

SPLITES, REGIONAL METAMORPHISM AND SUBDUCTION IN THE IBERIAN PYRITE BELT: SOME SOMMENTS

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ABSTRACT

The Devonian-Lower Carboniferous eugeosynclinal succession of the Iberian Pyrite Belt was deformed and regionally metamorphosed during the Hercynian orogeny. Metamorphic grade decreases from lower greenschist facies in the north to pumpellyite facies in the south. The volcanics comprise a felsic and a mafic-intermediate suite, the latter including spilite lavas and tuffs with intrusive diabases. Spilites precede, follow, and are contemporaneous with felsic eruptions. The parent magma of the spilites and albite diabases was not altered notably by contamination and assimilation. The felsic volcanics are sodic to potassic quartz keratophyres, mostly tuffs. Rio Tinto stratigraphy is reviewed. The possibility that the compositional bimodality of the volcanics might be related to subduction seems remote.

1. INTRODUCTION

A paper by Soler (1973) on volcanic, orogenic and metamorphic relationships in the Iberian Pyrite Belt reaches conclusions that are at odds with the views held by other workers in this area, and this prompts me to offer some critical comments in the hope to help clarify the points at issue.

First a rapid review of Pyrite Belt geology (fig. 1). The strata enclosing the massive sulphide bodies of this metallogenetic province were laid down in an external geosyncline flanking the central Hercynian block of Iberia. This basin displays the hallmarks of a classic eugeosyncline: its fill consists of sediments and volcanics, with a thick flysch pile at the top. As in many other geosynclines the composition of the volcanics is bimodal (felsic and mafic-intermediate). The outcropping strata are of Devonian and Lower Carboniferous age; the base of the succession is not exposed and the first orogenic phase is of Middle Westphalian age. This geosyncline — including the actual Pyrite Belt, a metalliferous zone 230 km long by up to 40 km wide — covered South Portugal and Southwest Spain up to the central Iberian block on the northeast (Schermerhorn, 1971a). Soler shuns the term "geosyncline", using in its stead "flexure préorogénique", and equates the geotectonic setting with a cordil-

leran-type mountain belt originated by oceanic subduction.

The Pyrite Belt succession comprises three lithostratigraphic groups. Oldest is the Phyllite-Quartzite Group (abbreviated PQ); its downward extent is not known but it contains Famennian fossils at the top. PQ is overlain by the Volcanic-Siliceous Complex (symbol: VS), named for its most characteristic lithofacies: felsic to mafic volcanics associated with prominent jaspers. The assemblage further includes various types of sedimentary rocks, and sulphidic and manganese ores. VS is of Tournaisian and Lower Viséan age (possibly including the very latest Devonian). At the top of the succession is the Culm Group, a thick flysch sequence of turbidite greywackes and mudstones. Its deposition starts with the Upper Viséan.

Vulcanicity was active throughout VS times though much varying in intensity and in composition from place to place. In particular, spilitic extrusives are known in VS from its base up to the top (Schermerhorn, 1970b). Soler contends that the mafic volcanism was confined to the end of VS times and always later than the felsic volcanism.

The stratiform sulphide deposits of this ore province appear at varying horizons in VS. They consist of pyritite, a name for sulphidic rock (Schermerhorn, 1970a, 1971b, 1971c), and are of volcanic origin, being related, like most of the manganese ore deposits (primary rhodochrosite and rhodonite), to the felsic volcanism. Part of the pyritite sheets accumulated by resedimentation; they exhibit various kinds of sedimentary structures. The most famous deposits are at Rio Tinto, where Soler presents a columnar section that differs markedly from established stratigraphy in that area.

The first and main orogenic phase threw the strata into folds verging to the south and southwest, with thrusting and wrench-faulting. Axial-plane cleavage was developed. Related to this phase occurred low-grade regional metamorphism. Soler denies its existence.

2. REGIONAL METAMORPHISM

Geologists working in the Pyrite Belt have generally recognized the presence of Hercynian regional metamorphism of a

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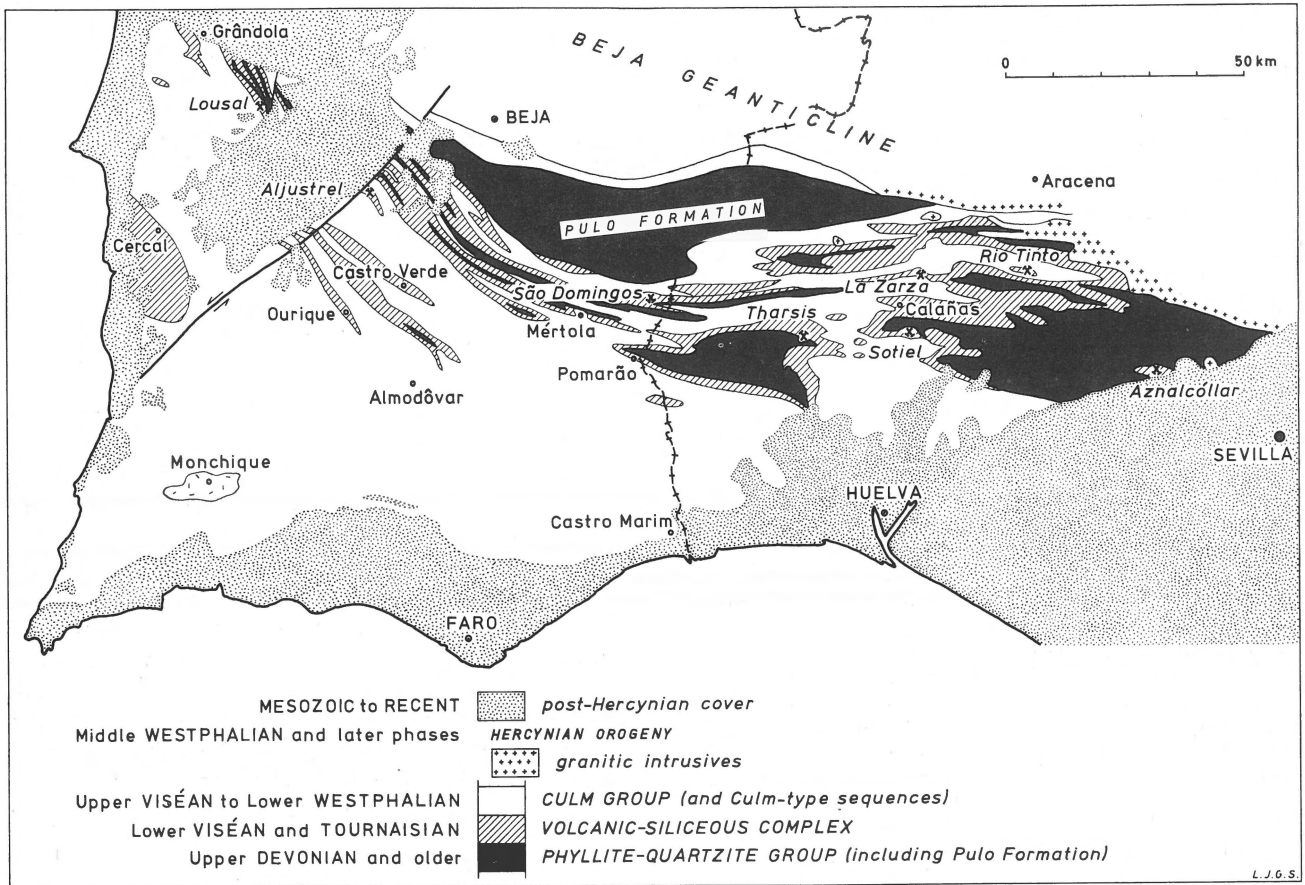


Fig. 1
Outline geology of the Iberian Pyrite Belt.

low grade, not surpassing chlorite-zone conditions (Schermerhorn, 1971a). Metamorphic recrystallization was evident to them in Pyrite Belt rocks from PQ through VS to Culm. In the Pomarão anticlinorium in South Portugal, van den Boogaard (1967) established the existence of prehnite-pumpellyite facies metamorphism.

