971).

The orebody worked at Valdarcas was a lens up to 8 m thick and about 100 m long that contained two kinds of ore: calc-silicate-rich scheelite ore ('skarn ore') and sulphidic (mainly pyrrhotite) wolframite ore with scheelite and ferberite after scheelite. The sulphide ore formed the main mass; calc-silicate ore occurred as a bed 1.5–2 m thick in the hanging wall of the sulphide lens. The grade of the Valdarcas ore was 2.5–5 kg wolframite-scheelite per tonne ore, or 0.18–0.25% WO_3 (Bayer, 1968), that is 0.14–0.20% W.

Bayer (1968) envisages the sequences of mineralization at Valdarcas as follows: an intercalation of either marl or a mafic eruptive rock was folded and metamorphosed to the amphibolite facies with the crystallization of diopside, hornblende, plagioclase and possibly also grossular-rich garnet and vesuvianite. The tungsten is of syngenetic origin, preconcentrated as volcanogenic scheelite in a mafic lava or tuff; below this lava a submarine iron-sulphide containing some copper, zinc and arsenic existed. During metamorphism and aplite-pegmatite emplacement the iron sulphide became pyrrhotite and the scheelite in the metamafic horizon was mobilized in part and redeposited as wolframite in the sulphide lens.

In contradistinction, Conde et al. (1971) consider the skarn to be metasomatized limestone, since borings revealed its passage to crystalline limestone, and this limestone by its reactivity channeled the mineralization which is a normal wolframitic mineralization generated by the Covas granite. Thadeu (1973) further remarks that volcanics are lacking in this region.

The Covas tungsten deposits are unique in the Peninsula. Their association with Ordovician calc-silicates sets them apart from the scheelite orebodies in Cambrian or Silurian host rocks. The presence of a vast amount of pyrrhotite is unusual. The lack of cassiterite in the very closely associated aplitic pegmatites seems curious, especially as cassiterite does occur in pegmatites near the Covas granite.

Tin and tungsten lodes of granitic descent are characterized by apical deposition, that is, during consolidation of the parent granite (which need not be their actual granitic host rock) the metalliferous fluids move upwards, not sidewards, and accumulate in the roof of the pluton. Hence the richest tin-tungsten concentrations are found over unexposed shallow granites or over gently dipping granite contacts. Deeply eroded granites with steep contacts are unfavourable exploration targets in contradistinction to contact-metamorphosed sedimentary or older granite areas with little or no younger granite outcropping.

Tin and tungsten became concentrated by crystal fractionation processes in the granite magmas in volatile-rich halide-enriched hydrothermal fluids, together with several other late-stage constituents, including those causing greisenization. However, greisenized granite lacking mineralization is frequent, and greisenization and metallization are linked to a certain extent but not inseparably.

When fissures crossing the granite-wall rock contact are opening at this late stage they become mineralized, producing high-grade low-volume lodes. Tungsten penetrates farther into the wall rocks than tin which is less mobile in this situation. Retention of stanniferous fluids within the granite roof zone, in the absence of late-stage fissuring or because the earlier veins close, originates disseminated low-grade high-volume cassiterite deposits in greisenized granite, as at Penouta in northwest Spain or Golpejas in western Spain.

Lastly, the Iberian tin-tungsten province presents certain characteristics setting it apart. For instance, though topaz does occur, it is distinctly rare in comparison to other Hercynian tin-tungsten belts in Europe (Schermerhorn, 1956, p. 536, 560; Conde et al., 1971, p. 41). Tourmaline, on the other hand, is abundant, mostly in the wall rocks of the lodes where tourmalinites may have formed. Also, unmineralized tourmaline pegmatites and aplites are widespread, and tourmaline frequently occurs in quartz veins and on joint planes. Apatite is very frequent too in the tin and tungsten deposits. Among the sulphides, marmatite appears to be more plentiful than it is, for instance, in the lodes of Cornwall.

Some occurrences show enrichment in certain elements. Thus the Borralha tungsten deposit is relatively rich in molybdenum as compared to other occurrences. Lithium is locally plentiful in cassiterite veins, for example in western Spain (see next section).

Autunian beryl, columbite-tantalite, lithium and feldspar

Minerals containing Be, Nb-Ta and Li, together with feldspar, are won from pegmatites in the Iberian tin-tungsten province. A great variety of iron-manganese and other phosphates occur in many of these pegmatites (Correia Neves, 1960).

Beryl and columbite-tantalite can be by-products of tin mining but pegmatites rich in beryl or columbite-tantalite and manganese phosphates contain little or no cassiterite (Correia Neves, 1960; Thadeu, 1965, 1973, 1977; Conde et al., 1971). The Mangualde pegmatite in northern Portugal has been worked for beryl and feldspar; it carried a little cassiterite and iron, manganese and lithium phosphates (Jesus, 1933).

Lithium minerals appear in pegmatites and aplites in Galicia, north Portugal and western Spain. Moreover, amblygonite and montebrasite-bearing cassiterite-quartz veins occur locally in western Spain (Weibel, 1955).

Portugal produced 1000 t lepidolite (1.5% Li_2O) in 1977 as a byproduct of feldspar exploitation (Bol. Minas (Lisboa), Dec. 1977).

Usually lepidolite is the common Li-mineral in pegmatites (Conde et al., 1971), and lepidolite pegmatites, often cassiterite-bearing as well, are widespread, but amblygonite and other minerals also occur. Hensen (1967) described an extensive zone 600 m wide by 15 km long of metalliferous pegmatite veins in micaschist in Galicia. These have been mostly mined for cassiterite (up to 0.2 wt %) and some for beryl; in addition, Li-minerals (spodumene and petalite), columbite-tantalite, phosphates and many other minerals occurred.

Pegmatites are furthermore the most important source of feldspar and quartz in the Peninsula. Spain produced 70,000 t feldspar in 1974 (no. 9 in world production) and Portugal 28,000 t. However, in 1977 Portuguese production had declined to 10,803 t feldspar (and 103,565 t quartz).

Autunian gold and silver

Gold and silver, like copper, tin and lead, have been won in the Iberian Peninsula since before the Romans. Both vein and placer gold were exploited. No gold or silver mines are active anymore in Spain; these metals are still being won as by-products from other ores (gold and silver from cupriferous sulphide deposits and their gossans (Cerro Colorado in the Pyrite Belt accounts for 90% of Spanish gold output) and silver from lead sulphide deposits; with the planned increase of production of these ores gold and silver output will mount too: from these sources Spain produced 320 kg Au and 67 t Ag in 1970 and expects (Prado Calzado, 1973) to produce 6 t Au and 170 t Ag in 1980). The Jales mine in northern Portugal is the only active gold-silver mine in the Iberian Peninsula, and Portuguese output of gold and silver as sulphide concentrate grading 133.8 g/t Au and 377.8 g/t Ag was 2,021 t in 1977 (Bol. Minas (Lisboa), December 1977). This is a very old mine remarkable for the enormous extent of the Roman workings.