This notwithstanding, according to Soler (1973) there only took place extremely weak anchimetamorphic recrystallization in the mafic rocks while the felsic volcanics and the sediments remained unmetamorphosed.

The mafic (to intermediate) volcanics of VS are diabase (s.l.) intrusives and spilitic (s.l.) extrusives. The Pyrite Belt rocks were affected by shearing to varying extents. Even very thick (hundreds of metres) spilite piles are cleaved throughout. Thick diabase sills show massive interiors but their borders are sheared, and thin sills may be completely schistose. Deformation can go as far as destroying the original igneous (ophitic or intergranular) textures of these pigeonite dolerite intrusions, producing a very fine-grained schistose aggregate of chlorite, sericite, leucoxene and calcite. Original amygdules have been strongly flattened to chlorite or calcite streaks. Now, in sheared diabases and in some

spilites there appear ragged actinolite poikiloblasts that have grown across the cleavage. This represents metamorphic recrystallization, separated by a phase of orogenic deformation from the magmatic crystallization of the rocks. As the latest mafics (the spilites in upper VS) are of Lower Viséan age while the orogenic phase causing the first cleavage was of Middle Westphalian age, the postkinetic actinolite blastesis was certainly much later than magma emplacement: it was not caused by deuteric or similar processes but is metamorphic.

Similar evidence is afforded by the felsic volcanics and the sedimentary rocks. Mudstones and shales have been transformed into phyllites. Very fine-grained felsic volcanics (dust tuffs) have been converted into siliceous slates and phyllites. Volcanic-derived sericite schists may enclose plentiful transverse chlorite porphyroblasts reaching over 1 mm length (fig. 2). In cleaved felsic tuffs the shear planes, winding around quartz and feldspar phenocrysts, are marked by trains of white mica scales, sometimes accompanied by chlorite. In some tuffs unstrained post-cleavage epidote is present, in part forming radiating aggregates.

Metamorphic recrystallization is also discernible in the

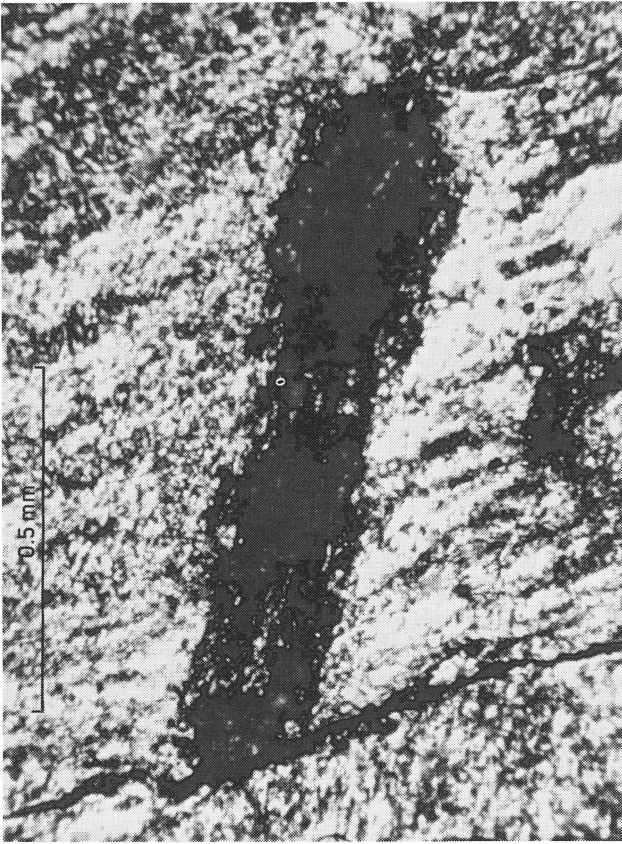


Fig. 2
Chlorite porphyroblast in sericitic siliceous phyllite (dust tuff) of Gavião Formation, southwest of Aljustrel. Crossed nicols.

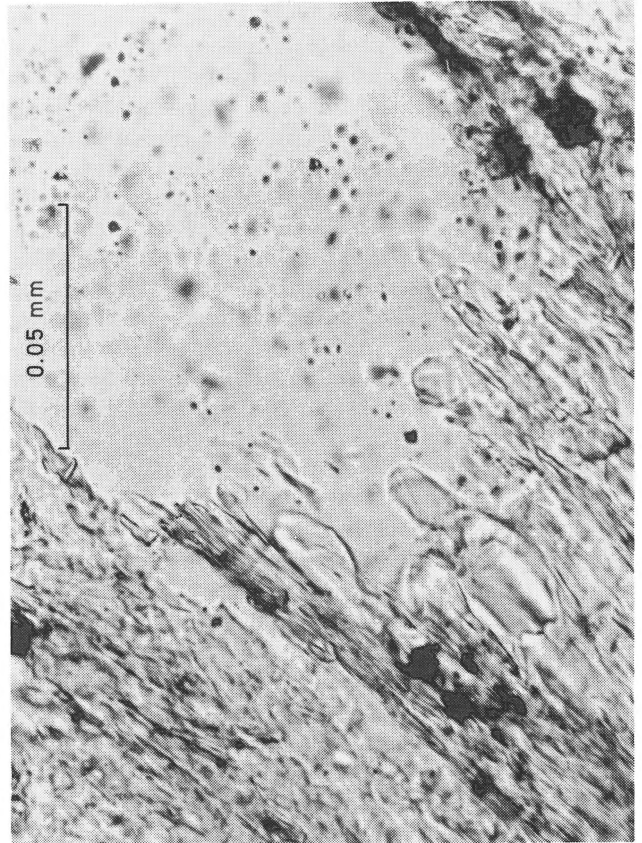


Fig. 3
Chlorite corroding clastic quartz grain in Culm greywacke, south of Aljustrel.

Culm. The sandstones of this group are albite-lithic greywackes composed of clastic grains of quartz, albite and various rock types, mostly fine-grained felsic (quartz-keratophytic) volcanics, with sedimentary clasts, detrital muscovite and altered biotite, and sparse heavy accessories. These greywackes are sheared to varying extents, and the interstitial matrix has recrystallized to fine-grained chlorite (including oxychlorite) with sericite and occasionally some calcite. These new-grown minerals are oriented parallel to the cleavage, are undeformed (unlike the clastic micas) and penetrate into adjacent quartz, albite and other grains, producing microserrated outlines (especially in the pressure shadows), as figure 3 shows. Again, this blastic phase is posterior to (or rather, outlasts) the folding and cleavage caused by the first orogenic phase.

The metamorphic grade decreases to the south. In the southernmost Culm, at Castro Marim, the rocks are folded but not yet cleaved. The composition of the greywackes is the same as farther north but the matrix consists of an ill-defined clayey-silty paste in which recrystallization has barely started (a little extremely fine-grained chlorite can be made out). The clastic grains have well-preserved sharp outlines.

Summing up, the Pyrite Belt was affected by low-grade regional metamorphism originating at the end of the first folding phase and thus of Middle Westphalian age. It caused the growth of late to post-deformational blastic minerals, including actinolite and chlorite porphyroblasts transecting the cleavage.

The mineral assemblage characteristic of the metamafic-intermediate rocks of much of the Pyrite Belt is: albite-chlorite-epidote (and clinozoisite)-sphene (leucosene)-calcite (-sericite-actinolite-quartz). Pyroxene occurs as relics in varying stages of replacement: as its amount diminishes chlorite increases. Ore is always present, and potash feldspar occasionally. Strauss (1965) lists this assemblage for diabases and spilites in the area around Lousal pyrite mine, and it has also been recorded from elsewhere in the Pyrite Belt and from the Cercal anticline to the southwest (Kleyn, 1960).

In the felsic metavolcanics the assemblage is: albite-quartz-white mica (-epidote, zoisite, clinozoisite-chlorite-potash feldspar). In the metasediments it is: quartz-white mica-chlorite (-albite).

The regional metamorphism producing these assemblages can therefore be assigned to the quartz-albite-muscovite-

chlorite subfacies of the *greenschist facies*.

At Pomarão appears a lower-grade facies, the *pumpellyite facies* (as defined in Schermernhorn, 1975). Here the mudstones of PQ, VS and Culm have recrystallized to phyllites with sericite and chlorite scales oriented parallel to the cleavage, occasionally accompanied by larger aggregates of these minerals grown at an angle to the cleavage. The Culm greywackes show abundant chlorite as a new-formed mineral. The felsic metavolcanics enclose muscovite, chlorite, albite and epidote as post-cleavage blastic minerals. This is like the greenschist metamorphism of the same rock types elsewhere. However, the diabases (including "keratophyres") contain pumpellyite and prehnite in considerable quantities (van den Boogaard, 1967) and this association is critical. Van den Boogaard reports: albite, pyroxene, chlorite, sphene, leucoxene, ilmenite, magnetite, sericite, quartz, carbonate, potash feldspar, brown and green mica (which appear to be oxychlorite), epidote, five varieties of amphibole, prehnite and pumpellyite. The amphiboles, ranging from brown to green, colourless and violet-blue, are partly primary minerals, partly later overgrowths on pyroxene or primary amphibole.

A violet-blue amphibole called "crossite-like" by van den Boogaard is present in two of his samples; he thought it might be magmatic, judging unlikely a metamorphic origin. This is the basis for Bard's (1971) statement that glaucophane appears in the Pyrite Belt, and the suggestion that blueschist metamorphism took place (Bard et al., 1973). However, a look at the Pomarão material, kindly made available by Dr. van den Boogaard, confirmed that the crossite-like amphibole is found in two diabases as scattered minute crystals, too small for reliable optical determination. It might be magnesioriebeckite, a mineral forming over a wide range of conditions (Lajoki and Ojanperä, 1973). Definite identification must await microprobe analysis. It is extremely rare: later searches at the sample localities failed to turn up any more blue amphibole (nor has it so far been found anywhere else in the Pyrite Belt). Anyhow, these slightly recrystallized rocks are not at all like true glaucophane or crossite schists (blueschists) showing advanced metamorphic reconstitution. No blueschist metamorphism took place in the Pyrite Belt.

East of Pomarão pumpellyite-facies metamafics continue into Spain, extending south of the zone of greenschist-facies metamorphism (Schermernhorn, 1975). In South Portugal most of the Pyrite Belt is in the greenschist facies. In the Iberian Pyrite Belt there is a northern zone metamorphosed to the lower greenschist facies and a southern zone metamorphosed to the pumpellyite facies; south of the Pyrite Belt the regional metamorphism fades out.

3. STRATIGRAPHIC POSITION OF THE SPILITES

Mafic to intermediate lavas and tuffs, the spilitic suite of VS, show their largest development in the west of the Pyrite

Belt, in South Portugal. Here they form the upper part of VS, separated by up to a few tens of metres of slates, siliceous slates and jaspers from the Culm. Locally spilites are overlain by *Posidonia becheri*-bearing Culm slates. In Spain spilites occur at the base of VS (Schermernhorn, 1970b) and basal spilites have recently been discovered in Portugal, southeast of the region where they appear in upper VS (unpublished work). Elsewhere felsic volcanics are found at the base of VS and at its top. This seems sufficient evidence to demonstrate that the mafic volcanism, though independent from the felsic extrusions, is broadly contemporaneous. Yet Sole (1973) maintains that the mafic volcanism is always later than the felsic volcanism.

In the north limb of the large anticlinorium stretching south of Rio Tinto (fig. 1) an extensive level of mafics crops out, overlying PQ and overlain by felsic volcanics which are followed by thin VS slates and siliceous slates, covered by Culm. The mafic volcanics in the lower part of the VS succession, in contact with PQ, comprise diabase sills and spilite flows (Williams, 1934, 1966; Rambaud, 1969; Moncada et al., 1970; Armengot, 1972). Williams (1966) refers specifically to pillowed spilites with shale and chert interbeds and with radiolarian chert infilling the interstices between pillows. Rambaud (1969) describes spilites, spilitic andesites and associated tuffs in a large area around Rio Tinto, occurring at the base or in the lower part of VS.

Farther south of Rio Tinto, mafic lavas are found at the base of VS and higher up (Armengot and Vázquez, 1972).

For an area farther west, between Tharsis and Sotiel, Fabel (1967) reported that the mafic lavas and tuffs are at the base of VS, being followed by felsic volcanics. In the Tharsis district (Strauss and Madel, 1974) spilite lavas and tuffs form a horizon resting on basal VS slates with some tuffs, and are overlain by thick felsic tuffs.

Spilites also occur near the top of VS in Spain, as at La Zarza (Strauss and Madel, 1974) or south of Calañas.

In South Portugal, north-northeast of Almodôvar, basal spilites overlie PQ phyllites with limestone lenses (containing conodonts for which a Famennian age was determined by Dr. M. van den Boogaard), sometimes in contact with limestone (so these mafics might be of latest Devonian age). The spilites are overlain by felsic volcanics, covered by VS slates and Culm.

There can therefore be no doubt that spilitic effusions took place throughout VS times, from the beginning until the end, though their stratigraphic positions vary between regions.

A related point is the supposed migration in time of the volcanic eruptions: earlier in the north, later and attenuated in the south (Sole, 1973). The evidence from South Portugal indicates otherwise: here the southernmost Pyrite Belt volcanics occur (fig. 1), and these are certainly of earliest Carboniferous age, or even latest Devonian. Also, here are the thickest, most extensive spilites and felsic volcanics. In the

Cercal anticline volcanism may be even older (Kleyn, 1960).

4. COMPOSITION AND ORIGIN OF THE SPILITES AND ALBITE DIABASES

A study of mafic volcanism in the Pyrite Belt geosyncline in South Portugal (Schermerhorn, 1970b) established the following:

1. there occur effusive spilites and intrusive albite diabases;
2. they are not ophiolites as ultramafics are lacking;
3. though the mafics are juxtaposed to felsic volcanics in space (VS in the Pyrite Belt) and time (Tournaisian and Lower Viséan) the two volcanisms are separate; no lithological transitions occur, the eruptive centres are distinct, and the two magmatic suites are not linked by differentiation.

It was therefore suggested that:

4. the felsic and mafic magmas originated and evolved separately;
5. the source of the mafic magma must be sought in the lower crust or the upper mantle, most likely in fractionated ultramafic mantle material;
6. the more variable felsic volcanism derives from local magma chambers developed by melting in the deep crust, possibly by heat supplied by rising mafic magma;
7. the mafic magma was not significantly affected by contamination and assimilation of sialic crust;
8. if the composition of spilites and albite diabases is not a primary feature it can only be due to modification during ascent of the magma, not on contact with seawater, since the diabases which intruded into sediments show the same composition as the spilites which are submarine lavas and tuffs;
9. the carbonate in the mafics might be a primary feature of the magma.

Soler (1973) is in tacit accord with points 1-6 and 8 but ascribes the sodium and water content of the spilites to contamination by connate water in geosynclinal sediments, derives their silicon from assimilated sialic crust, and proposes a new theory of spilite formation by intratelluric quenching of basaltic magma.