A formerly important gold-silver district is situated in northwest Spain (Galicia, Asturia and León) and northern Portugal. In this area numerous gold-silver-arsenopyrite veins have been worked. Gold veins in Galicia occur within and around granitic plutons and contain gold and arsenopyrite in association with pyrite, pyrrhotite, chalcopyrite, bismuth and bismuthinite; silver is also present (Armengot de Pedro & Campos Juliá, 1971).

The Jales mine in northern Portugal has been described by Brink (1960) and Portugal Ferreira (1971). Here gold-silver-arsenopyrite-quartz vein systems occur in Beira Schists, Ordovician, Silurian and Older Granite country rock, in an area where this granite has been potash-feldspathized, that is, plentiful potash feldspar megacrysts have developed in it, which is ascribed to the contact-metamorphic effects of an underlying porphyritic Younger Granite. The Gralheira veins, no longer in exploitation, trend WNW and are sited in basal Ordovician metaquartzites and schists. Somewhat more northwards is the Três Minas occurrence of disused gold workings in Silurian schists. To the south is the long Campo vein (Minas dos Mouros) which produces gold and silver. It trends NE across the contact between Older Granite and Beira Schists and reaches 1 m thickness. In the same area of Beira Schists are small cassiterite deposits.

The grade of the ore extracted from the Campo vein was 14 g/t Au in 1970 (Portugal Ferreira, 1971). The gold is mostly associated with arsenopyrite, also with pyrite and to a lesser extent with younger sulphides (chalcopyrite, galena and sphalerite). Jales production of silver is two or three times gold production and the silver is contained in tetrahedrite associated with the galena.

The Older Granite wall rock of the Campo vein has suffered microclinization of plagioclase, and microcline has been introduced into the schist wall rocks as well: Brink attributes this to metasomatic introduction of potash feldspar from the ore-bearing solutions which also caused muscovitization, silicification, sulphidization and tourmalinization.

The gold-silver veins of the northwest Meseta are generally regarded as Hercynian deposits but their precise age is still not clear. Some authors (Cerveira, 1952) see these veins as directly related to the Hercynian tungsten mineralization, based on the presence of wolframite or scheelite in gold mines. Brink's (1960) opinion was that there exists no connection between the two mineralizations: the gold and silver belong to an earlier, quite separate phase of ore deposition possibly related to the Older Granites. He argued that the older minerals in the Campo vein are deformed while the younger minerals are not; also, the wolframite veins in the area show no signs of deformation. Brink therefore concluded that the cataclastic gold-bearing older ores were deposited during fracturing of the Older Granite, and the undeformed later minerals formed during a posterior episode of vein dilation.

Still, the Campo vein was emplaced along a dextral wrench fault according to Portugal Ferreira (1971), and recurrence of minor movements during mineralization could well account for the deformation of the older minerals. Because the gold mineralization is clearly associated with wall rock alterations of the type generally caused by Younger Granites and because it occurs in an area affected by potash feldspathization generated by underlying Younger Granite, it seems more likely that this mineralization too is genetically linked to the Younger Granites.

Another type of gold mineralization is represented by disseminated gold, as described by Harris (1978) for the highly altered portions of the Salave granodiorite stock in Asturia, a Hercynian intrusion of Younger Granite age (Harris, pers. comm.). The gold is associated with arsenopyrite, pyrite, stibnite, some molybdenite and traces of sphalerite and galena. Alteration of the granodiorite produced a complex zonation: from the unaltered granodiorite towards the highly altered mineralized rock, igneous texture is progressively destroyed, Na and CO_2 increase and SiO_2 decreases. Peripheral molybdenite-bearing quartz veins may represent remobilized Mo and SiO_2 from the altered auriferous bodies. This occurrence is a prospect and this means that the Salave mineralization may prove to be economic and further that this type of gold deposit may be present elsewhere too.

Autunian antimony and gold

Antimoniferous veins occur in northern Portugal, northwest and western Spain and, to a lesser extent, in south Portugal. They are most abundant in the so-called 'antimony belt' (faixa antimonífera) in northern Portugal, a southeast-trending zone lying east and southeast of Oporto that reaches some 30 km in length by a width of up to 10 km. Here auriferous antimony veins, of which the largest was the Alto do Sobrido mine, have been worked for gold by the Romans. The veins were rather poor in gold but rich in stibnite and the district was an important antimony producer during the second half of the last century; it is now inactive.

The host rocks of the veins are mostly Beira Schists and Ordovician quartzites and slates, more rarely Silurian slates and Carboniferous conglomerates, in the limbs of the large Valongo anticline. The Sb-Au vein mineralization is of Younger Granite age (Portugal Ferreira, 1971; Portugal Ferreira et al., 1971). In this area occur furthermore Sb-Pb and Pb-Zn-Ag vein mineralizations..

The Sb-Au mineralization occurs in quartz veins of varying trends, most frequently ENE-WSW (following dextral wrench faults), that reach 2.5 m thickness. The mineralization consists of stibnite, berthierite, gold, pyrargyrite, pyrite, and locally galena, sphalerite and arsenopyrite (Portugal Ferreira, 1971).

The origin of the mineralization is uncertain: it has been ascribed to the Younger Granites, to later hydrothermal activity and to remobilization of antimony in older sediments or volcanics.

Stibnite veins carrying some gold and silver, emplaced in the local equivalent of the Beira Schists, have also been worked in the Valle de Alcudia in southern Spain (Crespo, 1972).

Gumiel et al. (1976) described an occurrence of 110°-trending stibnite-scheelite veins in Devonian limestone, the San Antonio mine near Albuquerque in western Spain; these authors assigned a juvenile epigenetic, late Hercynian origin to this mineralization, which they related to an antimoniferous belt traversing the western Meseta in a NW–SE direction from Oporto to Valdepeñas.

Other antimony vein mineralizations in northern and southern Portugal and western Spain are much less important.

Quartz-stibnite veins containing some sulphides in trace amounts are fairly frequent in southern Spain where they have been mined on a small scale in Ciudad Real province and elsewhere, and locally in northwest Spain.

Autunian and later lead, zinc and silver

Epigenetic lead and lead-zinc veins, in part argentiferous, are widespread in the Meseta. Their age is not always evident and some are certainly post-Hercynian.

The richest and most famous lead province of the Meseta is in the eastern Sierra Morena of central southern Spain (Fig. 2). Here metalliferous lodes, especially lead and argentiferous lead veins, that have been worked since pre-Roman times are found in the Los Pedroches batholith and the area north and east of it. This batholith is a very large elongated, WNW-trending massif composed of various types of intrusive granite and granodiorite (Ovtracht & Tamain, 1973). Its precise age is not known; some 50 km to the north is the small Fontanosas stock, a late or post-tectonic granodiorite (containing andesine and some cordierite) that has been dated at 302 ± 10 Ma with an initial $^{87}Sr/^{86}Sr$ ratio of 0.7099 (Leutwein et al., 1970). This age accords well with the age and initial strontium isotope ratios measured for Older Granites elsewhere in the Meseta. However, this result neither dates the Los Pedroches rocks nor the ore veins in the area!