The origin of spilites and associated albite diabases is in dispute. Four main groups of theories on spilite genesis can be distinguished: (1) postmagmatic alteration of existing rock: spilites are metabasalts; (2) late-magmatic alteration of crystallizing rock: spilites are autometasomatized basalts; (3) magmatic alteration: spilites are contaminated basalts (that is, they crystallized from basaltic magmas that assimilated seawater during or after emplacement, directly or from marine sediments); (4) no alteration: spilites are spilites (they crystallized from spilitic magmas or from hydrous andesitic or basaltic magmas, possibly already differentiating towards mugearite). Adherents of theories (3) and (4) sometimes

postulate primary magmatic crystallization of albite and chlorite, the typical spilite minerals.

Though under appropriate circumstances any of the above processes could play a dominant role in spilite formation, the most satisfactory (to me) petrogenetic model for the origin of spilites and albite diabases (the latter at times neglected by spilite theorists) involves crystallization from an original spilitic magma rich in volatiles, and combines features of theories (2) and (4). This is supported by textural and compositional evidence.

The relation between *texture* and mineral content in spilites and albite diabases appears contradictory: a high-temperature basaltic or doleritic texture is expressed in a low-temperature mineralogical assemblage (albite, chlorite, calcite, etc.). Though it simply denotes disequilibrium, this state of affairs is sometimes explained, as does Soler, by assuming that the spilite magma, a hydrous melt, held off bulk crystallization until primary precipitation of low-temperature minerals could take place. Such an explanation appears unlikely on various grounds, not the least compelling of which is that the same spilitic magma, when consolidating as intrusive diabase, produces plagioclase and pyroxene in abundance. It seems rather that the low-temperature assemblage pseudomorphoses and otherwise replaces an older primary assemblage of high-temperature minerals that crystallized in a high-temperature fabric. The paradoxical appearance of spilites and albite diabases, their fabric not fitting the mineral content, is caused by the telescoping of mineral facies and textures (Schermerhorn, 1960).

The field relations (lava flows, in part pillowed; intrusions with chilled margins and thermally metamorphosed wall rocks) prove that the mafics were emplaced as high-temperature melts. It is doubtful that such erupting magmas could delay crystallization long enough (except for some pyroxene and calcic plagioclase, found as unstable relics in the present assemblage) to enter the low-temperature field of primary albite and chlorite: chilled pillow and sill edges alone would argue against it. A magma forming pillow lavas or injected as thin sills must needs have been in a mobile state as a hot fluid, certainly not quenched before emplacement, for viscosity increases rapidly in magmas, however hydrous, on cooling (Carron, 1969). It needs no elucidation that albite and chlorite could not have crystallized at high magmatic temperatures. Further, it seems scarcely possible that primary crystallization of albite would mimic the high-temperature igneous-consolidation texture expressed by random laths of calcic plagioclase in basalts, andesites, mugearites and dolerites. Nor does it seem plausible that the microlitic fabric of albite in spilites would be indicative of slow crystallization, as proposed by Soler; on the contrary, as is evidenced by lavas, it must represent rapid crystallization. In the slower crystallizing diabase sills such fine-grained textures only appear at chilled borders.

For albite diabases, with their mostly coarse fabric, it can often be shown that albite pseudomorphoses earlier calcic plagioclase (relict zonal structures, saussuritization, remnants

of more calcic plagioclase), and pyroxene, in part pigeonitic, may be still present, in various stages of replacement by chlorite and other late minerals. In spilites this is difficult because they are fine-grained (and thus were more reactive), and here the deduction must often rest on analogy with albite diabase.

Though basalt fabrics vary, those basalts that possess microlitic fabrics, and especially tholeiitic basalts with an intergranular texture lacking olivine phenocrysts, are identical *as to texture* with spilites. "The analogy of the fabric of spilites with that of basalt and to a certain extent also with that of andesites is almost complete" (A m s t u t z, 1968). This contradicts Soler's assertion that spilite and basalt textures are different.

Because the Pyrite Belt mafics have been deformed and regionally metamorphosed to mineral facies in which albite is stable, it is difficult to distinguish between a basaltic/andesitic and a spilitic composition before metamorphism. Also, occurrences are reported of basaltic and other rocks bearing plagioclase more calcic than albite (F e b r e l, 1967; R a m b a u d, 1969). Spilites that have been completely retextured to greenschists (or even chlorite-sericite schists lacking albite) offer few petrographic clues as to their origin. From the study of the least deformed spilites and diabases (best represented in boreholes) it appears that part at least of the Pyrite Belt mafics could be original spilites and albite diabases, modified to varying extents by later metamorphism.

The spilite flows show a texture constituted by a framework of tiny laths of albite or altered plagioclase up to 0.5 mm long, generally in random arrangement, with the interstices filled by fine-grained chlorite and leucoxene (or sphene); calcite, sericite and scattered epidote and oxidic ore may be present too; red spilites owe their colour to hematite dust. Spilites still containing pyroxene are rare. Pumpellyite is not often seen. In addition, superposed on this basic texture, there may occur albitic phenocrysts up to a few mm long, and amygdules filled by hydrothermal late-stage minerals, mostly chlorite and calcite, sometimes quartz, epidote, albite or hematite.

The spilite tuffs, forming an important part of the spilite sequences in Spain and Portugal, display a fragmental fabric. They range from pillow breccias to coarse spilitic tuffs in which the clasts do not obviously derive from pillows, probably formed as hyaloclastite or through similar fragmenting processes. The spilite fragments in these tuffs are set in a finer-grained tuffaceous spilite matrix.

The diabases are coarser than the spilites, with plagioclases reaching 10 mm length, occasionally, and pyroxenes up to 15 mm size (S t r a u s s, 1965). Like the spilites they consist of a plagioclase-lath framework, though on a larger scale, with the interspaces occupied by chlorite, sphene or leucoxene, magnetite, ilmenite, epidote, clinzoisite and calcite, variously accompanied by pyroxene, amphibole, biotite, quartz, potash feldspar, and locally pumpellyite and prehnite. Chlorite and/or calcite amygdules also occur and

may be concentrated at the tops of sills. The diabases are not hematitic, unlike the spilites. Most are aphyric but some carry plagioclase phenocrysts. They are shallow intrusions emplaced in soft sediments, as shown by irregular, plastically deformed mudstone contacts, with loadcast-like structures. Some sills were autobrecciated locally during intrusion.

On these textures are superposed the effects of shearing and metamorphic recrystallization, producing metaspilites and metadiabases. Many spilites are highly cleaved greenschists. Pillow lavas are the least sheared but they make up a minor proportion of the spilites. A typical greenschist consists of a schistose-textured aggregate of sericite and chlorite in much varying proportions, with calcite grains and patches; it is dusty with much fine leucoxene (possibly including (clino)zoisite), and red metaspilites also contain finely dispersed hematite. Epidote and other minerals may be present in minor amounts. Scattered phenocrysts of more or less altered (sericitized) albite occur sometimes. The chlorite and calcite amygdules have been strongly flattened. Some metaspilites consist only of a very fine-grained aggregate of sericite, chlorite and leucoxene. Less strongly sheared metaspilites may contain small acicular actinolite crystalloblasts. Leucoxene or sphene is always present.

Spilites show *chemical compositions* comparable to those of many basalts, mugearites and andesites. Characteristic however are a relatively high Na₂O content together with low K₂O and often high CO₂ (though spilites contain less CaO on the whole than do basalts; in this respect they are like andesites). But these tenors are variable from spilite to spilite occurrence, and A m s t u t z (1968) has claimed that bulk spilite chemistry is not essentially different from basalt chemistry, the divergences noted being due to sampling bias.

Chemical analyses of mafic and intermediate volcanics in the Pyrite Belt have been published by many authors (such as C o l l i n s, 1922; C a r v a l h o s a, 1961; C r u z G a s p a r, 1961; S t r a u s s, 1965; R a m b a u d, 1969). Unlike Soler they indicated the localities of their samples.

Though the spilites and diabases of this metallogenic province are casually and conveniently subsumed under the designation of "mafics", they in fact embrace both *mafic* and *intermediate* compositions. Silica contents range widely, and whether one draws the upper limit of mafic rocks at 52 or at 55 percent SiO₂, part of the Pyrite Belt spilites and diabases are intermediate rocks. This is apparent in thin section from their low colour index and, sometimes, the amount of late-stage quartz. Such has been known for a long time. Thus R e g o L i m a (1890), in describing ophitic diabase not far from Caveira mine, near Grândola, termed this rock "andesitic diabase". C o l l i n s (1922), using chemical analyses, distinguished three types among his basaltic and dolerite rocks (spilites and diabases), in the Spanish Pyrite Belt: an average basic type ("typical diabase"), a most basic type, and a more acid type ("diabase porphyries" which Collins thought could be called "andesite-porphyrite"). In the Cercal region of South Portugal K l e y n (1960) distinguished albitites and keratophyres among his diabase-like

intrusives, in which he measured over 50 percent of feldspar (often over 70 percent).

Rambaud's (1969) analyses of dolerites (albite diabases) and spilitic andesites have over 52 percent SiO_2 , generally over 55 percent, while his spilites show less than 50 percent SiO_2 . Strauss (1965) presents analyses of spilites and spilitic tuffs ("Schalstein") that have well over 55 percent SiO_2 . One analysis of a porphyritic amygdaloidal spilitic lava even reaches 64.72 percent SiO_2 , near the upper limit of intermediate rocks (at 66 percent silica), approaching a felsic composition. Similarly Cruz Gaspar (1961) reports a spilitic with 65.40 percent SiO_2 . Of Soler's eight analyses, one has 55.28 percent SiO_2 and four are around 52 percent.

There is thus no need to apply the term "keratophyre" to intermediate spilites with up to 66 percent SiO_2 , expressed in abundant feldspar and interstitial quartz. Keratophyre, according to definition, is an intermediate leucocratic rock affiliated with quartz keratophyre (Schermerhorn, 1973), and employing this designation to mislabel spilites or diabases, however rich in silica, leads to confusion.

Then, some analyses of spilites are known (Collins, 1922; Strauss, 1965; Rambaud, 1969) that descend below 45 percent SiO_2 , to ultramafic compositions.

Soler's analyses all show alumina contents well over 15 percent Al_2O_3 , higher than usual for spilites or albite diabases. One "dolérite" (an albite diabase) even has 23.62 percent Al_2O_3 (but this analysis, like others, has been totalled incorrectly). This is also expressed by analyses provided by other authors, and a fairly high alumina content appears to characterize the Pyrite Belt spilites and diabases. It is independent from the silica content, mafic and intermediate spilites and diabases being equally aluminous (and generally more so than the highly silicic felsic volcanics), so the alumina is not likely to derive from sialic assimilation.

Late quartz is fairly common in the spilites and diabases, while olivine pseudomorphs are extremely rare or absent. In these respects the Pyrite Belt rocks are like tholeiitic basalts and quartz dolerites. The diabases often are albite-quartz diabases, with up to several percent of late-stage interstitial quartz. Excess silica is associated with the magma for jasper lenses are found interbedded with, or covering, spilites. These are reasons for considering the Pyrite Belt mafic-intermediate rocks to be of tholeiitic affinity. (Hermann et al. (1974) showed that Hercynian spilites in Germany possess a rare-earth elements distribution characteristic of continental tholeiites.)

Lastly some remarks on *petrogenesis*. Disregarding the evidence from the comagmatic albite diabase intrusions, Soler (1973) posits that the spilitic magma derived through intratelluric chilling from a basaltic parent magma, with subsequent contamination leading to enrichment in water, sodium, silicon and possibly CO_2 . This represents a novel and rather startling hypothesis about the origin and deposition of spilitic. Under this view quenching ("trempe", "refroidissement brutal") of the original basaltic melt intervened while it was rising through the crust, before acquiring its final

spilitic composition from contamination with sediments and connate water at the top of the crust. The resulting spilitic magma, in which minute crystals of albite were slowly separating, was extruded only afterwards onto the sea bottom. This can scarcely be considered a viable mechanism. For one thing, it does not explain why spilitic magma should originate in this way, via early intratelluric chilling (if such were possible), whereas normal basalt and andesite magmas emplaced under similar conditions in other geosynclines do not show spilitization. The magma producing spilitic pillow lavas was extruded in a highly fluid condition, and the chilled edges of pillows indicate that bulk crystallization only started during or shortly after extrusion, on rapid cooling in seawater. Barring occasional intratelluric phenocrysts, plagioclase precipitation is thus contemporaneous with effusion. The same applies to the diabase sills and their chilled borders. I fail to see how a rising basaltic magma could be quenched at all before attaining upper crustal levels. Such a melt consolidates rapidly once its temperature falls below 1100 or 1050°C (Carroll, 1969), yet is supposed to start assimilating quartzite and connate water. The range from mafic to intermediate is more plausibly explained by slight differentiation of an original tholeiitic magma. Nor does it seem necessary, for rocks with up to 5 or 6 percent Na_2O , to invoke a non-magmatic supply of sodium: it might well be juvenile.

The spilitic mystique has led petrologists to quest for external sources of sodium, it being held that Na_2O tenors rising to several percent could not be an original feature of the parent magma. There seems to be no logical necessity for such an assumption. Also, it entails the construction of some sort of makeshift mechanism enabling introduction of sodium into magma wholesale, and this leads to difficulties. This appears to be a case where Occam's razor might well be applied, to eliminate superfluous assumptions.

The sodium content of the Pyrite Belt spilitic-diabase association could not have been derived from the sea into which the spilites were extruded, because the comagmatic intrusives, equally rich in sodium, did not enter into the sea (Schermerhorn, 1970b). Let us see whether the sodium could stem from seawater entrapped in geosynclinal sediments traversed by the ascending magma. Sodium transfer could then take place only in the sedimentary layers at the top of the crust. When the magma arrived in this top zone it was in movement, prior to eruption: there can have been little time to extract significant sodium from saline pore water in the wall rocks, sufficient to raise the Na_2O content in basaltic magma by a few percent to spilitic values. If such happened there still remains a considerable quantity of chlorine to account for. The sodium must pass from the wall rocks bordering the magma conduits into the magma itself (for the latter could not assimilate enough sediments carrying NaCl-bearing pore water to raise its sodium tenor appreciably). However, it seems likely that the conduit walls would have been metamorphosed (as happened around diabase sills) and thus rendered impervious to circulating connate water. Admitting that sediments can preserve their

porosity and permeability down to a depth of 10 or 15 km (a minimum thickness if sufficient sodium from entrapped seawater is to be supplied to spilite-diabase piles many hundred metres thick) it is still doubtful whether the local stratigraphy did in fact comprise such a great thickness of uncompact sediments. The base of the Pyrite Belt succession is not exposed, but if it is like the remainder of Hercynian Iberia (and not, say, oceanic crust) there should be less than 7 km down to the base of the Cambrian, and it is not certain that sediments would stay uncompact for such a long period.

It thus appears that there is not enough seawater extractable from the underlying sediments to spilitize an ascending basalt magma (apart from the problem what happened to the remaining chlorine).

It may therefore be reasonably inferred that the sodium present in spilites and albite diabases was already contained in the magma: it must be a primary, juvenile feature, like the CO₂ content. (Schidlovski et al. (1970) obtained evidence for a magmatic, not a marine, origin of the carbonates in the Verrucano spilites, from the isotope compositions of the carbon and oxygen.) It would seem that the composition of the magma producing the spilites and diabases of the Pyrite Belt was not significantly modified by contamination, thus reflecting mantle differentiation or at any rate a deep-seated process of magma generation.