The Los Pedroches batholith contains several tungsten and some tin occurrences, mined in former days. Cassiterite-wolframite-arsenopyrite-quartz veins, cassiterite greisens, wolframite pegmatites and scheelite pegmatites are found in the central part of the massif; farther west only tungsten occurs and the east part of the massif has some wolframite, in places accompanied by scheelite, along the southern contact (Ovtracht & Tamain, 1973).

The batholith also encloses a large number of copper veins (chalcosite-chalcopyrite-carbonate-quartz veins) especially in its eastern part, and a lesser number of lead veins (galena poor in silver: 170–250 g/t Ag). Lead veins are more frequent in the Lower Carboniferous slates and greywackes on the north side of the batholith. Along the contact occur Bi-Ni-Co-Ag-Au veins.

North of the batholith and parallel to it extends the Valle de Alcudia, a broad plain where the Alcudiense (Beira Schists) outcrops. Here lead veins are abundant (Crespo, 1972).

Just south of the central part of the batholith is the important El Soldado-Las Morras orefield, a vein system following NE and ESE to E–W directions; only the NE veins are richly mineralized and they reach several kilometers length. The mineralization consists of galena and sphalerite, with pyrite, chalcopyrite, carbonates and quartz (Ovtracht & Tamain, 1973).

The richest lead district, between 1875 and 1920 the most productive leadfield in the world, lies at the east end of the Sierra Morena: the Linares and La Carolina orefields. At the beginning of the century, when Spain was the second largest lead producer in the world (after the U.S.A.), with for instance 265,500 t lead ore (60–80% Pb) in 1905, this region supplied nearly half of the Spanish lead ore output from some 1300 mines (Ahlburg, 1907; Azcárate & Argüelles, 1971). At present three veins are still being worked in the Linares field and three others in the La Carolina field (Rios Aragües, 1977).

North of Linares a granite massif in the eastern continuation of the Los Pedroches batholith contains NE to NNE trending veins of great extent (12 km for the Arrayanes vein, one of the richest lead veins known anywhere). The veins carry galena (with 160–250 g/t Ag) and lesser amounts of sphalerite, pyrite and chalcopyrite in a quartz-barite-carbonate gangue. A vertical zonation exists: from the surface down to 100 m depth lead and copper are present; 100–150 m: a barren zone; 150–300 m: a zone very rich in lead; 300–400 m: a poor zone; below 400 m: the lead tenor increases again (Ovtracht & Tamain, 1973).

North of La Carolina extends a mineralized area consisting of veins in the Santa Elena granite (the Santa Elena deposits), and to the west, first, the Los Guindos deposits and then the El Centenillo deposits, where ESE, respectively NNE trending lead veins cross Ordovician or Silurian country rock contact-metamorphosed by underlying granite. Argentiferous galena (up to 1500 g/t Ag) constitutes the ore, accompanied by sphalerite, pyrite and quartz-ankerite-barite gangue (Ovtracht & Tamain, 1973).

Except for the Valle de Alcudia, all these mineralizations are associated with granite; however, an important lead field, the San Quintín (Villamayor de Calatrava) vein system, is in Alcudian country rock north of the Valle de Alcudia, far from granite (Crespo, 1972). The veins trend E–W and are mineralized by highly argentiferous galena (up to 3500 g/t Ag), black sphalerite, pyrite and some chalcopyrite in quartz gangue with rare carbonate and very rare barite (Ovtracht & Tamain, 1973).

Ovtracht and Tamain (1973), as does Crespo (1972), relate the Cu, Pb and Sb mineralizations in the Sierra Morena to the Los Pedroches batholith, more particularly one of its phases, a porphyritic biotite granite (rather a granodiorite, as andesine exceeds potash feldspar in amount). The mineralization is zoned, with Cu nearest to the granite and Pb farther out. Sn and W are related to a fine-grained two-mica granite.

These are late Hercynian mineralizations: the veins are covered by barren Triassic sediments (Tamain, 1968; Azcárate & Argüelles, 1971). The Linares mineralization is thought to be associated with the Linares granite, and the Santa Elena, La Carolina and neighbouring mineralizations with the Santa Elena granite (Tamain, 1968; Ovtracht & Tamain, 1973).

In northern Portugal two types of lead-zinc mineralization of different ages are distinguished: the first type occurs in the area southeast of Oporto where several mines worked argentiferous lead-zinc veins of small extent, mostly emplaced in metamorphic Beira Schists. The best known of these is the Terramonte mine, active until recently (see hereafter). This type is characterized by a very complex mineral assemblage, by high silver tenors, by the presence of high-temperature ferriferous sphalerite (marmatite) which is earlier in paragenetical sequence than galena, and by the common presence of cataclastic textures; it is considered to be of late Hercynian age.

The second type, occurring over a large area in northern Portugal, has a simple mineral assemblage, shows low silver tenors, has sphalerite later than galena in sequences and does not display deformed textures although the mineralization generally follows faults; it is of post-Hercynian, probably mostly Mesozoic, age (Thadeu, 1973, 1977). For completeness' sake this type, though post-Hercynian, is briefly reviewed afterwards.

The first type is exemplified by the T e r r a m o n t e deposit southeast of Oporto, worked between 1966 and the middle of 1973. The orebody was a steep NE vein, 6–14 m thick, and much faulted and brecciated due to wrench fault movements. At first the ore graded 4.16 % Pb, 3.35 % Zn and 180 g/t Ag but these values decayed in later years to 2.0 % Pb, 3.0 % Zn and 100 g/t Ag (Portugal Ferreira, 1971).

The ore was mostly galena and sphalerite (marmatite rich in Cd: 3 kg/t Cd in the zinc concentrate) accompanied by minor amounts of pyrite, marcasite, chalcopyrite, arsenopyrite, bournonite and sulphosalts such as jamesonite and boulangerite in a quartz gangue with carbonates. The main silver-bearing ore minerals were freibergite and pyrargyrite with some miargyrite, polybasite and argentite (Gaspar, 1967).

Based on sulphur isotope ratios determined on some galena and sphalerite samples, indicating a juvenile source for the sulphur, Gaspar (1967) suggested a plutonic hydrothermal origin for this deposit.

Lead-zinc veins of the second type are numerous in the Beira Baixa region of northern Portugal, emplaced in Beira Schists and granite host rocks. They consist of galena and sphalerite (the latter not everywhere present) and varying small amounts of pyrite and chalcopyrite, with quartz, barite and carbonate gangue (Thadeu, 1951, 1977). Silver contents were low. None of these deposits was large and exploitation has long since come to an end.

The veins follow steep to vertical faults and fractures mainly trending NNE to NE and ENE to E.

The old C e i f e mine worked a lead vein near Penamacor in north Portugal. Here a NE crush zone in Beira Schists was mineralized with the development of four stockworks. The largest was 90 m long, 20 m wide and slightly over 100 m deep. These stockworks were formed by veinlets chiefly of galena and small amounts of barite, pyrite, dolomite, calcite and some sphalerite. No quartz was present (Thadeu, 1951).

The tungsten-tin veins at Panasqueira are cut by later faults trending N–S and ENE–WSW, and the fault gouge and breccia have been mineralized by lenses of undeformed dolomite-calcite-ankerite containing galena and sphalerite (Thadeu, 1951). Kelly and Wagner (1977) estimated from the proportion of gas and liquid in the fluid inclusions in the quartz and dolomite of these veins that they were deposited from hot fluids at temperatures above 100 °C.

This type of epigenetic hydrothermal mineralization is tectonically controlled: ore deposition took place in fault zones belonging to a regional system and is not linked to granites (Thadeu, 1951, 1973). Evidently the veins are post-Younger Granite in age, meaning that they are Permian or younger. Because they are related to regional fracturation they were assigned an Alpine age by Thadeu (1951). From fission track dating of the thermal history of the apatite in the tungsten-tin veins of Panasqueira, Kelly and Wagner (1977) infer that the lead veins are either late Jurassic (152 Ma) or late Cretaceous (79 Ma) in age. Either a Jurassic thermal event would have reheated the apatite to temperatures above 150 °C followed by slow steady cooling or this event would have been followed by rapid cooling and a second mild thermal event (not exceeding 150 °C) in Cretaceous times. Kelly and Wagner prefer the last interpretation and correlate the lead mineralization with it, while they relate the late Jurassic event to the 'early rifting of the Atlantic'. The early history of the opening of the North Atlantic Ocean started with continental stretching, fracturing and rifting. The age of this initial period was discussed by Schermerhorn et al. (1978) who concluded that rifting may well have begun in Early to Middle Triassic times. Continental separation accompanied by seafloor spreading and the generation of oceanic crust started during the Early Jurassic, about 180 Ma ago. These events are too early to be correlated with the fission-track timings of lead vein emplacement. However, somewhat later during the spreading stage, the Messejana Dolerite intruded, a very long dyke system (Fig. 2) emplaced by multiple intrusion mainly during the Early and Middle Jurassic (Schermerhorn et al., 1978). The Late Jurassic event recorded by Kelly and Wagner may be correlatable with later dolerite activity; also, during the late Cretaceous much igneous activity took place in Portugal, especially along the Atlantic coast, and post-Hercynian lead mineralization in the interior of the Meseta might well date from this time. More study is needed.

S i l v e r was an important byproduct of many leadmines in Spain, rarely in Portugal. Owing to supergene enrichment, the upper parts of lead veins such as those of Guadalcanal and Cazalla northeast of Seville yielded important quantities of native silver and silver sulphides.

Only one vein system consisted of true silver ore: at H i e n d e l a e n c i n a, about 90 km NE of Madrid (Fig. 2), three silver veins in gneiss and schist have been mined; they contained silver sulphosalts such as freieslebenite, miargyrite, stephanite and argyrite in a gangue of barite, quartz and siderite (Ahlburg, 1907; Gavala y Laborde, 1953).

Autunian phosphate

Epigenetic phosphate deposits occur in many places in western Spain (Estremadura) and adjacent areas of east-central Portugal (Alto Alentejo and Beira Baixa). In Estremadura, between Cáceres and Logrosán, these deposits have been worked since 1864 until fairly recently; the largest mine was Aldea Moret. There occur quartz veins mineralized by fine-grained apatite (often called phosphorite, that is, collophane, essentially cryptocrystalline apatite) in granite, slate or limestone host rocks, and furthermore pipe-like bodies in limestones (near Cáceres cassiterite-bearing amblygonite veins (Weibel, 1955) have been mined; these clearly belong to the tin paragenesis). Farther southeast, the Namurian limestones of the Belmez coalfield in the Sierra Morena northwest of Córdoba contain phosphorite-calcite veins and pockets (no quartz) which have been exploited during the second half of the last century.

The Aldea Moret mine southwest of Cáceres worked a large quartz-apatite vein down to a depth of 180 m. The vein, 800 m long and 1–12 m thick, cut Devonian limestone (Weibel, 1955).

Elsewhere the veins formed swarms. Near Logrosán a NE–SW quartz-apatite (phosphorite) vein swarm in Beira Schists reached 4 km width; one of the veins, the Constanza vein, 0.60–7 m thick, was over 6 km long.

As these veins in Portugal and Spain are known to cut Younger Granites, they must be of Autunian or younger age. There can be little doubt that many of the quartz-apatite veins are of hydrothermal origin and related to the Hercynian granites because of their paragenesis. From such veins in Portugal, Cotelo Neiva et al. (1952) describe perthitic orthoclase, muscovite, arsenopyrite and other minerals. These authors propose a pegmatitic-hydrothermal origin from residual fluids derived from the Hercynian granite magmas. Also, Schneider (1951) distinguished pegmatitic, pneumatolytic and hydrothermal phases during the formation of the Portuguese apatite-quartz veins. In favour of a directly granite-related origin is the fact that apatite is a common mineral in the Younger Granite pegmatites and tin-tungsten veins.

On the other hand, Ahlburg (1907) refers to the finding of phosphorite-encrusted bone breccias in the Constanza vein, to the alteration and impregnation by apatite of the granite wall rocks of the veins, and to the occurrence of phosphate nodules in slates associated with the limestones in the Cáceres district. In his opinion the phosphate mineralization is very young.

It seems to me that circulation of meteoric waters could well have remobilized and redeposited apatite at some later date, when erosion and planation had cut down to the level of the primary veins.

Late Palaeozoic chromium and nickel

Apart from the mafic-ultramafic complexes in Galicia and Trás-os-Montes, discussed under Early Palaeozoic chromium, much smaller ultramafic occurrences are found as differentiates of mafic rocks locally in the west, southwest and south Meseta. Unimportant chromium and nickel mineralizations are associated with some of these rocks.

North of Córdoba in southern Spain, Upper Palaeozoic, possibly Lower Carboniferous slates contain an alignment of serpentinite bodies associated with spilites and diabases; the serpentinites contain some disseminated chromite, magnetite and pentlandite (Crousilles et al., 1976).

Small ultramafic bodies, mostly serpentinized, occur as local differentiation products among the Early Palaeozoic (Cambrian and Silurian) spilites in the southwestern Meseta (Gaspar, 1971), and the Lower Carboniferous spilites and diabases of the Iberian Pyrite Belt (Strauss et al., 1977).