5. NATURE AND ORIGIN OF THE FELSIC VOLCANICS

The felsic volcanics in the Pyrite Belt are for a large part quartz keratophyres, varying in texture from felsites and felsophyres to coarse-grained tuffs and tuff-breccias or agglomerates; tuffs are dominant. Associated rock types comprise potassic quartz keratophyre (quartz kalikeratophyre) or rhyolite and some keratophyre (Schermerhorn, 1971a). The quartz keratophyres are of two textural types, as is often the case in quartz keratophyre provinces: type I contains feldspar and quartz phenocrysts, and type II only feldspar phenocrysts, though it is equally silicic (Schermerhorn, 1973). In some quartz keratophyres and in the potassic varieties potash feldspar phenocrysts appear in addition, sometimes attaining large sizes, as in the Megacryst Tuff at Aljustrel (Freire d'Andrade and Schermerhorn, 1971). It is therefore not correct to state that potash feldspar never forms phenocrysts. Even in the fine-grained volcanics studied by Soler plagioclase phenocrysts may be accompanied by phenocrystic potash feldspar, a fact long known (Finlayson, 1910; Rambaud, 1969).

Though the quartz keratophyres were at first regarded as albitized dacites, they are more leucocratic and silicic than would accord with dacite, and it is now thought that they represent a special type of magma, a sodic analogue of rhyolite, producing volcanics in which plagioclase crystallized

and was albitized soon afterwards during the evolution of the rock (Schermerhorn, 1973), providing a felsic counterpart to spilite crystallization.

Like the mafics, the felsic volcanics have been regionally metamorphosed under albite-producing conditions and it is hard to decide whether the albite is related to a late or post-magmatic stage of autometasmatism induced by residual magmatic fluids or whether it is due to metamorphic processes unrelated to magmatic evolution. Nonetheless one observation is in favour of the first hypothesis (which applies to non-metamorphic quartz keratophyres generally): the plagioclase phenocrysts in the quartz-keratophyric detritus in the Culm greywackes, conglomerates and tilloids are always albitic. The clasts in the Culm rocks stem from the Beja Geanticline north of the Pyrite Belt, an emergent source area during Culm deposition that provided great quantities of freshly eroded detritus largely composed of felsic volcanics like those in the Pyrite Belt (but not deformed and recrystallized), possibly even more sodic in view of the great rarity of potash feldspar in the Culm. South of the metamorphic facies covering the Pyrite Belt, the plagioclase phenocrysts in quartzkeratophyric clasts in Culm rocks and the plagioclase clasts derived from large phenocrysts are albite. This appears to support a pre-metamorphic albitic composition of the plagioclase in the Pyrite Belt volcanics.

Nevertheless, intratelluric plagioclase phenocrysts in a silicic magma can never have crystallized as primary albite: their present composition is due to albitization, with albite pseudomorphosing earlier oligoclase or andesine (and the potash feldspar phenocrysts mostly no longer consist of sanidine).

The petrography of the felsic volcanics is a great deal more complex than would appear from Soler's account which only treats of quartz keratophyres with a fine-grained texture supposedly emplaced as lavas, ignoring the more widespread tuffs (also quartz keratophyres) described by all other authors (this use of the compositional term "quartz keratophyre" in an arbitrary textural sense is not to be recommended). Nor is it true that devitrification did not occur; on the contrary, felsitic devitrification textures are common, and many tuffs are made of vitric lapilli, now devitrified. Very widespread all over the Pyrite Belt is granular tuff composed of small grains arranged in a "dilute greywacke" texture, with the components floating (to almost touching) in a fine-grained volcanic matrix; the grains are generally glassy particles devitrified to felsite, with quartz and feldspar fragments. Such rocks may be aphyric or albite-phyric, more rarely quartz-phyric. When the size of the component grains (really microlapilli) increases such tuffs grade into lapilli tuffs. The granular and lapilli tuffs were deposited in a marine environment; they may display graded bedding and other sedimentary structures, and were transported and laid down as volcanoclastic sediments. The very finest volcanic dust went to make up felsic muds, now siliceous slates, and these too are widely distributed as televolcanic deposits. Tuff-breccias and agglomerates, of vitric and vitric-

lithic composition, form a very small proportion of the felsic volcanics.

Notwithstanding Soler's belief that lavas are emplaced explosively as if they were pyroclastics ("la mise en place explosive des laves acides") it could be reasonably expected that at least a few felsic lavas, meaning true, non-explosive effusives, might be present. Strauss and Madel (1974) refer to a few autobrecciated lavas. Other accounts are controversial: the felsic volcanics at Cerro Colorado (Rio Tinto) are lavas to Pryor et al. (1972) but tuffs to Moncada et al. (1970). Submarine felsic lavas should show chilled brecciated tops and bases, should be thick and steep-sided, and should display lava-type contorted fluxion banding (Schermerhorn, 1970a). If there exist protrusions or lava domes, they could not be detached lenses (like tuffs) but must have remained attached to their feeders: in this strongly folded province there should be exposed volcanic necks linked to the protrusions.

Chemical analyses of the felsic volcanics indicate that they are for the most part highly silicic, with low Fe_2O_3 , FeO , MgO and CaO (Strauss, 1965; Rambaud, 1969). This composition, typical for quartz keratophyres (sodic) and quartz kalikeratophyres (potassic) the world over, does not accord well with average sialic crust. It has been suggested that the Pyrite Belt felsics could represent precursors, the precocious extrusive equivalents, of the extensive Hercynian granitic plutonism in Iberia (Schermerhorn, 1970b) that most likely originated by anatexis of sialic crust. If so, the felsic volcanics might be early quartzofeldspathic exudates. On the other hand, felsic, more especially sodic to sodipotassic quartz-keratophyre magmas could also form by differentiation of hydrous tholeiitic to andesitic magmas in the upper mantle. The dichotomy of the erupting magmas, felsic and mafic-intermediate, is not an argument against derivation from a deep-seated common source.

6. RIO TINTO STRATIGRAPHY

The celebrated Rio Tinto deposits occur mantling a west-northwest trending anticline of VS felsic volcanics, 6 km long by 1 km wide, surrounded by a synclorium of Culm slates. Two km to the north is a VS-PQ anticlinorium which near its southern contact with the Culm encloses two pyrite deposits, the old mines of Chaparrita and Peña de Hierro, north of Rio Tinto. One km south of the Rio Tinto anticline is another VS-PQ anticlinorium and here only one small pyrite occurrence, Valle, is known, at the top of VS. The geology of this whole area is well known (Williams, 1934, 1966; Rambaud, 1969; Moncada et al., 1970; Armengot, 1972). It is therefore surprising to find Soler (1973) presenting aberrant views on Rio Tinto stratigraphy. The following discussion aims at setting the record straight (fig. 4).

1. The base of VS in the Rio Tinto anticline is not known;

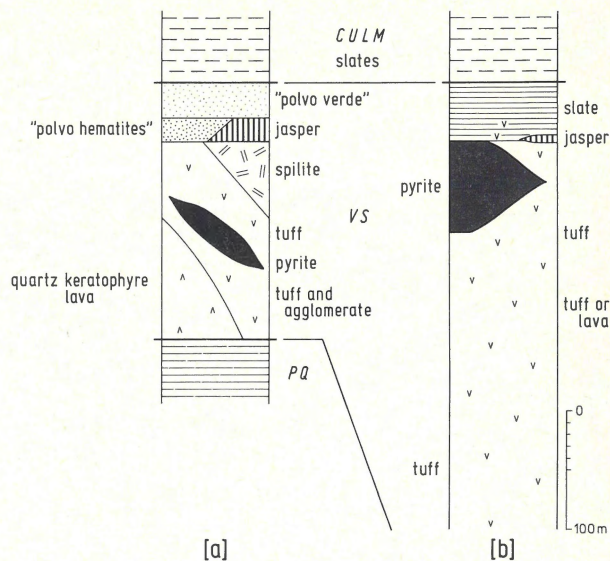


Fig. 4

Rio Tinto stratigraphy: (a) after Soler (1973); (b) after Williams (1934, 1966), Rambaud (1969), Moncada et al. (1970), Pryor et al. (1972). VS: Volcanic-Siliceous Complex; PQ: Phyllite-Quartzite Group.

the deepest level cut in the tunnel below Cerro Colorado, traversing the anticlinal core, still has volcanics (Moncada et al., 1970; Pryor et al., 1972). In the anticlinorium south of Rio Tinto PQ is exposed but is here covered by thick mafic volcanics, in turn overlain by felsic volcanics. Soler's stratigraphic section does not show any mafic rocks at the base of VS.