The southern part of the Beja eruptive complex in south Portugal consists mainly of gabbros and diorites with volcanic rocks; associated with these rocks occur serpentinites. In this area folded Upper Devonian limestones show minor copper and nickel mineralizations that are considered to derive from leached ultramafics (Batista et al., 1976). Pyroxenite, peridotite and dunite associated with gabbro intrusive into Cambrian carbonates occurs in east-central Portugal (NE Alentejo) near the frontier with Spain (Gonçalves, 1971).

Nickeliferous veins are furthermore known to occur in the Cabo Ortegal area in Galicia (Armengot de Pedro & Campos Juliá, 1971). None of these mineralizations are of economic significance.

Late to post-Hercynian epigenetic copper, cobalt, nickel and bismuth

Epigenetic metalliferous veins and replacements unaffected by orogenic deformation and often following Hercynian faults are widespread in the Meseta. They frequently show no obvious relationship to granite and they tend to occur in more or less well-defined metalliferous provinces. Pb, Zn and Cu deposits are most frequent. Co-Ni-Bi-quartz-carbonate veins are found along the borders of the Los Pedroches batholith in southern Spain (Crespo, 1972; Arribas, 1978). Veins carrying Co-Ni arsenides in Galicia and elsewhere are considered related to Hercynian granite magmatism.

Co-Ni-Cu veins and replacement deposits in dolomitized Namurian limestones in León, formerly mined, contain villamaninite (Cu, Ni, Co, Fe) $(S, Se)_2$, named after Villamanín in this area, together with chalcopyrite, linnaeite, pyrite, bravoite, niccolite and other sulphides and secondary minerals (Ypma et al., 1968; Arribas, 1978).

Copper veins are widespread especially in south Portugal and southwest Spain, as in the Los Pedroches batholith or in the Iberian Pyrite Belt, discussed on previous pages. Their mineral assemblage is simple: chalcopyrite may be accompanied by pyrite, by tennantite/tetrahedrite or by galena and sphalerite; the gangue consists of quartz and carbonate. In south Portugal several copper veins have been exploited, for instance the Aparís occurrence, a system of veins following shear faults in greywackes and slates. The fissures were filled by carbonates (mainly dolomite, ankerite and calcite), quartz and some chlorite. The mineralization consists of chalcopyrite with some pyrite rich in arsenic and minor arsenopyrite, pyrrhotite, marcasite, sphalerite, galena and other minerals (Gaspar, 1968). Copper veins containing uranium minerals are known in central and southern Spain (Ahlburg, 1907; Arribas, 1978).

These veins postdate the main folding and regional metamorphism, so are post-Westphalian, but their exact age remains unknown.

Late Hercynian ferromanganese

Southwest of the Pyrite Belt proper in south Portugal lies a large structure, the Cercal anticlinorium (Fig. 5), in Lower Carboniferous volcanics and sediments much like the Volcanic-Siliceous Complex of the Iberian Pyrite Belt. Here occur a large number of ferromanganese veins, of which two are still being worked. These veins reach 18 m thick by up to 5 km long and they follow NE–SW fractures dipping 40–70° SE. They consist of quartz with iron-manganese oxides (lepidocrocite and pyrolusite mostly) which are thought to derive by superficial oxidation from primary manganiferous siderite. The average grade is 8% Mn and 43% Fe. In addition the veins carry barite and in places base-metal sulphides as well. In 1977, 30,250 t ferromanganese (7.7% Mn, 40.5% Fe) and 590 t barite were produced.

This hydrothermal epigenetic paragenesis is ascribed to late Hercynian magmatic activity (Carvalho et al., 1971b; Carvalho, 1976a, b).

Autunian (?) mercury

In the Central Asturian coalfield several small mercury-arsenic occurrences, formerly mined, are found between Mieres and Pola de Lena, south of Oviedo. Cinnabar associated with realgar, orpiment, arsenopyrite (in considerable amounts: arsenic was formerly produced in this field), various other sulphides and carbonate or quartz gangue occur along fractures and joints and also as impregnations in mostly Westphalian sandstones, conglomerates, limestones and even coal seams. Ore grades averaged 0.2% Hg and 0.5–1.2% As. These occurrences, of which the most important is the Mieres mineralization, form a NNE alignment some 16 km long (Anger et al., 1968). Their age and origin is uncertain.

Autunian (?) epigenetic fluorite and barite

Quartz veins carrying sphalerite, galena and other sulphides and rich in fluorite are found in the southern Meseta. The fluorite mine El Chaparral near Cerro Muriano in south Spain exploits two long steep veins reaching 30 m thick and trending 110°–120° and 060°, respectively. The veins run in granite and the surrounding gneiss country rock. They are quartz veins with much fluorite and some sulphides, mainly sphalerite with pyrite, galena and a little chalcopyrite; barite is lacking.

Barite veins formerly worked in small mines are likewise most abundant in the southern Meseta but also occur in central and northwest Spain.

These veins are undeformed and cut deformed strata and Hercynian granites, hence their age is late Hercynian or possibly post-Hercynian.

Remobilized Autunian uranium

Uranium deposits in Hercynian rocks of the Iberian Meseta are numerous in the eastern parts of north and central Portugal and west and south Spain.

In 1907 the Urgeiriça deposit in northern Portugal was discovered and it was worked for radium from 1913 until 1944 when uranium extraction took over (Matos Dias, 1976). Systematic prospecting for uranium in Portugal began in 1955 by the Junta de Energia Nuclear, a state organization, leading to the discovery of nearly four hundred deposits of which about one hundred offer economic potential. In Spain extensive prospecting has likewise resulted in the discovery of promising uranium deposits.

In 1977 the Portuguese uranium ore reserves recoverable at up to \$130/kg U (\$50/lb U_3O_8) were assessed at 9150 t U (6800 t U reasonably assured reserves (at a cutoff grade of 425 ppm U) minable at up to \$80/kg U (\$30/lb U_3O_8), 850 t U recoverable at up to \$80/kg U by in situ leaching of worked-out mines, and 1500 t U exploitable at up to \$130/kg U (\$50/lb U_3O_8) by leaching of low-grade ores grading less than 425 ppm U) (Uranium – Resources, Production and Demand, December 1977, OECD Nuclear Energy Agency and IAEA, OECD, Paris). Portugal thus holds 0.4% of the world's uranium reserves. The 6800 t U reserves embrace 112 deposits; however, only 15 deposits contribute 80% of the reserves and one, Nisa, has 1884 t U or 27.7% of the total (Matos Dias, 1976).

In Portugal uranium extraction is in the hands of the Junta de Energia Nuclear which operates several mines in the Urgeiriça and Guarda areas, with a total output of 87.6 t U in 1976 and 66.3 t in 1977; the production is being stockpiled.

Spanish uranium production was 207 t U_3O_8 in 1976 and 201 t U_3O_8 in 1977. Two state entities exploit uranium mines in Spain. The main producer is a state company, Empresa Nacional del Uranio S.A. (Enusa) which in 1977 produced 132 t U_3O_8 from its mine near Ciudad Rodrigo (which has reserves in excess of 10,000 t). The remainder of Spanish uranium output stems from mines belonging to the Junta de Energía Nuclear.

In 1977 the Spanish uranium ore reserves at up to \$80/kg U (\$30/lb U_3O_8) stood at 6800 t U in the reasonably assured category plus another 8500 t U estimated additional resources, totalling 15,300 t U; for the cost range \$80–130/kg U no figures were given (Uranium – Resources, Production and Demand, December 1977, OECD Nuclear Energy Agency and IAEA, OECD, Paris). Like Portugal, Spain has 0.4% of world reserves.

The majority of the uranium deposits of the Meseta are situated within or near Younger Granites. The intragranite deposits are veins, and the deposits around the granites, mostly within the contact aureoles, are veins and disseminations. There is generally a clear tectonic control in that the veins and disseminations follow shears, faults and shatter zones. As for the Pb-Zn veins this fracturation has been assigned an 'Alpine' age and the mineralization was considered to be Alpine too, that is, Tertiary. It now seems likely that it is largely of Mesozoic age.

The main zone of uranium deposits of economic importance stretches from northern and central Portugal over western Spain into southern Spain (where deposits in the Pedroches batholith have been worked) up to the Guadalquivir fault.

The Urgeiriça mine in north Portugal exploited uraniferous jasper veins in granite; ore extraction has ceased and the deposit is now worked solely by leaching (the applicability of this method was discovered when 5000 tonnes of ore grading 0.5 % U_3O_8 were stockpiled in the open during four years, when it had lost about 70% of the contained uranium due to the winter rains). Mineralization is controlled by an ENE wrench fault 7.5 km long with a crush zone 1–10 m thick. Workable uranium mineralization extends over a distance of 1300 m along this zone and consists mainly of pitchblende in jasper gangue with some pyrite or marcasite (Cameron, 1960; Portugal Ferreira, 1971).

The Nisa prospect, discovered in 1957 during a scintillometer survey, is by far the largest uranium deposit in Portugal. It consists of irregular disseminations along crush zones in the contact-metamorphosed Beira Schist wall rock of the coarse porphyritic Nisa granite, a Younger Granite, which contains several other uranium deposits. The uranium orebodies in the crush zones extend to a depth of 25 metres. Autunite predominates, accompanied by other uranium phosphates and rare pitchblende and black oxides (Portugal Ferreira, 1971).

Of the same type are the orebodies in the Ciudad Rodrigo region (Mina Fé, Mina Esperanza and others) in western Spain, which represent the largest economic uranium concentration in Spain. The principal ore mineral is pitchblende, accompanied by its alteration products (gummite, etc.), some pyrite and a gangue consisting of quartz, jasper, chalcedony, barite, calcite and fluorite (Arribas, 1960, 1975; Fernández Polo, 1970).

The origin and mode of deposition of these uranium deposits has been diversely interpreted. Originally regarded as Hercynian endogenic hydrothermal mineralizations it was soon realized that exogenic factors must have played a role.

Matos Dias and Soares de Andrade (1970) classify the Portuguese uranium ore occurrences as concordant and disconcordant (discordant according to Portugal Ferreira, 1971) deposits; so far the only known concordant deposit consisted of uranium phosphate impregnations in lacustrine sandstones in the Urgeiriça region. The disconcordant deposits make up the remainder: they are veins and disseminations (with Portuguese reserves evenly distributed between the two – Matos Dias, 1976) and are associated with the Younger Granites. The vein deposits embrace veins with jasper gangue, veins with milky and smoky quartz gangue, quartz-limonite breccias and weathered mafic dykes; they show mainly NE trends. In dissemination deposits uranium minerals impregnate altered granite or metasedimentary wall rock. Wall rock alteration (sericitization, hematization and limonitization) is characteristic. Hematized granite may have suffered desilicification too (Junta de Energia Nuclear, 1968). This classification seems applicable to the Spanish Meseta deposits as well.

The uranium minerals are pitchblende, black oxides, coffinite, gummite and diverse coloured silicates such as uranophane, phosphates (autunite, torbernite, sabugalite etc.) and sulphates (uranopilite and zippeite). Pyrite and/or marcasite are frequent, other sulphides appear in minor amounts and calcite and other carbonates are infrequent.

None of the known U mineralizations extends to any great depth, not even pitchblende (max. 50 m generally).

Apart from their close spatial association with Younger Granites the uranium deposits also display a relationship to old planation surfaces in central and northern Portugal and they are controlled by shears, faults and fractures marked by tectonized zones.

Matos Dias and Soares de Andrade (1970) present the following genetical model: uranium was leached by circulating meteoric waters from the Younger Granites which are rich in radioactive elements (Cameron, 1960) and accumulated in the soil on Mesozoic planation surfaces. Afterwards uraniferous solutions from these preconcentrations infiltrated joints, fractures, breccias and mafic dykes in the bedrock and deposited uranium minerals in open spaces.

According to calculations, removal of a quarter of the original uranium content (stated to average about 47 ppm U_3O_8) by sericitization and the breakdown of biotite would release enough U and Fe to account for the uranium deposits and the wall rock alteration (Cameron, 1960). Radiometric dating (using Pb^{206}/U^{238}, Pb^{207}/U^{235} and Pb^{207}/Pb^{208} methods) of some Portuguese pitchblendes resulted in two groups of ages: 80 ± 20 Ma, 83 ± 8 Ma and 190 ± 10 Ma, and this fits the Upper Cretaceous and Triassic planation periods (Matos Dias & Soares de Andrade, 1970; Portugal Ferreira, 1971). In post-Hercynian times strong erosion cut down to expose the Hercynian granites, and during a long phase of planation conditions were favourable for prolonged leaching and redistribution of uranium. Another major episode of erosion and planation occurred during the Cretaceous.

It thus appears likely that while the deposition of the uranium orebodies took largely place in Mesozoic times, their source is to be sought in the Younger Granites; hence these deposits consist of remobilized Autunian uranium.

SYNTHESIS

Various metallogenetic provinces and epochs have been outlined in the foregoing; especially significant and productive because of exceptionally strong concentration of ore elements are (Fig. 2):

1. the skarn iron province of the magnetite belt in the Ossa-Morena Zone, in Lower Cambrian carbonates – as stated, these deposits may be syngenetic, as reaction skarns, and/or epigenetic, as metasomatic skarns.
2. the Lower Ordovician marine-sedimentary ironstone province in the northwest Meseta, fringing an emergent landmass – these are syngenetic deposits.
3. the Lower Silurian mercury province of the Almadén area in south Spain – a syngenetic deposit that may be wholly volcanogenic or may have a sedimentary source with deposition activated by volcanic heat.
4. the Lower Carboniferous polymetallic pyritite province of the Iberian Pyrite Belt in the southwest Meseta – syngenetic volcanogenic massive sulphides.
5. the Lower Permian tin-tungsten province of the northwest and central Meseta – mostly epigenetic granite-bound deposits.
6. the Permian lead province of the eastern Sierra Morena in south Spain – epigenetic lead-zinc veins.