2. Soler draws a thick lens of quartz keratophyre lava (protrusion) at the base of VS but no Rio Tinto geologist refers to such a rock or such a geometry: only tuffaceous rocks are mentioned.

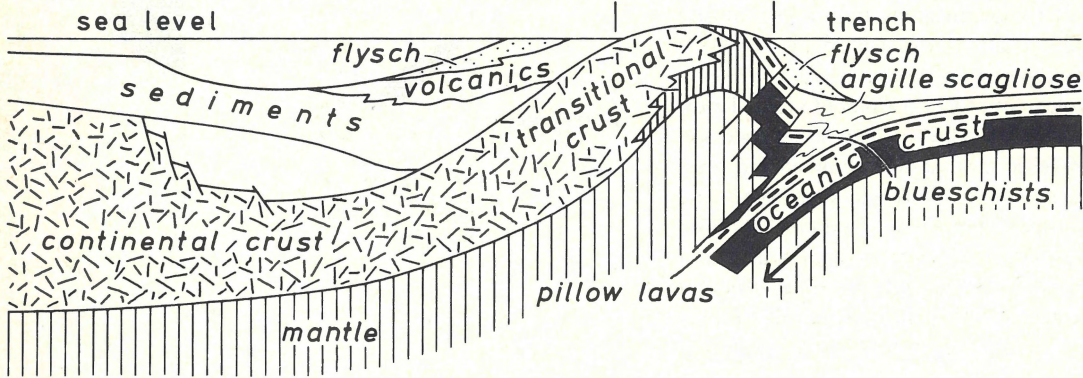
3. The pyrite bodies of the Rio Tinto district are at the top of the felsic volcanics, certainly not in their middle. In fact, at Rio Tinto itself they are separated from the Culm by up to 70 m only of topmost VS slates with some tuffs.

4. At the top of the felsic volcanics, below jasper, Soler places a spilite level. Jasper occurs at the west end of the Rio Tinto anticline but spilite in upper VS is unknown from Rio Tinto and the surrounding region.

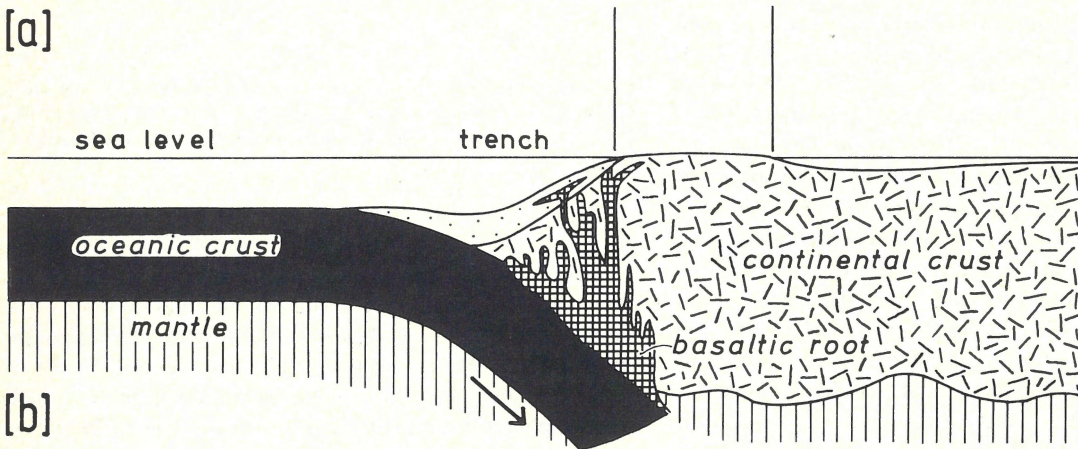
5. According to Soler the jasper passes laterally to "polvo hematites" (literally: hematite powder, an outmoded name for red slate), and: "Dans la région de Rio Tinto, ce polvo hematites est lié spatialement aux sommets rouge foncé des coulées spilitiques." Not so, for the red slates are here related to the felsic volcanism. They overlie felsic tuffs as "a narrow zone of reddish hematitic slates which probably represent

PYRITE BELT GEOSYNCLINE | CENTRAL IBERIAN BLOCK

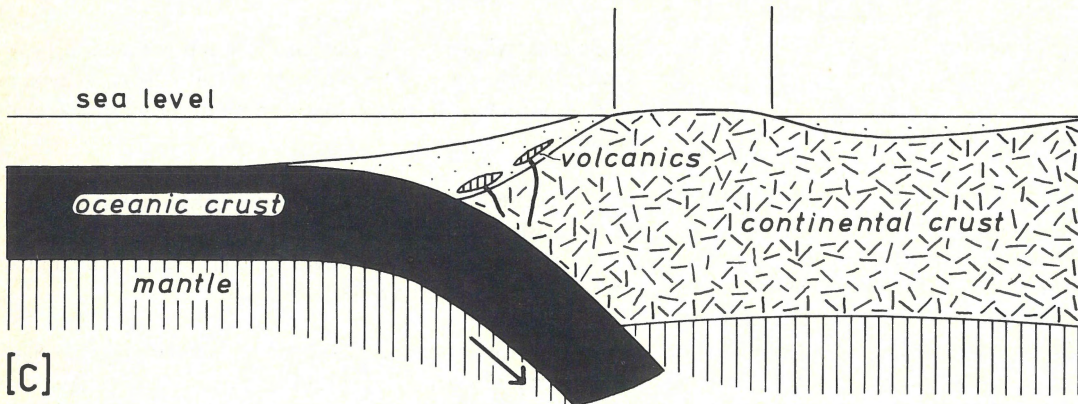
PYRITE BELT

BEJA
GEANTICLINE

[a]



[b]



[c]

Fig. 5

Three versions of subduction and volcanism in the Iberian Pyrite Belt: (a) after Soler (1973); (b) after Bard (1971) and Bard et al. (1973); (c) after Carvalho (1972).

cleaved volcanic dusts" (Williams, 1966). Rambaud, 1969) states that these "pizarras de hematites" (hematitic

slates) occur generally but not always on top of the felsic volcanics. The red slates should not be regarded as a Pyrite Belt-wide marker horizon because their position varies from region to region and such rocks can also appear at other stratigraphic levels in the Pyrite Belt (Rambaud, 1969). This is borne out by Strauss and Madel (1974) and by work in the Portuguese half of the Pyrite Belt.

7. GRAVITY SLIDES?

Folded overthrusts were first described from Aljustrel (Schermerhorn and Stanton, 1969) and have since been found, as predicted, in other regions as well (unpublished work; see also Pffeferkorn, 1972); an instructive example was recorded from Mértola by Faninet (1971). Schermerhorn and Stanton explained the folded thrusts in the Aljustrel district as original low-angle overthrusts generated in the early stages of folding, with subsequent compression and the development of axial-plane cleavage.

An origin as gravity slides does not appear likely. The Aljustrel thrust developed out of an overthrust anticline, cutting out the adjacent limb of a large syncline to the south. This is characteristic of compression tectonics but not of gravitational thrusts in which this limb should still be present, nor of gravitational gliding nappes which are no longer attached to their root zone.