In terms of tectonic environment we find much varying settings of mineralization. The iron ore of the magnetite belt, if syngenetic, would have been deposited on a Lower Cambrian carbonate shelf, developed during a major lull in the clastic infilling of the Cambrian geosyncline covering the Meseta. The Lower Ordovician iron ores accumulated in shallow water along a coastline at the end of a transgression phase. The Almadén mercury was deposited on a Silurian platform during an episode of mafic volcanism. The Pyrite Belt is the only major orefield developed outside the central block, in the southwestern geosyncline, when tensional movements during the early Carboniferous set off intense felsic and mafic volcanism. The tin-tungsten ores are associated with the second episode of major granite intrusion. The lead-zinc veins followed late- to post-orogenic shear faults.

There is no obvious metallogenetic zonation, and the distribution of these major ore provinces, let alone the immense variety of minor deposits, cannot be adequately explained by plate-margin tectonics involving subduction, collision, arcs etc. Local strong subsidence, block-faulting, geanticlinal uplift and emergence, as well as well-developed volcanism ranging from mafic to felsic, characterize the history of the Meseta and indicate differential vertical movements and crustal tension. The tectono-sedimentary facies and their palaeogeographical control plus the symmetrical structure of the Meseta (as expressed in two outer geosynclines or foredeeps flanking a large rising central block, in outwards verging folds and thrusts and in the abundance of granites in the central block) produce a pattern consistent with an intraplate environment where orogenic evolution is controlled by vertical tectonics and great magmatic activity that find their origin in the underling mantle. The broad framework of Hercynian mineralization is within-plate, not plate-margin.

Mantle activity below the Meseta all through the Hercynian cycle is indicated by strong volcanicity in almost all periods from the Late Precambrian on, by repeated intrusion of granitic magmas, and by tectono-sedimentary differentiation into a mosaic of subsiding, stable and rising palaeogeographical units.

I interpret this in the following terms. Mantle upwellings below the Hercynian orogen affected the crust by producing uplifts and by generating fractionates that rose through the crust as volcanic and plutonic magmas. Some mantle diapirs came to pierce the crust, forming the mafic-ultramafic complexes of the northwest corner of the Meseta (van Calsteren, 1977).

One period of uplift and related tectonism took place towards the end of the Cambrian. Over most of the Meseta the Cambrian strata (including Beira Schists and equivalents) emerged and became variably eroded; locally they were folded. It seems likely that this resulted from heightened mantle activity: the ultramafic diapirs of the northwest date from this time, as does the uplift of this corner of the Meseta and its emergence during the Ordovician (Fig. 4). Thus the Early Ordovician iron deposits laid down around the emerged area ultimately owe their origin to mantle convection. As mantle diapirism and the ascent of fractionated magmas continued in the northwest, the surrounding zone subsided strongly, and the Upper Ordovician, Silurian and Devonian flysch was deposited, a facies contrasting with the platform facies dominant elsewhere in the Peninsula.

At the beginning of the Carboniferous the Meseta became differentiated into a central block with rising tendencies and two marginal basins. The emergence of the central Iberian block coincides with the emplacement of the Tournaisian granites at ca. 350 Ma and the eruption of the extensive Pyrite Belt felsic volcanism during the Tournaisian and earliest Viséan. The central block was the site of metamorphism and granite intrusion. This appears to indicate the slow rise of magma underneath this block, while the adjacent secondary geosynclines of the Cantabrian and South Portuguese Zones subsided.

The uplift of the central block eventually gave rise to folding, with the folds spilling over into the marginal troughs. As a result, vergence is to the north in the northern Meseta and to the southwest in the southwest. Most of the folding took place during the main orogenic phase, during the Westphalian. Though some folding and refolding occurred afterwards, later tectonism was mainly of the nature of wrench-faulting and block-faulting.

The ascent of hot mantle material into the lower crust caused extensive remelting of lower to middle crustal levels, resulting in the generation of the Older Granite magmas. At a higher level, in the middle to upper crust, the rise of geotherms over the mantle domes originated regional metamorphism, locally passing to migmatization and anatexis. The Older Granite magmas gained the higher crustal levels somewhat later. Almost no important mineral deposits of metamorphic origin are known and none derive from migmatization and anatexis, nor are the Older Granites associated with much mineralization. This means that the crustal level where they originated lacked ore elements in sufficient concentration.

Twenty million years later, however, a new granite magma intruded the earlier granites and their country rock: the Younger Granites which are distinguished from the Older Granites by a higher content in radioactive minerals and by their mineralization.

Shortly afterwards the consolidated crust of the Meseta, composed of folded and variably metamorphosed strata and vast granitic massifs, was faulted and fractured. No large offsets occur and the faults probably arose from uneven stresses set up in a rigid crustal slab in response to slight movements in the underlying mantle. Thus channelways were opened to post-granitic hydrothermal solutions from which epigenetic vein minerals precipitated, some of them of granitic derivation, the others produced by near-surface recycling.

This completes a broad outline of framework and evolution of mineralization during the Hercynian cycle in the Iberian Meseta. Now let us look at some deposits and their possible sources in more detail.

First the part played by volcanism in metallic sulphide, sulphide-oxide and oxide mineralization.

The origin of the Cambrian lead-zinc sulphides that occur as disseminations in carbonates in the northwest and southwest Meseta, is obscure since they are not obviously related to volcanism, unlike the massive base-metal sulphides of the pyritite type. For the Pb-Zn mineralizations of the northwest Meseta, such as Rubiales, Guillou (1971a) infers an exogenic source of the metals, on an adjacent weathered continent. The same explanation may also hold for the lead-zinc belt in south Portugal, though it may be noted that volcanism took place in the southwest Meseta during the deposition of the carbonates. The ore of the magnetite belt, if Cambrian and synsedimentary, may well be volcanogenic, as its composition (high S, low P) and its association with copper sulphides in economic amounts does not favour a purely sedimentary origin.

The other stratiform base-metal sulphide mineralizations – the Late Precambrian (or Silurian) copper sulphides in Galicia, the Silurian deposits in the northwestern and southern Meseta and the Lower Carboniferous Iberian Pyrite Belt – were deposited during important volcanic episodes and are submarine exhalative-sedimentary (and resedimented) ores. The volcanism was attended by hydrothermal activity on the seafloor and sulphides were precipitated around submarine vents.

The Late Precambrian deposits in Galicia and the Silurian sulphides are associated with mafic volcanics, and the felsic volcanism taking place in these epochs is barren. However, the much more important Lower Carboniferous sulphides (and the manganese) of the Pyrite Belt are related to felsic volcanism, and the contemporaneous mafic volcanism is practically barren. There is thus an evolution from mafic to felsic magmas as ore-element carriers. Even though a greater or smaller part of the sulphur in these sulphides may be derived from seawater sulphate, the metals must be largely or entirely of juvenile origin. I deduce this from the contrast between contemporaneous ore-bringing and barren volcanics. On a larger scale there is contemporaneity or nearly so between felsic and mafic magmas but their metalliferous potential is completely different in any one epoch. On a smaller scale, mineralization occurs during only one or at most a few episodes during a volcanic period and is linked to only a few of the erupting volcanics. The other rocks in the volcanic piles are barren. If volcanic heat were all that is needed to mobilize and redistribute metals contained in seawater, in contemporaneous volcanics or in basement rocks, the resulting ore would be dispersed through the volcanics, not highly concentrated at one particular level in one particular place. Hence the metals must have a deep-seated origin and have become concentrated in one particular magma phase, by differentiation.

Next the role of plutonism. Leaving aside the Bragança, Cabo Ortegal and other mafic-ultramafic complexes with their minor chrome deposits, most endogenic mineralization is linked to only one of the four or more Hercynian granite suites: the Younger Granites (the magnetite belt constitutes a doubtful case: if the iron ore deposits are not syngenetic but epigenetic, they are skarns generated by metasomatic introduction of ore elements from the granitic magmas in that area – the age of this possible granite-bound metallization would be the age of the granites, and this age is still unknown).

The Younger-Granite-bound endogenic mineralization can be divided into the following groups: Sn-W and related elements (Nb, Ta, Mo, Be, Li); Au-Ag; Sb-Au. Part of the Pb-Zn-Ag mineralization belongs here too, and the U mineralization derives from these granites. Why is this metallization linked to the Younger Granites and what is the source of this magma and the associated ore elements?

First, the contrast between the Younger and the Older Granites as regards associated mineralization indicates that their sources are different. This provides important constraints on the source region of the Younger Granites and their ores. If the Older Granites represent recycled middle crust, then the Younger Granites must derive from still deeper crustal levels or, ultimately, from the mantle. Though the Younger Granite magma may have been water-undersaturated, it was not a dry one. The ore elements and volatiles (especially F and Cl to act as tin carrier ions) in the Younger Granites can hardly be explained by assimilation from middle and upper crustal levels, since the Older Granites which consist of remelted middle crust – and also the Hercynian geosynclinal rocks and their basement (which we know from its detritus in the geosyncline), that is, the upper crust – are poor in these constituents. Besides, if the Older Granites could not extract these elements from their parent material or from the overlying crustal levels through which they rose, how could the Younger Granites do so?

Second, the Younger Granites are a very diverse suite, much more so than the Older Granites, and range from mafic rocks to calcalkaline and alkaline granites, including greisens. This indicates strong differentiation in the magma and supports an origin by fractional crystallization from more mafic material without significant crustal contamination. This material may have been derived by partial melting from the ultramafic upper mantle or it may have constituted the lower crust. The latter possibility is considered unlikely because mafic or intermediate crustal materials such as basalts would be too poor in ore elements of the Sn-W group. Besides, the enormous heat transfer leading to regional metamorphism and granite intrusion over the width of the Meseta orogen – 600–700 km – requires a vast source such as can only be provided by the mantle. It appears that mantle material, possibly a derived mafic magma, ascending into the crust and introducing heat caused melting on a vast scale of the crust above it, thus generating the Older Granite magmas. The Younger Granites are then to be seen as later contaminated fractionates of the mafic magma, slowly growing into large magma blisters in the deep crust. The Younger Granites, on this view, may be of mantle derivation, including assimilated crustal material.

Next, the source of the metals, which has been disputed. Notably the main source of the tin and tungsten and of the accompanying metals has been sought in sedimentary preconcentration in geosynclinal sediments, with the intruding granite magmas extracting, concentrating and redepositing the metals. Thus Ypma (1966) suggested that tin and other metals would have been present in sedimentary formations in the crust, entering the ascending magmas. For tungsten the matter is more complicated than for tin since sedimentary tungsten is known in the Meseta: it seems likely that the tungsten disseminations in Cambrian and Silurian metamorphic marly beds are syngenetic stratiform deposits, because of their extent; later contact or regional metamorphism induced local concentration into skarn-type pockets. The source of this tungsten remains problematical: the scheelite may be a mechanical sediment of placer type, eroded from pre-existing deposits, or else it may have formed as an exhalative product, trapped in marly sediments.

Cambrian and Silurian syngenetic tungsten has been proposed as a preconcentrated source for the later epigenetic deposits associated with the Hercynian granites, for instance through anatexis of tungsten-bearing Cambrian or Silurian strata to yield a mineralized magma (Noronha, 1976).

However, Older Granites are mostly not associated with tin and tungsten mineralization but Younger Granites breaking through them are. When the tin and tungsten associated with the latter granites have been deposited in granite – that is to say, the ore-bearing Younger Granite has not been in contact with Cambrian or Silurian sediments – then these metals must certainly derive from deeper levels even than the middle crust where the Older Granites originated by anatexis.

If the Younger Granites stem from fractionated mantle-derived magma, and if the Sn and associated ore elements could not have been collected from crustal rocks, because the Older Granites failed to do so, then these metals too must have a mantle source. Evidently an exception must be made for that part of the tungsten minderalization that represents Cambrian and Silurian syngenetic stratabound scheelite deposits. Still, most of the Sn, W, Nb, Ta, Mo, Be, Li, U, Au, Ag and related elements may have been brought up direct from the mantle. On this heretical note we conclude this section.

To sum up, ore elements have been derived 1) by weathering, leaching or other exogenic processes from the surface of the crust – this was probably the origin of the large ironstone deposits of Ordovician age in the northwest Meseta and possibly of the Cambrian lead-zinc ores; 2) by volcanic or plutonic heat from pre-existing rocks in the upper crust – this has been proposed as a possible origin for the Almadén mercury; 3) by volcanic or plutonic magmas from lower crustal or upper mantle levels – this is the case for the base-metal sulphide deposits of Late Precambrian, Silurian and Carboniferous age, for the volcanogenic iron deposits, and for the granite-bound mineralization of the Sn-W province.

It seems very likely that lineaments were instrumental in localizing mineralization. Overall structural control of mineral deposits in the Meseta may have been exerted – in this intraplate setting – by deep crustal zones of structural weakness that acted as channelways to magmas and ore solutions. Local structural control became pronounced in the case of the late Hercynian and post-Hercynian vein deposits.

If it is considered that of the major, economically important deposits of the Peninsula only the mercury of Almadén is associated with mafic volcanism, while metals of the Pyrite Belt and the tin-tungsten province are related to felsic volcanism and plutonism, it is clear that felsic magmas played a more important role than mafic magmas.

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