Soler's sweeping statement that the first tectonic movements produced gravity slides down the slope of the "sialic flexure": "des nappes par gravité venues du Nord reposent sur les importantes accumulations schisto-grauwackeuses du Sud" (cf. fig. 5a), appears to be a too hasty extrapolation. In actual fact the known overthrusts are of local derivation for their roots can still be traced. Furthermore, the Culm reaches several km thickness but is always regularly stratified, not showing any incipient recumbent folding, as would be indicative of gravity sliding.

8. SUBDUCTION?

To explain the compositional bimodality of the Pyrite Belt magmas in geotectonic terms, Soler (1973) situates the province in a plate-tectonic context, choosing as a model Dewey and Bird's (1970) "cordilleran-type mountain belt" (fig. 5a, but Soler arbitrarily puts continental crust in the "sialic flexure" where the Pyrite Belt would be located, unlike Dewey and Bird who indicate oceanic crust here).

Dewey and Bird based their cordilleran type of orogen on the Cordilleras and the Andes along the western margin of the Americas. Such orogens arise when an oceanic lithosphere plate underthrusts a continental margin. Thus, in figure 5a the oceanic plate to the right (the central Iberian block) would be equivalent to the East Pacific Plate, the Pyrite Belt would be placed on the eastern slopes of the Andes, and the Beja Geanticline would represent the Andes. (The central Iberian block (the main part of the Meseta), a sialic entity since at least Late Precambrian times, is the site of voluminous Hercynian granites.) There is no sign of an oceanic trench over the subduction zone, filled by blueschist mélanges ("argille scagliose"), along the northeast side of the Beja Geanticline. It is true that there is here an internal basin filled by flysch exactly like the Culm of the Pyrite Belt, the Terena Greywackes (Schermerhorn, 1971a; Carvalho, 1972), but this flysch rests on Silurian and

Devonian sediments. Nor is Soler's diagram of distension fissuring of the "sialic flexure" to allow igneous eruption very apt: a crustal sheet at least 100 km across, including thick sediments, is not likely to develop dilatational cracks opening downwards just by sagging one or two kilometres: the radius of curvature is too large. It seems improbable that this could have occurred during a compressive phase when a lithosphere plate was actively underthrusting another plate.

Bernard and Soler (1974) equate the zone of supposed Carboniferous subduction with the upthrust forming the south border of the Beja Geanticline and place the geanticlinal flexure in the Pulo Formation (fig. 1), basing this on the alleged absence of volcanics in the narrow Ficalho zone running north of the Pulo Formation. This is even more unlikely: (1) a south-dipping subduction zone conflicts with the northerly dip of the upthrust (with Cambrian overriding Upper Devonian or Carboniferous flysch of the Ficalho zone) and the south to southwest vergence on either side of the thrust; (2) the Pulo Formation disappears in Spain (fig. 1), i.e. the "flexure sialique" essential to their volcanic scheme fades out, though the Pyrite Belt volcanism continues; (3) ophiolites, mélanges and high-pressure metamorphism are absent in the supposed subduction zone; (4) the Culm-type flysch on either side of the Beja Geanticline, in Pyrite Belt, Ficalho zone and Terena Greywackes, is identical and derives from rocks eroded on the Beja Geanticline (which cannot have been separated by an ocean from the Pyrite Belt); (5) Pyrite Belt-type volcanism is known in at least two localities in the Ficalho zone; (6) the upthrust loses importance and disappears east of Aracena, and in this area granites cross the supposed subduction zone into the Pyrite Belt; (7) in a cordilleran-type mountain belt the underthrust geanticlinal zone becomes an orogenic welt with abundant granite and other plutons, and high-temperature deformation and metamorphism, completely unlike the Pulo Formation.

Other authors, basing themselves on a wider knowledge of the regional geology, have proposed a more reasonable plate-tectonic model involving a subduction zone dipping in the opposite direction. Bard (1971) interpreted the Pyrite Belt in terms of an oceanic plate underthrusting the central Iberian block (fig. 5b). Bard et al. (1973), rightly stressing that the Hercynian orogen of Iberia was chiefly supra-sialic, suggest that the southwest part, comprising the Pyrite Belt geosyncline, developed on an oceanic floor. During the Lower Carboniferous a subduction zone would have operated here, with oceanic crust plunging northeastward below the continental plate of central Iberia and a "basaltic root" forming above the subducted plate. Thus the Pyrite Belt would be a marginal oceanic trench.

Carvalho (1972) uses essentially the same model but suggests that the mafic rocks were derived from upper-mantle slices brought up along fractures (fig. 5c). At the beginning of the Carboniferous the oceanic plate in the southwest intensified its underthrusting and this caused the volcanism in the Pyrite Belt eugeosyncline.

Disregarding minor differences, it is seen that Soler's

model is the mirror image of the model of Bard and co-authors, and Carvalho. If plate tectonics operated at all in this area, the latter authors are likely to be right in proposing a northeast-dipping subduction zone.

However, there are discrepancies between the Hercynian geology of southwest Iberia and the geology of those Mesozoic to recent regions elsewhere where subduction beneath a continental margin can be demonstrated. There is the lack of trench mélanges (argille scagliose and similar chaotic gravity-slide sediments), of blueschist metamorphism, and of ophiolites in the Pyrite Belt. Also, the presence of oceanic crust underlying the Pyrite Belt geosyncline is still entirely hypothetical.

A longitudinal seismic profile across South Portugal (Mueller et al., 1973) established the following crustal structure. An upper level 9-10 km thick shows seismic velocities increasing gradually from 4.25-4.40 km/sec at the surface to 6.45-6.55 km/sec at the bottom. Mueller et al. interpret this as Paleozoic sediments passing down to intermediate or basic rocks. As the profile runs through Culm rocks and as the thickness of the Paleozoic succession is at least 4 km, and strongly folded, it appears that the upper crustal level consists of geosynclinal sediments and volcanics. A middle low-velocity level 7-10 km thick (between 9-10 km and 17-21 km depth) has velocities of 5.3-5.6 km/sec but it encloses a thin (1.5 km) layer where velocity increases to 6.7 km/sec. The authors do not interpret the nature of this zone of velocity inversion. Its low velocity indicates a "granitic" (felsic) rather than a mafic composition, i.e. no oceanic crust. As the seismic profile is far SW of the presumed subduction zone, it does not seem likely that the low-velocity level could represent subduction-generated magma. The lowermost crustal level, between 17-21 km and 30-35 km depth, has a velocity of 7.07 km/sec and according to Mueller et al. this is "not much different to what has been found in continents elsewhere". Below it is the upper mantle with 8.15 km/sec velocity. These results are based on a profile parallel to the orogenic strike (a cross-section would be more useful to geological interpretation of deep crustal structure); the indications, so far, are not favourable to the hypothesis of oceanic crust flooring the Pyrite Belt geosyncline.

Thus subduction of an oceanic plate below a continental margin in this area is still unproven. The association of non-ophiolitic spilites and quartz keratophyres is not typical of volcanism at convergent plate margins. Continental margins under such conditions are characterized, primarily, by voluminous calc-alkaline andesites, and no niche has yet been found in plate-tectonic theory for non-ophiolitic spilites and associated felsics.

Bard et al. (1973) recognized the problems raised by trying to apply plate tectonics to the very wide (over 800 km) Hercynian orogen in Iberia, especially the difficulty of explaining the thermal doming responsible for the metamorphism and the abundant granites far from any subduction zone. It would be awkward to invoke an origin by subduction-generated crustal anatexis for the granites (s.l.) of

the Beja Geanticline (which in Spain spill over into the Pyrite Belt, for instance the large Campo Frio massif north of Rio Tinto — compare fig. 1) and a completely different genesis for the granites in the remainder of Hercynian Iberia. This is a pressing problem, for the older granitic suite of the central Iberian block consists largely of albitic granodiorites and quartz diorites (or trondhjemites), and these might be regarded as the plutonic equivalents of quartz keratophyre.

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